**THE GEOLOGY AND MID-JURASSIC AMALGAMATION OF TRACY ARM
TERRANE AND STIKINIA OF NORTHWESTERN BRITISH COLUMBIA**

by

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A thesis submitted to the Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Doctor of Philosophy

Ottawa-Carleton Geoscience Centre
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ABSTRACT

Tracy Arm terrane is a metamorphosed assemblage of mainly clastic continental margin type strata in the northern Coast Mountains of British Columbia that occupies a surprising position in the collage of terranes that comprise the northern Cordillera. It is separated from the nearest continental margin-type rocks (ancestral North America) by younger oceanic arc-dominated terranes, such as Stikinia, that record little continental influence. Tracy Arm terrane could be allochthonous and far-travelled, a fragment rifted from North America and later accreted, or part of North America that has been translated by strike-slip movement.

A key to the tectonic history of Tracy Arm terrane is the age of the well exposed Wann River shear zone, which juxtaposes the Tracy Arm terrane hangingwall on the west, against metamorphosed Paleozoic arc-like rocks of Stikinia, in the Tagish Lake area. This shear zone is recognised on the basis of new 1:50 000 scale geologic mapping and U-Pb zircon geochronology that indicate that some metamorphic rocks in the Tagish Lake area (previously included in Tracy Arm terrane) belong to Stikinia. It is a >4 km thick zone that is characterized by a mylonitic, penetrative transposition fabric and a gently plunging, south-southeast to north-northwest trending mineral-elongation lineation that formed under upper greenschist to lower-middle amphibolite facies conditions. It has an upper member to the south-southeast sense of displacement, and involves a local metamorphic inversion that has resulted in partial recrystallization of the Wann River shear zone mylonitic fabric. The time of thrusting is bracketed by 184-193 Ma intrusives of Stikinia that are involved in the Wann River shear zone, and a 177.5 ± 1.5 Ma U-Pb age for metamorphic monazite from Tracy Arm terrane schist, which is interpreted as the age of metamorphic minerals that overprinted the Wann River shear zone transposition fabric. Thus, Tracy Arm terrane was accreted to Stikinia between 184 and 176 Ma, during or just before the accretion of Stikinia to the western margin of ancestral North America, and coeval with intense deformation and terrane accretion throughout much of the Canadian Cordillera.

Above the shear zone, in Tracy Arm terrane, metamorphic fabrics and mineral assemblages reflect at least 2 phases of deformation and 2 upper greenschist to amphibolite facies metamorphic episodes, and below the shear zone, in Stikinia, there is evidence for 2, and locally 3 phases of deformation and one stable amphibolite facies metamorphic mineral assemblage; all phases are considered older than the Wann River shear zone because they have been transposed within the shear zone and cannot be followed across the shear zone from one terrane to the other. The large-scale geometry and regional extent of the Wann River shear zone has been obscured by folding, faulting, intrusion of Cretaceous to Tertiary plutons, and Tertiary volcanic rocks.

The rocks that comprise the Wann River shear zone hangingwall (Florence Range suite) are interpreted as a continental margin assemblage and part of Tracy Arm terrane on the basis of thick, continuous carbonate layers, quartzites, fine-grained

clastic rocks, Precambrian detrital zircons, their evolved isotopic character, and the scarcity of rocks with igneous protoliths. Rocks of the Wann River shear zone footwall (Boundary Ranges suite) are herein regarded as part of Stikinia on the basis of lithologic and isotopic similarities with coeval rocks of the Paleozoic Stikine assemblage of Stikinia. The Boundary Ranges suite comprises: metapelitic rocks, felsic and mafic metavolcanic rocks, and marble that are in part Mississippian or older, and a metamorphosed Permian volcanic rock. Like the Stikine assemblage, the Boundary Ranges suite is intruded by plutons of Early to Middle Jurassic age. These plutons comprise the igneous Tagish Lake suite.

Correlations proposed in this thesis between the Boundary Ranges suite and terranes in Yukon (parts of Yukon-Tanana terrane and/or Nisutlin Subterrane of Kootenay Terrane), southeast Alaska (parts of Taku terrane) and west-central British Columbia (undivided metamorphic rocks) imply that Stikinia's Paleozoic basement is more widespread than previously thought, forming a semi-continuous belt along the western margin of Stikinia. Assemblages included in Tracy Arm terrane generally lie to the west of this belt; they include the Florence Range suite and lithologically similar assemblages from southeastern Alaska (Tracy Arm assemblage; Kah Shakes sequence), British Columbia (quartz-rich metasedimentary rocks near Bella Coola), and southwest Yukon (Nisling assemblage). Some pendants of "undivided metamorphic rocks" in the Coast Belt may also belong to Tracy Arm terrane. Depositional contact relationships between Tracy Arm terrane and assemblages of Paleozoic volcano-sedimentary assemblages of probable Stikinia affinity have not been demonstrated, and in most cases these contacts could be tectonic.

The origin of Tracy Arm terrane is uncertain because it has not been unequivocally correlated with parautochthonous North American terranes that lie to the east of Stikinia. Tectonic models involving obduction, tectonic enclosure, dextral or sinistral strike-slip movement, and rifting of a continental ribbon from the trailing edge of the land mass (Australia?) that formed the rifted complement to the western margin of Pangea during, or soon after the Neoproterozoic break-up of Laurentia, are all possible. Models that assume Tracy Arm-Stikinia accretion occurred in the Triassic, or that Tracy Arm formed the depositional basement to Stikinia are not supported by the geometry or mid-Jurassic age of the Stikinia-Tracy Arm terrane boundary in the Tagish Lake area.

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INTRODUCTION

This thesis is concerned with metamorphic rocks in the northwestern Canadian Cordillera that belong to two terranes, Tracy Arm terrane (previously referred to as Nisling terrane) and Stikinia, and in large part concentrates on the terrane boundary between them. It represents a significant contribution to our understanding of tectonic history of the Canadian Cordillera because the previously little studied Tracy Arm terrane occupies an unexpected position in the Cordillera; it is composed of metamorphosed continental margin-like sediments, but is separated from the nearest continental margin rocks (ancestral North America) by arc-like terranes, including Stikinia, that have little continental affinity.

METHOD OF PRESENTATION

Chapter 1 reviews the controversy surrounding the relationship between Tracy Arm terrane and Stikinia and refutes previous suggestions that Tracy Arm terrane was the depositional basement to Stikinia. It was published in Geology as a joint paper with R.R. Parrish (Currie and Parrish, 1993; see also Brew et al., 1994 and Currie and Parrish, 1994). The remaining chapters have not been published, but will be submitted to journals or the Geological Survey of Canada for publication as individual papers.

Within the last five years, the assemblage and terrane terminology for the northern Canadian Cordillera has changed significantly, but no taxonomy has been

universally accepted. Table 1 provides the reader with a quick reference for the assemblage and terrane names used by other authors, and those used in this thesis for the areas shown in Figure 1. When Chapter 1 was published, the continental margin-like assemblage of the Tagish Lake area was included in Nisling terrane, rather than Tracy Arm terrane, reflecting interpretations made before 1994, but on the basis of arguments presented in Chapter 2, this assemblage is included in Tracy Arm terrane and the term "Nisling terrane" is not used in this thesis. In addition to distinguishing between Nisling terrane and Tracy Arm terrane, Chapter 2 describes the continental margin-like rocks exposed in the Tagish Lake area and discusses their depositional setting and evaluates possible correlations with other quartz-rich metasedimentary assemblages in the Cordillera. In Chapter 3, some of the metamorphic rocks that were previously included in Tracy Arm terrane are correlated with Paleozoic and Mesozoic rocks of Stikinia on the basis of new U-Pb ages and the similarity between a composite lithologic section for the Tagish Lake area and one for the type Stikinia area. Chapter 4 documents the age of the Stikinia-Tracy Arm terrane boundary, which has been determined from intrusive and structural relationships, and U-Pb crystallization and cooling ages. It also describes the terrane boundary and structures that pre-date and post-date the terrane boundary. In conclusion, Chapter 5 discusses possible correlations, evaluates previously proposed tectonic models, and presents some new tectonic models for the tectonic history of Tracy Arm terrane. Appendix A was published by the

Geological Survey of Canada (Currie, 1992a). It presents U-Pb zircon ages for undeformed intrusive rocks in the Tagish Lake area.

ORIGINAL CONTRIBUTION

The original proposal for this project evolved from discussions with M.G.Mihalynuk of the British Columbia Geological Survey.

The thesis research includes 10 months of field observations. I worked for the British Columbia Geological Survey mapping the 1:50 000 scale 104M/9W and 104M/10E map-areas with M.G.Mihalynuk in 1988. During the following two and one third field seasons I independently mapped the southwestern third of 104M/8, the central third of 104M/1, and re-mapped parts of 104M/10 in greater detail. While in the field I visited the Stikine River area with Bob Anderson, Bill McLelland and Derek Brown, part of southeastern Alaska (north of Juneau) with George Gehrels and Jay Jackson, and parts of southwest Yukon with Craig Hart, Steve Johnston and Jay Jackson.

To help support the field work component of this study I applied for and received research grants from the Geological Society of America (1990, 1991), the Department of Indian and Northern Affairs (Northern Science Training Grant; 1989, 1990, 1991), and the British Columbia Geological Survey (1989).

I performed all stages of mineral separation and selection for all geochronology samples, and performed chemistry on ~50% of the zircon and

monazite fractions. The remaining zircon and monazite fractions and all titanite and rutile fractions were analysed by isotope technicians at the geochronology laboratory of the Geological Survey of Canada (G.S.C.). Mass spectrometry was performed by technicians at the G.S.C.

I performed all microprobe analyses.

As mentioned above, Chapter 1 was co-written with R.R.Parrish. The interpretations and ideas presented in the remainder of this thesis are my own, but have benefitted from discussions with my advisors and my colleagues.

LOCATION AND ACCESS

The Tagish Lake area is on the eastern flank of the Coast Mountains, in northwestern British Columbia. It has up to 1600 metres of relief and is moderately rugged, although most areas can be reached by foot. Areas within walking distance of the shores of Atlin and Tagish lakes were accessed by zodiac from the town of Atlin. The zodiacs used in the Fantail Lake and Nelson Lake areas were flown in by helicopter or small fixed-wing aircraft. All field work in the remaining areas was done from helicopter-supported fly-camps.

Figure 1. Terrane map of parts of British Columbia, Yukon and southeast Alaska, as used in this thesis (modified from Wheeler et al., 1991). Locations shown on map correspond with those used in Table 1.

Table 1. Terrane and assemblage nomenclature for northern British Columbia, Yukon and southeast Alaska.

	Central Yukon	Southwest Yukon	Tagish Lake, B.C.	Taku River, B.C.	Stikine River, B.C.	Juneau	Ketchikan	Skeena River	Bella Coola
CONTINENTAL MARGIN- LIKE ROCKS Metamorphosed pelite, marble, basite, and quartzite. Metaclastic rocks commonly contain detrital Precambrian zircon.	TRACY ARM (Coney et al., 1980) NISLING (Wheeler et al., 1988; 1991) "lower unit" of YTT (Mortensen, 1992)	YTT (Coney et al., 1980) NISLING (Wheeler et al., 1988; 1991) "lower unit" of YTT (Mortensen, 1992) "lower package" of Aishihik metamorphic suite of YTT (Johnston and Timmerman, 1994a, b, c)	TRACY ARM (Coney et al., 1980) NISLING (Wheeler et al., 1988; 1991)	Whitewater suite of YTT (Mihalynuk et al., 1994)	Stikine assemblage of STIKINIA (McClelland, 1992)	Tracy Arm assemblage of YTT (Gehrels et al., 1990)	Kah Shakes sequence of YTT (Rubin and Saleeby, 1990)		Eastern unnamed assemblage of TRACY ARM OR YTT (Gehrels et al., 1992b) or possibly STIKINIA
This Thesis	Yukon assemblage of NISUTLIN SUB- TERRANE, or NORTH AMERICA	Nisling assemblage of TRACY ARM	Florence Range suite of TRACY ARM (Currie, 1990)	STIKINIA (see below)	STIKINIA (see below)	Tracy Arm assemblage of TRACY ARM	Tracy Arm assemblage of TRACY ARM		STIKINIA OR TRACY ARM
ARC-LIKE ROCKS Paleozoic to Mesozoic volcanic, sedimentary and plutonic rocks that commonly include Devonian plutonic protoliths and Permian volcanic or plutonic protoliths, and are intruded by Early to Middle Jurassic plutons.	"middle and upper units" of YTT (Mortensen, 1992)	"middle unit" of YTT (Mortensen, 1992) "lower package" of Aishihik metamorphic suite of YTT (Johnston and Timmerman, 1994a, b, c)	Boundary Ranges suite (Mihalynuk and Rouse, 1988)	Boundary Ranges and Mt. Stapler suites of STIKINIA (Mihalynuk et al., 1994)	Asitka assemblage of STIKINIA	Port Houghton and Endicott Arm assemblages of YTT (Gehrels et al., 1994)	Alava sequence of TAKU (Rubin and Saleeby, 1990)	Scotia-Quall metamorphic belt of NISLING (Gareau, 1991a, b)	Western unnamed assemblage of YTT or STIKINIA (Boghossian et al., 1993)
This Thesis	Nisutlin and Nasina assemblages (respectively) of NISUTLIN SUB- TERRANE	assemblage of STIKINIA and considered equivalent to the Stikine assemblage	Boundary Ranges suit of STIKINIA and considered equivalent to the Stikine assemblage	Boundary Ranges, Mt. Stapler and Whitewater suites of STIKINIA (considered equivalent to the Stikine assemblage)	Stikine Assemblage of STIKINIA	Stikine Assemblage of STIKINIA, Alava sequence of TAKU, or Sicker Group of WRANGELLIA	Alava sequence of TAKU	Stikine Assemblage of STIKINIA	STIKINIA or TRACY ARM

Terrane names are in italicized upper case letters, assemblage names are in lower case letters. Locations are shown in Fig. 1. Note that the term "Nisling terrane" is not used in this thesis. YTT = Yukon-Tanana terrane.

CHAPTER 1: JURASSIC ACCRETION OF NISLING TERRANE ALONG THE WESTERN MARGIN OF STIKINIA, COAST MOUNTAINS, NORTHWESTERN BRITISH COLUMBIA

ABSTRACT

A metamorphosed assemblage of continental margin type strata, termed Tracy Arm terrane, exists as a narrow terrane sliver in the northern Coast Mountains of British Columbia. It could be a fragment rifted from North America and later accreted, or, alternatively, it could be allochthonous and far traveled. Tracy Arm terrane is tectonically bounded by younger, oceanic, arc-dominated terranes on both sides, the Alexander terrane to the west and metamorphosed Paleozoic rocks of Stikinia to the east. Tracy Arm terrane was accreted to Stikinia along a sinistral transpressive ductile shear zone between 185 and 176 Ma, during or before the accretion of Stikinia to the ancestral margin of North America. Along this tectonic boundary, intense deformation was accompanied by lower amphibolite facies metamorphism. Previous inferences by other workers that Tracy Arm-Stikinia accretion occurred in the Triassic are incorrect, as is the supposition that Tracy Arm terrane formed the depositional basement to Stikinia. Although Tracy Arm terrane exists for ~1000 km along strike on the west side of Stikinia, its connection to either North America or another continent remains uncertain and tectonically puzzling. The collision in the mid-Cretaceous of the Tracy Arm-Stikinia terrane composite with the Alexander terrane overprinted the Jurassic deformational features and produced a complex belt of tectonic slivers involving all of these terrane fragments.

INTRODUCTION

The western Canadian Cordillera is a collage of allochthonous terranes that were accreted to the western margin of North America mostly during the Mesozoic and Tertiary (Coney et al., 1980, Monger et al., 1982; Fig. 1.1). Tracy Arm terrane, a narrow terrane interpreted as representing a continental margin setting, is separated from the continental margin rocks of ancestral North America by oceanic rocks of the Slide Mountain and Cache Creek terranes and by primitive arc rocks of Stikinia and Quesnellia south of latitude 61°N (Fig. 1.1). The location of Tracy Arm terrane, west of oceanic and arc-like terranes, pertains to the tectonic evolution of the Cordillera. Three possible scenarios that attempt to explain the present position of Tracy Arm terrane have been proposed: (1) Tracy Arm terrane was rifted from the western margin of North America, possibly transported latitudinally by strike-slip motion, and later accreted west of the oceanic and arc-like terranes; (2) oroclinal folding of North American continental margin rocks and adjacent arc-related terranes around the Cache Creek ocean placed continental margin rocks (Tracy Arm Terrane) west of Stikinia, Cache Creek and Quesnellia (Nelson and Mihalynuk, 1993; Mihalynuk et al., 1994b); or (3) Tracy Arm terrane is entirely allochthonous to ancestral North America. These models are discussed in detail in Chapter 5.

Until now the original contact between Tracy Arm terrane and the closest terrane to the east, Stikinia, has not been described and the relation between these two terranes has been a matter of speculation. We present evidence from the Tagish Lake area of northwestern British Columbia for a major tectonic boundary between Tracy

Arm terrane and the western margin of Stikinia, and we restrict the age of their juxtaposition along the boundary. In doing so we dispute previous claims that Tracy Arm terrane was contiguous with Stikinia during Triassic time and may have been the depositional basement to Stikinia (Gehrels et al., 1990; Hart and Radloff, 1990; Hansen et al., 1990; Jackson et al., 1991; McClelland, 1992).

WHAT IS TRACY ARM TERRANE?

The metamorphic rocks in the Tagish Lake area were originally included in Tracy Arm terrane by Coney et al. (1980), and later included in Nisling terrane by Wheeler et al. (1991). For reasons outlined in Chapter 2 the term "Tracy Arm terrane" is used in this thesis (Table 1). Wheeler et al. (1988) defined Nisling terrane as a "metamorphosed Proterozoic to lower Paleozoic (?) passive continental margin assemblage and partly metamorphosed carbonaceous and siliceous offshore sediments" (see also Wheeler et al., 1991). In Yukon and southeast Alaska, on the west side of the Coast belt, rocks that have been included in both Tracy Arm terrane (Coney et al., 1980) and Nisling terrane (Wheeler et al., 1991) contain detrital zircons with Precambrian U-Pb ages (Gehrels et al., 1990) and are correlated with the Yukon Crystalline terrane (Gehrels et al., 1990) or Yukon-Tanana terrane (McClelland et al., 1991; Gehrels et al., 1992a; "lower unit" of Yukon-Tanana terrane in the Aishihik Lake area, Mortensen, 1992). In places they are spatially associated with Paleozoic felsic and mafic rocks and related intrusives, leading many authors to regard these Paleozoic magmatic rocks and the continental margin-like assemblages as part of the

same terrane (Oldow et al., 1989; Gehrels et al., 1990; McClelland et al., 1991; Samson et al., 1991a; Mortensen, 1992). We do not advocate these views because Tracy Arm terrane metasedimentary rocks, which we correlate with the lithologically similar rocks from southeast Alaska and Yukon, are not associated with Paleozoic magmatic rocks in the Tagish Lake area, where they are well exposed; the Tracy Arm terrane strata are almost entirely of sedimentary protolith and lack major magmatic rocks as an essential component.

TRACY ARM TERRANE OF THE TAGISH LAKE AREA

The metamorphic rocks in the Tagish Lake area (Fig. 1.1) were previously included in Nisling terrane in their entirety (Wheeler et al., 1991). However, there are two distinct suites of metamorphic rocks in this area, one of which (termed the Florence Range suite by Currie, 1990a; Chapter 2; Fig. 1.1) is lithologically consistent with the definition of Tracy Arm terrane. This suite comprises a metasedimentary suite of biotite-muscovite schist ($\pm$ garnet, fibrolite, andalusite, and staurolite), marble, calcsilicate, and minor quartzite and amphibolite. It is considered bona fide Tracy Arm terrane, and lies in the hanging wall of a shear zone that juxtaposes it against the second suite of metamorphic rocks, which clearly resembles Stikinia (Fig. 1.1; Chapter 3; see below).

STIKINIA OF THE TAGISH LAKE AREA

Rocks of Stikinia are exposed as unmetamorphosed Upper Triassic volcanic and sedimentary rocks (Stuhini Group) east of the Llewellyn fault (Fig. 1.1) and as greenschist and amphibolite facies metamorphic rocks west of the Llewellyn fault, where they lie in the footwall of the Wann River shear zone, which juxtaposes them against the Tracy Arm terrane rocks (Currie, 1992b). This second metamorphic suite includes Mississippian and older polydeformed chlorite-actinolite schist and chlorite schist, with minor pyroxene-chlorite schist, biotite schist ($\pm$ muscovite, garnet, kyanite, staurolite, and graphite), discontinuous marble layers, chlorite-hornblende schist, and felsic orthogneiss (Boundary Ranges suite; Currie, 1992b), a Permian (270 ± 5 Ma; Currie, 1992b) intermediate quartz-plagioclase-hornblende ($\pm$ minor biotite) gneiss (Wann River gneiss), and variably deformed Early Jurassic plutonic rocks (including the Hale Mountain granodiorite), which locally intrude the Permian gneiss (Currie, 1992b; Chapter 3). These are interpreted as mafic and felsic igneous rocks and sedimentary rocks formed largely in an island-arc setting, and we interpret them to be part of Stikinia.

Stikinia comprises Devonian to Mississippian bimodal arc volcanic rocks (including pyroxene-phyric mafic flows; Brown and Gunning, 1989), siliceous and carbonaceous shale, and limestone members of the Stikine assemblage (Anderson, 1989; Brown et al., 1991); Permian volcanic rocks (Anderson, 1989; Brown et al., 1991); Middle and Early Triassic volcanic rocks; and Early Jurassic plutons (Anderson, 1989; Anderson and Bevier, 1990; Brown et al., 1991). Metamorphic rocks

of the second suite in the Tagish Lake area contain many lithologic elements of the Paleozoic part of Stikinia and the Mississippian and Early Jurassic igneous suites. The Boundary Ranges suite and the Wann River gneiss may also be correlated with the "middle and upper units" of the Yukon-Tanana terrane of Mortensen (1992), respectively (Currie, 1992b). They are distinctly different from Tracy Arm terrane, which is almost entirely sedimentary. They are also distinct from the Alexander terrane because they lack coarse clastic rocks and thick Devonian to Pennsylvanian carbonate rock layers, and because no rocks correlative with Mississippian felsic orthogneisses of the Tagish Lake area are represented in the Alexander terrane.

STIKINIA-TRACY ARM TERRANE BOUNDARY

The Stikinia-Tracy Arm terrane boundary is an imbricated shear zone that is referred to as the Wann River shear zone. It is exposed in three places: on White Moose Mountain where it is subhorizontal, in the upper Wann River valley where it dips $\sim 50^\circ$ to the south, and in the Hoboe Creek area where it is subvertical (Fig. 1.1). Within 1 km of this contact, all fabrics in Tracy Arm terrane, including fabrics formed during at least two previous phases of deformation, are transposed into parallelism with the contact (Fig. 1.1). The resulting planar shear fabric continues, concordant to the contact, into adjacent Stikinia rocks. The Early Jurassic pluton adjacent to the shear zone (Hale Mountain granodiorite; 185 Ma; Currie, 1992b) preserves quartz ribbons and shear sense indicators (including rotated feldspar porphyroclasts and C-S fabrics), indicating that movement of Tracy Arm Terrane hanging wall was toward the

south where the shear zone is subhorizontal to steep, and sinistral where it is subvertical. Hornblende-oligoclase stable mineral assemblages and dynamic recrystallization of feldspar indicate that deformation took place when the rocks were at transitional greenschist to amphibolite facies temperatures and pressures.

The shear fabric is younger than the 185 Ma Hale Mountain granodiorite and older than ~176 Ma, on the basis of U-Pb titanite, rutile, and other dates (Currie, 1992b; Chapter 4), which indicate postmetamorphic cooling. We relate shear fabric formation to juxtaposition and imbrication of Tracy Arm and Stikinia metamorphic suites.

IMPLICATIONS

The post-Early Jurassic age for accretion of Tracy Arm terrane to Stikinia disagrees with three lines of evidence that have been used to infer a pre-Late Triassic age for accretion of "Nisling terrane" (ie. Tracy Arm terrane) to "Stikinia" in northern British Columbia. Bultman (1979) and Jackson et al. (1991) have cited the presence of a schist clast in a Late Triassic conglomerate within Stikinia as evidence for a "Nisling" source nearby at that time. Second, Werner (1978) and Jackson et al. (1991) argued that dykes similar to the Late Triassic Stuhini Group volcanic rocks of Stikinia intruded "Nisling" rocks. Finally, Hart and Radloff (1990), Hansen et al. (1990), and Jackson et al. (1991) claimed that the two terranes were together before Late Triassic time on the basis of discordant U-Pb data from the Bennett granite, of presumed 220 ± 5 Ma age (Doherty and Hart, 1988), which is interpreted as intruding the boundary

between "Nisling terrane" and "Stikinia" in southern Yukon (Tally Ho shear zone, Hart and Radloff, 1990; Hansen et al., 1990). We offer alternative views of these lines of evidence.

The schist clast referred to by Jackson et al. (1991) is described as a quartz-chlorite schist, a lithology that has not been observed in Tracy Arm terrane rocks (Florence Range suite; Currie, 1990a, 1991). The Boundary Ranges suite is probably correctly interpreted as the source of the clast by Jackson et al. (1991) and is also the most likely source of the metamorphic clasts described by Bultman (1979); schists (from chlorite-actinolite to garnet-kyanite grade) are known in Mississippian or older metasedimentary rocks of the Boundary Ranges suite (Currie, 1991). However, in contrast to Jackson et al. (1991), who include Boundary Ranges suite in Nisling terrane (ie. Tracy Arm terrane), we consider it part of Stikinia and do not consider the metamorphic clasts in the Late Triassic conglomerate indicative of a link between Tracy Arm terrane and Stikinia. Jackson et al. (1991) used Sr and Nd isotopic signatures to infer that the schist clast and Triassic sandstones were derived from "Nisling terrane," but these signatures are not necessarily unique to either Nisling terrane or Tracy Arm terrane since a reasonable characterization of Paleozoic metasedimentary rocks from Stikinia has yet to be carried out (there are published isotopic data for only one Paleozoic sedimentary rock from Stikinia, a 270 Ma greywacke with primitive isotopic signatures; $\epsilon_{\text{Nd}}(0) = +4.6$, $\epsilon_{\text{Nd}}(\text{T}) = +6.4$, and $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7045$; Samson et al., 1989).

The "Stikinia" dykes referred to by Werner (1978) and Jackson et al. (1991) are near the south end of Atlin Lake, an area known to be underlain by metavolcanic and metasedimentary rocks that are regarded as Stikinia (Boundary Ranges suite; Fig. 1.1). It is no surprise that such dykes exist there. What is more significant is their absence in metasedimentary rocks of Tracy Arm terrane (Florence Range suite) to the west.

The interpretation of U-Pb zircon data for the Bennett granite by Doherty and Hart (1988) is incorrect and not consistent with a 220 Ma age. The original interpretation was made on the basis of a regression of three discordant analyses that yields a 220-Ma, upper-intercept age; the fourth discordant analysis was ignored. This interpretation is erroneous because $^{207}\text{Pb}/^{206}\text{Pb}$ and Pb/U ages for the four analyses range from 241 to 906 Ma and from 130 to 310 Ma, respectively, indicating that the zircon fractions contained a component of inherited zircon much older than the interpreted 220 Ma upper intercept. These data provide no useful limits on the crystallization age of the Bennett granite. It is most likely that the granite is considerably younger than 200 Ma and therefore these data do not provide limits for the age of the accretion of Tracy Arm terrane to Stikinia.

The oldest confirmed link between Tracy Arm terrane and Stikinia is the shear zone fabric that formed during juxtaposition of these terranes, which otherwise share no features older than 185 Ma. Tracy Arm terrane lacks Paleozoic and Mesozoic (early Jurassic and younger) plutonic and volcanic rocks characteristic of Stikinia and consequently cannot be the basement to Stikinia, which helps explain the primitive

isotopic signatures characteristic of Stikinia (Samson et al., 1989). The sinistral sense of shear observed in the Hoboe Creek shear zone is consistent with available Middle Jurassic plate motion data for the Farallon and Izanagi plates relative to the North American plate and its adjacent terranes (Engelbrechtsen et al., 1985).

The reasonably close spatial association in southeast Alaska and Yukon of metasedimentary strata bearing Precambrian zircons (Tracy Arm assemblage, Gehrels et al., 1992a; "lower unit" of the Yukon-Tanana terrane, Mortensen, 1992; quartzose turbiditic strata in the Iskut River area, McClelland, 1992) and Paleozoic metaigneous rocks (most of the Endicott Arm assemblage, Gehrels et al., 1990, 1992a; Ruth assemblage, McClelland et al., 1991; Alava sequence, Rubin and Saleeby, 1991; "middle unit" of Yukon-Tanana terrane, Mortensen, 1992; Stikine assemblage, McClelland, 1992), as alluded to earlier in this chapter, requires explanation. We correlate the metasedimentary rocks with Tracy Arm terrane and the metaigneous rocks with the Paleozoic part of Stikinia (Chapters 2; 3). In the Tagish Lake area, Tracy Arm and Stikinia rocks were juxtaposed during the Jurassic along a major shear zone, faulted, and subsequently folded. In Yukon, the unraveling of younger structures is hampered by the poor exposure of critical geologic relations. In southeast Alaska, rocks of both terranes were severely deformed, metamorphosed, and sheared by middle Cretaceous and/or Paleogene tectonic events resulting from their interaction with the Alexander terrane (e.g., Crawford et al., 1987; Rubin et al., 1991; Gehrels et al., 1992a). This subsequent event reformed what was probably a previously imbricated and metamorphosed composite terrane of Tracy Arm and Stikinia fragments. The

considerable tectonic mixing that resulted from this superposition of structures has no doubt obscured the primary relations between these originally distinct disparate terranes. I suggest that this tectonic mixing along younger structures has produced a lithologic association that has led to the confused nomenclature of the "Nisling"- "Tracy-Arm"- "Yukon-Tanana" terrane affiliations in southeast Alaska and Yukon.

Figure 1.1. Geologic map and cross section of the Tagish Lake area. Cross section transects the imbricated contact between Tracy Arm terrane and Stikinia, which is called the Wann River shear zone (WRSZ). This boundary formed during Early to Middle Jurassic time (184 to 176 Ma; Chapter 4). Geology is based on work by Mihalynuk et al. (1989a, b; 1990) and Currie (1990, 1991, 1992a, b). Inset shows location of Tracy Arm terrane and adjacent terranes (modified from Wheeler et al., 1991). Location of Tracy Arm terrane in southeast Alaska is based on presence of quartz-rich metaclastic rocks, some of which are known to contain detrital zircon of Proterozoic age. Ak - Alaska, BC - British Columbia, YT - Yukon Territory.

CHAPTER 2: THE DESCRIPTION, DEPOSITIONAL SETTING AND CORRELATION OF THE FLORENCE RANGE SUITE, TRACY ARM TERRANE, NORTHEASTERN COAST MOUNTAINS, BRITISH COLUMBIA.

ABSTRACT

Tracy Arm terrane is a narrow terrane sliver in the northern Coast Mountains that is composed of discontinuous fragments of metamorphosed continental margin-type sedimentary rocks. One of these fragments, the Florence Range suite, occurs in the Tagish Lake area of northwestern British Columbia. It is primarily composed of multiply deformed, amphibolite facies, well-layered metasedimentary rock that are in some cases continuous over many kilometres. These layers include biotite and muscovite-biotite schist ($\pm$ garnet $\pm$ kyanite $\pm$ sillimanite $\pm$ andalusite $\pm$ staurolite), semipelitic rocks, marble, and minor quartzite and calc-silicate. The Florence Range suite also contains minor metaigneous rocks, including hornblende amphibolite ($\pm$ garnet), hornblende-plagioclase schist, and felsic gneiss. This suite is interpreted as a continental margin assemblage on the basis of its thick, continuous carbonate rock layers, quartzites, fine-grained clastic rocks, Precambrian detrital zircons, evolved isotopic character, and the scarcity of rocks with igneous protoliths. The depositional age of the Florence Range suite is only constrained to between ca. 1.1 Ga and 175 Ma.

Assemblages in the western Cordillera that are lithologically similar to the Florence Range suite are included in Tracy Arm terrane. They include the Tracy Arm

assemblage and Kah Shakes sequence of southeastern Alaska, quartz-rich metasedimentary rocks near Bella Coola, and the Nisling assemblage of southwest Yukon. Some pendants of "undivided metamorphic rocks" in the Coast Belt may also belong to Tracy Arm terrane. These assemblages are referred to as Tracy Arm terrane, rather than Nisling terrane, because (1) they are lithologically distinct from some of the metamorphic rocks exposed in the Nisling Range, for which Nisling terrane was named, and (2) they have not been unequivocally correlated with the quartz-rich rocks metasedimentary rocks exposed in central Yukon that were included in Nisling terrane by Wheeler et al. (1991). Tracy Arm terrane resembles the Dillinger, Nixon Fork, Kilbuck, and Idono terranes of mainland Alaska.

INTRODUCTION

When terrane theory first emerged, metamorphic rocks in the Tagish Lake area were included in Tracy Arm terrane (Fig. 2.1; Coney et al., 1980). Later, they were described as a "displaced continental margin assemblage of unknown origin" and included in Nisling terrane (Wheeler et al., 1988; 1991). More recent detailed mapping has shown that although some of the metamorphic rocks in the Tagish Lake area have protoliths consistent with a continental margin setting, and could belong to Tracy Arm terrane or Nisling terrane, others were primarily derived from igneous rocks and are more reasonably included in Stikinia (Currie, 1992b; Chapter 3). This chapter describes the well-exposed and well-preserved rocks from the Tagish Lake area that have been included in Tracy Arm terrane. A continental margin depositional setting is

proposed for these rocks, consistent with previous interpretations (Coney et al. 1980, Wheeler et al., 1991; Gehrels et al., 1990), and possible correlations with similar assemblages and terranes from Alaska, British Columbia and Yukon are evaluated. The tectonic relations between the continental margin-like assemblages and adjacent arc-like assemblages are discussed in Chapter 5.

GEOLOGY OF THE TAGISH LAKE AREA

Tracy Arm terrane rocks in the Tagish Lake area have been referred to as the Florence Range metamorphic suite (Currie 1990a; Mihalynuk et al., 1990), and are hereafter referred to as the Florence Range suite for brevity. They are equivalent to the micaceous quartzite, amphibolite, mica schist and "limestone" of Christie's (1957) "unit 1a", and the "coast metamorphic rocks" mapped by Werner (1977; 1978; unpublished map). The Florence Range suite has only been recognised west of the Llewellyn fault zone, a steep, north-northwest striking fault zone that was active after ~100 Ma, and possibly earlier (Appendix A; Fig. 2.2). It is preserved in the core of the Willison synform, and as a klippe on top of White Moose Mountain (Fig. 2.2). In both areas the contact between the Florence Range suite and the underlying metamorphic rocks of Stikinia is sheared and imbricated, and is interpreted as the Stikinia-Tracy Arm terrane boundary (Fig. 2.2; Currie 1990a; 1991; Chapters 1 and 4). Metamorphosed Stikinia rocks are also exposed east of the Llewellyn fault zone, but there they are structurally overlain by relatively unmetamorphosed Mesozoic volcanic and sedimentary rocks of Stikinia (Fig. 2.2). On both sides of the Llewellyn fault

zone, mid-Mesozoic and older rocks from both terranes have been intruded by Early Cretaceous and younger plutons of the Coast Mountains (Appendix A) and are overlain by Tertiary volcanic rocks (Sloko Group and Mt. Switzer volcanics; M.G. Mihalynuk pers. comm., 1994; Figs. 2.2; 2.3).

The Florence Range suite provides an excellent opportunity to study Tracy Arm terrane because it is well exposed, and relative to other metamorphic rocks in the Coast Mountains, it is only mildly affected by Cretaceous and younger Coast Belt plutonism and deformation. However, detailed stratigraphic analysis of this suite has been hindered by (1) the lack of fossils and isotopic data that reflect its age, (2) the absence of way-up indicators, (3) structural complication, and (4) metamorphic modification. The structurally lowest ~1 km of the Florence Range suite comprises strongly transposed straight gneisses and schists that have been significantly modified by the strain associated with the mid-Jurassic thrusting of the Florence Range suite over Stikinia (Chapter 4). The Florence Range suite rocks that lie above the Wann River shear zone preserve evidence for at least two pre-accretionary phases of deformation and amphibolite facies metamorphic events, and the entire suite has been disrupted by post-accretionary faults and folds (Chapter 4). Where the displacement across one of these structures is unknown, correlation of lithologic units between lithologic sections cannot be resolved without additional information, such as the ages of the metasedimentary rocks. However, despite structural complication and disruption by plutons, it is possible to document first-order relative abundances of rock-types, and lithologic associations in the Florence Range suite, and construct lithologic, but not

stratigraphic, sections for this suite (Fig. 2.4). Individual lithologic layers are generally continuous over 100's of metres, and are continuous for kilometres where they have not been structurally truncated or "pinched out" by a change in facies conditions. In fact, some lithologic sections appear to be continuous and without structural repetition or omission. For example, the Willison Glacier section may be >10 km thick (Figs. 2.2; 2.4).

The three lithologic sections shown in Fig. 2.4 were chosen because they contain only minimal structural repetition; all sections were measured perpendicular to strike (section locations are shown in Fig. 2.2). The present thicknesses of both these sections, and the individual lithologic layers they are composed of, are not considered equal to their depositional thickness because all layers have been affected by multiple phases of deformation.

Rock-Types of the Florence Range Suite (Tracy Arm Terrane) and Their Relative Abundances

The Florence Range suite primarily comprises multiply deformed, amphibolite facies, cm- to 10 m-scale metasedimentary layers of uncertain age (see below; Currie, 1990a). They include muscovite-biotite schists (>20% micaceous minerals, $\pm$ garnet $\pm$ kyanite $\pm$ sillimanite $\pm$ andalusite $\pm$ staurolite), semi-pelitic rocks (10 to 20% micaceous minerals), marble, and minor quartzite (>90% quartz) and calc-silicate layers (Figs. 2.3 and 2.4; Plates 2.1a to 2.1c; see Chapter 4 for descriptions of metamorphic mineral assemblages and textural relations; Table 4.1). Metaigneous

rocks are minor; they include 0.1 to 10 m thick, medium to fine-grained hornblende amphibolite layers, and rare, <1 m thick quartzo-feldspathic gneisses that may have felsic igneous protoliths (Table 4.1).

The Mt. Caplice section

The Mt. Caplice section is exposed west of Hoboe Creek and north of the Caplice Glacier (Section A-A'; Fig. 2.4a). In general, lithologic layers in this section are sub-vertical and strike north-south. South of this section they have been offset by minor east-west striking structures with sinistral offsets, but in general appear to be continuous with layers exposed ~4.5 km to the south along the northern edge of the Llewellyn Glacier (Fig. 2.3). Approximately 3.5 km farther south, the same lithologic units appear to be exposed on a nunatak (Fig. 2.2). The base of the section is tentatively placed at the contact between a marble and a Stikinia hornblende-plagioclase gneiss (the Wann River gneiss; Chapter 3). Directly overlying the marble are interlayered, cm- to m-scale quartzite and pelitic layers (Plate 2.1a) that are definitely considered part of the Florence Range suite. These layers are interpreted as facies equivalents of a ~400 m thick, medium-grained, mature quartzite layer exposed along the edge of the Llewellyn Glacier. The quartzite weathers white to tan, with cm-scale colour-banding (Plate 2.1b), and comprises >90% quartz, with minor muscovite and chlorite, and accessory zircon and apatite. The pelitic layers are medium- to fine-grained and are medium to dark grey on weathered surfaces (Plate 2.1c). They commonly contain cm-scale leucosome lenses that locally form layers that are

continuous over many metres. Garnet is common, but kyanite, sillimanite, andalusite and staurolite are rare.

The quartzite and pelite layers are overlain by m-scale pelitic, semipelitic, and marble layers that are up to 20 m thick. The semipelitic layers are light to medium grey and are characterized by compositional layering defined by variable concentrations of quartz (30-50%) and feldspar (15-30%; Plate 2.1d). They also contain biotite and muscovite, and locally contain garnet. Marble layers are generally coarse grained and bleached white with mm-scale grey graphitic (?) laminae (Plate 2.1e). Calc-silicate layers are often characterized by cm-scale compositional banding, with alternating calcite-, diopside-, grossular garnet-, vesuvianite-, and/or titanite-rich layers. Layers of different composition vary in colour and resistance to weathering (Plate 2.2a). In this section calc-silicate layers are up to 20 m thick, but elsewhere they are generally <30 cm thick and associated with amphibolite layers (Plate 2.2b). Approximately 2500 m above the quartzite layers in this section, the first amphibolite layers are preserved as 1 m to 10 m long, layer-parallel boudins that are 10 cm to 2 m thick. They are medium- to coarse-grained, dark green to black hornblende amphibolite to hornblende-plagioclase gneiss, and commonly contain biotite and accessory titanite, epidote, and chlorite. The western end of the Mt. Caplice section is truncated by the Willison Creek fault zone.

The Willison Glacier section

The Willison Glacier section is exposed north of Willison Glacier, where it is bounded by the Willison Creek fault zone along its eastern margin. This section has been mapped for ~5 km, and on the basis of airphoto interpretation, appears to continue for at least 5 km (Section B-B'; Figs. 2.2; 2.4c; Plates 2.2c; 2.2d). No evidence of structural repetition has been documented in this section. It comprises thick marble layers (≤ 120 m wide) separated by schist and amphibolite rocks. The pelitic layers are almost exclusively biotite and garnet-biotite schist with little or no semipelitic component. Amphibolite layers are commonly garnet-bearing, with garnets up to 20 cm in diameter, and are generally thicker than in other sections (≤ 40 m wide). As in the Mt. Caplice section, calc-silicate layers are commonly associated with amphibolite layers. This section lacks quartzite layers, but possible quartzite clasts (10 cm-scale) were observed within some of the marble layers (Plate 2.3a).

The Willison Bay section

The Willison Bay section is exposed on the east side of the Willison Creek fault zone, and west of a hornblende diorite pluton exposed along the western shore of Atlin Lake (Section C-C'; Figs. 2.2; 2.4c; Plate 2.3b). Despite evidence for minor structural duplication within this section, it has not been affected by major structural complication (compare with section W-W' in Fig. 4.5; Chapter 4). Marble layers in this section are generally thinner (< 10 m wide) and more discontinuous than elsewhere. Amphibolite layers have not been observed, but quartzite layers are

relatively common. Both quartzite and schist layers are commonly dark grey to black due to the presence of fine-grained graphite (Plates 2.3c; 2.3d). The original grain size of the quartzite layers has been altered by recrystallization, and is controlled by graphite content (Plate 2.3e).

Other Rock-Types

A few distinctive rock-types of the Florence Range suite have not been recognised in the sections described above. They include a deformed carbonate pebble conglomerate exposed east of the Willison Creek headwaters (Plate 2.3f), and two felsic gneiss layers that are preserved in the Wann River valley. Both felsic gneisses are ~1 m thick and are parallel to layering and foliation in the adjacent rocks. They comprise quartz and feldspar, with minor biotite and accessory zircon and apatite.

The Age and Isotopic Character of the Florence Range Suite

The Florence Range suite lacks both macro- and microfossils (six carbonate rock samples from the Tagish Lake area analysed for microfossils were barren; M.J. Orchard, pers. comm., 1990; 1991) and isotopic age data that reflect the age of its deposition. It is older than Middle Jurassic because it was metamorphosed at ca. 177 Ma and contains fabrics that formed during displacement on the 184-176 Ma Wann River shear zone (Chapter 4). Rb-Sr, K-Ar and Ar-Ar mineral ages provide information about the cooling histories of metamorphic rocks and provide minimum ages for their protoliths. The oldest metamorphic mineral age from the Florence Range

suite is an unpublished K-Ar age of 244 ± 16 Ma for hornblende in an amphibolite from the Florence Range suite (unpublished data, R.L. Armstrong, pers. comm., 1991; Table 2.1). If this age is correct, some, if not all Florence Range suite rocks must be >228 Ma old. The oldest dated, massive pluton that intrudes the Florence Range suite is Cretaceous in age (127.2 ± 0.6 Ma; Appendix A).

Isotopic characterization of metasedimentary Florence Range suite rocks indicates that they were primarily derived from an isotopically evolved source. $^{87}\text{Sr}/^{86}\text{Sr}$ values for muscovite schist from the Florence Range suite are between 0.720 and 0.750 (R.L. Armstrong, pers. comm., 1991; Table 2.1; Fig. 2.5). In contrast, $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for metabasalts (amphibolites) from the Florence Range suite are much lower (0.7044-0.7061; R.L. Armstrong, pers. comm., 1991; Table 2.1). An Rb-Sr isochron plot for the Florence Range suite indicates that analyses of the metasedimentary rocks resemble those of Late Proterozoic autochthonous North American rocks (Windermere Supergroup; Fig. 2.5). Discordant single grain U-Pb ages for detrital zircons from a Florence Range suite quartzite indicate that the ultimate source of the zircons was Precambrian in age (L.D. Currie, unpublished data). Precambrian inheritance is also reflected in the regression of U-Pb analyses of zircon from Florence Range suite leucosome, which yields an upper intercept of ca. 1.85 Ga (Chapter 4), and by zircon xenocrysts in a felsic gneiss from within the Florence Range suite (L.D. Currie, unpublished data).

The Inferred Depositional Setting of the Florence Range Suite

The Florence Range suite is primarily composed of rocks of sedimentary protolith. With the exception of lithologic contacts that are considered transposed bedding planes, no sedimentary structures have been observed. The marble layers are interpreted as limestone layers; in most cases they are thick and continuous. Schistose and semi-pelitic rocks are regarded as metamorphosed fine-grained clastic rocks. Most of the quartzite layers are also considered metaclastic rocks, with the possible exception of the fine-grained graphitic quartzite layers in the Willison Creek section which may represent metamorphosed chert layers. Metasedimentary layers are regarded as transposed beds because their contacts are coplanar with the predominant foliation, except in the hinge regions of rare folds that have axial planar foliations (Chapter 4).

The amphibolite layers are generally interpreted as flows or sills, though some could be sedimentary in origin. Because their original contacts have been transposed, it is not possible to determine whether the metaigneous layers are intrusive or extrusive. Major element geochemistry cannot be used to determine the original composition of the amphibolite layers because the alkali elements show evidence of remobilization, which presumably occurred during metamorphism (Nazarchuk, 1990). However, systematic variations in trace element chemistry for seven Florence Range suite amphibolite samples from near the Wann River indicate that the amphibolite layers could be cogenetic, and on the basis of two discrimination diagrams (the Zr/Y vs. Zr (ppm) diagram of Pearce and Norry, 1979; and the Ti/100, Zr, and Yx3 ternary diagram of Pearce and Cann, 1973) they are tentatively considered within plate basalts

that formed in either a continental or ocean island setting (Nazarchuk, 1990). The applicability of discrimination diagrams for these metamorphosed rocks is questionable because it is not possible to determine whether their protoliths were fractionated during crystallization. Similar amphibolite layers that occur in southeastern British Columbia (Horsethief Creek Group, Windermere Supergroup) are interpreted as metamorphosed dykes, sills, tuffs and reworked tuffs that formed in a crustally thinned Atlantic-type continental margin setting (Sevigny, 1987). The Horsethief Creek Group amphibolite layers resemble those of the Florence Range suite in composition, are of like thickness and continuity, and are associated with comparable rock-types. A similar continental margin setting interpretation for the Florence Range suite amphibolite layers is consistent with the isotopic ratios for the Florence Range suite amphibolite layers, which are less evolved than the ratios of the associated metasedimentary rocks (Table 2.1). Centimetre-scale calc-silicate layers along the contacts between marble and amphibolite layers are considered the products of regional metamorphism, suggesting that their present composition does not reflect their original composition (Chapter 4). Thicker (1 m to 30 m thick) calc-silicate layers, however, probably reflect an originally silicious carbonate composition.

The Florence Range suite is considered a continental margin assemblage, and probably an outer shelf sequence, on the basis of the presence of thick, continuous carbonate rock layers, quartzites, fine-grained clastic rocks and Precambrian detrital zircons, its evolved isotopic character, and the scarcity of rocks with igneous protoliths. However, its depositional setting cannot be unequivocally determined

because it has been modified by deformation and metamorphism, and because the character of its depositional basement is unknown (the structurally lowest Tracy Arm terrane rocks exposed in the Tagish Lake area are metasedimentary rocks that are in sheared contact with the underlying Stikinia terrane; Chapters 1; 4). By analogy with modern continental margin deposits, the Florence Range suite was probably deposited on continental or transitional crust. Speculatively, the amphibolites may be related to igneous intrusion concomitant with the initial extension along the rifted margin (e.g. Keen and de Voogd, 1988).

Structural complexity, and the lack of age control (see below) and paleocurrent indicators preclude recognition of facies changes between sections that are not contiguous. The northward transition from the thick quartzite layer exposed along the northern margin of the Llewellyn Glacier, to thinner interlayered quartzite and pelitic layers in the Mt. Caplice section suggests that, at least locally, the source may have been to the south (present coordinates). Thinning of marble layers toward the north is consistent with this interpretation (compare Mt. Caplice and Willison Bay sections). If the graphitic quartzite is metamorphosed chert, its presence in only the northernmost section (Willison Bay) would also support this interpretation.

The Extent of the Florence Range Suite

The Florence Range suite has been recognised as far north as White Moose Mountain, where it forms a klippe, and as far west and south as the Llewellyn Glacier, where it is bounded by Cretaceous, and younger, plutons of the Coast Belt (Figs. 2.2;

2.3). Pendants in the Coast Belt plutons are referred to as "undivided metamorphic rocks" (Fig. 2.6; Gehrels and Berg, 1984; Wheeler et al., 1991), and comprise both metasedimentary and metavolcanic rocks that have not yet been mapped in detail. Some of these rocks may be extensions of the Florence Range suite, whereas others may belong to Stikinia. Within one pendant, the post-Triassic to pre-60 Ma Meade Glacier fault juxtaposes metasedimentary rocks (part of the Florence Range suite or Stikinia?) against stratified, fossiliferous Triassic rocks that have been interpreted as belonging to Stikinia (Brew and Ford, 1987; Brew et al., 1994), but which could also be part of Taku terrane. Brew et al. (1994) have suggested that the Meade Glacier fault may be the continuation of the Wann River shear zone or the Coast Ranges megacolineament, and the locus of significant displacement in this area. Until more information regarding the age, orientation and sense of motion on this fault is available, its influence on the present distribution of Florence Range suite will not be known.

In the Tagish Lake area the Florence Range suite has only been recognised west of the Llewellyn fault zone (Fig. 2.2); the Florence Range suite may not be exposed east of the fault zone, or it may have never extended beyond the present trace of the fault zone. The Llewellyn fault zone was active after ~100 Ma, and possibly before (Appendix A) and extends from at least as far south as the Taku River (Mihalynuk et al., 1994), to north of the British Columbia-Yukon border, where it is truncated by an Early Eocene pluton (Hart and Radloff, 1990). Donelick and Dickie (1991) have suggested on the basis of ~60 Ma apatite fission track ages from west of

the fault zone that contrast with ~100 Ma ages from the east side, that some of the offset on the Llewellyn fault zone is west-side-up. However, it is not clear whether vertical motion was predominant, or the fault zone was the locus of significant dextral or sinistral strike slip displacement. Because the magnitude and sense of displacement on the Llewellyn fault zone are unknown (Chapter 4), it is not possible to restore the offset on the fault zone and locate the eastern equivalent of the Florence Range suite.

TRACY ARM TERRANE, NISLING TERRANE AND YUKON TANANA-TERRANE: AN HISTORICAL PERSPECTIVE OF TERRANE TERMINOLOGY

Quartz-rich metasedimentary rocks that were ultimately derived, at least in part, from a Precambrian source occur as disparate fragments within and along the margins of the Coast Belt from as far south as Bella Coola (~53.5°N latitude; Wheeler et al., 1991; Gehrels et al., 1992b, Boghossian et al., 1993), through southeastern Alaska and northwestern British Columbia, to central Yukon and Alaska (Fig. 2.1). In general, they are exposed west of Stikinia and east of Wrangellia and Alexander terrane. The presence of an evolved Precambrian component is reflected in their isotopic compositions, and the ages of detrital zircon crystals in the quartz-rich metasedimentary rocks and inherited xenocrystic zircons in the plutons that intrude them (Inset of Fig. 2.1; LeCouteur and Tempelman-Kluit, 1976; Morrison et al., 1979; Armstrong, 1988; Arth et al., 1988; Samson et al., 1991a; 1991b; Jackson et al. 1991; Jackson, 1992; Gehrels et al., 1990; 1991a).

Originally, the quartz-rich metasedimentary rocks of the Coast Mountains were correlated with metamorphic rocks in central and eastern Yukon, and considered part of the Yukon Group of Cairnes (1913; Kindle, 1952; Christie, 1957; Aitken, 1959; Wheeler, 1961; Souther, 1971) or Yukon Crystalline complex (Tempelman-Kluit, 1974). Later, the metamorphic rocks of the Coast Mountains and southwestern Yukon were referred to as the Tracy Arm terrane and considered distinct from the metamorphic rocks of central and eastern Yukon, which were included in Yukon-Tanana terrane. The term "Yukon-Tanana terrane" was reserved for metamorphic rocks between the Tintina and Denali faults in Alaska, and the metamorphic rocks exposed east of Stikinia and west of the Tintina Fault in Yukon (Fig. 2.6a; Coney et al., 1980; see also: Monger and Berg, 1984; 1987; Berg et al., 1978; Howell et al., 1985; Plafker, 1990).

In 1988 Wheeler et al. published a 1: 2 000 000 scale terrane map of Yukon, British Columbia and southeastern Alaska (also see Wheeler and McFeely, 1991; Wheeler et al. 1991), in which they divided the metamorphic rocks into two terranes: Nisling terrane and the Nisutlin subterrane of Kootenay terrane (Fig. 2.6b). Nisling terrane included both the metasedimentary rocks of Tracy Arm terrane, and quartz-rich metasedimentary rocks of Yukon-Tanana terrane that are exposed in inferred tectonic windows. These tectonic windows lie to the east of the Triassic to Early Jurassic Klotassin suite plutons of Stikinia (in part now referred to as the Aishihik plutonic suite by Johnston and Erdmer, in press; Fig. 2.6b). Nisling terrane was subdivided into Cambrian to Devonian "carbonaceous and silicious offshore sediments" of the Nasina

Formation, and the underlying Nisling continental margin assemblage of presumed Upper Proterozoic to Lower Cambrian age. The Nisutlin subterrane of Kootenay terrane consists of metamorphic rocks that form the hanging wall to the tectonic windows (Fig. 2.6b).

The term "Nisling terrane" has also been defined on the basis of Early Cretaceous $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages (Hansen, 1990), and Howell et al. (1985) have suggested that some terranes, including Yukon-Tanana terrane, be defined as "metamorphic terranes", rather than tectonostratigraphic terranes. However, these definitions are not used here because the terranes that this chapter is concerned with comprise reasonably coherent strata of discernible protolith, and a tectonostratigraphic terrane definition based on characteristic rock-types is considered more useful.

Recently, metamorphic rocks with Precambrian detrital zircons from the Coast Mountains and southwestern Yukon have once again been correlated with Yukon-Tanana terrane on the basis of lithologic and isotopic similarities, comparable U-Pb single grain detrital zircon ages and their association with metavolcanic rocks and Devonian orthogneisses (e.g. Gehrels et al., 1990; Rubin and Saleeby, 1992; Mortensen, 1992). Mortensen (1992) criticized Wheeler et al.'s (1991) interpretation because the Triassic to Early Jurassic plutons of the Klotassin (and Aishihik Plutonic suites) were considered part of Stikinia, but their country rocks were not. Instead, Mortensen (1992) divided the metamorphic rocks from both sides of the Klotassin and Aishihik suite plutons into the "upper, middle and lower units" of Yukon-Tanana

terrane, and suggested that the Mesozoic plutons be treated as a "stitching" assemblage between Stikinia, Nisling and possibly Cache Creek terranes.

I prefer Coney et al.'s (1980) interpretation, which involves both Tracy Arm and Yukon-Tanana terrane, over Mortensen's (1992) interpretation, which includes all of the metamorphic rocks in Yukon-Tanana terrane, because a Stikinia origin for the country rocks of the Klotassin and Aishihik suite plutons is permissible (Currie, 1992b; Chapter 3); like the Paleozoic Stikine assemblage of Stikinia, the "middle and upper units" of the Yukon-Tanana terrane include metavolcanic rocks (in part Permian in age) and Devonian felsic orthogneisses that have relatively unevolved isotopic signatures (Mortensen, 1992). In addition, they form the northern continuation of a semi-continuous belt of lithologically similar rocks that are considered part of Stikinia (ie. the Paleozoic Stikine assemblage) that runs from the Stikine River area to the British Columbia-Yukon border, and probably beyond (Chapter 3). No Jurassic or older geologic links between the quartz-rich metasedimentary rocks of the "lower unit" and the "middle or upper units" have been demonstrated, making the relationship between these units and the "lower unit" uncertain.

The term "Tracy Arm terrane" (Coney et al., 1980) is favoured over "Nisling terrane" (Wheeler et al., 1991) for two reasons. First, there are metamorphic rocks in the Nisling area that contain a significant Paleozoic igneous component (Johnston, 1993) and do not conform to the "metamorphosed Proterozoic to lower Paleozoic (?) passive continental margin assemblage and partly metamorphosed carbonaceous and siliceous offshelf sediments" definition of Nisling terrane given by Wheeler et al.

(1991); they may instead belong to Stikinia (Chapter 3). Secondly, the continuity between quartz-rich metasediments from west and east of the Klotassin and Aishihik suite plutons has not been demonstrated. The "prong" of Mesozoic plutonic rocks with "intermediate" initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios ($\sim 0.705\text{--}0.707$) roughly coincides with the western margin of Stikinia, and *appears* to be continuous with a belt of plutons of similar isotopic character in eastern and central Yukon (inset of Fig. 2.1; Armstrong, 1988). However, it is not possible to determine whether the isotopically evolved country rock that influenced the isotopic ratios of the Mesozoic plutons is continuous between the southwestern "prong" and rocks to the north and east on the basis of isotopic ratios alone. The country rock could consist of a composite, isotopically evolved basement separated by faults (compare Figs. 2.6a and b). Ideally, the continuity of the country rock should be determined by detailed mapping, but this has proven difficult because many of the metamorphic country rocks in central Yukon are structurally complex and poorly exposed. Because the correlation of the continental margin-like rocks from west of Stikinia and those to the east that are parautochthonous to North America has not been demonstrated, the term "Tracy Arm terrane", which does not presuppose such a link, is used throughout this thesis, instead of "Nisling terrane".

TRACY ARM TERRANE AND THE CORRELATION OF THE FLORENCE RANGE SUITE

The Florence Range suite is considered part of Tracy Arm terrane, and by inference, is tentatively correlated with other assemblages that have been included in

Tracy Arm terrane. Correlations are based on lithologic similarities because (1) the Florence Range suite is isolated from other continental margin-like assemblages by plutons of the Coast Belt and large-scale faults, (2) it is in large part covered by the Juneau icefield, and (3) its age is not well known. Until the ages of Tracy Arm terrane strata are known it will be impossible to identify facies-equivalent rocks with certainty, and all correlations will be questionable. Assemblages that are correlated with the Florence Range suite should primarily comprise rocks of sedimentary protolith, including thick, continuous carbonate rock layers, clastic rocks, such as quartzite, pelitic and semipelitic rocks that are in part derived from a Precambrian source, and may also include amphibolite, or lower grade equivalents (basites).

Although correlations with the Florence Range suite could be strengthened by detrital zircons that are the same age as those found in this suite, quartz-rich metasedimentary rocks with a component of inherited Precambrian detritus cannot be used as a hallmark for Tracy Arm terrane because similar rocks exist in other terranes within the Coast Belt. For example, a component of inherited Precambrian zircon has been documented in plutons of Stikinia (Bevier and Anderson, 1991; Thorkelson, et al., in review), and the structurally lowest (oldest) Stikinia rocks may include a component of quartz-rich strata (Chapter 2).

The metamorphosed continental margin-like assemblages with Precambrian detrital zircons in Yukon and in the Coast Mountains of British Columbia and southeastern Alaska have been included in Tracy Arm terrane. These assemblages form a semi-continuous belt that is exposed along the western flank of the Coast

Mountains, between Bella Coola (~53°N) and Skagway (~59°N), and east of the Coast Belt from the Tagish Lake area to western Yukon (Figs. 2.1 and 2.6a; Coney et al., 1980; Monger and Berg, 1984). They are described below, from north to south.

Southwest Yukon (Aishihik Lake area)

An unfortunate outcome of the terrane terminology as it stands, is the uncertainty of the terrane affiliation of metamorphic rocks in the Nisling Range of central Yukon (Fig. 2.6; they have been considered part of Tracy Arm terrane by Coney et al. (1980), part of Nisling terrane by Wheeler et al. (1991), and have been correlated with the Nisutlin subterrane by Mortensen (1992) and Johnston and Timmerman (1994a). It has also been suggested that they may be in part equivalent to Paleozoic Stikinia (Currie, 1992b; Chapter 3; Fig. 2.6). They have been referred to as the Nisling assemblage by Johnston (1993), the Aishihik assemblage by Erdmer (1991), and more recently, they have been termed the Aishihik metamorphic suite by Johnston and Timmerman (1994a).

Johnston and Timmerman (1994a) divided the Aishihik metamorphic suite into "lower and upper packages" (see also Johnston and Timmerman, 1994b and 1994c), which correspond with the "lower and middle units" of Yukon-Tanana terrane described by Mortensen (1992); no rocks corresponding to the "upper unit" of Yukon-Tanana terrane have been recognised in Aishihik Lake area (Mortensen, 1992). The "upper package" includes metabasite, metagabbro, metafelsite, an Early Mississippian two-mica granitic gneiss, and hornblende diorite gneiss. It has been correlated with the

Nasina assemblage by Mortensen (1992) and Johnston and Timmerman (1994a). The "upper package" is in sheared intrusive contact with an Early Jurassic calc-alkaline pluton of the Aishihik plutonic suite (Aishihik batholith). The "lower package" comprises quartzofeldspathic-micaschist, marble, quartzite, metachert, and rare metabasite. It has been correlated with the Nisling assemblage of Wheeler et al. (1991) by Mortensen (1992), and Johnston and Timmerman (1994a), and is herein considered part of Tracy Arm terrane. It is referred to as the "Nisling assemblage" because it is exposed in the Nisling Range, and conforms to the original definition of the Nisling assemblage (Wheeler et al., 1991). The term "Aishihik assemblage" is retained for the "upper package", which is interpreted as the northern extension of Stikinia (Currie, 1992b; Chapter 3). It follows from this interpretation that the Stikinia-Tracy Arm terrane boundary is exposed in the Aishihik Lake area (Chapter 5).

Juneau

South of Juneau, Alaska, ~75 km west and ≥ 100 km southwest of the Tagish Lake area, metamorphic rocks of the Tracy Arm, Endicott Arm, and Port Houghton assemblages are preserved as pendants in, and along, the western margin of Late Cretaceous to Paleocene granitoids of the Coast Belt (Gehrels et al., 1992a; Fig. 2.6). All of these have been correlated with Yukon-Tanana terrane by Gehrels et al. (1990, 1992a), but only the Tracy Arm assemblage is considered equivalent to Tracy Arm terrane (this chapter). The Endicott Arm assemblage more closely resembles Paleozoic rocks of Stikinia (Chapter 3). The terrane affiliation of the Port Houghton assemblage

is unclear because it includes conglomerates with metarhyolite and metabasalt clasts, a rock-type that has not been observed elsewhere in Tracy Arm terrane, and is more likely associated with an arc assemblage. The Tracy Arm assemblage comprises amphibolite facies quartz-rich metasedimentary rocks, and subordinate metapelite, metabasalt, metarhyolite, and marble that are characterized by an evolved isotopic signature ($\epsilon_{\text{Nd}}(0)$ values are from -26 to 0, and T_{DM} are from 850 to 2450 Ma; Samson et al., 1991a). U-Pb ages for detrital zircons from the Tracy Arm assemblage are similar to those of detrital zircons from the western North American miogeocline, the western Canadian shield and Yukon-Tanana terrane in eastern Alaska, and Yukon (Gehrels et al., 1991a). Thus, only the Tracy Arm assemblage strongly resembles the Florence Range suite and is regarded as part of Tracy Arm terrane.

Ketchikan, Alaska

South of the Juneau area, near Ketchikan, Alaska, lower Paleozoic meta-silicic tuff, quartzite, marble, metabasalt, calc-silicate, pelitic schist, silicious quartzofeldspathic schist and orthogneiss of the Kah Shakes sequence crop out (Rubin and Saleeby, 1992). These rocks were included in both Tracy Arm terrane and Taku terrane by Berg et al. (1988). Mid-Cretaceous thrusts have imbricated the Kah Shakes sequence with Alexander terrane, Taku terrane, and upper Jurassic to lower Cretaceous Gravina sequence rocks (Rubin and Saleeby, 1992). The Kah Shakes sequence is tentatively considered part of Tracy Arm terrane because it is lithologically similar to the Florence Range suite.

Bella Coola area

Schist, marble, metapelite and quartzite in the Bella Coola area are separated from fossiliferous Mesozoic volcanic and sedimentary rocks of Stikinia to the east by the Menzies Point shear zone (Rusmore et al., 1992). ϵ_{Nd} values and $^{87}Sr/^{86}Sr$ ratios for these rocks are variable (-10.3 to +6.6, and 0.7042 to 0.7193, respectively), and preliminary single grain detrital zircon U-Pb ages are Late Archean and Early Proterozoic in age (Boghossian et al., 1993). It is not clear whether these rocks are related to the Paleozoic Stikine assemblage or Tracy Arm terrane. However, they were included in Tracy Arm terrane and correlated with the "lower" Yukon-Tanana terrane by Gehrels et al. (1992b) and considered part of the lower Stikine assemblage by Boghossian et al. (1993). These rocks are bounded along their western margin by a ~83 Ma granodiorite, west of which there are lower grade rocks, including a granitoid clast conglomerate with late Silurian detrital zircons, siliciclastics, marble, and metapelite that are considered part of Alexander terrane. Farther west, a second sliver of metasedimentary rocks is preserved between the Alexander terrane and Wrangellia (Boghossian et al., 1993). This western sliver of supracrustal rocks has been interpreted as either an exotic crustal fragment (Tracy Arm terrane?), or part of the quartz-rich assemblage exposed to the east (Paleozoic Stikinia) that has been imbricated with the adjacent Alexander and Wrangellia terrane rocks by strike slip faults (Boghossian et al., 1993). Both interpretations are considered reasonable.

THE AGE OF TRACY ARM TERRANE

Characterization of Tracy Arm terrane would be greatly improved by a knowledge of its age, but unfortunately there are no known fossils or reliable isotopic age data that reflect the age of deposition for any of the assemblages that comprise Tracy Arm terrane, including the Florence Range suite, Nisling assemblage, Tracy Arm assemblage, Kah Shakes sequence, and the metamorphic rocks exposed near Bella Coola. If fossils existed in the protolith of the Florence Range suite, they have likely been altered beyond recognition by multiple phases of deformation and amphibolite facies metamorphism (Chapter 4). The minimum age limit for the Florence Range suite indicates that the Tracy Arm terrane is at least in part older than 176 Ma, and possibly 228 Ma (Chapter 4). Because detrital zircon ages for the metaclastic rocks provide information about the minimum age, or average age of provenance material, they only provide a maximum age for the metasediments. Projected ages of discordant analyses of single zircon grains from west of the Coast Belt have been interpreted to suggest that the sources of quartzites from the Tracy Arm assemblage in western margin of the Coast belt were 1.05-1.40 Ga, 1.75-2.00 Ga, ~2.3 Ga, 2.5-2.7 Ga, and ~3 Ga (Gehrels et al., 1991a). On the basis of the maximum age of the youngest projected detrital zircon ages from the Tracy Arm assemblage (~1.1 Ga), the Tracy Arm terrane is tentatively considered, at least in part, younger than 1.1 Ga. Rb-Sr whole rock ages are in agreement with this interpretation (Fig. 2.5).

Nowhere has the depositional basement to Tracy Arm terrane been preserved; either the base of the supracrustal section is not exposed, or the structurally lowest

preserved supracrustal rocks are in fault contact with underlying rocks. If Tracy Arm Terrane includes rocks that are Precambrian in age, the only exposed rocks in the Coast Mountains that could conceivably form their depositional crystalline basement are those of the Yellow Aster Complex, which is exposed in the Cascade Mountains of northwestern Washington, and comprises Precambrian to Lower Proterozoic "gabbroic to trondhjemitic rocks and their metamorphic derivatives" (Misch, 1977; Mattinson, 1972).

POSSIBLE TRACY ARM TERRANE EQUIVALENTS IN MAINLAND ALASKA

Tracy Arm terrane is generally exposed east of Stikinia, Cache Creek terrane, Quesnellia and parautochthonous terranes, and west of Wrangellia and Alexander terrane, and terranes accreted to the western margin of North America during the Tertiary. It is not correlated with parautochthonous terranes in Alaska, such as the Arctic Alaska, Coldfoot, Hammond, and Ruby terranes. Some of these terranes contain a significant component of metavolcanic rocks (in part Devonian in age), and occupy a different location relative to other terranes. Nor is it correlated with Yukon-Tanana terrane (shown as Nisutlin subterrane of Kootenay terrane in Fig. 2.1), for reasons outlined above. Major terranes in mainland Alaska that are considered allochthonous, and that have continental margin-like Precambrian and Paleozoic rocks are the Nixon Fork, Kilbuck and Dillinger terranes (Churkin and Whitney, 1983; Silberling et al., 1984; Fig. 2.1); these terranes are considered potential equivalents of Tracy Arm terrane.

The Dillinger terrane occurs in southwestern Alaska, south of the Denali fault (>800 km away from the Florence Range suite), and south of part of Yukon-Tanana terrane (Nisutlin subterrane), and north of terranes that were accreted to North America during the Cenozoic. It comprises a complexly folded and faulted sequence of lower Paleozoic graptolitic shale, basinal carbonate rocks, turbiditic sandstone, shale and grit, unconformably overlain by Early Jurassic fossiliferous calcareous sandstone and siltstone. The unmetamorphosed state of the Early Jurassic sediments indicates that unlike the Florence Range suite, the Dillinger terrane was not metamorphosed during mid-Jurassic time. It can only be tentatively correlated with Tracy Arm terrane on the basis of lithologic similarity because it has not been characterized isotopically, and no detrital zircon ages have been reported from this terrane.

The Nixon Fork terrane is exposed north of the Denali fault, and northwest of the Nisutlin subterrane (Yukon-Tanana terrane). It is composed of a stratigraphic sequence of Paleozoic reefal carbonate rocks (Ordovician to Devonian), deposited on Precambrian basement and overlain by Permian, Triassic and Cretaceous fossiliferous sedimentary rocks (Silberling et al., 1984). The basement rocks are mainly pelitic and calc-schists with minor marble, quartzite, felsic metavolcanic rocks. Churkin et al. (1984) interpret this terrane as a rifted continental margin sequence derived from Western North America (Canadian Cordillera) because the transgressions and regressions are coeval with those of the North American stratigraphic record.

The Kilbuck terrane comprises gneiss and schist exposed in the Kanektok River region, including biotite-hornblende gneiss, garnet amphibolite, quartz-mica schist,

marble and ca. 2.0 Ga orthogneisses (Box et al., 1990; Kanektok metamorphic complex of Hoare and Coonrad, 1979). Radiogenic ages for these rocks are indicative of Precambrian metamorphism (Hoare and Coonrad, 1979). A possible outlier of the Kilbuck terrane is the Idono complex of Miller and Bundtzen (1985), which is exposed to the west of the Nixon Fork terrane, and has yielded zircons with U-Pb ages of 2.03 Ma. The Idono complex has also been included in Nixon Fork terrane (Howell et al., 1985). If Tracy Arm terrane metasedimentary rocks are related to those of the Kilbuck terrane in Alaska (Fig. 2.1), the ca. 2.0 Ga orthogneisses in the Kilbuck terrane (Box et al., 1990) could represent the depositional basement of Tracy Arm terrane.

CONCLUSIONS

Despite the structural and metamorphic overprints, which largely conceal the original stratigraphy of the Florence Range suite, the depositional setting of this suite can be regarded as an Early Jurassic, or older, stable Atlantic-type continental margin. It retains sufficient protolith characteristics to distinguish it from adjacent terranes. It is included in Tracy Arm terrane, rather than Nisling terrane, because (1) they are lithologically distinct from some of the metamorphic rocks exposed in the Nisling Range, for which Nisling terrane was named, and (2) they have not been unequivocally correlated with the quartz-rich Nisling terrane metasedimentary rocks exposed in central Yukon. Because the Florence Range suite is the best exposed, most extensive, and least disrupted by Cretaceous and younger deformation and plutonism,

of all of the assemblages interpreted as part of Tracy Arm terrane, it should be considered the type assemblage for Tracy Arm terrane.

Before the entire tectonic history of Tracy Arm terrane can be resolved and definitive correlations made, the depositional age of the Tracy Arm terrane strata, and the age(s) of structures and metamorphic events in Tracy Arm terrane that predate its accretion to the western margin of Stikinia must be determined, and the offsets on large structures, such as the Llewellyn fault zone, need to be constrained. Correlations between the Florence Range suite and other possible Tracy Arm terrane assemblages could be strengthened by (1) additional concordant detrital zircon ages for all quartz-rich assemblages, (2) isotopic characterization of single detrital grains, such as ϵ_{Nd} values for monazite, and Hf isotopic compositions for zircon grains, and (3) determination of crystallization ages, the model ages for derivation from the mantle (T_{DM}), and isotopic and geochemical characters of the metabasites.

Lithologic comparison indicates that Tracy Arm terrane includes the Florence Range suite, Tracy Arm assemblage, Kah Shakes sequence, and quartz-rich metasedimentary rocks near Bella Coola, in southeastern Alaska, and Nisling assemblage in Yukon. These assemblages form a semi-continuous north-northeast trending belt in the northwestern Cordillera, that in general lies to the east of Stikinia, and west of Alexander and Wrangellia terranes, and terranes accreted to the western margin of North America during the Tertiary. Tracy Arm terrane may be related to the Dillinger, Nixon Fork, Kilbuck, and Idono terranes of mainland Alaska. Like Tracy

Arm terrane, the origin of these terranes is a mystery because they are separated from autochthonous continental margin rocks by oceanic and arc-like terranes (Fig. 2.1).

Figure 2.1. Terrane map of British Columbia, Yukon and Alaska (modified from Plafker, 1990) showing the location of Tracy Arm terrane (TA) and the Tagish Lake area. Van = Vancouver; Anc = Anchorage. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio contours are shown in the inset (modified from Armstrong, 1988).

Figure 2.2. Geological map of the Tagish Lake area showing locations of sections A to C, and large-scale structures (modified from Mihalynuk et al., 1989a, b; 1990; Currie, 1990a). CF = Caplice fault; STTB = Stikinia-Tracy Arm terrane boundary; WCFZ = Willison Creek fault zone; WS = Willison synform; WF = Wann fault.

Figure 2.3. Geology map of Florence Range suite showing marble layers in the Florence Range suite that outline medium-scale structures. CF = Caplice fault; WCFZ Willison Creek fault zone.

Figure 2.4. Lithologic sections for the Mt. Caplice (A-A'), Willison Glacier (B-B') and Willison Bay (C-C') areas. Section locations are shown in Fig. 2.2.

Mt Caplice

Willison Glacier

Willison Bay

Figure 2.5. Rb-Sr isochron plot for samples from the Florence Range suite and Aishihik assemblage. Data are presented in Table 2.1. Uncertainty for $^{87}\text{Rb}/^{86}\text{Sr}$ values is 2% and for $^{87}\text{Sr}/^{86}\text{Sr}$ is ± 0.0001 (see Armstrong et al., 1991 for analytical methods). Fields for autochthonous North American rocks (Paleozoic, Windermere Supergroup and Purcell Supergroup rocks) are from Armstrong et al. (1991).

Figure 2.6. Terrane maps of northwestern British Columbia, southeastern Alaska and Yukon showing the locations of, and terminology for continental margin-like assemblages discussed in text (modified from Wheeler and McFeely, 1991). Figures 2.6a and b contrast the geographic extents of rocks included in Tracy Arm terrane (modified from Coney et al., 1980) and Nisling Terrane (Wheeler et al., 1991), respectively. FFZ = Finlayson Fault Zone; L. Fault = Llewellyn Fault zone. Numbers in parentheses refer to references: (1) = Mortensen (1990a; 1992); (2) = Wheeler et al. (1991); (3) = Coney et al., 1980; (4) = this chapter; (5) = Johnston and Timmerman (1994a, b, c); (6) = Gehrels et al., (1992a); (7) = Rubin and Saleeby (1991; 1992). Terrane and assemblage names used in this thesis are shown in bold.

TERRANES AND PLUTONS

- (V) CRETACEOUS AND TERTIARY VOLCANICS
- (CB) CRETACEOUS AND TERTIARY PLUTONS
- (+) LATE TRIASSIC TO MIDDLE JURASSIC PLUTONS

- STIKINIA
- DEVONIAN GNEISS
- UNDIVIDED METAMORPHIC ROCKS
- AX ALEXANDER
- CC CACHE CREEK
- CY CHUGACH AND YAKUTAT

- (D) DORSEY
- (GN) GRAVINA-NUTZOTIN
- (NA) ANCESTRAL NORTH AMERICA
- (Q) QUESNELLIA
- (SM) SLIDE MOUNTAIN
- (TU) TAKU
- (WR) WRANGELLIA
- (WM) WINDY-MCKINLEY

- Plate 2.1a. Quartzite and schist layers of the Florence Range suite from Section A-A'. These metasedimentary layers are facies equivalents of a ~400 m thick quartzite layer exposed ~4.5 km to the south, along the Llewellyn Glacier (Fig. 2.3). Pen is 15 cm long.
- Plate 2.1b. Quartzite of the Florence Range suite from Section A-A' (Figs. 2.2; 2.4). This quartzite exhibits colour-banding. Hammer head is 16 cm long.
- Plate 2.1c. Crenulated muscovite schist of the Florence Range suite from the Wann River valley. Hammer head is 16 cm long.
- Plate 2.1d. Semipelitic rock of the Florence Range suite with continuous cm-scale trondhjemitic leucosome layers. Pen is 15 cm long.
- Plate 2.1e. White marble of the Florence Range suite with dark grey graphitic laminae, from the toe of the Mt. Caplice glacier. The graphitic laminae are interpreted as bedding surfaces, and have been folded by upright open folds with shallowly plunging fold axes. Book is 20.5 cm long.

Plate 2.2a. Variably weathering calc-silicate rocks of the Florence Range suite from near the base of the Willison Creek fault zone hangingwall, in the Willison Creek valley. The more resistant layers are quartz rich and the recessive layers are calcite-rich and may also contain diopside or grossular garnet.

Plate 2.2b. Folded garnet-bearing calc-silicate and amphibolite of the Florence Range suite from east of the Willison Creek fault zone, near the Willison Creek headwaters.

Plate 2.2c. View northwest across the Willison Creek headwaters to marbles and schistose rocks of the Florence Range suite exposed in the Willison Glacier section (see Section B-B' in Figs. 2.2; 2.4). The metasedimentary layers are truncated by the Willison Creek fault zone (see Fig. 2.3; Plate 2.2d).

Plate 2.2d. Sketch of Plate 2.2c.

b

a

p

c

Plate 2.3a. Marble of the Florence Range suite with a possible quartzite clast, from near the base of the Willison Creek fault zone hangingwall, in the Willison Creek valley.

Plate 2.3b. View north to ridge above Willison Glacier (Section C-C'; Figs. 2.2 and 2.3)

Plate 2.3c. Outcrop of graphitic quartzite of the Florence Range suite from the Wann River valley. Black laminae represent graphite-rich layers (see Plates 2.3d; 2.3e). Black quartzite has only been observed east of the Willison Creek fault zone.

Plate 2.3d. Hand sample of graphitic quartzite of the Florence Range suite from the Wann River valley.

Plate 2.3e. Photomicrograph of fine-grained black quartzite of the Florence Range suite from the Wann River valley. Grain size appears to increase with decreased graphite content. Field of view is 3 mm wide (Sample LC-91-449).

Plate 2.3f. Deformed carbonate pebble conglomerate of the Florence Range suite from east of the Willison Creek fault zone. Hammer head is 16 cm long.

TABLE 2.1 $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the Florence Range suite in the Tagish Lake area of Northern British Columbia and the Aishihik assemblage in Yukon.

Sample	Lithologic Unit	Easting	Northing	Sr (ppm)	Rb (ppm)	$\frac{^{87}\text{Rb}}{^{86}\text{Sr}}$	$\frac{^{87}\text{Sr}}{^{86}\text{Sr}}$	Age (Ma)	Age ¹
METAPELITIC ROCKS²									
LW76-a8B	Muscovite schist	561648	6547391	32.6	192	17.1	0.7400		
LW76-a40C	Quartz-feldspar muscovite schist	542087	6570142	224	117	1.51	0.7319		
LW76-a57A	Muscovite schist	547690	6547563	186	89.6	1.393	0.7223		
LW76-a71C	Muscovite schist	544227	6557730	97.3	143	4.27	0.7392		
LW76-a72B	Quartz-muscovite schist	546026	6558680	61.2	156	7.40	0.7464		
LW76-b70A	Muscovite schist	544051	6556429	150	120	2.33	0.7357		
LW76-c52C	Muscovite schist	546089	6553483	154	163	3.061	0.7369		
AMPHIBOLITES²									
LW76-a41D	Amphibolite	541162	6567904	334	35.6	0.025	0.7057	214 ± 8 ³	
LW76-a69C	Biotite amphibolite	543377	6556978	425	27.7	0.189	0.7045		
LW76-b17A	Hornblende biotite amphibolite	535790	6562839	436	5.2	0.034	0.7044		
LW76-c34A	Amphibolite schist	542071	6571627	35.6	1.0	0.082	0.7240	244 ± 16 ³	
LW77-b42A	Migmatitic biotite-hornblende amphibolite	541617	6561042	341	43.9	0.373	0.7061	144 ± 5 ³	
LW77-c25B	Amphibolite	541216	6562894	748	30.1	0.116	0.7053		
AISHIHIK ASSEMBLAGE⁴									
89-AT-163	Schist (εNd = -20.0)	471900	6671500	67.6	86.3	3.71	0.7448		
89-AT-155	Gneiss (εNd = -15.6)	491400	6674500	186	93.9	1.46	0.7156		

¹ decay constants are from Steiger and Jager (1977); ² unpublished data collected at University of British Columbia, 1970-1985 (R.L. Armstrong, pers comm., 1991); ³ K-Ar hornblende age; ⁴Jackson et al., 1992.

CHAPTER 3: PALEOZOIC AND MESOZOIC ROCKS OF STIKINIA EXPOSED IN THE NORTHEASTERN COAST MOUNTAINS, BRITISH COLUMBIA

ABSTRACT

Stikinia is a tectonostratigraphic terrane in the Canadian Cordillera that formed in an arc environment during the Paleozoic and Mesozoic; it is composed of volcanic rocks and associated plutonic and sedimentary rocks, with the Paleozoic rocks (the Stikine assemblage) forming a semicontinuous belt along its western margin. Previously, the Paleozoic rocks have only been recognised in a restricted area in northern British Columbia (between the Stikine and Taku river areas), but a metamorphosed Paleozoic volcanic assemblage exposed immediately south of the Yukon-British Columbia border, referred to as the Boundary Ranges suite, is herein correlated with the Stikine assemblage and regarded as part of Stikinia on the basis of similar rock-types, lithologic associations, and new U-Pb zircon ages. The Boundary Ranges suite comprises: metapelitic rocks, felsic and mafic metavolcanic rocks, and marble that are in part Mississippian or older; Devono-Mississippian gneisses, including the weakly foliated Bighorn Creek granite (366 ± 9 Ma) and two Mississippian dykes (335 ± 10 Ma; 352 ± 5.5 Ma); and a metamorphosed Permian volcanic rock (Wann River gneiss; 270 ± 5 Ma). Like the Stikine assemblage, the Boundary Ranges suite is intruded by plutons of Early to Middle Jurassic age; these plutons are included in the Tagish Lake suite, which comprises the Mt. Caplice quartz monzodiorite (180.9 ± 3.1 Ma), Hale Mountain granodiorite (186 ± 1.5 Ma), Hoboe quartz syenite (189.5 ± 3 Ma) and coeval dykes, including a 191 ± 1 Ma tonalite dyke

that intruded the Wann River gneiss. The recognition of the Boundary Ranges suite and Tagish Lake suite as part of Stikinia is important because (1) proposed correlations with terranes in Yukon (parts of Yukon-Tanana terrane and/or Nisutlin Subterrane of Kootenay terrane), southeast Alaska (parts of Taku terrane) and west-central British Columbia (undivided metamorphic rocks) imply that Stikinia's Paleozoic basement may be more widespread than previously thought; and (2) it has implications for the Tracy Arm-Stikinia terrane boundary because these rocks form the footwall of a major shear zone that carries Tracy Arm terrane in its hanging wall.

INTRODUCTION

Metamorphic rocks in the Tagish Lake area of northwestern British Columbia are exposed along the western edge of arc-like Stikinia (e.g. Wheeler et al., 1991), the largest of the allochthonous terranes accreted to the western margin of North America during the Mesozoic (e.g. Coney et al., 1980; Fig. 3.1). Until recently, all of the layered metamorphic rocks in the Tagish Lake area were included in Tracy Arm Terrane, a "metamorphosed Proterozoic to lower Paleozoic (?) passive continental margin assemblage and partly metamorphosed carbonaceous and siliceous offshore sediments" (Wheeler et al., 1988, 1991), but detailed mapping has shown that many of the metamorphic rocks have volcanic or plutonic protoliths (Mihalynuk and Rouse, 1988a, b; Mihalynuk et al., 1989a; Currie, 1990a; 1991; 1992b). This metamorphosed, primarily volcanic assemblage is considered distinct from Tracy Arm terrane because it is lithologically dissimilar to Tracy Arm terrane, is separated from Tracy Arm

terrane by a major shear zone of mid-Jurassic age, and shares no stratigraphic, intrusive or structural links with Tracy Arm terrane until after ca. 180 Ma (Chapter 1). The description, relative apparent stratigraphic relations, U-Pb zircon and titanite ages, and inferred protoliths of the metamorphosed volcanic assemblage are presented in this chapter. On the basis of these observations and interpretations, it is argued that the volcanic assemblage is best correlated with similar rocks in Stikinia.

GEOLOGY OF THE TAGISH LAKE AREA

The Tagish Lake area is located in northwestern British Columbia along the Stikinia-Tracy Arm terrane boundary (Fig. 3.1). Allochthonous and parautochthonous terranes separate Stikinia from autochthonous rocks of North America to the east, and to the west the metamorphic rocks of the Tagish Lake area are intruded by Cretaceous to Tertiary plutonic rocks of the Coast Belt. West of the Coast Belt, Taku, Wrangellia and Alexander terranes, and the overlying Gravina-Nutzotin overlap assemblage are exposed (Fig. 3.1).

In the Tagish Lake area, unmetamorphosed Mesozoic volcanic and sedimentary rocks of Stikinia are well exposed east of the Llewellyn fault zone, a north-northwest striking, steep fault zone that was active after ~100 Ma, and possibly earlier (Appendix A; Figs. 3.1; 3.2). Greenschist to amphibolite facies, polydeformed metamorphic rocks occur west of the Llewellyn fault zone and are also exposed east of the fault zone, south of Mt. Mussen (Fig. 3.2). Rocks on both sides of the fault zone are overlain by

Tertiary volcanic rocks and intruded by Cretaceous and Tertiary, undeformed plutonic rocks of the Coast Belt (Figs. 3.2; 3.3).

Stratigraphic characterization of the metamorphic rocks in the Tagish Lake area is hampered by metamorphism and polyphase deformation of original contact relations, the absence of way-up indicators, and a lack of age control due to the absence of fossils and, until now, reliable isotopic dates. The age of the metamorphic rocks has been loosely constrained by several known geologic relations. An undeformed Cretaceous pluton that intruded the metamorphic rocks has been dated at 127.2 ± 0.6 Ma, providing a lower limit for the age of the metamorphic rocks (Appendix A), and the metamorphic fabric associated with the 184 to 176 Ma Wann River shear zone that deforms the metamorphic rocks supports an even lower limit of 184 Ma for the age of these rocks (Chapters 1; 4; Currie, 1992b).

The metamorphic rocks were first mapped by Christie (1957) who divided them into two units: one characterized by micaceous quartzite and hornblende-quartz-feldspar gneiss, with amphibolite, mica schist and "limestone" (unit 1a); the other (unit 1b) was defined as metamorphic rocks "generally bearing chlorite and finer grained than those of unit 1a." Bultman (1979) and Werner (1977, 1978) mapped the metamorphic rocks as "coast metamorphic rocks". Wheeler et al. (1991) included them in Tracy Arm Terrane due to their apparent similarity with schist, quartzite and limestone in the type Tracy Arm area of western Yukon. Recent detailed mapping in the Tagish Lake area (Mihalynuk et al., 1989b; 1990; Currie, 1990a; 1991; 1992b) has shown that the metamorphic rocks include two distinct lithologic assemblages that are

juxtaposed along a sheared and imbricated boundary (Currie, 1990a; 1991; Fig. 3.2), but this subdivision differs from those of Christie (1957) and Wheeler et al (1991). The hanging wall is dominated by well exposed metaclastic and metacarbonate rocks that are considered "bona fide" Tracy Arm terrane and are termed the Florence Range metamorphic suite (Currie, 1990a; Mihalynuk et al., 1990); as in Chapter 2, they are referred to as the Florence Range suite for brevity (Fig. 3.2). The footwall of the Wann River shear zone is primarily composed of polydeformed upper greenschist to amphibolite facies Paleozoic metasedimentary and metaigneous rocks, referred to as the Boundary Ranges metamorphic suite (hereafter referred to as the Boundary Ranges suite) , and deformed Mesozoic plutonic rocks of the Tagish Lake suite (Currie, 1990a; 1992b). These footwall rocks are the focus of this chapter.

The contact between these two assemblages lies within the Wann River shear zone, a >4 km wide zone of rock, that has been altered by mid-Jurassic, shear-related fabrics that are interpreted as having formed during the accretion of Tracy Arm terrane to the western margin of Stikinia (Chapters 1; 4).

Lithologic units

Boundary Ranges suite

The Boundary Ranges suite, defined by Mihalynuk and Rouse (1988a; originally referred to as the Boundary Ranges metamorphic suite), extends over 130 km, from the Taku River area to the British Columbia-Yukon border (Fig. 3.1), and consists of polydeformed biotite schist, muscovite schist, graphitic schist and phyllite,

chlorite schist, chlorite-actinolite schist, actinolite schist, hornblende-plagioclase schist, biotite-plagioclase-quartz schist, fine-grained felsic orthogneiss that is commonly porphyritic, and marble (Table 3.1; Fig. 3.3; Mihalynuk and Rouse, 1988b; Mihalynuk et al., 1989a; b; 1990; Currie, 1990a; 1991; Smith and Mihalynuk, 1992). These rocks are interpreted as greenschist to amphibolite facies metamorphosed siltstones, basalts, intermediate pyroclastic rocks, felsic tuffs and intrusive rocks, and carbonate rocks (Mihalynuk and Rouse, 1988a, b).

Previously, the ages of Boundary Ranges suite rocks were poorly constrained because (1) these rocks are isolated from formerly contiguous rocks by faults and plutons; (2) polyphase deformation (Plate 3.1a; c.f. Mihalynuk et al., 1989a) and metamorphism have resulted in poor fossil preservation (all carbonate rock samples processed for microfossils were barren; M.J. Orchard, pers. comm., 1990; 1991), and (3) no reliable isotopic ages were available. The age of this suite could only be limited to older than ~176 Ma, the age of the metamorphic fabric that deforms them (Chapter 1; 4). U-Pb zircon ages for felsic orthogneisses presented here are Devonian-Mississippian in age (335 ± 10 Ma; 352 ± 5.5 Ma; 366 ± 9 Ma; Appendix 3.1). One of these orthogneisses is a 335 ± 10 Ma dyke that cross-cuts layering in actinolite schist, limiting the age of some of the layered rocks to 325 Ma or older (Plate 3.1b; Appendix 3.1). The youngest dated lithologic unit in the Boundary Ranges suite is the Permian Wann River gneiss (270 ± 5 Ma; Appendix 3.1), which is considered to be a metamorphosed volcanoclastic rock on the basis of its graded layering, composition, and local association with metapelitic and metacarbonate rocks (Table 3.1; Plate 3.1c).

The Boundary Ranges suite was previously recognised only as far north as the British Columbia-Yukon border (Mihalynuk and Rouse, 1988), but a ~320 Ma U-Pb zircon age for a felsic dyke that intruded actinolite schist near the B.C.-Yukon border suggests that the Boundary Ranges suite extends into Yukon (C.Hart and J.K. Mortensen pers. comm., 1993).

Nd and/or Sr isotopic data for this suite are limited to four samples: one quartzite and two schistose rocks have $\epsilon_{Nd(0)}$ values from +0.5 to -4.5 and $^{87}Sr/^{86}Sr$ ratios from 0.7060 to 0.7123 (Jackson, 1992; $^{87}Sr/^{86}Sr_i$ calculated for 360 Ma are from 0.7047 to 0.7078, and may be lower if the rocks are significantly older); and a sample of Wann River gneiss from west of Hoboe Glacier has an initial $^{87}Sr/^{86}Sr_i$ ratio of 0.7062 (calculated at 270 Ma; Appendix 3.2, unpublished data, R.L. Armstrong, pers. comm., 1991). On the basis of this limited data set the Boundary Ranges suite appears to be in part characterized by relatively unevolved isotopic signatures.

Early Jurassic Tagish Lake Suite Plutonic Rocks

The Early Jurassic Tagish Lake suite is a strongly differentiated intermediate to mafic, alkaline to calc-alkaline plutonic suite. All plutons are hornblende-bearing, commonly contain epidote and titanite, and may contain biotite (Table 3.2). Those recognised in the Tagish Lake area are variably deformed. They include the Hale Mountain granodiorite (A in Fig. 3.2; 186 ± 1.5 Ma; Plate 3.1d), Mt. Caplice quartz monzodiorite (B in Fig. 3.2; 180.9 ± 3.1 Ma), and Hoboe quartz syenite (C in Fig. 3.2; 189.5 ± 3 Ma). The Tagish Lake suite also includes hornblendite ($\pm$ pyroxenite) bodies,

one of which has a hornblende K-Ar age of 187 ± 7 Ma (M.G. Mihalynuk, pers. comm. 1993), and coeval dykes, including one that intruded the Wann River gneiss and was dated at 191 ± 1 Ma (Fig. 3.2; Plate 3.2a; Appendix 3.1).

Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7049 for a sample of Mt. Caplice quartz monzodiorite, and 0.7052 for a sample of Hale Mountain granodiorite are relatively low and do not reflect the presence of an evolved source (Appendix 3.2, unpublished data, R.L. Armstrong, pers. comm. 1991).

Cross-Sections and Stratigraphic Sections

Nowhere is there a complete section of the Boundary Ranges suite exposed, but by linking four lithologic sections from different areas, it is possible to construct a composite section of the mappable lithologic units that comprise the footwall of the Wann River shear zone. Contact relations and relative positions of lithologic units are documented in Figure 3.3 in five cross sections: Mt. Lawson (V-V'); White Moose Mountain to Wann River (W-W'); Mt. Switzer (X-X'); Mt. Caplice (Y-Y'); and west of Hoboe Glacier (Z-Z'; see Fig. 3.2 for the locations of these sections). The lithologic sections of Figure 3.4 are derived from, and facilitate comparison between these cross-sections. It is reasonable to treat these sections (Fig. 3.4) as structurally transposed "stratigraphic sections" because the first order lithic components, their abundances, and mutual relations are consistent and distinctive, even though their original depositional thicknesses have been modified by metamorphism and multiple phases of deformation. It is possible to compare the composite section for the Tagish Lake area Boundary

Ranges suite with stratigraphic sections considered representative of other areas (terrane) because the ages of some units are constrained by the U-Pb age data presented in Appendix 3.1.

Mt. Lawson (Section V-V')

Some of the structurally lowest rocks in the Tagish Lake area are exposed on Mt. Lawson, because southeast of Mt. Lawson, on Hale Mountain, lithologic contacts dip slightly to the southeast (Fig. 3.5) allowing structurally lower rocks to be exposed in the Mt. Lawson area than are exposed to the south (Figs. 3.3; 3.4). On Mt. Lawson the steep, west-side-down Chicken Creek fault juxtaposes moderately eastward-dipping amphibolite and greenschist facies metamorphic rocks to the east against similar rocks that form a close, upright antiform to the west (Figs. 3.2; 3.3). Folding about the antiform is reflected in the variation in the orientation of layering and foliation west of the fault (Fig. 3.5; the interlimb angle of the antiform appears larger than the true interlimb angle because section V-V' has been constructed at an oblique angle to the fold axis). The orientation of the cross-section in Figure 3.3a was chosen to include the structurally lowest rocks and show the relation between the Hale Mountain granodiorite with the underlying rocks. The metamorphic rocks are intruded by Cretaceous, or younger, granite of the Coast Belt to the north, south and west (Figs. 3.2; 3.3; Mihalynuk et al., 1989b; Currie 1992b).

Metamorphic rocks west of the Chicken Creek fault include biotite schist, muscovite schist, well-layered felsic orthogneiss, minor carbonate rock layers (1-5 m

thick) and a distinctive garnet-biotite schist with 1 to 2 cm diameter garnets that have inclusion trails that are at an angle to schistosity (LC-90-286b; Table 3.1). The schistose rocks are overlain by a ~200 m thick felsic orthogneiss with well-developed layering, which is in turn overlain by muscovite and biotite schist with minor felsic gneiss. Interlayered with the uppermost biotite schist layers are 5 cm to 2 m thick layers of actinolite schist, and pyroxene- and/or hornblende-phyrlic actinolite schist (Plate 3.2b). The actinolite schist is almost everywhere interlayered with mm- to cm-scale layers of feldspar-rich leucocratic layers (Plate 3.2c). Both the felsic and the mafic layers are continuous over many metres. A variably foliated hornblende granodiorite, similar in composition and texture to the Hale Mountain granodiorite of the Tagish Lake suite (TLS in Fig. 3.3), intrudes the uppermost actinolite schists and is the structurally highest deformed rock west of the Chicken Creek fault.

Similar lithologic units are exposed east of the Chicken Creek fault, but confident correlation of individual layers is not possible with the available data. East of the fault are 250 m of biotite schist, muscovite schist, biotite-muscovite schist and graphite schist ($\pm$ garnet, kyanite, staurolite; Table 3.1), with minor carbonate rock (0.5 to 3 m thick) and felsic gneissic layers (0.5 to 3 m thick; Figs. 3.3; 3.4), and a garnet-biotite schist layer identical to the distinctive garnet-biotite schist observed east of the Chicken Creek fault. Above these rocks is the well-layered, felsic Mt. Lawson gneiss (Plate 3.2d; Table 3.1; Currie, 1991), which may be equivalent to the well-layered felsic gneiss exposed west of the Chicken Creek fault (Fig. 3.3). The Mt. Lawson gneiss is overlain by the Devonian-Mississippian Bighorn Creek granite

(Appendix 3.1; Table 3.1; Plate 3.3a) which crops out on the southern flank of Mt. Lawson and in the Bighorn Creek valley (Figs. 3.2; 3.3; Mihalynuk et al., 1989b). In contrast, the felsic gneiss from west of the Chicken Creek fault is overlain by biotite and muscovite schists; the absence of the Bighorn Creek granite west of the Chicken Creek fault may be a function of its original geometry (Fig. 3.4).

Neither the contact between the base of the Bighorn Creek granite and the underlying Mt. Lawson gneiss, nor its contact with the overlying biotite schist have been observed. Biotite schist crops out above the Bighorn Creek granite on the south side of Bighorn Creek, (Figs. 3.3; 3.4). There is a gradual upward change in composition of the schist, from primarily biotite schist, through a 500 m-thick section of interlayered, 10 cm to 50 cm thick, biotite schist and actinolite schist layers, to a 400 m thick section consisting almost exclusively of actinolite schist with thin felsic layers, similar to that seen west of the Chicken Creek fault (Plate 3.2c). Displacement on the Chicken Creek fault is considered west-side-down and in the order of 0.5 to 1 km, on the basis of the offset of the biotite schist-actinolite schist transition, but the exact magnitude of this displacement is poorly constrained due to the indistinct nature of the biotite schist-actinolite schist contact; this fault may also be the locus of strike-slip motion.

Near the top of the Hale Mountain-White Moose Mountain ridge, 10 cm to 1 m thick layers of the Hale Mountain granodiorite (Plate 3.1d), are interlayered with the uppermost 200 m of the actinolite schist-dominated section. Uninterrupted Hale Mountain granodiorite plutonic rocks are exposed above the uppermost actinolite schist

layer, and intruded by undeformed, coarse-grained, pink and white, felsic dykes of unknown age.

In the Bighorn Creek-Hale Mountain area there is an overall upward progression from amphibolite facies pelitic rocks with well-layered felsic gneiss, to greenschist facies mafic schist (biotite schist and then actinolite schist), to the overlying Hale Mountain granodiorite of the Tagish Lake suite (Figs. 3.3; 3.4).

White Moose Mountain to Wann River (Section W-W')

The White Moose Mountain to Wann River section is constructed using two sections, both sections intersect the Florence Range suite (Tracy Arm terrane)-Wann River gneiss contact (Fig. 3.5), and the southern section also crosses an imbricate sliver of Wann River gneiss that is bounded both sides by Florence Range suite rocks (Fig. 3.2). The apparent thicknesses of lithologic units are considered approximately equal to their true lithologic thicknesses because their contacts dip shallowly on Mt. Lawson and Mt. Switzer, and because the cross section is oriented nearly perpendicular to strike near the Wann River.

Together these two sections show that the Boundary Ranges suite, the overlying Hale Mountain granodiorite, and the contact between them can be extrapolated from Hale and White Moose mountains where lithologic contacts and the dominant foliation are relatively flat-lying, across Tagish Lake to the Mt. Switzer and Wann River areas where the contacts and foliation dip moderately toward the southwest (Figs. 3.3; 3.5). Along the western and southern shores of Tagish Lake,

chlorite-actinolite and actinolite schist with minor marble and felsic orthogneiss crop out below the Hale Mountain granodiorite. Centimetre- to metre-scale interlayering along the schist-Hale Mountain granodiorite contact has been observed on the east side of Hale Mountain and the north side of Mt. Switzer; it is analogous to that seen on the north side of Hale Mountain (section V-V'; Fig. 3.3), and may also occur on the east side of Mt. Switzer, above Nelson Lake (see below). Near the top of Mt. Switzer, the Hale Mountain granodiorite is strongly deformed within discrete layers, and massive between these layers (Plate 3.1d). Locally, it is coarse-grained, with hornblende crystals up to 5 cm long surrounded by a matrix of fine-grained feldspar (see Plate 3.1-19-1 in Currie, 1990b). Near the top of Mt. Switzer, the metamorphic rocks are unconformably overlain by the Tertiary Mt. Switzer volcanic rocks (Figs. 3.2; 3.3) and intruded by Tertiary plutons (Appendix A). The normal fault exposed south of Mt. Switzer offsets the contact between the Hale Mountain granodiorite and the Wann River gneiss. It may be coeval with the Tertiary Mt. Switzer volcanic rocks because it forms the southern wall of a probable caldera structure, within which there are large blocks of metamorphic country rock, up to 10's of m in diameter, surrounded by Mt. Switzer volcanic rocks. On the south side of Mt. Switzer, the upper margin of Hale Mountain granodiorite is in sheared contact with the overlying Permian Wann River gneiss (Appendix 3.1; Plate 3.3b). Farther south, within the Wann River shear zone, the Wann River gneiss is juxtaposed against biotite schist and semipelitic rocks of the Florence Range suite. Approximately 475 m above this contact there is a ~30 m thick

tectonic sliver of Wann River gneiss bounded on both sides by Florence Range suite rocks (Figs. 3.2; 3.3).

The White Moose Mountain to Wann River section is characterized by an upward transition from Boundary Ranges suite actinolite schist, with minor felsic orthogneiss and marble, at its base, to the Hale Mountain granodiorite of the Tagish Lake suite, and the overlying Permian Wann River gneiss (also of the Boundary Ranges suite). The top of the Wann River gneiss is in sheared and imbricated contact with the overlying Florence Range suite (Tracy Arm terrane), within the Wann River shear zone (Figs. 3.3; 3.4).

Approximately 1 km west of the trace of the White Moose Mountain to Wann River cross-section (W-W'), a strongly lineated biotite granite crops out along the southern shore of the Wann River for ~1 km (Fig. 3.2; LC-90-437; Table 3.1). The granite is interpreted to be older than 170 Ma and younger than 275 Ma; deformation fabrics in the granite parallel the penetrative fabric of the 184-176 Ma Wann River shear zone and are considered to have formed during displacement on the shear zone, and the granite contains xenoliths thought to be derived from the 270 ± 5 Ma Wann River gneiss (Appendix 3.1). However, the granite could be older than Permian if the inclusions were derived from older Boundary Ranges suite rocks (possibly Mississippian or older).

Mt. Switzer, West of Nelson Lake (Section X-X')

West of Nelson Lake, a section of Boundary Ranges suite actinolite schist, Hale Mountain granodiorite and Wann River gneiss dips moderately to the southwest (Fig. 3.5). Section X-X' is constructed perpendicular to strike, and the thicknesses of lithologic units are considered true thicknesses. The metamorphic rocks are intruded by hornblendite of uncertain age to the north (Figs. 3.2; 3.3); it could be coeval with an Early Jurassic hornblendite body exposed north of Fantail Lake (Mihalynuk pers. comm., 1993; Mihalynuk et al., 1989b; Fig. 3.1), or could be younger (Fig. 3.2). Along the western shore of Nelson Lake, at the base of this section, only the uppermost actinolite and chlorite-actinolite schists of the Boundary Ranges suite are exposed, consistent with this section being structurally higher than the Mt. Lawson section (Section V-V'; Figs. 3.3; 3.4). The actinolite schists are interlayered with thin, fine grained felsic layers and resemble those seen on Mt. Lawson, Hale Mountain, and north of Mt. Switzer (Plate 3.2c). Their contact with the overlying Hale Mountain granodiorite is not exposed. On the ridge west of Nelson Lake, the Hale Mountain granodiorite and the overlying Wann River gneiss are separated by a ~15 m thick section of 1 to 3 m thick layers of carbonate rock, calcsilicate and pelite (Figs. 3.3; 3.4); layering in the sedimentary rocks parallels foliation in both the Hale Mountain granodiorite and Wann River gneiss. Farther south, above the Wann River gneiss, marble and schist of the Florence Range suite are exposed (Figs. 3.2; 3.3).

Mt. Caplice (Section Y-Y')

Wann River shear zone fabrics, the Florence Range suite-Wann River gneiss contact and lithologic units in the Boundary Ranges suite are folded around a major southwest plunging synform, referred to as the Willison synform, that results in the rotation of these contacts from moderately south-dipping in the Wann River area to steep and north-south striking in the Mt. Caplice area (Figs. 3.2; 3.5). The geometry of the Willison synform must be inferred from the correlation of lithologic units and the attitude of foliation in Florence Range suite rocks (which can be followed almost continuously around the core of the synform), because the continuity of the Florence Range suite-Wann River gneiss contact and lithologic units in the Boundary Ranges suite is disrupted between the Mt. Switzer and Mt. Caplice areas by the Llewellyn fault zone to the east and a mid-Cretaceous pluton to the south (Fig. 3.2).

Lithologic units and contacts are also disrupted by the northwest-southeast striking Caplice fault, which offsets geologic contacts (including the Wann River shear zone) in the Mt. Caplice area (Figs. 3.2; 3.6). The Caplice fault has been observed in one place, south of Caplice Creek and ~3 km east of Hoboe Glacier, where it dips ~30° toward the southwest, juxtaposes actinolite schist of the Boundary Ranges suite (northeast) and probable Tagish Lake suite hornblende granodiorite (southwest), and is associated with cataclasite and mylonite. Elsewhere it is either covered, or difficult to recognise because possible Tagish Lake or Boundary Ranges suite rocks occur on both sides of the fault. Immediately west of the Llewellyn fault zone the Caplice fault is overlain by Tertiary Sloko Group volcanic flows. West of Caplice Creek and east of

Caplice Glacier, where the Caplice fault juxtaposes Florence Range suite rocks (southwest) against the Mt. Caplice quartz monzonite (northeast), the fault follows a stream bed, but is not exposed (Fig. 3.6). Within the first ~300 m of this contact the Florence Range rocks are intruded by <1 m to 100 m wide, undeformed, felsic dykes that comprise ~20 to 50% of the exposed rock. The Caplice fault presumably continues to the west, underneath the Caplice Glacier (Fig. 3.6), but it has not been recognised northwest of the Caplice glacier because the disruption of geologic contacts by the Willison Creek fault zone (not shown here; see Chapter 4; Currie, 1991) makes its recognition difficult, if not impossible.

Displacement on the Caplice fault is inferred to have components of sinistral and/or northeast-side-up motion on the basis of C-S fabrics, rotated feldspar clasts, the juxtaposition of actinolite schist of the Boundary Ranges suite (northeast) with the Wann River gneiss, and the juxtaposition of the Caplice quartz monzodiorite (Wann River shear zone footwall; northeast) against the Florence Range suite rocks (Wann River shear zone hanging wall; southwest; Fig. 3.6). Deflection of the generally north-south striking foliation and lithologic contacts near the Caplice fault (Fig. 3.6) could be due to the formation of drag folds during movement on the Caplice fault, or folding that occurred prior to movement on the Caplice fault.

The easternmost metamorphic rocks exposed on the north side of the Caplice fault are the structurally lowest rocks in the Mt. Caplice section; structurally lower rocks may be hidden underneath the almost flat-lying Sloko Group volcanic rocks to the east (Figs. 3.2; 3.3). The metamorphic rocks include chlorite-actinolite, actinolite,

and feldspar- and pyroxene-phyrlic actinolite schists (e.g. Plates 3.2b; 3.3c), and subordinate leucogabbro, marble, and felsic Mississippian dykes (Table 3.1; Appendix 3.1); all are regarded as part of the Boundary Ranges suite. To the west, the Boundary Ranges suite rocks are intruded by the Early Jurassic Hoboe quartz syenite (189.5 ± 3 Ma) of the Tagish Lake suite (Figs. 3.2; 3.6; Appendix 3.1). Actinolite schist and minor marble of the Boundary Ranges suite are interlayered (1 m to 5 m scale) with Hoboe quartz syenite over ~ 100 m. To the northwest, across the steep Caplice Creek valley, and north of the Mt. Caplice cross-section, the Early Jurassic Mt. Caplice quartz monzodiorite (180.9 ± 3.1 Ma) is exposed (Appendix 3.1; Fig. 3.6). Its contact with the Florence Range suite to the west is considered an extension of the Wann River shear zone (Fig. 3.6).

South of the Caplice fault and east of its sheared and imbricated contact with the Florence Range suite, the Wann River gneiss is associated with 1-3 m thick biotite schist and carbonate rock layers (Figs. 3.3; 3.6). East of the Wann River gneiss and southwest of Caplice Creek, biotite schist, marble and a well-layered, felsic gneiss of unknown age are exposed (Fig. 3.4; Table 3.1). East of Caplice Creek, the Wann River gneiss is interlayered with 1 to 20 m thick layers of foliated, hornblende-bearing, intermediate plutonic rocks (Tagish Lake suite; Fig. 3.6). Lithologic units observed south of the Caplice fault continue to the south where they intersect the Hoboe Glacier section (see below).

Hoboe Glacier (Section Z-Z')

Metamorphic rocks are exposed on both sides of the Llewellyn fault zone in the Hoboe Glacier section (Figs. 3.2; 3.3); those that crop out east of the Llewellyn fault zone are described in the following section. Metamorphic rocks from west of the Llewellyn fault zone are south of, and approximately along strike with the part of Mt. Caplice section that is exposed south of the Caplice fault (Fig. 3.2). Sinistral offsets of lithologic contacts south of the Caplice fault are shown as possible folds in Fig. 3.6, but have not been mapped in detail. Alternatively, they could be faults analogous to the Caplice fault, or could reflect the discontinuous nature of lithologic units caused by imbrication within the Wann River shear zone. Lithologic contacts and foliation are generally steep and strike ~north-south (Fig. 3.5).

The easternmost rocks exposed west of the Hoboe Glacier comprise a ~100 m wide section of 1 m- to 10 m-wide slivers of a variety of undeformed to highly strained rock types (including shale, carbonate rock, hornblendite, and potassium feldspar megacrystic granite) that are juxtaposed along anastomosing, steep, ~north-south striking faults. This shear zone is regarded as part of the Llewellyn fault zone which, where it is exposed to the north, is characterized by anastomosing faults that juxtapose a variety of rock types, some of which have not been penetratively metamorphosed or deformed (Mihalynuk et al., 1989b). In contrast, the Wann River shear zone is characterized by penetrative deformation fabrics that formed at upper greenschist to lower amphibolite facies conditions (Chapters 1; 4).

Actinolite schist of the Boundary Ranges suite is preserved as a thin sliver (<200 m thick) west of the Llewellyn fault zone and east of the Hale Mountain granodiorite, and as slivers (up to 15 m thick) within the Hale Mountain granodiorite (Figs. 3.2; 3.5). Like the schistose rocks to the north, actinolite schist in the Hoboe Glacier section is associated with pyroxene-, hornblende- and feldspar-phyrlic actinolite schist, and intruded by felsic dykes; one dyke has a Mississippian U-Pb zircon age of 352 ± 5.5 Ma (Fig. 3.2; Appendix 3.1). Actinolite schist is also associated with deformed, medium-grained leucogabbro (Fig. 3.6; Table 3.1). To the west, the Hale Mountain granodiorite is exposed for ~1.25 km (Figs. 3.4; 3.6). Along its western margin, it is in sheared contact with the Wann River gneiss.

Within the hornblende-plagioclase gneiss of the Wann River gneiss, metre-scale layers of carbonate rock and biotite schist are preserved (Figs. 3.3; 3.4). They may be metamorphosed sedimentary rocks that are coeval with the volcanic rocks, or infolds or fault-bounded slivers of the Florence Range suite (Figs. 3.3; 3.5). The Wann River gneiss is intruded by a 190.6 ± 1 Ma felsic dyke that is considered part of the Tagish Lake suite (Appendix 3.1). The dyke, the Wann River gneiss, and its foliation have been folded (Plate 3.2a; Fig. 3.3). Farther west, the Wann River gneiss is juxtaposed against marble of the Florence Range suite (Figs. 3.2; 3.3; 3.4).

The sequence of lithologic units in the southern Tagish Lake area (Hoboe Glacier and Mt. Caplice sections) is similar to that of the Tagish Lake area to the north (e.g. Mt. Switzer cross-section): in general, actinolite schist and the Wann River gneiss of the Boundary Ranges suite are separated by plutons of the Tagish Lake suite,

and the Wann River gneiss is juxtaposed against Florence Range suite rocks within the Wann River shear zone.

Other Localities of Boundary Ranges Suite Rocks

Polydeformed metasedimentary and metaigneous rocks exposed at the headwaters of the Swanson River, along the headwaters of the Wann River, and south of Mt. Mussen (east of the Llewellyn fault zone) are interpreted as belonging to the Boundary Ranges suite, but have poorly constrained stratigraphic relations because they are isolated from the metamorphic rocks described above by plutons or faults (Fig. 3.2). Metamorphic rocks of the Swanson River headwaters are preserved as a pendant in Cretaceous granite of the Coast Belt, and those exposed in the northern Wann River headwaters and east of the Llewellyn fault zone are separated from adjacent metamorphic rocks by faults (Fig. 3.2).

Swanson River

At the Swanson River headwaters, polydeformed, layered metamorphic rocks are surrounded by probable Cretaceous potassium feldspar megacrystic granite (Figs. 3.2; 3.7). Interlayered hornblende-plagioclase gneiss and biotite schist are exposed on a ridge, north of the Swanson River. They overlie polydeformed feldspathic and quartzofeldspathic, biotite-bearing semipelitic rocks, felsic gneisses, and minor actinolite schist, biotite schist and marble exposed to the south (Fig. 3.7). Marble layers are generally less than 3 m thick, with the exception of one structurally

thickened 20 m thick layer. The layered rocks are intruded by hornblendite (Fig. 3.7) and mafic dykes that have experienced at least one phase of deformation. The contact between the hornblende gneiss and the underlying rocks to the south has not been mapped in detail and its nature is not understood.

The Swanson River rocks are considered part of the Boundary Ranges suite because they most closely resemble rocks exposed on Mt. Lawson (lithologically similar hornblende-plagioclase gneiss and well-layered felsic orthogneiss exposed south of Mt. Lawson and Bighorn Creek, and west of the Chicken Creek fault, are described above; Fig. 3.2). The hornblende-plagioclase gneiss could be the Wann River gneiss, or an amphibolite facies equivalent of the greenschist facies mafic rocks of the Boundary Ranges suite, such as those seen in the Bighorn Creek area to the north. The underlying, dominantly felsic layered rocks are similar in lithology to rocks below the Bighorn Creek granite, and the hornblendite may be part of the Tagish Lake suite. The Swanson River rocks are not considered part of the Florence Range suite because they lack the characteristic thick carbonate rock layers, amphibolite and quartzite and because they include hornblende-plagioclase gneiss and abundant felsic gneisses, which are almost non-existent in the Florence Range suite.

Northern Wann River headwaters

Polydeformed metamorphic rocks, similar to those at the Swanson River headwaters, crop out near the northern headwaters of the Wann River, west of a steep fault that separates them from the Florence Range suite (Tracy Arm terrane). There,

hornblende-plagioclase gneiss is associated with polydeformed felsic orthogneiss (Plate 3.3d), minor carbonate rock and minor pelite. In general, the metamorphic rocks have a steep, north-south striking foliation (Fig. 3.5). On the basis of their lithologic similarity and proximity to the Swanson River headwater rocks (described above), they are also regarded as part of the Boundary Ranges suite and therefore the Wann River shear zone footwall (Fig. 3.2). The hornblende plagioclase gneiss is thought to be equivalent to the Wann River gneiss or mafic schistose rocks of the Boundary Ranges suite and the felsic gneiss and metasedimentary rocks are interpreted as possible equivalents of the Mississippian or older (?) Boundary Ranges suite. Correlation of these rocks with Wann River shear zone footwall rocks implies that the relative vertical motion on the steep fault that separates these rocks from the Florence Range suite to the east (hanging wall of the Wann River shear zone) is west-side-up (Fig. 3.2).

East of the Llewellyn fault zone

Metamorphic rocks are exposed east of the Llewellyn fault zone, on the south flank of Mt. Mussen, where they are separated from the overlying Late Triassic Stuhini Group volcanic rocks (Stikinia) by the Mt. Mussen shear zone (Fig. 3.2, 6). They consist primarily of steep, southwest-dipping (Fig. 3.5), actinolite schist that is locally feldspar-phyrlic, and locally interlayered on a centimetre-scale with feldspar-rich layers. The actinolite schist resembles mafic and felsic layers of the Boundary Ranges suite exposed on Mt. Lawson, along the west shore of Tagish Lake, and west

of Hoboe Creek. Like other mafic schistose rocks of the Boundary Ranges suite, they are associated with minor, discontinuous layers of marble and graphitic schist (<5 m thick; Fig. 3.6). Quartzofeldspathic muscovite schist (see Table 3.1), chlorite schist, and graphitic schist crop out farther east, near the south end of Atlin Lake.

Summary of Sections

In all sections the overall lithologic sequence is similar; Mississippian and possibly older rocks of the Boundary Ranges suite are structurally lower than the Permian Wann River gneiss, and the structurally highest rocks are juxtaposed against the Florence Range suite within the Wann River shear zone. Foliated ca. 190 Ma granitic rocks of the Tagish Lake suite commonly intervene between the Wann River gneiss and other rocks of the Boundary Ranges suite.

The oldest rocks of the Boundary Ranges suite in the Tagish Lake area crop out in the Bighorn Creek area, where quartzofeldspathic biotite semipelitic rocks, biotite and muscovite schist, well-layered felsic orthogneisses, the Bighorn Creek granite and minor marble layers are exposed. The Bighorn Creek granite is a Devonian-Mississippian metamorphosed granite; its lack of strong, penetrative, mesoscopic fabrics suggests that it acted as a rigid block while the adjacent, less competent rocks were deformed. The orthogneisses are interpreted as metamorphosed felsic tuffs on the basis of composition and their well developed millimetre-scale layering (the Mount Lawson orthogneiss was previously interpreted as a quartzite (PPMq); Mihalynuk et al., 1989a, b). The associated quartzofeldspathic biotite semipelitic rocks could be

metaclastic rocks primarily derived from felsic volcanic rocks, but may also have a component of continentally derived sediments. Biotite and muscovite schists had pelitic protoliths, such as mudstones and shales. Marble layers are generally discontinuous (e.g. Mihalynuk et al., 1989b), but may have once been continuous limestone beds that have since been dismembered and transposed during deformation. Similar rock-types are also exposed in the Swanson River and northern Wann River headwater areas.

Structurally above, and presumably younger than the pelitic rocks and felsic gneisses are actinolite and chlorite-actinolite schist interpreted as metamorphosed mafic volcanic rocks (in part pyroxene-, hornblende-, and/or feldspar-phyric). They are interlayered with leucocratic, feldspar-rich layers thought to be metamorphosed felsic tuffs based on their composition and thin, but continuous nature. The mafic schists are interpreted as tuffs where they are interlayered with felsic layers on a millimetre to centimetre scale (Plate 3.2c); where they are more massive and porphyritic (e.g. Plate 3.2b), they are interpreted as metamorphosed flows, sills and transposed dykes; in places they may be reworked volcanic rocks. The upward change from pelitic and metaclastic rocks with felsic tuffs to dominantly mafic metamorphosed volcanoclastic rocks is reflected in the decreased abundance of pelitic rocks and felsic gneiss in the structurally higher Mt. Switzer, Mt. Caplice and Hoboe Glacier cross sections to the south (Figs. 3.2; 3.4).

The Wann River gneiss (270 ± 5 Ma; Appendix 3.1) is interpreted as a metamorphosed volcanoclastic rock on the basis of composition (Table 3.1), well-

defined millimetre to centimetre-scale compositional layering, the locally preserved graded nature of this layering (Plate 3.1c), and its association with metapelitic and metacarbonate rocks. It is important to note that due to the emphasis on its mineralogy, some rocks mapped as Wann River gneiss may not be Permian in age, but instead might be amphibolite facies equivalents of the older Boundary Ranges suite actinolite schist. For the same reasons, Permian Boundary Ranges suite rocks (ie. Wann River gneiss equivalents) may be more widespread than shown in Figure 3.2 if they contain greenschist facies mineral assemblages. Minor carbonate and metapelitic rocks associated with the Wann River gneiss (e.g. Mt. Switzer and Hoboe Glacier sections; Fig. 3.3) are in depositional contact with the Wann River gneiss in some cases, and may also be Permian, but in others, they may be imbricate slivers of Florence Range suite that were interleaved with the Wann River gneiss during movement on the Wann River shear zone, which is known to be an imbricate structure (Chapters 1; 4).

The original geometries and distribution of the Tagish Lake suite plutons are not known. The Hoboe quartz syenite and Mt. Caplice monzodiorite are truncated by the Caplice fault, and the Hale Mountain granodiorite is truncated by the Llewellyn fault zone and by mid-Cretaceous plutons of the Coast Belt (Fig. 3.2). The Hale Mountain granodiorite is a sheet-like body, possibly in part due to shearing associated with displacement along the Wann River shear zone. It is not exposed north of Bighorn Creek, where chlorite-actinolite schists of the Boundary Ranges suite are unconformably overlain by Tertiary Teepee Peak volcanic rocks (Mihalynuk et al.,

1989a, b). Its absence may be due to an unconformity beneath the volcanic rocks, or may indicate that the northwestern margin of the pluton lies in the vicinity of Bighorn Creek. The southern margin of the Hale Mountain granodiorite lies to the south of the Tagish Lake area.

Most primary geologic contacts within the Boundary Ranges suite are difficult to interpret because original contact relations have been modified by metamorphism and polyphase deformation, including folding, shearing and imbrication. The nature of the contact between the pre-Permian Boundary Ranges suite and the Wann River gneiss is not clear because it is obscured by deformed Early Jurassic plutonic rocks of the Tagish Lake suite. Contacts between the Tagish Lake suite plutons and the Boundary Ranges suite are interpreted as sheared intrusive contacts. The contact between actinolite schist of the Boundary Ranges suite and the Hale Mountain granodiorite exposed on Hale Mountain and Mt. Switzer is gradational, as is the contact between the Boundary Ranges suite and the Hoboe quartz syenite. Some of the plutonic layers that alternate with actinolite schist layers may have been apophyses, but textural evidence implies that the interleaved character of these contacts most likely reflects shearing and possibly folding, along intrusive contacts (Chapter 4). Slivers of Boundary Ranges suite rocks within the Hale Mountain granodiorite exposed west of Hoboe Glacier could represent pendants or xenoliths of country rock, or imbricated tectonic slivers. The interpretation of the contact between the Wann River gneiss and the Hale Mountain granodiorite as a sheared intrusive contact (Plate

3.3b) is supported by their relative ages and the intrusive relationship between the Tagish Lake suite dyke and the Wann River gneiss (Figs. 3.3; 3.4; Plate 3.2a).

TECTONIC SETTING OF THE BOUNDARY RANGES SUITE AND TAGISH LAKE SUITE

The footwall of the Wann River shear zone comprises the Boundary Ranges suite (a sequence of Paleozoic volcanic, pelitic, carbonate and intrusive rocks) and the Early Jurassic plutons of the Tagish Lake suite that intruded them. Assuming that the lithological sections described in this chapter are essentially right-way-up and without major structural inversions, the Boundary Ranges suite can be interpreted as recording progressive deposition of Mississippian (or older) to Permian (and possibly younger?) rocks in a volcano-sedimentary tectonic environment that was metamorphosed during, and possibly prior to, the Early to Middle Jurassic (Chapter 4). The lowermost pelitic and quartzofeldspathic semipelitic rocks and felsic tuffs of the Boundary Ranges suite were deposited on an unknown basement, presumably overlain by felsic to intermediate volcanic rocks (tuffs and flows), and intruded by Devonian-Mississippian plutons and dykes, possibly related to the volcanic rocks. During the Permian, additional volcanic rocks (Wann River gneiss protolith) and sedimentary rocks were deposited above the felsic to intermediate volcanic rocks. Near the end of the Early Jurassic, the Paleozoic rocks were intruded by plutons of the Tagish Lake suite. The abundance of volcanic and plutonic rocks is suggestive of an arc environment, and the primitive isotopic signature of these rocks (Jackson, 1992; Appendix 3.2) is indicative of a relatively

juvenile source for this magmatism. However, it is important to note that the Boundary Ranges suite has not been rigorously isotopically characterized, and some rocks, especially the metaclastic rocks of the Bighorn Creek and Swanson River areas may have been in part derived from an evolved, continental source.

CORRELATION OF THE BOUNDARY RANGES SUITE AND TAGISH LAKE SUITE

Potentially equivalent rocks of the Boundary Ranges suite and the Tagish Lake suite can be found in numerous terranes of the North American Cordillera, including Stikinia, Taku terrane, Yukon-Tanana terrane (in part equivalent to the Nisutlin subterrane of Kootenay terrane) and Wrangellia; they may also be preserved intermittently as pendants of metasedimentary and metavolcanic rocks in plutons of the Coast Belt to the west, (referred to as undivided metamorphic rocks; Wheeler et al., 1991). On the basis of the stratigraphy outlined above, potential correlations and the rejection of correlations with other terranes (Tracy Arm and Alexander terrane) are discussed below.

Stikinia

Stikinia is an allochthonous terrane composed of mid-Paleozoic to mid-Mesozoic assemblages of volcanic, plutonic and sedimentary rocks that have lithologic, geochemical and isotopic characteristics indicative of an arc environment that formed above a young, geochemically and isotopically primitive crust (e.g.

Monger and Berg, 1984; Samson et al., 1989; Wheeler et al., 1991). It is presently separated from autochthonous North America by pericratonic terranes, arc-like Quesnellia, and the allochthonous oceanic Cache Creek and Slide Mountain terranes (e.g. Coney et al., 1980; Monger and Berg, 1987; Wheeler et al., 1991; Fig. 3.1).

Paleozoic (Devonian to Permian) rocks of Stikinia are referred to as the Stikine assemblage (Monger, 1977; Brown et al., 1991; Asitka Group of Wheeler and McFeely, 1991; Fig. 3.4), and form a semi-continuous southeast-northwest trending belt that extends from south of the Stikine River to the Taku River (Fig. 3.1). Stikine assemblage rocks are well exposed and best studied in the Stikine River area where they comprise: (1) Permian and older sediments, tuffs and intermediate volcanic rocks; (2) a Mississippian limestone; and (3) a Permian limestone (Fig. 3.4; Souther, 1972; Monger, 1977; Anderson 1989; 1993; Brown et al., 1991, and references therein). The Permian and older sediments include polydeformed chloritic, graphitic and maroon phyllite with interbedded quartz-sericite schist and recrystallized limestone, and the structurally overlying intermediate to mafic volcanic package of greenstones and chlorite schists that was derived from intermediate flows (including feldspar- ± augite-phyric flows; Brown and Gunning, 1989), sills and tuffs. They are considered older than Devonian because they are intruded by two Devonian plutons (369 ± 5.1 Ma and 370.7 ± 6.6 Ma; U-Pb zircon ages; McClelland et al., 1993). The oldest sedimentary unit of known age is a penetratively deformed Devonian coralline limestone (Anderson, 1989; Read et al., 1989; Brown et al., 1991).

Locally, two Mississippian limestone members (50 to 150 m thick) conformably overlie the older sedimentary and volcanic rocks of the Stikine assemblage (e.g. Logan and Koyanagi, 1989). They are intercalated with centimetre to metre-scale purple and green ash-tuff beds, and between them there is a wedge (less than 250 m thick) of phyllite, tuff and intraformational limestone-pebble conglomerate (Logan and Koyanagi, 1989). These Mississippian rocks are not regionally extensive; within the Stikine River area ~1500 m of Mississippian strata disappear over 5 km (Monger, 1970), possibly reflecting the presence of a profound post-Upper Mississippian, pre-Lower Permian unconformity.

Lower Permian strata include thin-bedded, coralline limestone and volcanic arc-like mafic to felsic volcanic flows, tuffs and volcanoclastic rocks (Anderson, 1989). They are locally in apparent conformable contact with the overlying Middle Triassic shale (e.g. Read et al., 1989; Logan et al., 1992). Elsewhere, they are unconformably overlain by Early Triassic Stuhini Group volcanic rocks (e.g. Anderson, 1989; Read et al., 1989), which are in turn overlain by Early to Middle Jurassic Hazelton Group volcanic rocks (e.g. Anderson, 1989). The plutonic equivalents of the Hazelton Group are the calc-alkaline (hornblende ($\pm$ biotite) granodiorite and quartz monzodiorite) and alkaline (gabbro-syenite complexes, monzogranite, monzonite, and potassium felspar porphyry) plutons of the Texas Creek plutonic suite (195-189 Ma), and ca. 186-185 Ma plutons of an unnamed suite, and the calc-alkaline Three Sisters suite (179-176 Ma; Anderson, 1993). These plutonic suites are roughly coeval with the Klotassin and Aishihik suites (212-185 Ma; Woodsworth et al., 1991 and references therein;

Johnston and Erdmer, in press), which has been considered the main diagnostic characteristic of Stikinia in Yukon and eastern Alaska (Wheeler et al., 1991; Woodsworth et al., 1991). The Klotassin suite is compositionally and isotopically similar to the Three Sisters, unnamed and Texas Creek plutonic suites (all are characterized by "low" $^{87}\text{Sr}/^{86}\text{Sr}_i$ ratios; Samson et al., 1989; Mortensen, 1992), but it lacks known volcanic equivalents. In northern British Columbia and southern Yukon sedimentary rocks of the Early to Middle Jurassic Laberge Group (Hettangian to Bajocian; 208-166 Ma; Tempelman-Kluit, 1985) unconformably overlie Late Triassic Stuhini Group equivalent rocks (referred to as Lewes River Group in Yukon; e.g. Wheeler, 1961).

Comparison of Tagish Lake area rocks with Stikinia

The Boundary Ranges suite is correlated with the Stikine assemblage of Stikinia on the basis of age, lithology and characteristic isotopic similarities (Fig. 3.4). In both areas there is an upward transition from dominantly metasedimentary rocks to dominantly mafic metavolcanic rocks, including pyroxene- and feldspar-phyric mafic rocks. Both assemblages include Devonian-Mississippian orthogneisses and rocks that are Mississippian (or older) to Permian in age, and include rocks with relatively primitive isotopic signatures.

The ages of marble lenses in the Boundary Ranges suite are unknown; they could be equivalent to Devonian, Mississippian, Pennsylvanian (?), and/or Permian carbonate rock layers of the Stikine assemblage. The absence of thick Mississippian

and Permian limestone layers in the Tagish Lake area may reflect a paleogeography with reefs that formed around volcanic highs, and/or omission due to later faulting or erosion, as indicated by the variable thickness of limestone layers in the Stikine River area.

The Permian metavolcanic Wann River gneiss and its associated metasedimentary rocks are correlated with coeval foliated chloritic tuffs, carbonate rock and siltstones of the Stikine assemblage. The component of inherited Precambrian zircon in zircon fractions from the Wann River gneiss (D; Appendix 3.1; Fig. 3.7d) is consistent with a Stikinia origin, even though Stikinia is in general characterized by juvenile isotopic signatures (Samson et al., 1989). Some Stikinia rocks also contain components derived from older and possibly more evolved sources; zircons from the Triassic Zippa Mountain syenite contain inherited cores (Bevier, pers. comm. 1992), as do zircons from the Early Jurassic Spatsizi River Stock and Cold Fish Volcanics (Thorkelson, 1992).

Stratified, fossiliferous Triassic rocks to the west of the Tagish Lake area (described in Brew et al., 1985) have been interpreted as belonging to Stikinia (Brew et al., in press), but rocks of Triassic age have not been recognised in the Boundary Ranges suite. However, metamorphic equivalents of rocks as young as Early Jurassic in age could be included in this suite.

The Tagish Lake suite comprises plutons that are similar in age and lithology to the Texas Creek and the Three Sisters suites of the Stikine River area, and the

Klotassin suite of Yukon. All suites are characterized by hornblende-bearing, alkaline to calc-alkaline, intermediate to mafic plutons that are isotopically unevolved.

The proposed correlation of Boundary Ranges suite rocks with Stikinia is consistent with four proposed geologic links between this suite and rocks previously recognised as part of Stikinia. North of the Tagish Lake area, the Boundary Ranges suite is unconformably overlain by possible Middle to Upper Jurassic volcanic rocks that are considered probable Hazelton Group equivalents (M.G. Mihalynuk pers. comm., 1993) and a conglomerate that has been correlated with conglomerates of the Early Jurassic Laberge Group that unconformably overlie Stikinia rocks elsewhere (Mihalynuk and Rouse, 1988a). Augite-phyric mafic dykes that intrude the Boundary Ranges suite east of the Llewellyn fault zone have been interpreted as feeders to the overlying Late Triassic Stuhini Group volcanic rocks of Stikinia (Werner, 1978). Finally, metamorphic clasts in Late Triassic Stuhini Group conglomerates of Stikinia (Souther, 1971; Bultman, 1979; Jackson et al., 1991) were likely derived from the Boundary Ranges suite (Chapter 1).

The Boundary Ranges suite is interpreted as a northern extension of the belt of Paleozoic Stikinia rocks that extends from the Stikine River area to the Taku River area. Boundary Ranges suite rocks recognised by Smith and Mihalynuk (1992) and Mihalynuk et al. (1994) north of the Taku River, geographically link the Boundary Ranges suite of the Tagish Lake area to the Stikine assemblage to the south (Fig. 3.1; Chapter 5).

Paleozoic Stikinia rocks may also occur as far south as the Skeena River area (Fig. 3.1), where metamorphic rocks of the Scotia-Quaal metamorphic belt are exposed along the western margin of Stikinia. They are composed of north-northwest trending bands of metavolcanic, metaplutonic and metasedimentary rocks, including the Devonian Big Falls orthogneiss (385 $\pm$ 4.2/-3.7 Ma) a quartz dioritic orthogneiss (336.8 $\pm$ 17.7/-7.1 Ma) and Early Jurassic plutons (188-202 Ma), Paleozoic (or older?) metavolcanic and metasedimentary rocks (Gareau, 1991a; b). A 191.7 $\pm$ 0.6 Ma orthogneiss has intruded the contacts between all of the major lithologic units, indicating that they were all together before Middle Jurassic time. Although the Scotia-Quaal metamorphic belt has been correlated with Tracy Arm terrane (Nisling terrane; Gareau, 1991a; 1991b), its metavolcanic and metaplutonic components, especially those of Devonian age, suggest that these metamorphic rocks could represent the southernmost exposures of the Stikine assemblage (Chapter 2).

Comparison with Yukon-Tanana Terrane (Kootenay Terrane)

Terrane nomenclature for metamorphic rocks in Yukon is confusing due to poor exposure, complex structure, metamorphism, and poor age control (for a review of Yukon terrane nomenclature see Mortensen, 1992). Two recent compilations differ significantly: Mortensen (1992) retains the definition of Yukon-Tanana terrane proposed by Monger and Berg (1987) and divides Yukon-Tanana terrane into "upper, middle and lower units", with the "lower unit" comprising quartzofeldspathic metasediments and associated rock-types; in contrast, Wheeler et al. (1991), in general,

regard the quartzofeldspathic metasediments and associated rock-types that occur east of Klotassin Suite plutons (Stikinia) and southwest of the Tintina Fault as the Nisutlin subterrane of the pericratonic Kootenay terrane, and consider those that crop out west of Stikinia as Nisling terrane (Fig. 3.1).

Equivalents of all of the regionally significant lithologic units in the Boundary Ranges suite are represented in the "middle and upper units" of the Yukon-Tanana terrane (as defined by Mortensen, 1992). In some cases, rocks included in the "lower unit" of Yukon-Tanana terrane are overlain by Late Devonian to mid-Mississippian volcanic rocks and intruded by rocks of the same age; they may be correlative with the lowermost Boundary Ranges suite rocks. The "middle unit", which "consists of roughly equal proportions of carbonaceous metasedimentary rocks interlayered with mafic and felsic volcanic rocks" (Mortensen, 1992), is lithologically similar to, and coeval with the pre-Permian metamorphic rocks of the Boundary Ranges suite. As in the Tagish Lake area, they are intruded by Late Devonian to Early Mississippian granitic orthogneisses (364-343 Ma, e.g. Pelly Gneiss; Mortensen, 1992; referred to as Mink Creek suite by Woodsworth et al., 1991). However, no plutons corresponding to the Late Devonian to Early Mississippian hornblende-bearing orthogneisses of the Yukon-Tanana terrane (e.g. Simpson Range plutonic suite, Selwyn Gneiss, and Fiftymile batholith; Tempelman-Kluit, 1974; Tempelman-Kluit and Wanless, 1980; Mortensen, 1986) have been recognised in the Tagish Lake area, although some undated metaplutonic rocks tentatively included in the Tagish Lake suite may be Devono-Mississippian age.

The "upper unit" of Yukon-Tanana terrane consists of mid-Permian (263-256 Ma) felsic metaplutonic, and felsic and minor mafic volcanic rocks exposed in western Yukon (e.g. Sulphur Creek orthogneiss, Klondike schist; Mortensen, 1992, and references therein) that are similar in age to the Permian Wann River gneiss (Appendix 3.1) and associated metasedimentary rocks. Mortensen (1992) argues that the "upper unit" (possible Wann River gneiss equivalent), was deposited above the "middle unit" (possible Boundary Ranges suite equivalent) on the basis of a mid-Permian dyke that intrudes the "middle unit" and is lithologically similar to mid-Permian plutonic rocks of the "upper unit". An unequivocal link between the Wann River gneiss and older Boundary Ranges suite rocks has not been documented in the Tagish Lake area because they are separated by plutons of the Tagish Lake suite. However, if the correlation of these units with the "middle and upper units" of Yukon-Tanana terrane is correct and the link between the "upper and middle units" of Yukon-Tanana terrane is valid, then the presumed depositional link between the Boundary Ranges suite and the Wann River gneiss would also be valid.

If the "middle and upper units" of Yukon-Tanana terrane are indeed correlative with Paleozoic Stikinia, then it is not surprising that they are intruded by Late Triassic to Middle Jurassic Klotassin suite plutons (included in Stikinia by Wheeler et al., 1991). Mortensen (1992) has suggested that the Klotassin plutonic suite should be regarded as a "stitching" assemblage that links Yukon-Tanana terrane and Stikinia.

Stikinia-equivalent rocks may also be preserved in pendants in Cretaceous or younger plutons of the Coast Belt. Within one pendant, the post-Triassic to pre-60 Ma

Meade Glacier fault (Brew and Ford, 1987) juxtaposes metasedimentary rocks (part of the Florence Range suite or Stikinia?) against stratified, fossiliferous Triassic rocks that have been interpreted as belonging to Stikinia (Brew et al., 1994), but which could also be part of Taku terrane.

Comparison with Taku Terrane

The Taku terrane occurs as a thin, ~700 km long strip along the western margin of the Coast Belt (Fig. 3.1; e.g. Monger and Berg, 1987; Wheeler et al. 1991) and comprises quartzite and felsic metatuff, upper Paleozoic basaltic metatuff and agglomerate, Permian crinoidal marble intercalated with pelitic phyllite and felsic metatuff, Middle and Upper Triassic basalt, limestone, shale, phyllite and basaltic breccia, and Jurassic to Cretaceous metamorphosed sediments and volcanics similar to those of the Gravina belt (Monger et al., 1991). It includes the Ruth assemblage of McClelland et al. (1991), Endicott Arm assemblage of Gehrels et al. (1992a), and the Alava sequence of Rubin and Saleeby (1991).

The Ruth assemblage comprises metapelitic rocks, quartzite, marble, felsic metatuff, mafic metavolcanic rocks and orthogneiss, including a 367 ± 10 Ma felsic metatuff and a 345 ± 13 Ma orthogneiss (McClelland et al., 1991), and may be equivalent to the Boundary Ranges suite. It is correlated with Yukon-Tanana terrane and the Endicott Arm assemblage by McClelland et al. (1991). The Endicott Arm assemblage comprises mid-Paleozoic metavolcanic rocks and subordinate metaclastic rocks, including quartz-rich layers, most of which were derived from rhyolitic and

dacitic volcanic rocks (Gehrels et al., 1992a), and some of which were likely derived from evolved sources (Samson et al., 1991a). Within the Endicott Arm assemblage the proportion of metavolcanic rocks, relative to metapelitic and metaclastic rocks increases upsection (Gehrels et al., 1992a), similar to the Boundary Ranges suite. Gehrels et al. (1992a) correlate the Endicott Arm assemblage with the Yukon-Tanana terrane of Mortensen (1992), and Nisutlin subterrane of Kootenay Terrane as defined by Wheeler et al. (1991).

The upper Paleozoic portion of the Alava sequence may be correlative with the Permian Wann River gneiss; it is characterized by metamorphosed basalt, mafic tuff, and Permian crinoidal limestone, and is locally associated with quartzite (Rubin and Saleeby; 1991). Permian rocks are inferred to be unconformably overlain by the lower Mesozoic portion of the Alava sequence which are thought to represent a "metamorphosed last vestige of Stikinia" by Rubin and Saleeby (1991). They comprise carbonaceous phyllite, fine-grained argillaceous and silicious limestone, metabasalt, and mafic breccia and tuff that are at least in part Triassic in age (Rubin and Saleeby, 1991).

Thus, the Taku terrane resembles both mid-Paleozoic and late Paleozoic Boundary Ranges suite rocks, and includes possible equivalents of early Mesozoic Stikinia rocks, but lacks documented Early Jurassic plutons (Tagish Lake suite equivalents).

Comparison with Wrangellia

The Paleozoic Sicker Group of Wrangellia could be equivalent to the mid-Paleozoic part of the Boundary Ranges suite. It is exposed in the Vancouver Island area, and includes marine calc-alkaline volcanic rocks overlain by Carboniferous and Early Permian limestone (Muller, 1980). The volcanic sequence is made up of the Nitinat Formation, which consists of more than 3000 m of augite-bearing agglomerate and tuff, and the overlying McLaughlin Ridge Formation (Myra Formation of Muller, 1980), which is composed of volcanic sandstone, conglomerate, and local pillow lavas and felsic pyroclastic rocks. Like the pyroxene-bearing schists of the Boundary Ranges suite, the augite-bearing agglomerate of the Nitinat Formation is intruded by Late Devonian quartz porphyry and granodiorite (Saltspring Intrusions; 362 ± 2 Ma and 367 ± 2 Ma; Parrish and McNicoll, 1991). Similar in age to the Saltspring Intrusions is a felsic volcanic rock of the McLaughlin Ridge Formation (366 ± 4 Ma; Parrish and McNicoll, 1991) that could be coeval with the felsic metatuffs of the Boundary Ranges suite. Mid-Paleozoic rocks of Wrangellia, like the coeval rocks of the Boundary Ranges suite, are isotopically primitive (Samson et al., 1989; 1990).

Permian and younger rocks of Wrangellia differ from those of the Tagish Lake area: the Permian strata comprise limestone and chert, but lack a volcanic component (Muller, 1980), and thick Triassic tholeiitic basaltic rocks characteristic of Wrangellia are unknown in the Tagish Lake area. Plutons of Wrangellia that are coeval with the Tagish Lake suite are exposed on Vancouver Island (late Early Jurassic to early Middle Jurassic Island Intrusions; Isachsen et al., 1985); north of Vancouver Island,

Jurassic plutons of Wrangellia are younger than those found in Stikinia (Middle to Late Jurassic; e.g. Anderson et al., 1991; Wheeler and McFeely, 1991; Anderson, 1993). In conclusion, the mid-Paleozoic rocks of the Tagish Lake area are similar to those of Wrangellia, but the Permian to Early Jurassic geologic histories of the Tagish Lake area and Wrangellia are significantly different.

Comparisons with Tracy Arm and Alexander terranes

Rocks in the footwall of the Wann River shear zone are primarily composed of volcanic and igneous rocks of the Boundary Ranges suite and Tagish Lake suite; they are distinctly different from Tracy Arm terrane rocks of the hanging wall, which are almost entirely sedimentary. Tracy Arm terrane comprises metamorphosed pelite and semipelitic rocks, marble, and quartzite, with widespread, but volumetrically minor amphibolite (Currie, 1990a, 1991; Mihalynuk et al., 1990a; b). In addition, the Boundary Ranges suite lacks the quartzites and extensive carbonate rocks of Tracy Arm terrane. As mentioned earlier, the footwall of the Wann River shear zone shares no stratigraphic, intrusive or structural links with the Tracy Arm terrane hangingwall prior to ca. 184 Ma (Chapters 1; 4).

Alexander terrane (Fig. 3.1) consists of Cambrian (and older?) metamorphosed volcanic and sedimentary rocks and comagmatic intrusives, Ordovician to Devonian volcanic, sedimentary and carbonate rocks, Mississippian to Pennsylvanian sedimentary and carbonate rocks (including extensive, thick limestone layers), and Permian and Upper Triassic volcanic, clastic, and carbonate rocks (Gehrels and

Saleeby, 1987). The Boundary Ranges suite does not resemble coeval rocks of Alexander terrane: it lacks coarse clastic rocks and thick Devonian to Pennsylvanian carbonate rocks, and no rocks correlative with Devono-Mississippian felsic orthogneisses of the Tagish Lake area are represented in the Alexander terrane.

SUMMARY OF TERRANE COMPARISONS

The Paleozoic (and older?) Boundary Ranges suite and Mesozoic Tagish Lake suite of the Tagish Lake area formed in an arc environment and most closely resemble coeval rocks of Stikinia. Coeval equivalents of all rock-types found in these suites exist in Stikinia, in volumetrically similar quantities, and with similar geologic relations. Furthermore, it is logical to consider the Boundary Ranges suite part of the semi-continuous belt of Paleozoic Stikinia rocks that extends from the Stikine River area to the Taku River area (Fig. 3.1), and relate the Tagish Lake suite plutons to lithologically similar coeval plutons of Stikinia which exist from the Stikine River area to northern Yukon (Wheeler and McFeely, 1991). Correlations with Taku terrane, parts of Yukon-Tanana terrane (Nisutlin subterrane and Klotassin suite), undivided metamorphic rocks in the Skeena River area, and possibly Wrangellia, are also considered reasonable, on the basis of lithologic similarities and associations. The Boundary Ranges suite and Tagish Lake suite are distinct from, and are not correlated with Tracy Arm or Alexander terranes.

IMPLICATIONS FOR TERRANE NOMENCLATURE

The incorporation of the Boundary Ranges suite and Tagish Lake suite in Stikinia is important because until recently Mesozoic Stikinia rocks were recognised from southern British Columbia to beyond the Yukon-Alaska Border, but the Paleozoic portion of Stikinia was thought to exist only as far north as the Taku River area (Fig. 3.1; Wheeler and McFeely, 1991). This lead some authors to question the nature of the unexposed basement of the Mesozoic Stikinia arc north of the Taku River and, in some cases, propose that in some areas the basement to Stikinia may be the dominantly quartzofeldspathic metasediments that in part contain detrital Precambrian zircons (correlated by some authors with Nisling terrane (ie. Tracy Arm terrane); e.g. Mihalynuk et al., 1989c; Jackson et al., 1991; Monger et al., 1991; Mortensen, 1992; Rubin et al., 1990; McClelland, 1992). However, at least in the Tagish Lake area, if not in Yukon, Tracy Arm terrane could not have been the basement to Stikinia, because these terranes were not juxtaposed until after Early Jurassic time (Chapter 1). Three possibilities remain: the pre-Devonian basement of Stikinia is not exposed, is not preserved, or has not been previously recognised north of the Taku River. The latter is most likely: this chapter argues strongly that Paleozoic Stikinia is exposed in the north, but has not been recognised because metamorphism, deformation (including imbrication along terrane boundaries), and a lack of age control have prevented confident correlation of pre-Triassic rocks. Paleozoic rocks considered part of Stikinia are preserved north of the Taku River, where they have been included in the Boundary Ranges suite by Smith and Mihalynuk (1992) and Mihalynuk et al. (1994), in the

footwall of the Wann River shear zone in the Tagish Lake area (this chapter), and may also exist in Yukon where they have been included in Yukon-Tanana terrane (c.f. Mortensen, 1992), southeast Alaska where they have been interpreted as part of Taku terrane, and the Skeena River area of west-central British Columbia.

Numerous authors have discussed the similarities between the geologic history of Stikinia and the histories of Yukon-Tanana, Taku, Quesnellia, Chilliwack, Kootenay, Olds Ferry, Grindstone, Eastern Klamath, and Northern Sierra terranes. Due to a shared Permian fauna, termed the McCloud fauna, they have been referred to as the McCloud belt terranes (e.g. Miller, 1987). Attributes common to terranes of the McCloud belt are: Devonian to mid-Mississippian metavolcanic and/or metaplutonic rocks; Early Permian "McCloud-type" fauna; Early and Middle Permian mafic to felsic metavolcanic rocks; evidence of post-mid Permian and pre-Late Triassic tectonism; and Late Triassic and Early Jurassic plutonism and/or volcanism (Miller, 1987; Wernicke and Klepacki, 1989; Oldow et al., 1989; Rubin et al., 1990; Mortensen, 1992). These terranes were either proximal to one another from mid-Paleozoic to mid-Mesozoic time, perhaps existing as a fringing arc in the eastern Pacific Ocean (e.g. Miller, 1987); or they occupied similar, but disparate tectonic settings contemporaneously. Wrangellia also appears to have occupied a similar tectonic setting during the mid-Paleozoic, but lacks Permian volcanic rocks and McCloud fauna.

The inferred Stikinia rocks of the Tagish Lake area share some attributes of the McCloud Belt, such as Devonian to mid-Mississippian metaplutonic and/or metavolcanic rocks, Permian mafic to felsic metavolcanic rocks, and Early Jurassic

plutonism. The remaining McCloud Belt attributes that have not been identified in the Tagish Lake area are permissible: (1) the apparent absence of McCloud fauna may be due to obliteration by deformation and recrystallization (no fossils have been found in the metamorphic rocks of the Tagish Lake area); (2) a post-mid Permian unconformity may not have been recognised in the Tagish Lake area, due to a lack of age control, and/or because it has been obscured by Jurassic deformation and/or plutonism; and (3) the lack of a Triassic component in the Boundary Ranges suite (with the possible exception of pyroxene-phyric dykes of proposed Triassic age that intrude Boundary Ranges suite rocks south of Atlin Lake; Werner, 1978) may be the result of erosion or tectonic removal, or may not be recognised due to sparse age control.

CONCLUSIONS

Paleozoic rocks of Stikinia were previously recognised only between the Stikine and Taku river areas, but are now thought to be more extensive. Correlations presented in the chapter indicate that some of the polydeformed greenschist to amphibolite facies metamorphic rocks from north of the Taku River area (Tagish Lake area) comprise the northern continuation of the Paleozoic Stikine rocks. These rocks also resemble coeval parts of Yukon-Tanana (Nisutlin subterrane of Kootenay terrane) and Taku terranes and some of the metamorphic rocks exposed south of the Stikine River, suggesting that previously unrecognised Paleozoic rocks of Stikinia may also exist elsewhere.

The Boundary Ranges suite and the Tagish Lake suite were juxtaposed against Tracy Arm terrane along the Wann River shear zone during the mid-Jurassic; their recognition as part of Stikinia is therefore critical to the understanding of the Tracy Arm-Stikinia terrane boundary (see Chapter 1) and the tectonic history of the northern Cordillera.

APPENDIX 3.1 - U-Pb ANALYSES OF BOUNDARY RANGES SUITE AND TAGISH LAKE SUITE ROCKS

Age Constraints for the Boundary Ranges Suite and Tagish Lake Suite

In the past, ages of the Boundary Ranges suite and Tagish Lake suite were limited to older than 175 Ma, the minimum age of their metamorphic fabric (Chapter 1; 4). Samples from two mappable gneiss bodies within the Boundary Ranges suite were dated to determine their ages: the Bighorn Creek granite (a metamorphosed granite) and the Wann River gneiss (a metamorphosed volcanic rock; this chapter). In addition, two dykes were collected; one because it provides a youngest age limit for some of the schists, and the other in hopes of determining the minimum age of movement on the Llewellyn fault zone.

Zircon and titanite fractions from three deformed plutons of the Tagish Lake suite were analysed to determine the maximum age for deformation caused by displacement on the Wann River shear zone. The plutons are included in the Tagish Lake suite because they are alkaline to calc-alkaline, hornblende-bearing, intrude the Boundary Ranges suite, and have similar ages (190 to 180 Ma). A dyke, which is also similar in age, was collected because of its intrusive relationship with the Wann River gneiss and because the dyke, the Wann River gneiss country rock and the Wann River shear zone foliation are folded (Plate 3.2a); the age of the dyke therefore provides a

maximum age for folding that postdates movement on the Wann River shear zone and a link between the Tagish Lake suite and the Wann River gneiss.

U-Pb Analytical Methods

U-Pb analyses of zircon and titanite were performed at the Geochronology Laboratory of the Geological Survey of Canada in Ottawa using analytical procedures summarized by Parrish et al. (1987; 1991). Blanks for U were 5 pg or less, and blanks for Pb ranged from 7-20 pg for zircon and from 20-300 pg for titanite analyses.

Some zircon analyses are discordant due to Pb-loss, ^{206}Pb deficiency, and/or the use of an incorrect initial common Pb composition. Discordance caused by inheritance of an older component of zircon has affected some zircon analyses, and is discussed below where appropriate.

Some analyses of zircon are discordant due to Pb-loss, even though most fractions had low to moderate U concentrations (Tables 2; 3) and all zircons were strongly abraded prior to dissolution (Krogh, 1982). Because the age of Pb-loss is uncertain, analyses of zircon from orthogneisses of the Boundary Ranges suite are difficult to interpret; if Pb-loss occurred in the past, the maximum age of crystallization of the zircon could be older than its $^{207}\text{Pb}/^{206}\text{Pb}$ age. Pb-loss in zircons from metamorphic rocks in the Tagish Lake area most likely occurred during a period of greenschist to amphibolite facies metamorphism and plutonism between 184 and 176 Ma (Chapter 1; 4); and the $^{207}\text{Pb}/^{206}\text{Pb}$ ages of zircon analyses for Boundary

Ranges suite orthogneiss samples have therefore been recalculated assuming a Pb-loss event at 175 Ma (Table 3.2).

Some zircon analyses exhibit an apparent deficiency of ^{206}Pb caused by exclusion of ^{230}Th from zircon during crystallization (Mattinson, 1973). Since ^{230}Th is an intermediate daughter product of the ^{238}U - ^{206}Pb decay chain (half life = $7.52 \times 10^4\text{Ma}$), the exclusion of ^{230}Th will result in a deficit of ^{206}Pb , and the $^{206}\text{Pb}/^{238}\text{U}$ age will be slightly less than the $^{207}\text{Pb}/^{235}\text{U}$ age. The method for correcting ^{206}Pb deficiency outlined in Schärer (1984) has been successfully applied to zircon analyses (Coleman and Parrish, 1991) and is applied to all analyses younger than 200 Ma presented here (Table 3.3); the correction is negligible for older zircon fractions and for titanite fractions (Table 3.3).

In many cases, even after correction for ^{206}Pb deficiency, the error ellipses for zircon fractions from the Tagish Lake suite are offset slightly to the right of concordia, possibly reflecting Pb-loss or the use of an inappropriate common Pb isotopic composition. In an attempt to improve the correction for common Pb, initial isotopic Pb compositions of feldspars from an approximately coeval pluton thought to have formed in a similar environment were used; the Katete Mountain quartz monzodiorite is from the Iskut River area, is ~192 Ma and is considered part of Stikinia (M.L. Bevier, pers. comm., 1991; Table 3.3)). However, these Pb isotope ratios are close to those determined using the Stacey and Kramers (1975) Pb isotope growth curve for 192 Ma, and do not significantly affect the common Pb correction for zircon analyses, which are apparently more strongly affected by Pb-loss (see below).

Discordance in titanite analyses is interpreted to be primarily due to the incorrect initial isotopic Pb composition (titanite incorporates more common Pb during crystallization than zircon), which outweighs the effect of Pb loss, even though titanite fractions were not abraded. The $^{206}\text{Pb}/^{238}\text{U}$ ages for titanite are considered more realistic than the $^{207}\text{Pb}/^{235}\text{U}$ ages; they are less sensitive to the initial common Pb composition than the $^{207}\text{Pb}/^{235}\text{U}$ ages because common Pb contains much less ^{206}Pb than ^{207}Pb . Errors for titanite analyses determined using the initial common Pb isotopic composition for the Katete Mountain quartz monzodiorite are smaller than those determined using the Stacey and Kramers (1975) Pb isotope growth curve because analytical errors associated with the Katete Mountain quartz monzodiorite values are smaller.

Mississippian Orthogneisses of the Boundary Ranges Suite

Bighorn Creek granite (A)

The Bighorn Creek granite is a light pink-weathering, garnet-bearing, meta-granite exposed in the Bighorn Creek valley (Table 3.1; Fig. 3.2; Plate 3.3a; Mihalynuk et al., 1989a; b). It is a moderately ($\sim 30^\circ$) southwest dipping, ~ 100 m thick tabular body that is exposed over an area of at least 1 km by 3 km. On Mt. Lawson it overlies felsic orthogneiss and in the Bighorn Creek valley, at the base of Hale Mountain, it is overlain by biotite and actinolite schist (Fig. 3.3, 4). The nature of its contacts with these rocks are unknown because they are not exposed. Microscopically, the orthogneiss exhibits abundant deformation textures, including irregular quartz and

feldspar grain boundaries, subgrain development within quartz grains and reduced grain size along feldspar grain boundaries, undulose extinction within quartz subgrains, and exsolution textures in feldspar. Penetrative macroscopic fabrics (e.g. foliation) are poorly, and only locally developed (Plate 3.3a). Relict garnet crystals were originally euhedral and are very fine-grained (<1 mm diameter, but visible in handsample); it is not known whether they are primary or metamorphic.

Four fractions of clear, euhedral prismatic zircons plot beside concordia near ~350 Ma (fractions B, C, D and G; Fig. 3.8a; Table 3.2); they are thought to have experienced minor Pb-loss. Fractions F and H have larger errors with younger Pb/U ages and are interpreted as having experienced the most Pb-loss; their intersection with concordia is considered fortuitous. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages for all of these fractions are rather consistent, yielding a ca. 360 Ma age, but the crystallization age of the zircons could be slightly older if they experienced Pb-loss during the intrusion of the nearby 187.7 ± 1.1 Ma Hale Mountain granodiorite (see below; Fig. 3.2) and/or during Jurassic metamorphism. Assuming the zircons experienced Pb-loss at ~175 Ma, they could be as old as 375 Ma (the maximum age for fraction B if Pb-loss occurred at 175 Ma; Table 3.2), which is regarded as the maximum age for the Bighorn Creek granite. The most reasonable estimate of a minimum crystallization age for zircons from the Bighorn Creek granite is derived from their $^{207}\text{Pb}/^{206}\text{Pb}$ ages because there is no evidence for a component of inherited zircon in any of the fractions. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages for the most precise analyses (B and C; Table 3.1) provide an interpreted minimum age for the Bighorn Creek granite of 357 Ma. An estimate of 366 ± 9 Ma

(latest Devonian to earliest Mississippian) is regarded as a reasonable age and error assignment.

Felsic dyke from the Wann River shear zone, west of Hoboe Creek (B)

A 30 cm wide, well foliated, feldspar-phyrlic, felsic tonalite dyke was collected from ~2 km west of Hoboe Creek and ~1 km north of the Caplice fault (Table 3.1; Figs. 3.2; 3.3; 3.6b). The dyke transects layering in actinolite schists of the Boundary Ranges suite and provides a youngest possible age for some schists of the Boundary Ranges suite (Plate 3.1b).

One concordant and two slightly discordant zircon analyses are dispersed along concordia between ~290 and 300 Ma (Table 3.2; Fig. 3.8b). The error ellipse for fraction A, which comprised large, subhedral, equant crystals, barely overlaps concordia between 324.9 Ma and 326.1 Ma, and provides a minimum age of ~325 Ma for the dyke. Assuming Pb-loss occurred at 175 Ma (the most likely age of a Pb-loss event, see above), the age of this fraction is calculated to be 332 ± 13 Ma and the maximum possible age is older than the $^{207}\text{Pb}/^{206}\text{Pb}$ age of this fraction (328.8 ± 6 Ma; Table 3.2). The crystallization age of this fraction, and therefore the dyke, could be as old as 345 Ma.

Crystals in fraction B comprised euhedral, clear, colourless prisms (length:width = <2:1), with minor inclusions. The age of this fraction, assuming Pb-loss at 175 Ma, is older than that of fraction A (Table 3.2), probably reflecting a component of inherited zircon. Zircons crystals in fraction C were euhedral, flat,

prisms (length: width = 2:1), and smaller than crystals of the concordant fraction (A); they are thought to have experienced more Pb-loss than crystals in fraction A due to their shape and smaller size.

The interpreted age of the dyke is 335 ± 10 Ma (Mississippian); it is derived from the overlap of the error ellipse of the concordant fraction with concordia (325 Ma) and the maximum permissible age for fraction A (345 Ma) calculated assuming Pb-loss occurred at ~ 175 Ma. This age limits the age of some schists of the Boundary Ranges suite to 325 Ma or older.

Felsic dyke with sinistral shear sense indicators (C)

A sample of a 30 cm wide, pink, foliated, porphyritic granitic dyke was collected to limit the age of movement along the Wann River shear zone. The dyke contains rotated porphyroclasts indicative of sinistral shear and a shear fabric that are interpreted to have formed during Early to Middle Jurassic movement on the Wann River shear zone (the Stikinia-Tracy Arm terrane boundary; Chapter 1). This sense of shear is consistent with that determined from rotated porphyroclasts and C-S fabrics in nearby rocks. The dyke intrudes chlorite-actinolite schists of the Boundary Ranges suite 0.5 km west of the Hoboe Creek Glacier and 1.5 km east of the contact between the Boundary Ranges suite and the Hale Mountain granodiorite (Fig. 3.2).

Four fractions of clear, euhedral, prismatic zircon crystals extracted from this sample (Table 3.2; Fig. 3.8c) have discordant U-Pb analyses that plot next to concordia between ~ 320 Ma and 340 Ma. The interpreted age for this dyke is derived

from their $^{207}\text{Pb}/^{206}\text{Pb}$ ages because the discordance is considered primarily due to Pb-loss, consistent with the correlation between the degree of discordance and U content. The youngest possible age for the dyke is 346.4 Ma (the minimum permissible age for fraction D). If Pb-loss occurred ~175 Ma ago (see above), the dyke could be as old as 357.3 Ma (the oldest permissible age for fraction B; Table 3.2). The age for this dyke is therefore interpreted to be 352 ± 5.5 Ma (early Mississippian); implicit in this interpretation are the assumptions that the fractions do not contain inherited zircon and all discordance is due to Pb-loss. This age supports a similar U-Pb zircon age for the Bighorn Creek granite, but provides no new data regarding the age of movement on the Llewellyn fault zone which is known to have been active after 102 Ma (Appendix A).

Permian Volcanic Rocks of the Boundary Ranges Suite

Wann River Gneiss (D)

The Wann River gneiss is an amphibolite facies volcanoclastic rock exposed in the footwall of the Wann River shear zone. It appears, on the basis of U-Pb data to be the stratigraphically highest, as well as the structurally highest exposed unit of the Boundary Ranges suite. It is recognised in the upper Wann River valley, on White Moose Mountain, on Mt. Switzer, and west of the Hoboe Glacier (Fig. 3.2). It may be more widespread, but at lower metamorphic grade it would be difficult to distinguish from older chlorite or actinolite schists of the Boundary Ranges suite.

The sample of Wann River gneiss used for U-Pb analysis was collected from a ~30 m thick sliver of Wann River gneiss that was imbricated with Tracy Arm Terrane rocks within the Wann River shear zone ~475 m above the main body of the Wann River gneiss and north of the upper Wann River (Figs. 3.2; 3.3). Zircon fractions from the Wann River gneiss comprised clear, elongate (length:width; 2-3:1) crystals with rounded edges, synneusis twins, and minor inclusions, but lacked visible cores (Table 3.2; Fig. 3.8d). Their analyses plot below concordia, reflecting inheritance and possible Pb-loss (Table 3.2; Fig. 3.8d). Regression of the data yields a lower intercept of 262.7 ± 0.8 Ma (MSWD=0.84), which can only be viewed as an approximate age of crystallization because of potential Pb-loss. The 270 ± 5 Ma $^{207}\text{Pb}/^{206}\text{Pb}$ age of the youngest fraction (Early Permian; Table 3.2) is regarded as a better estimate of the age of crystallization. The assigned age and error estimate take into account the $^{207}\text{Pb}/^{206}\text{Pb}$ ages of fractions A, C, D, and E (excluding error), and allow for the possibility that these fractions lacked an inherited component of zircon, but experienced minor Pb-loss.

Early Jurassic Plutonic Rocks: Tagish Lake Suite

Hale Mountain Granodiorite (E)

The Hale Mountain granodiorite is exposed on Hale Mountain, White Moose Mountain, at the base of Mt. Switzer, and west of Hoboe Glacier (Fig. 3.2). Typically, it is a medium- to coarse-grained hornblende ($\pm$ biotite) granodiorite with feldspar phenocrysts in a fine-grained matrix, which is green due to the presence of epidote and

actinolite (Table 3.1; Plate 3.1d). Locally, the Hale Mountain granodiorite has a penetrative foliation, in part mylonitic, that is concordant with the Wann River shear zone near its margins and along localized layers within the pluton.

A sample of the Hale Mountain granodiorite was collected near the top of Hale Mountain (Fig. 3.2). Six fractions of clear, euhedral, and elongate (length:width of 2-3:1) zircon crystals with minor rod-shaped, colourless inclusions, and no visible cores were analysed. With one exception, all of the fractions cluster between 184 and 186 Ma and plot in a linear array, slightly below concordia (Fig. 3.9a; Table 3.3). The zircon analysis above concordia was relatively poor and its reverse discordance is probably due to additional, unspecified analytical error in the Pb analysis. The slight normal discordance of the remaining zircon fractions can be accounted for in part by ^{206}Pb deficiency and in part may be due to Pb-loss. After correction for ^{206}Pb deficiency the zircon analyses plot closer to concordia and have $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 186.6 and 188.8 Ma. The age of the Hale Mountain granodiorite is interpreted to be 187.7 ± 1.1 Ma (Early Jurassic).

Analyses of two fractions of pale yellow, anhedral titanite crystals with minor inclusions have almost identical $^{206}\text{Pb}/^{238}\text{U}$ ages of 184.5 ± 0.8 Ma and 184.6 ± 0.8 Ma (Table 3.3; Fig. 3.9a). The U-Pb titanite age for the Hale Mountain granodiorite is therefore interpreted as 184.5 ± 0.9 Ma, which is only ~3 Ma younger than the interpreted U-Pb zircon age, indicating that the Hale Mountain granodiorite cooled through $\sim 600 \pm 50^\circ\text{C}$, the closure temperature for U and Pb in titanite (Tucker et al., 1987; Heaman and Parrish, 1991), soon after the granodiorite crystallized. As

expected, this titanite age is ~17 Ma older than a 165 ± 3 Ma - Bultman, 1979) K-Ar age for hornblende from the Hale Mountain granodiorite exposed ~10 km to the southeast, on the north shore of Tagish Lake, south of White Moose Mountain (the closure temperature for diffusion of Ar in hornblende is estimated to be lower than that of Pb in titanite ($500 \pm 50^\circ\text{C}$; Harrison, 1981). The difference in ages could reflect a difference in closure temperatures between titanite, which could be closer to 550°C and hornblende, which could be closer to 450°C ; or 17 Ma could have elapsed between the times that different parts of the Hale Mountain granodiorite cooled through $\sim 500^\circ\text{C}$; or one or both ages could be incorrect. The titanite age is considered more reliable than the K-Ar age because two Pb/U ages determined from two independent isotopic systems are in agreement. The K-Ar age is considered suspect because the Hale Mountain granodiorite was deformed after cooling to greenschist facies temperatures ($<450^\circ\text{C}$) and hornblende is known to lose Ar during deformation.

Felsic dyke that intrudes the Wann River gneiss (F)

West of the Hoboe Creek glacier, a 1 to 2 m wide, felsic, fine-grained tonalite dyke intrudes the Wann River gneiss (Table 3.1; Figs. 3.2; 3.4). The dyke, the country rock, and the foliation within the country rock are folded around a horizontal, upright, close fold (Plate 3.2a). The dyke lacks a well developed foliation, but grain size decreases along mm-scale shear surfaces, and quartz grains exhibit undulose extinction and have irregular grain boundaries.

U-Pb analyses for four zircon fractions from this dyke range from only slightly discordant to very discordant after correction for ^{206}Pb deficiency (Table 3.3; Fig. 3.9b). The two most discordant fractions (D and E) consist of euhedral crystals and fragments of prismatic crystals that are slightly metamict and obviously contain inherited zircon. Their U concentrations (>1000 ppm) are higher than those of less discordant fractions (<550 ppm; Table 3.3), probably reflecting the presence of older, uranium-rich, inherited zircon. Regression of data from all of the analyses yields an upper intercept of 2.1 Ga, which is interpreted as indicating that some of the inherited zircon is at least 2.1 Ga (Fig. 3.9b). This is not surprising because inherited Precambrian zircon has been documented in zircons from the Wann River gneiss country rock (see above).

The two least discordant fractions (A and B) consist of clear, colourless, euhedral, prismatic zircon crystals. Analyses of these fractions plot to the right of concordia due to Pb-loss, inheritance of older zircon, or a combination of both. It is likely that only minor recent Pb-loss affected these fractions because they were strongly abraded, contained only minor fractures, and had moderate U concentrations (Table 3.3). The $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$ ages for the most concordant fraction (A) are 191.7 ± 0.3 and 192.4 ± 0.6 Ma, respectively, and the lower intercepts of the regression lines between fractions A and B and fractions A and D are 191.3 Ma and 190.8 Ma, respectively. If a large portion of the inherited zircon component in fraction A was 270 Ma (the age of the Wann River gneiss), a lower intercept age younger than 190 Ma would be possible, but in the absence of strong evidence for 270 Ma inherited

zircon and in the light of the Pb/U ages for the most concordant fraction, it is assumed that most of the inherited zircon is Precambrian and the age of the gneiss is not younger than 190 Ma. The interpreted age of the dyke, relies on the linear regression of analyses A and B, and A and D, and is 191 ± 1 Ma.

The Early Jurassic age for this dyke constrains the age of some of the folding to older than 190 Ma. This age limit is reasonable because the foliation in the Wann River gneiss, which is also folded, is considered younger than 180 Ma because it is thought to have formed during displacement along the Wann River shear zone. Perhaps more importantly, the age of the dyke provides an intrusive link between the Early Jurassic Tagish Lake suite and the Boundary Ranges suite.

Hoboe Quartz Syenite (G)

A ~250 m by 500 m, lineated, weakly- to well-foliated hornblende quartz syenite is exposed southeast of Mt. Caplice and 3.5 km west of Hoboe Creek (Fig. 3.2; Table 3.1), where it is bounded on both sides by gradational contacts with actinolite schist of the Boundary Ranges suite (Fig. 3.6). Four zircon fractions from the Hoboe quartz syenite were analysed (Table 3.3; Fig. 3.9c); three were composed of equant to slightly elongate, multifaceted euhedral crystals (fractions A, B and C), and one consisted of euhedral, flat or prismatic, elongate (length:width of 2:1) crystals (fraction D). Analyses of two zircon fractions (B and D) plot near concordia between 186 Ma and 187.5 Ma, but are displaced to the right relative to concordia, after correction for ^{206}Pb deficiency. This discordance is likely partially due to inheritance

and in part caused by Pb-loss. Regression of these analyses and the analysis for fraction A indicates that fraction A, and possibly fractions B and D contained a component of older zircon, perhaps older than 1 Ga. The relatively older $^{207}\text{Pb}/^{206}\text{Pb}$ age of fraction C may also be due to the presence of inherited zircon, but its young Pb/U ages are interpreted to be the result of Pb loss (Table 3.3).

The youngest possible age of crystallization for the Hoboe quartz syenite is interpreted as the minimum overlap of the concordant fraction B with concordia (186.6 Ma), and its oldest age limit is derived from the $^{207}\text{Pb}/^{206}\text{Pb}$ age of fraction D (192.3 Ma), which is not thought to contain a significant inherited component. The interpreted age of the Hoboe quartz syenite is therefore 189.5 ± 3 Ma (Early Jurassic).

Titanite crystals from the Hoboe quartz syenite are slightly frosted, pale amber, anhedral and lack inclusions. The $^{206}\text{Pb}/^{238}\text{U}$ ages of two fractions overlap between 186.5 ± 0.4 Ma (Fig. 3.9c) and the interpreted age for titanite from this rock is 186.5 ± 0.4 Ma. This titanite date is within error of the interpreted crystallization age of the Hoboe quartz syenite (189.5 ± 3 Ma), indicating that this pluton also cooled relatively quickly following emplacement.

Mt. Caplice Quartz Monzodiorite (H)

The Mt. Caplice quartz monzodiorite is a medium-grained, weakly foliated and lineated, epidote-hornblende quartz monzodiorite (Table 3.1) that is exposed over an area of ~ 5 km², southeast of Mt. Caplice and 4.5 km west of Hoboe Creek (Fig. 3.2). The southern margin of the quartz monzodiorite is unexposed in a stream bed and is

considered an extension of the Caplice fault; its northern margin is in intrusive contact with an undeformed, mid-Cretaceous quartz monzodiorite (Currie, 1992a); and its western margin is interpreted as an extension of the Wann River shear zone because it is juxtaposed against the Florence Range suite (Figs. 3.2; 3.6). Its contact with the Boundary Ranges suite to the east has not been observed.

Three zircon and two titanite fractions from the Mt. Caplice quartz monzodiorite were analysed (Table 3.3; Fig. 3.9d). The zircon fractions comprised large, colourless, very clear, euhedral, prismatic crystals with minor inclusions. All of the analyses plot to the right of concordia. Pb-loss, rather than inheritance of older zircon, is regarded as the most likely cause of this discordance, because there is no textural evidence for inheritance. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages (after correction for ^{206}Pb deficiency) are considered most reliable. The maximum age for the quartz monzodiorite is 184.1 Ma (the maximum $^{207}\text{Pb}/^{206}\text{Pb}$ age for Fraction C), its minimum age is 177.8 Ma (the minimum age $^{207}\text{Pb}/^{206}\text{Pb}$ for Fraction A), and its interpreted age of crystallization is 180.9 ± 3.1 Ma (Early Jurassic).

Analyses for two fractions of anhedral, clear, pale to dark, amber titanite crystals from the Mt. Caplice quartz monzodiorite are concordant, overlap one another and have $^{206}\text{Pb}/^{238}\text{U}$ ages of 176.6 ± 1.2 Ma and 177.0 ± 0.6 Ma (Fig. 3.9d). On the basis of the best analysis (S2) the Mt. Caplice quartz monzodiorite is thought to have cooled through about $\sim 600 \pm 50^\circ\text{C}$ (the closure temperature for U and Pb diffusion in titanite; Tucker et al., 1987; Heaman and Parrish, 1991) at 177.0 ± 0.6 Ma, soon after the Mt. Caplice quartz monzodiorite crystallized.

An Rb/Sr isochron age of 235 Ma for a pluton in the Mt. Caplice area was reported by Werner (1978) and has been used to suggest that age of metamorphism in the rocks the pluton intrudes is pre-Late Triassic (Werner, 1978; Monger et al., 1991). Although the Rb/Sr sample location has not been revisited, it is considered part of the Mt. Caplice quartz monzonite on the basis of mapping in adjacent areas. The Rb/Sr date is incorrect and unsupported because it differs from the zircon date for the Mt. Caplice quartz monzodiorite (180.9 ± 3.1) and is derived from only two points (Werner, 1978): one from a biotite hornblende diorite and one from an aplite dyke that intruded the diorite (unpublished data; R.L. Armstrong, pers. comm., 1991).

APPENDIX 3.2. INITIAL $^{87}\text{Sr}/^{86}\text{Sr}$ RATIOS

Sample	Lithologic Unit	Easting	Northing	ppm Sr	ppm Rb	^{87}Rb ^{87}Sr	^{87}Sr ^{86}Sr	Age(Ma)	$^{87}\text{Sr}(\text{i})^1$ ^{86}Sr
LW76-a68A ²	Wann River Gneiss	547918	6552207	414	5.9	0.041	0.7064	262 ±5	0.7062
LW76-a76c ²	Mt. Caplice Granodiorite	545723	6560161	1081	95.9	0.257	0.7056	180.9 ±3.1	0.7049
LW76-b22B ²	East of LF Muscovite schist	560238	6564075	234	42.1	0.521	0.7071		
LW76-b76c ²	Mt. Caplice Felsic dyke	545723	6560161	449	113	0.728	0.7073	180.9 ±3.1	0.7054
T75-323-6 ²	Hale Mtn Granodiorite	538892	6588298	997	36.3	0.104	0.7055	185 ±1	0.7052

¹decay constants are from Steiger and Jager (1977); ²unpublished data collected at University of British Columbia, 1970-1985, R.L. Armstrong, pers comm., 1991).

Figure 3.1. Terrane map of northwest British Columbia, southwest Yukon and southeast Alaska (modified from Wheeler and McFeely, 1991 and Wheeler et al., 1991). Paleozoic rocks and Late Triassic to Middle Jurassic plutons of Stikinia, Cretaceous to Tertiary plutons and the location of Figure 3.2 are also shown. KSF = King Salmon Fault; LFZ = Llewellyn fault zone. Inset: AK = Alaska; BC = British Columbia; YT = Yukon Territory.

Figure 3.2. Geologic map of the Tagish Lake area (based on Mihalynuk et al., 1989a, b; 1990; Currie, 1990a; 1991; 1992b). For simplicity, the imbricated boundary between Tracy Arm terrane and Stikinia is labelled "WRSZ", even though the shear-related fabrics that characterize the Wann River shear zone affect rocks on either side of this contact. U-Pb analyses A to H are presented in Appendix 3.1, and cross-sections V-V' to Z-Z' are shown in Fig. 3.3. CCF = Chicken Creek fault.

Figure 3.3. Cross-sections for Mt. Lawson, (V-V'), white Moose Mountain to Wann River (W-W'), Mt. Switzer (X-X'), Mt. Caplice (Y-Y') and Hoboe Creek (Z-Z'). Section locations are shown in Figure 3.2. "line of section refers to the transect used to construct the apparent "stratigraphic sections" of Figure 3.4. CF = Caplice Fault; CCF = Chicken Creek fault; TLS in the Mt. Lawson section = homblende granodiorite tentatively included in the Tagish Lake suite. As in Figure 3.1, the imbricated contact between Tracy Arm terrane and Stikinia is labelled "WRSZ"; event though the shear-related fabrics that characterize the Wann River shear zone affect rocks on either side of this contact.

Figure 3.4. "Stratigraphic sections" for Mt. Lawson, Mt. Switzer, Mt. Caplice, the Hoboe Creek and Stikine River areas. The first four sections are derived from cross sections shown in Figure 3.3. Structural thickening and thinning have not been taken into account.

Figure 3.5. Structural domain map for Boundary Ranges and Tagish Lake suite rocks, with associated equal area stereonet projections of poles to foliation for each domain. Open circles represent mean poles to foliation. Grey areas on the location map represent different structural domains discussed in text. Patterns and structural symbols correspond to those of Figure 3.2. CF = Caplice fault; CCF = Chicken Creek fault; WRSZ = Wann River shear zone.

Figure 3.6. Geologic map of the Hoboe Glacier-Mt. Caplice area (see Fig. 3.2 for location). MMSZ = Mt. Mussen shear zone; WRSZ = Wann River shear zone.

Figure 3.7. Geologic map of the Swanson River headwaters (see Fig. 3.2 for location).

Figure 3.8. U-Pb concordia plots for zircon fractions from the Boundary Ranges suite: (a) the Bighorn Creek granite (A in Figure 3.2 and Table 3.2); (b) a dyke that intrudes Boundary Ranges suite (B in Figure 3.2 and Table 3.2; Plate 3.1b); (c) a felsic dyke with sinistral fabrics from west of Hoboe Creek and within the Wann River shear zone (C in Figure 3.2 and Table 3.2); and (d) the Wann River gneiss (D in Figure 3.2 and Table 3.2).

b

d

c

a

Figure 3.9. U-Pb concordia plots for zircon and titanite fractions from the Tagish Lake suite: (a) the Hale Mountain granodiorite (E in Figure 3.2 and Table 3.3; S1 and S2 are titanite fractions); (b) a dyke that intrudes Wann River gneiss (F in Figure 3.2 and Table 3.3; note that both the dyke and the country rock are folded; Plate 3.3b); (c) the Hoboe quartz syenite (G in Figure 3.2 and Table 3.3); and (d) the Mt. Caplice quartz monzodiorite (H in Figure 3.2 and Table 3.3; inset shows error ellipses for zircon fractions; S1 and S2 are titanite fractions). All analyses have been corrected for ^{206}Pb deficiency.

Plate 3.1a. Polydeformed chlorite-actinolite schist of the Boundary Ranges suite.

Multiple phases of deformation and metamorphism have hampered stratigraphic characterization and correlation of the Boundary Ranges suite.

Plate 3.1b. The right side of this Mississippian felsic dyke (B in Fig. 3.1 and

Appendix 3.1; 335 ± 10 Ma) truncates layering in actinolite schist of the

Boundary Ranges suite rocks, limiting the ages of some Boundary Ranges suite rocks to 325 Ma or older.

Plate 3.1c. The Permian Wann River gneiss (D in Fig. 3.1 and Appendix 3.1; 270 ± 5

Ma) is interpreted to be a metamorphosed volcanoclastic rock on the basis of its graded compositional layering, composition and association with metasedimentary rocks.

Plate 3.1d. Typical, weakly deformed, Hale Mountain granodiorite (Tagish Lake suite)

with feldspar phenocrysts in a fine-grained, green, hornblende-epidote matrix

(E in Fig. 3.1 and Appendix 3.1; 187.7 ± 1.1 Ma).

Plate 3.1~~d~~^e. Tagish Lake suite dyke (F in Fig. 3.1 and Appendix 3.1; 191 ± 1 Ma) that

intrudes the Wann River gneiss west of Hoboe Glacier. Both the dyke and the country rock are folded.

Plate 3.1~~e~~^f. Pyroxene-phyrlic actinolite schist of the Boundary Ranges suite from Mt.

Lawson, west of the Chicken Creek fault.

Plate 3.2a. Interlayered actinolite schist and fine-grained feldspar-rich layers,

characteristic of the Boundary Ranges suite, are interpreted as bimodal tuffs.

Plate 3.2b. The felsic Mt. Lawson orthogneiss is interpreted as a metamorphosed felsic

tuff on the basis of its composition and well-developed layering.

Plate 3.2c. The granitic Bighorn Creek granite (A in Fig. 3.1 and Appendix 3.1; 366 ± 9

Ma) lacks well-developed, penetrative, mesoscopic fabrics. It probably acted as a rigid block while adjacent rocks were deformed.

Plate 3.2d. Interlayered Permian Wann River gneiss and Early Jurassic Hale Mountain

granodiorite. Their contact is interpreted as a sheared intrusive contact.

Plate 3.2e. Plagioclase-phyric actinolite-chlorite schist of the Boundary Ranges suite

from west of Hoboe Glacier (Sample LC-90-322; Table 3.1).

Plate 3.2f. Folded felsic orthogneiss of the Boundary Ranges suite is associated with

hornblende-plagioclase gneiss, schist and marble in the northern Wann River headwaters.

Table 3.1. Lithologic descriptions of representative rocks belonging to the Boundary Ranges suite, the Wann River gneiss and the Tagish Lake suite of the Tagish Lake area.

LITHOLOGY ¹		MINERALOGY ²		CHARACTERISTICS ²	
Boundary Ranges suite					
muscovite-chlorite phyllite					
LC-90-351		25% Qtz, 20% Ms, 15% Chl, 10% 20% Pl, 7% Cal, 3% Py		Fine-grained schist with well-developed foliation defined by alignment of Ms and mm scale Qtz-Pl layers separated by thin Ms-rich layers. Within the micaceous layers shear fabrics are preserved and within Qtz-Pl layers cataclatic texture with Cc matrix is present. Mortar texture around Qtz and Pl grains and undulose extinction in Qtz.	
actinolite schist					
LC-88-14		50% Act, 30% Pl, 10% Ep, 5% Hb, 5% Cal		Fine-grained green schist with medium-grained Hb and Px porphyroclasts. Px is commonly replaced by Ep. Foliation defined by alignment of Act, annealed Qtz ribbons, Chl, and Ms. Cal occurs as mm scale veins. Pl is polycrystalline.	
north shore of Fantail Lake		50% Pl, 20% Qtz, 15% Chl, 10% Ms, 5% Op.		Fine-grained green schist with medium-grained Pl and Op porphyroclasts. Well developed foliation defined by Chl, Ms and recrystallized Qtz ribbons. Pressure shadows around Op grains filled by Qtz. Pl porphyroclasts have inclusion-rich cores and inclusion-free rims.	
LC-90-258b		60% Pl, 35% Hb, 3% CPx, 2% Ep, acc. Chl, Clinzoisite		Green, fine-grained actinolite schist with medium-grained Hbl and Px porphyroclasts. Strong shear fabric is defined by crystallographic alignment of Hbl and compositional layering. Hbl porphyroclasts have Hbl and Pl tails and some are twinned.	
Mt. Lawson		60% Pl, 30% Qtz, 10% Ms, acc. graphite, Op, Ap.		Melanocratic schist with well-developed foliation defined by alignment of muscovite. Qtz exhibits well-developed sub-grains and irregular grain boundaries.	
graphitic schist		35% Ms, 30% Qtz, 25% Bt, 3% Ky, 2% St, 2% Gar, acc graphite		Fine to medium-grained melanocratic schist. Composition varies from Qtz-rich to Qtz-poor on mm scale. Crenulated foliation defined by alignment of micas. Rotated 1-2 cm euhedral, altered, Grt with Qtz tails are wrapped by foliation. Ky folded by F2. St and some Bt overprint foliation	
LC-90-286d		50% Pl, 25% Chl, 20% Act, 5% Ep, acc. Ti, Rt.		Medium-grained Pl augen surrounded by fine-grained green matrix of Chl, Act, and Ep. Pl porphyroclasts altered, except in cores. Foliation defined by Chl, although some Chl cross-cuts foliation.	
Mt. Lawson		40% Qtz, 30% Or, 20% Pl, minor Hb, Ms, Chl, Ep, acc Zrn, Grt.		Massive, medium-grained, pink orthogneiss. Feldspar and mica crystals exhibit undulose extinction and Qtz is fine-grained and polycrystalline, with serrate grain boundaries. Relict Grt is fine-grained.	
chlorite schist		40% Pl, 40% Qtz, 20% Ms, acc. Zr, Ap		Fine- to medium-grained felsic gneiss with well developed foliation. Recrystallized ribbons of Qtz wrap around slightly altered, medium-grained Pl porphyroclasts. Qtz subgrains have serrate boundaries. Rotated porphyroclasts (3mm diameter) and C-S fabrics indicate sinistral sense of shear.	
LC-90-322		50% Pl, 30% Qtz, 20% Or, acc Zr.		Pink-weathering Or-porphyry gneiss. Or porphyroclasts (up to 0.5 cm) are slightly altered and are surrounded by medium-grained Pl porphyroclasts and a fine-grained Pl and Qtz matrix. Relict Qtz ribbons are preserved as ribbons of the porphyroclasts.	
west of Hoboe Glacier		55% Qtz, 30% Pl, 10% green Bt, 5% Ms, acc Ti, Ap, Zrn, Op.		Well layered, fine-grained leucocratic gneiss with well developed foliation defined by micas. Micaceous layers are interlayered with quartzofeldspathic layers on mm scale. Qtz exhibits undulose extinction and curved grain boundaries. Elongate Bt and Ms grains are parallel with foliation; blocky Bt grains x-cut foliation. Foliation is folded by open cm scale folds.	
felsic orthogneisses		50% Qtz, 30% Pl, 10% green Bt, 5% Chl, 5% Or, acc Ti.		Fine-grained felsic gneiss with poorly-developed foliation. Qtz and Pl exhibit undulose extinction and Qtz has formed subgrains. Relict Or porphyroclasts are anhedral, are slightly larger than other grains, and have Qtz inclusions.	
Bighorn Creek orthogneiss		50% Pl, 50% Hb, acc. Py.		Medium-grained, mottled, green and white leucogabbro with moderately developed foliation defined by flattened Pl grains. Hb has been replaced by Chl and Ep.	
LC-88-28-1 (A; 366 ± 9 Ma)					
Felsic gneiss					
(LC-90-304)					
(B; 335 ± 10 Ma)					
Pink orthogneiss					
(LC-90-267b)					
(C; 352 ± 5.5 Ma)					
Mt. Lawson gneiss					
LC-90-287					
Felsic orthogneiss from Swanson River headwaters					
LC-90-205					
leucogabbro from west of Hoboe Glacier					
LC-90-302c					

1. Capital letters (A, B, C, etc.) are referred to in Figure 2 and in Tables 2 and 3; 2. acc. - accessory mineral (comprises <1% of rock). Act - actinolite, Ap - apatite, Bt - biotite, Cal - Calcite, Chl - chlorite, Ep - epidote, Hbl - hornblende, Grt - garnet, Ms - muscovite, Pl - plagioclase, Py - pyrite, Op - opaque minerals, Or - orthoclase, Qtz - quartz, Rt - rutile, St - staurolite, Ti - titanite, Zrn - zircon.

Table 3.1 (continued). Lithologic descriptions of representative rocks belonging to the Boundary Ranges suite, the Wann River gneiss and the Tagish Lake suite of the Tagish Lake area.

LITHOLOGY AND AGE¹ MINERALOGY²

Boundary Ranges suite (continued)

Sliver of Wann River gneiss (D; 270 ± 5 Ma). LC-89-91b.	40% Hbl, 40% Pl, 17% Qtz, 2% Ep, 1% green Bt, acc. Ti, Zrn, Rt.	Green, medium-grained mesocratic gneiss with compositional layering on cm to m scale (Plate 2c). Mineral lineation defined by alignment of Hbl. Qtz and Pl are recrystallized and do not preserve evidence of deformation.
Wann River gneiss from north of upper Wann River. LC-90-458b.	55% Plg, 25% Hbl, 20% Bt, 1% Ti.	Green, medium-grained, mesocratic gneiss with compositional layering on mm to cm scale (Plate 2c). Granoblastic texture (recrystallized). Minor mm-scale fractures with reduced grain size along boundaries.

Tagish Lake suite

Hale Mountain granodiorite (E; 186 ± 1.5 Ma). LC-88-35-1.	60% Pl, 20% Qtz, 13% Ep, 5% Hbl, 2% Ti, acc. Zrn	Green, medium- to coarse-grained granodiorite with Pl phenocrysts (porphyroclasts where deformed; Plate 2b); green colour due to presence of Ep. Coarse grained equivalents have Hbl and Pl crystals up to 5 cm long. Pl is commonly sericitized, and where it is deformed, it is dynamically recrystallized and wrapped by polycrystalline Qtz (recrystallized ribbons). Ep replaces Hbl and does not appear to be igneous.
Hale Mountain granodiorite LC-90-269	55% Pl, 20% Hbl, 15% Qtz, 8% Chl, 2% Ep, acc. Ti, Rt.	Green, medium-grained, mesocratic Hbl monzoniorite. Well developed foliation defined by alignment of Hbl, recrystallized flattened Qtz grains, and Chl.
LC-90-272	50% Pl, 20% Hbl, 15% Qtz, 10% Bt, 5% Or, acc. Ti, Op.	Green and white granodiorite with well-developed foliation. Variable grain size (fine- to medium-grained), possibly due to grain-size reduction, rather than a primary texture. Pl present in the form of medium-grained porphyroclasts (may have Qtz tails). Qtz ribbons are partially annealed; in places subgrains have developed, and in others the ribbons are recrystallized.
Hoboe quartz syenite (G; 189.5 ± 3 Ma). LC-90-314.	45% Or, 30% Hbl, 10% Qtz, 5% Pl (An 24-31), 2% Ep, 2% Bt, 1% Ti, acc. Zrn, Rt	Mesocratic, medium grained Qtz syenite gneiss with weak mineral lineation defined by Hbl.
Folded felsic dyke (F; 191 ± 1 Ma). LC-89-273.	50% Qtz, 40% Pl, 10% Or, minor white mica and Chl, acc. Zrn	Leucocratic, fine-grained felsic dyke that intrudes the Wann River gneiss (Plate 2d). Both the dyke and the country rock are folded. Grain size is reduced along shear surfaces, but dyke lacks well developed foliation. Qtz has undulose extinction and irregular grain boundaries.
Mt. Caplice quartz monzoniorite (G; 180.9 ± 3.1 Ma). LC-90-343.	40% Qtz, 35% Pl (An 29), 15% Hbl, 9% Bt, 1% Ti, acc. Zrn, Rt	Medium-grained, mesocratic hornblende quartz monzoniorite with granitic texture. Subgrain development and undulose extinction in Qtz. Variable degree of alignment of Hbl.
Tagish Lake suite ? LC-91-437 Lineated granite from Wann River valley	40% Qtz, 25% Pl, 10% Or (minor alteration). 15% Bt (green), 5% Ep, acc. Ti, Ap.	Lineated (?), medium-grained, biotite granite with striped appearance (mm-scale). Lineation defined by elongation of Qtz and Pl, and composition. Granite has granitic texture, but curved to wavy grain boundaries perpendicular to lineation. Qtz, Pl and Or exhibit undulose extinction.

1. Capital letters (D, E,etc.) are referred to in Figure 2 and in Tables 2 and 3; 2. acc. - accessory mineral (comprises <1% of rock), Act - actinolite, Ap - apatite, Bt - biotite, Cal - Calcite, Chl - chlorite, Ep - epidote, Hbl - hornblende, Grt - garnet, Ms - muscovite, Pl - plagioclase, Py - pyrite, Op - opaque minerals, Or - orthoclase, Qtz - quartz, Rt - rutile, St - staurolite, Ti - titanite, Zrn - zircon.

Table 3.2. U-Pb data for zircon fractions from Mississippiian orthogneisses and the Permian Wann River gneiss of the Boundary Ranges metamorphic suite.

Fraction size ¹	Wt. μ g	U ppm	Pb ² ppm	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	Pbc ⁴ pg	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$ %	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	Corr. Coeff.	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	207/206 Age ⁶ age (Ma)	207/206 Age ⁷ age (Ma)
MISSISSIPPIAN ORTHOGNEISSES												
(A) Bighorn Creek orthogneiss (LC-88-28-1), medium-grained pale pink metagranite from north of Bighorn Creek, UTM 8v 531700E 6599400N.												
B (105)	24	374	21.50	2737	12	12.1	0.05572 $\pm$ 0.09%	0.4127 $\pm$ 0.14%	0.7746	0.05371 $\pm$ 0.09%	359.0 $\pm$ 3.9	367.4 $\pm$ 7.2
C (105)	25	456	26.36	3590	11	12.5	0.05582 $\pm$ 0.10%	0.4136 $\pm$ 0.13%	0.8344	0.05374 $\pm$ 0.07%	360.1 $\pm$ 3.3	369.0 $\pm$ 6.0
D (105)	20	386	22.44	422	68	13.0	0.05571 $\pm$ 0.14%	0.4124 $\pm$ 0.41%	0.6581	0.05369 $\pm$ 0.33%	358.0 $\pm$ 15.1	365.6 $\pm$ 28
F (105-149)	19	407	22.72	327	85	11.8	0.05436 $\pm$ 0.17%	0.4012 $\pm$ 0.51%	0.6445	0.05352 $\pm$ 0.42%	351.0 $\pm$ 18.9	360.2 $\pm$ 36
G (105-149)	13	449	26.12	1772	12	13.0	0.05586 $\pm$ 0.13%	0.4137 $\pm$ 0.19%	0.6228	0.05371 $\pm$ 0.15%	359.0 $\pm$ 6.9	366.7 $\pm$ 13
H (105-149)	12	483	26.93	373	55	11.9	0.05418 $\pm$ 0.21%	0.4010 $\pm$ 0.57%	0.5667	0.05368 $\pm$ 0.48%	357.5 $\pm$ 21.9	374.0 $\pm$ 41
(B) Felsic dyke (LC-90-304), felsic dyke that truncates layering in chlorite schists of Boundary Ranges metamorphic suite from west of Hoboe Creek glacier UTM 8v 549250E 6556150N.												
A (+149)	47	207	10.30	1628	19	5.9	0.05179 $\pm$ 0.11%	0.3785 $\pm$ 0.19%	0.6876	0.05300 $\pm$ 0.14%	328.8 $\pm$ 6.4	332.2 $\pm$ 13
B (+149)	44	167	8.62	869	29	7.2	0.05275 $\pm$ 0.10%	0.3885 $\pm$ 0.23%	0.6371	0.05342 $\pm$ 0.18%	346.7 $\pm$ 8.1	362.2 $\pm$ 16
C (105)	17	181	8.05	772	12	7.4	0.04546 $\pm$ 0.16%	0.3293 $\pm$ 0.41%	0.4716	0.05253 $\pm$ 0.37%	308.8 $\pm$ 16.6	340.1 $\pm$ 39
(C) Pink orthogneiss (LC-90-267b), pink mylonitized orthogneiss with sinistral shear sense indicators that intrudes chlorite schists of Boundary Ranges metamorphic suite from west of Hoboe Creek glacier, UTM 8v 550425E 6552000N												
A (+149)	171	600	32.27	1532	223	12.0	0.05226 $\pm$ 0.11%	0.3836 $\pm$ 0.16%	0.8175	0.05324 $\pm$ 0.10%	339.0 $\pm$ 4.3	349.9 $\pm$ 8.4
B (+149)	83	994	51.76	2600	102	11.3	0.05099 $\pm$ 0.10%	0.3738 $\pm$ 0.13%	0.8898	0.05316 $\pm$ 0.06%	335.7 $\pm$ 2.7	351.9 $\pm$ 5.4
C (+149)	33	774	42.21	1260	68	11.7	0.05313 $\pm$ 0.11%	0.3921 $\pm$ 0.19%	0.6957	0.05352 $\pm$ 0.14%	350.9 $\pm$ 6.2	367.9 $\pm$ 12
D (+149)	47	538	30.18	3119	28	12.3	0.05430 $\pm$ 0.09%	0.4004 $\pm$ 0.12%	0.8486	0.05348 $\pm$ 0.06%	349.3 $\pm$ 2.9	357.3 $\pm$ 5.4
PERMIAN												
(D) Wann River Gneiss (LC-89-918), compositionally layered hornblende-plagioclase gneiss from south of the upper Wann River, UTM 8v 6573525N 536750E.												
A (+74)	65	217	9.9	1970	19	16.9	0.04181 $\pm$ 0.09%	0.2978 $\pm$ 0.16%	0.6998	0.05166 $\pm$ 0.11%	270.4 $\pm$ 5.2	
B (+74)	52	185	9.43	3726	7	17.8	0.04606 $\pm$ 0.09%	0.3468 $\pm$ 0.13%	0.7750	0.05461 $\pm$ 0.08%	396.4 $\pm$ 3.7	
C (+74)	33	251	11.63	1974	11	18.1	0.04194 $\pm$ 0.12%	0.2993 $\pm$ 0.18%	0.6643	0.05177 $\pm$ 0.13%	275.2 $\pm$ 6.2	
D (+74)	22	218	10.07	1187	11	17.5	0.04211 $\pm$ 0.11%	0.3004 $\pm$ 0.26%	0.5959	0.05173 $\pm$ 0.22%	273.5 $\pm$ 9.9	
E (+74)	25	186	8.72	471	28	18.4	0.04224 $\pm$ 0.12%	0.3021 $\pm$ 0.45%	0.6086	0.05187 $\pm$ 0.39%	279.8 $\pm$ 17.9	

Notes: ¹sizes (i.e. +105) refer to length aspect of crystals in microns; ²radiogenic Pb; ³measured ratio, corrected for spike and fractionation; ⁴total common Pb in analysis corrected for spike and fractionation; ⁵corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; ⁶corrected for blank and common Pb, errors are two sigma in Ma; decay constants are those of Steiger and Jager (1977); for analytical details see Parrish et al. (1987). ⁷age is calculated assuming Pb-loss occurred 175 Ma ago.

Table 3.3. U-Pb data for zircon and titanite fractions from plutons of the Early Jurassic Tagish Lake suite.

Fraction size ¹	Wt. μ g	U ppm	Pb ² ppm	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	Pbc ⁴ pg	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	Corr. Coeff.	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$	207/206 Age ⁶ age (Ma)	207/206 Age ⁷ age (Ma)	
(E) Hale Mountain Granodiorite (LC-88-35-1): variably deformed hornblende granodiorite from Hale Mountain, UTM 6596500N, 531925E.												
zircon												
A (+149)	169	405	11.1	4267	30	11.2	0.02919 $\pm$ 0.08%	0.2008 $\pm$ 0.10%	0.8572	0.04989 $\pm$ 0.05%	190.0 $\pm$ 2.5	188.8
B (+149)	182	291	8.72	4299	23	12.6	0.02904 $\pm$ 0.09%	0.1996 $\pm$ 0.11%	0.8954	0.04984 $\pm$ 0.05%	187.4 $\pm$ 2.4	186.9
C (+149)	260	398	11.0	4010	48	11.6	0.02919 $\pm$ 0.15%	0.2001 $\pm$ 0.16%	0.8911	0.04972 $\pm$ 0.07%	181.8 $\pm$ 3.4	180.7
D (+149)	211	390	11.5	5311	28	11.6	0.02900 $\pm$ 0.09%	0.1994 $\pm$ 0.11%	0.9158	0.04987 $\pm$ 0.04%	188.9 $\pm$ 2.1	187.7
E (+149)	182	431	11.8	8711	16	11.6	0.02897 $\pm$ 0.11%	0.1991 $\pm$ 0.12%	0.9303	0.04986 $\pm$ 0.05%	188.4 $\pm$ 2.1	186.6
F (+149)	110	382	11.4	5756	13	11.6	0.02909 $\pm$ 0.09%	0.2000 $\pm$ 0.12%	0.8591	0.04987 $\pm$ 0.06%	188.8 $\pm$ 2.9	187.5
titanite												
S1 (+149)	256	222	8.24	161	736	34.4	0.02900 $\pm$ 0.20%	0.2019 $\pm$ 0.10%	0.7652	0.05048 $\pm$ 0.88%	217.3 $\pm$ 41	217.2
S2 (+149)	310	172	6.86	146	773	38.9	0.02901 $\pm$ 0.21%	0.2026 $\pm$ 0.11%	0.7664	0.05066 $\pm$ 0.94%	225.5 $\pm$ 44	224.6
(F) Felsic dyke (LC-90-273), folded felsic dyke that intrudes Wann River gneiss from west of Hoboe Creek glacier, UTM 8v 549225E 6553100N.												
zircon												
A(105-149)	49	524	15.0	2140	23	4.9	0.03018 $\pm$ 0.09%	0.2087 $\pm$ 0.18%	0.7105	0.05015 $\pm$ 0.13%	201.9 $\pm$ 6.0	201.7
B(105-149)	53	414	11.8	588	75	3.3	0.03040 $\pm$ 0.09%	0.2151 $\pm$ 0.21%	0.6729	0.05131 $\pm$ 0.17%	254.8 $\pm$ 7.7	254.5
D(105-149)	142	1120	47.4	723	576	13.4	0.03969 $\pm$ 0.19%	0.3914 $\pm$ 0.21%	0.9578	0.07151 $\pm$ 0.06%	972.1 $\pm$ 2.5	972.6
E(105-149)	30	1795	75.2	2652	51	13.9	0.03908 $\pm$ 0.09%	0.3937 $\pm$ 0.11%	0.9234	0.07307 $\pm$ 0.04%	1015 $\pm$ 1.8	1015
(G) Hoboe quartz syenite (LC-90-314), weakly foliated hornblende quartz monzoniorite from Mt. Caplice, UTM 8v 547850E 6556325N.												
zircon												
A(105-149)	165	131	4.04	1426	29	12.8	0.02970 $\pm$ 0.09%	0.2060 $\pm$ 0.20%	0.6833	0.05031 $\pm$ 0.15%	209.4 $\pm$ 7.2	208.3
B(105-149)	171	136	4.13	810	55	12.9	0.02936 $\pm$ 0.17%	0.2025 $\pm$ 0.24%	0.6048	0.05003 $\pm$ 0.19%	196.2 $\pm$ 8.8	195.1
C(105-149)	103	130	3.93	899	28	13.7	0.02878 $\pm$ 0.10%	0.1989 $\pm$ 0.25%	0.6407	0.05011 $\pm$ 0.21%	200.1 $\pm$ 9.5	199.0
D(105-149)	89	132	4.19	984	23	15.9	0.02943 $\pm$ 0.11%	0.2027 $\pm$ 0.28%	0.6217	0.04996 $\pm$ 0.23%	193.4 $\pm$ 10	192.3
titanite												
S1 (+149)	251	151	5.61	215	364	28.2	0.02935 $\pm$ 0.10%	0.2017 $\pm$ 0.30%	0.6149	0.04984 $\pm$ 0.25%	187.7 $\pm$ 11	186.9
S2 (+149)	322	193	7.28	229	558	29.3	0.02932 $\pm$ 0.10%	0.2022 $\pm$ 0.27%	0.6183	0.05002 $\pm$ 0.22%	195.9 $\pm$ 10	195.2
(H) Mt. Caplice (LC-90-343), weakly lineated hornblende quartz monzoniorite from east of Mt. Caplice, UTM 8v 564825E 6557750N.												
zircon												
A (+149)	196	294	8.65	2831	36	14.3	0.02796 $\pm$ 0.09%	0.1916 $\pm$ 0.13%	0.8401	0.04970 $\pm$ 0.07%	181.2 $\pm$ 3.4	180.8
B (+149)	97	293	8.65	1369	37	14.5	0.02791 $\pm$ 0.11%	0.1915 $\pm$ 0.19%	0.7134	0.04976 $\pm$ 0.13%	184.0 $\pm$ 6.2	182.9
C (+149)	180	327	9.64	4789	22	14.2	0.02796 $\pm$ 0.09%	0.1917 $\pm$ 0.12%	0.8729	0.04973 $\pm$ 0.06%	182.4 $\pm$ 2.8	181.3
titanite												
S1 (+149)	268	595	15.3	161	1995	2.2	0.02777 $\pm$ 0.12%	0.1911 $\pm$ 0.49%	0.6465	0.04991 $\pm$ 0.43%	190.9 $\pm$ 20	189.5
S2 (+149)	286	786	20.4	439	960	3.1	0.02784 $\pm$ 0.12%	0.1908 $\pm$ 0.18%	0.7859	0.04971 $\pm$ 0.11%	181.4 $\pm$ 5.3	180.2

Notes: ¹sizes (i.e. +105) refer to length aspect of crystals in microns; ²radiogenic Pb; ³measured ratio, corrected for spike and fractionation; ⁴total common Pb in analysis corrected using initial common Pb isotopic compositions from feldspars from Katete Mountain quartz monzonite from the Iskut River area, west-central B.C., $\frac{^{206}\text{Pb}}{^{204}\text{Pb}} = 19.015 \pm 0.1$, $\frac{^{207}\text{Pb}}{^{204}\text{Pb}} = 15.662 \pm 0.1$, $\frac{^{207}\text{Pb}}{^{206}\text{Pb}} = 38.714 \pm 0.2$ (M.L. Bevier, pers. comm., 1991) and corrected for spike and fractionation; ⁵corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; ⁶corrected for blank and common Pb, errors are two sigma in Ma; decay constants are those of Steiger and Jäger (1977); for analytical details see Parrish et al. (1987); ⁷corrected for initial Th exclusion, assuming whole rock Th/U=3/1 (see Scharer, 1984).

CHAPTER 4: THE EARLY TO MIDDLE JURASSIC ACCRETION OF TRACY ARM TERRANE TO THE WESTERN MARGIN OF STIKINIA: CONSTRAINTS FROM STRUCTURAL, METAMORPHIC AND COOLING HISTORIES

ABSTRACT

Tracy Arm terrane is a metamorphosed assemblage of polydeformed continental margin type strata (previously referred to as Nisling terrane) in the northern Coast Mountains of British Columbia. It forms a narrow tectonic sliver, bounded by younger, oceanic, arc-dominated terranes on both the west (Alexander terrane) and east (metamorphosed Paleozoic rocks of Stikinia). The Stikinia-Tracy Arm terrane boundary is well exposed in the Tagish Lake area, where it is an imbricate contact that lies within a wider, concordant zone of ductile strain termed the Wann River shear zone. Neither the regional extent of the shear zone nor the location of its root zone are known because it has been complicated by mid-Jurassic and younger folding and faulting and partially obscured by Cretaceous and Tertiary plutons of the Coast Belt.

The Wann River shear zone is >4 km thick and characterized by a mylonitic, penetrative transposition fabric and a gently plunging, south-southeast to north-northwest trending mineral-elongation lineation that formed under upper greenschist to lower-middle amphibolite facies conditions. Structures and metamorphic mineral assemblages that predate the Wann River shear zone, above the shear zone in Tracy Arm terrane and below the shear zone in Stikinia, cannot be followed across the shear

zone from one terrane to the other. The Wann River shear zone has a sense of displacement of Tracy Arm terrane hangingwall southeast relative to footwall Stikinia, and involves a local metamorphic inversion that has resulted in partial recrystallization of the Wann River shear zone mylonitic fabric. The time of shearing is bracketed by 184-193 Ma intrusives of Stikinia that are involved in the Wann River shear zone, and a 177.5 ± 1.5 Ma U-Pb age for metamorphic monazite from Tracy Arm terrane schist, which is interpreted as the age of metamorphic minerals that overprinted the Wann River shear zone transposition fabric. Thus, Tracy Arm terrane was accreted to Stikinia between 184 and 176 Ma, during or just before the accretion of Stikinia to the western margin of ancestral North America; coeval with intense deformation and terrane accretion throughout much of the Canadian Cordillera. U-Pb rutile ages indicate that by ~145 Ma ago, rocks from both Tracy Arm terrane and uppermost Stikinia cooled through ~430°C.

INTRODUCTION

The Canadian Cordillera is in part composed of terranes that were accreted to the western margin of North America during the Mesozoic (e.g. Coney et al., 1980). Where the character and age of a terrane boundary are known, they can be used to constrain tectonic models for the construction of the Cordillera. This chapter documents the age of the accretion of Tracy Arm terrane (also referred to as "Nisling terrane"; e.g. Wheeler et al., 1991; see Chapters 1; 2 and references therein) to the western margin of Stikinia (Stikine terrane) on the basis of the ages of structures and

metamorphic mineral assemblages. The Stikinia-Tracy Arm terrane boundary is of interest because Tracy Arm terrane is considered a displaced continental margin assemblage, but is separated from the nearest source of cratonic rocks, ancestral North America, by terranes that record very little continental influence, such as Stikinia and Cache Creek terrane (e.g. Monger, 1989; Fig. 4.1). It is also of interest because it lies between the Insular superterrane (Wrangellia and Alexander terrane) and the Intermontane superterrane (Stikinia, Cache Creek and Quesnellia; Figs. 4.1; 4.2); its age provides a maximum age limit for the disputed age of the accretion of the Insular superterrane to the western margin of Stikinia. The age of the Insular-Intermontane superterrane boundary is considered by some to be Middle Jurassic or older (e.g. Gehrels and Saleeby, 1985; McClelland et al., 1992; van der Heyden, 1992), and no older than Cretaceous in age by others (e.g. Monger et al., 1982; 1994).

Other workers have suggested that Tracy Arm terrane (Nisling) and Stikinia were together before Triassic time (e.g. Gehrels et al., 1990; Hart and Radloff, 1990; Hansen et al., 1990; Jackson et al., 1991; McClelland, 1992; Rubin et al. 1990), but this suggestion has not been demonstrated conclusively (Chapter 1; Currie and Parrish, 1994). Three lines of reasoning indicate that terrane accretion along Stikinia-Tracy Arm terrane boundary took place during mid-Jurassic time. First, the absence of Early Jurassic plutons in Tracy Arm terrane, in contrast with their abundance in Stikinia, argues strongly against a pre-Early Jurassic juxtaposition of these two terranes (Fig. 4.2). Second, in the Tagish Lake area the Stikinia-Tracy Arm terrane boundary is a shear zone (referred to as the Wann River shear zone) that formed between 184 and

176 Ma ago; it involves rocks as young as 184 Ma and has been overgrown by metamorphic minerals which formed late tectonically, during the thermal peak at ca. 177 Ma (this chapter). Finally, structures that pre-date the Wann River shear zone cannot be correlated across the shear zone between Stikinia and Tracy Arm terrane (this chapter).

This chapter reviews the multiple phases of deformation and metamorphism in both Tracy Arm terrane and Stikinia that preceded formation of the Wann River shear zone, documents the relationship between these older structures and the Wann River shear zone and presents new U-Pb age data that constrain the age of peak metamorphism, which outlasted displacement on the Wann River shear zone. It also presents U-Pb age data reflecting the subsequent cooling history of the area and results of Al-in-hornblende geobarometry for some of the plutons in the Tagish Lake area (Appendices 4.1 to 4.3), and garnet-biotite Fe-Mg exchange geothermometry for two metapelitic rocks from Tracy Arm terrane (Appendices 4.4 to 4.6). In conclusion, the implications of these data for possible mechanisms for the accretion of Tracy Arm terrane to Stikinia are discussed in the light of the regional geology.

GEOLOGIC SETTING OF THE TAGISH LAKE AREA

The Tagish Lake area is on the eastern margin of the Coast Belt, west of the main mass of Stikinia and east of Wrangellia, Alexander terrane, and terranes accreted to the western margin of North America during the Tertiary (Fig. 4.1). Retreating glaciers have provided exceptional exposure, which when coupled with 1400 m of

topographic relief and large areas unaffected by Cretaceous and younger Coast Belt plutonism and deformation, makes the Tagish Lake area an excellent place to study the Stikinia-Tracy Arm terrane boundary (Figs. 4.2; 4.3). Tracy Arm terrane lies structurally above metamorphosed rocks of Stikinia and is only recognised west of the Llewellyn fault zone (Figs. 4.1; 4.2). East of the Llewellyn fault zone metamorphosed Stikinia rocks of probable pre-Late Triassic age, and relatively unmetamorphosed ($\leq$ zeolite facies) Late Triassic and Early Jurassic volcanic and sedimentary rocks of Stikinia crop out (Stuhini and Laberge groups; Fig. 4.2). On both sides of the fault zone, mid-Mesozoic and older rocks are overlain by Tertiary volcanic rocks and intruded by undeformed Cretaceous to Tertiary plutonic rocks of the Coast Belt (Figs. 4.1; 4.2; Appendix A; Mihalynuk and Rouse, 1988a,b; Mihalynuk et al., 1989a,b; 1990; Mihalynuk and Mountjoy, 1990).

Metamorphic rocks of Stikinia and Tracy Arm terrane can be distinguished from one another because they differ lithologically and are juxtaposed against one another within the Wann River shear zone (Fig. 4.3). The Tracy Arm terrane rocks in the Tagish Lake area form the hangingwall of the Wann River shear zone and are referred to as the Florence Range suite (Chapter 2; they have also been called the Florence Range metamorphic suite; Currie, 1990a; Mihalynuk et al., 1990). They comprise multiply deformed, middle amphibolite facies rocks of uncertain age, including schist ($>20\%$ micaceous minerals), semipelitic rocks (10 to 20% micaceous minerals) and marble, with minor quartzite ($>90\%$ quartz), calc-silicate and hornblende amphibolite (Table 4.1; Fig. 4.4; Currie, 1990a; Chapter 2). Lithologic layers are

commonly continuous over many kilometres and vary from <1 cm to 40 m thick. The Florence Range suite is interpreted as a continental margin assemblage on the basis of its characteristic rock-types (marble, metapelitic rocks, quartzite and amphibolite; Chapter 2), Precambrian detrital zircons (Gehrels et al., 1991; Chapter 2), and evolved isotopic character (Samson et al., 1991a). It contrasts with Stikinia, which is an assemblage of mid-Paleozoic to mid-Mesozoic volcanic, plutonic and sedimentary rocks, in part metamorphosed, that have lithologic, geochemical and isotopic characteristics indicative of an arc environment that formed above a young, geochemically and isotopically primitive crust (e.g. Samson et al., 1989; 1991 b; Wheeler et al., 1991; Chapter 3; Fig. 4.1). West of the Llewellyn fault zone, Stikinia is interpreted to be represented by a ~6 km thick section of Paleozoic and possibly younger rocks of the Boundary Ranges suite that in general appears to be right-way-up, and the foliated ca. 185 Ma granitic rocks of the Tagish Lake suite that intrude them (see Chapter 3 for a detailed correlation of metamorphosed Stikinia rocks from the Tagish Lake area with coeval rocks from the Stikinia type-area to the south). Within the Boundary Ranges suite there is an upward change from predominantly felsic metavolcanic and metapelitic rocks to more mafic metavolcanic rocks that are at least in part Mississippian or older. The structurally highest rocks of this suite are Permian metavolcanic rocks, termed the Wann River gneiss (Currie, 1990a). These gneisses are juxtaposed against the Florence Range suite within the Wann River shear zone. Commonly intervening between the Wann River gneiss and other rocks of the Boundary Ranges suite are foliated Early Jurassic plutons of the Tagish Lake suite,

which includes the Mt. Caplice quartz monzodiorite, Hale Mountain granodiorite, Hoboe quartz syenite and coeval dykes (Currie 1992b; Chapter 3; Fig. 4.2).

Metamorphic grade in the footwall of the Wann River shear zone (Stikinia) varies upward from amphibolite facies in the structurally lowest rocks, to greenschist facies at intermediate levels, and amphibolite facies in the structurally highest Boundary Ranges suite rocks (cross-section U-U'; Fig. 4.5).

THE STIKINIA-TRACY ARM TERRANE BOUNDARY AND THE WANN RIVER SHEAR ZONE

The contact between Stikinia and Tracy Arm terrane is tentatively placed at the top of the Wann River gneiss, and along the western margin of the Mt. Caplice quartz monzodiorite (Fig. 4.2; Fig. 4.5a). The metapelitic and marble layers that overlie the Wann River gneiss south of the Wann River are interpreted as belonging to the Florence Range suite, as are similar rocks that are exposed west of the Wann River gneiss along the northern margin of the Llewellyn Glacier, and west of the Mt. Caplice quartz monzodiorite (Figs. 4.2 and 4.3). The Stikinia-Tracy Arm terrane boundary lies within the >4 km thick Wann River shear zone, and is exposed west of the Llewellyn fault zone, on top of White Moose Mountain, in the Wann River valley, and west of Hoboe Glacier (Figs. 4.3; 4.5; 4.7; Plate 4.1). North of White Moose Mountain, the terrane boundary and the shear zone project above the present surface (Fig. 4.5a). Farther south they are covered by the Llewellyn Glacier, but have been extrapolated ~3.5 km south of the edge of the glacier to a nunatak where, on the basis

of airphoto interpretation, they appear to crop out (Fig. 4.2). Their location south of the nunatak is not known, and their westernmost exposure in the Tagish Lake area is bounded by Cretaceous or younger plutons of the Coast Belt. It is not clear whether or not the shear zone or the terrane boundary reappear on the west side of the Coast Belt (Chapter 5). Because younger structures that post-date the Wann River shear zone have influenced its present geometry, the analysis of the Wann River shear zone is preceded by the following description of these younger structures.

The Geometry of the Wann River shear zone and Structures that Post-Date the Wann River Shear Zone

The present geometry of the Wann River shear zone is primarily controlled by structures that are thought to post-date its displacement because they involve both the shear zone and adjacent rocks. These younger structures include map-scale folds that involve the shear zone and faults that truncate the Wann River shear zone (the Llewellyn fault zone, Willison Creek fault zone, Caplice Fault and steep faults shown as normal faults in Figures 4.2, 4.3 and 4.4). North of the upper Wann River, the orientation of the Wann River shear zone varies from nearly horizontal on White Moose Mountain to south-southwest dipping in the Wann River valley due to folding about a broad, open, antiform shown in cross-section U-U' (Fig. 4.5). Farther south the shear zone is folded by the Willison synform (Figs. 4.3; 4.4; Plate 4.2a).

The Willison Synform

One of the largest structures in the Tagish Lake area is the shallow, southwest-plunging Willison synform (Fig. 4.3; Plate 4.2a). It has rotated lithologic contacts and the predominant foliation of the Florence Range, Boundary Ranges and Tagish Lake suites from nearly horizontal or moderately south-dipping in the Wann River valley (Figs. 3.5; 4.4; 4.5; domains 3 and 4 in Fig. 4.8), to steep and approximately north-south striking along the west side of the Llewellyn fault zone (domain 9 in Fig. 4.8; Chapter 1).

The western limb of the Willison synform is offset by steep faults that are pre-Early Cretaceous in age (≥ 127 Ma; Appendix A; Fig. 4.2). The vertical sense of displacement on these faults has been determined on the basis of the inferred offset of the Boundary Ranges suite-Florence Range suite contact, which is generally south-dipping in this area; the faults may also be the loci of undetermined amounts of strike-slip displacement. The westernmost fault is exposed west of the Wann River headwaters, has a west-side-up sense of displacement, and juxtaposes the Boundary Ranges suite (west) against the Florence Range suite (east; Figs. 4.3; 4.4; Plate 4.2b). East of this fault, compositional layering and foliation in the Florence Range suite have been folded about the recumbent Dolphin synform (Plate 4.2c; Figs. 4.3; 4.4). It is not known if the synform involves the Boundary Ranges suite, and consequently it is uncertain whether the synform is older or younger than the Wann River shear zone.

Approximately 3.5 km east of the Dolphin synform, an east-side-up fault truncates the Wann River shear zone, and juxtaposes the Florence Range suite (west)

against the Hale Mountain granodiorite (east) north of the Wann River, and a sliver of the Wann River gneiss (east) against the Florence Range suite (west) south of the Wann River (Figs. 4.2; 4.4). Approximately 4.5 km farther east, a third steep fault juxtaposes the Wann River gneiss against the Hale Mountain granodiorite north of the Wann River, and separates the Florence Range suite (west) from the Wann River gneiss (east) south of the Wann River (Fig. 4.3). Lithologic contacts and foliation are predominantly south-southwest-dipping on the west side of this fault (Figs. 4.3; 4.4; domain 3 in Fig. 4.8), and on the east side they mainly dip toward the south-southeast (Figs. 4.3; 4.4; domain 4 in Fig. 4.8). This fault does not significantly offset the Stikinia-Tracy Arm terrane boundary, suggesting that displacement on the fault decreases toward the north. Nor does it significantly offset the Mt. Switzer volcanic rocks; it may be covered by the Mt. Switzer volcanic rocks and connected to the fault exposed northwest of Mt. Switzer that juxtaposes the Wann River gneiss against the Hale Mountain granodiorite, or displacement across the fault near the Mt. Switzer volcanic rocks may be insignificant. The fault has not been mapped to the south, but the map pattern suggests that it may be related to the Willison Creek fault zone, described below (Figs. 4.3; 4.4).

The change in the orientation of lithologic contacts and foliation from roughly south-dipping on the western limb of the Willison synform to steep and approximately north-south striking on its eastern limb is best exposed west of the south end of Nelson Lake (Fig. 4.4; Plate 4.2a). The variable orientations of the foliation and fold axes in this area (domains 5, 6 and 7 in Fig. 4.8) are attributed to multiple phases of

deformation (see below), and may also reflect structural complications in the core of the synform. The Wann River gneiss-Florence Range suite contact (Stikinia-Tracy Arm terrane boundary) and the Wann River shear zone cannot be followed continuously around the Willison synform because they are truncated by the Llewellyn fault zone, and intruded by a mid-Cretaceous pluton (Fig. 4.3; Appendix A). However, they can be extrapolated from the south end of Nelson Lake, to south of the mid-Cretaceous pluton where compositional layering and foliation are also vertical to steeply west-dipping, strike approximately north-south, and continue to parallel the Wann River shear zone and Florence Range suite-Boundary Ranges suite contact (Fig. 4.4; domain 9 in Fig. 4.8).

South of the mid-Cretaceous pluton, the Wann River shear zone is offset by small sinistral structures (faults or folds) (Fig. 4.3). The largest of these structures is the northwest-southeast striking, mid-Jurassic or younger (<184 Ma) Caplice fault (Figs. 4.3; 4.4), which locally dips toward the south, and has an apparent sinistral displacement of ~1.5 km and an unknown component of dip-slip movement (Fig. 3.6; Chapter 3).

The Willison Creek Fault Zone

The Willison Creek fault zone truncates lithologic contacts and the Wann River shear zone transposition fabric on the western limb of the Willison synform. It is a ~1 km wide fault zone that is bounded by upper and lower décollements that dip ~45° toward the west (Plate 4.2d). Within the fault zone 1 to 200 m³ blocks of marble,

schist, quartzite, and undeformed plutonic rocks are juxtaposed by variably oriented faults (cross-section X-X' in Fig. 4.7; Plates 4.3 to 4.4c). The fault zone is exposed in the Wann River valley, on the ridge between the Wann River and Willison Creek (Plate 4.3), in the Willison Creek valley (Plates 4.4a, b) and on top of Mt. Caplice (Plate 4.3c). Airphoto interpretation and reconnaissance mapping suggest that it continues to the south (Figs. 4.3; 4.4).

Metamorphic rocks in the fault zone resemble those in the hangingwall and footwall, but because neither the magnitude of the displacement across the Willison Creek fault zone, nor displacements across faults within the fault zone are known, the inferred relations between structures and metamorphic fabrics preserved inside the fault zone blocks and those from outside of the fault zone (described below) are equivocal. The oldest igneous rocks recognised in the Willison Creek fault zone are felsic dykes that lack a penetrative fabric but have margins that have been offset by minor structures that are in most cases concordant with the shear fabric in the country rock (Plate 4.4d). The felsic dykes are preserved in fault-bounded blocks of metamorphic rock within the Willison Creek fault zone and within xenoliths in the younger quartz monzonite (Plate 4.5a). Obviously, it is not possible to determine a meaningful sense of shear from the minor structures that offset these dykes because the dykes have probably been rotated during later, and probably unrelated, displacement along the Willison Creek fault zone. However, in the immediate hangingwall of the shear zone lithologically and structurally similar dykes are almost entirely indicative of top-to-the-south sense of shear where their foliation dips

moderately and dextral sense of shear where foliation is sub-vertical. Rotated garnet porphyroclasts in the same rocks exhibit the same sense of shear (Plate 4.5b). The quartz monzonite is intruded by hornblende diorite and associated mafic dykes (Plate 4.5c). All of the intrusive rocks are bound by faults of the Willison Creek fault zone (Plates 4.4a, b; 4.5c).

The faults within and bounding the Willison Creek fault zone truncate compositional layering (Plate 4.5d, e), foliation and folds that involve the foliation within the fault blocks (Plate 4.5e); this has resulted in a scattered orientation of foliation surfaces (domain 10 in Fig. 4.8). Fold axes and mineral-elongation lineations within the fault zone are surprisingly consistent (Fig. 4.3; domain 10 in Fig. 4.8), suggesting that they may have formed during displacement on the Willison Creek fault zone. The latter interpretation is consistent with fabrics along the upper décollement of the fault zone that are indicative of top-to-the-east sense of displacement (Plate 4.2d). North of the Wann River the fault zone does not appear to significantly offset the Florence Range suite-Wann River gneiss contact, suggesting that displacement across the fault zone is not large at its northern end.

The Willison Creek fault zone is not a sedimentological feature, such as a talus breccia because it lacks a matrix, and faults within the fault zone truncate at least two generations of metamorphic fabrics (Plate 4.5e). Its age can be limited to <184 Ma old by the age of the Wann River shear zone, and might be further constrained by dating of igneous rocks cut by the fault zone. Two intrusives exposed near Mt. Switzer that have U-Pb zircon ages of ca. 56 Ma (Appendix A) are lithologically similar to those

exposed in the Willison Creek headwaters area, but their relation to the Willison Creek fault zone is not clear. One is a biotite quartz monzonite, which is exposed near the apparent northern end of the Willison Creek fault zone, and the other is a hornblende diorite that is exposed ~7 km northwest of the quartz monzonite. If the intrusives in the Willison Creek headwaters area are the same age as the dated intrusives exposed near Mt. Switzer, the Willison Creek fault zone would have to be younger than 56 Ma old.

The Llewellyn Fault Zone

The Wann River shear zone and the Willison synform are truncated by the Llewellyn fault zone (Figs. 4.1; 4.2). The eastern complement of the Wann River shear zone has not been recognised, possibly because it is not exposed at the surface, or because Tracy Arm terrane never extended far beyond the eastern limit of the fault zone. The Llewellyn fault zone (Souther, 1971; also referred to as the Llewellyn Fault system by Donelick and Dickie, 1991) consists of steep, anastomosing faults of unknown offset that juxtapose m- to km-scale lozenges of variably deformed shale, metavolcanic rocks, marble, granite, and hornblendite. Pervasive brittle fracturing along the eastern margin of a 103 ± 0.7 Ma granite from immediately west of the Llewellyn fault zone indicates that the fault zone was active after 102 Ma (Appendix A). It has been mapped from the Taku River (Mihalynuk et al., 1994a) to beyond the British Columbia-Yukon border (Mihalynuk and Rouse, 1988b; Mihalynuk et al., 1989b; 1990) where it overprints the Tally Ho shear zone and is plugged by an

Eocene pluton (Hart and Radloff, 1990). On some maps it is shown as a continuation of the mid-Jurassic King Salmon fault (e.g. Souther, 1971; Gabrielse, 1991a), but it is unlikely that this interpretation is correct because the Llewellyn fault zone cuts the mid-Jurassic Wann River shear zone and associated fabrics, which are about the same age as the King Salmon fault (see below; Ricketts et al., 1992).

The timing and the sense of displacement of the Llewellyn fault zone are uncertain. The fault zone may have accommodated vertical motion, strike-slip motion, or a combination of both, either contemporaneously or at different times. Two lines of evidence suggest that the Llewellyn fault zone may have been active after 58 Ma. First, possible deformed equivalents of the Sloko Group volcanic rocks crop out within the Llewellyn fault zone, on the west side of Hoboe Creek (Fig. 4.4). Second, gently tilted volcanic flows interpreted as part of the Sloko Group are preserved on the west side of the valley at 800 m above sea level, but on the east side of the Hoboe Creek valley mountain tops reach elevations of >2000 m above sea level, but lack Sloko Group volcanic rocks. In contrast, Souther (1971) interpreted the variable elevation of the base of the Sloko Group flows to paleotopography that existed before Sloko group deposition.

Donelick and Dickie (1991) have suggested that some of the offset on the Llewellyn fault zone is west-side-up on the basis of ~60 Ma apatite fission track ages from west of the fault zone that contrast with ~100 Ma apatite fission track ages from the east side of the fault zone. However, the Tertiary ages could reflect reheating due to plutonic activity, rather than cooling related to the offset on the Llewellyn fault

zone that postdates regional metamorphism. Metamorphic rocks exposed on either side of the Llewellyn Fault zone may have cooled before displacement on the fault zone occurred, and the juxtaposition of metamorphic rocks (west) against unmetamorphosed rocks (east) could have occurred during subsequent, primarily strike-slip, displacement on the fault zone.

THE WANN RIVER SHEAR ZONE

The Wann River shear zone is characterized by a partially annealed, penetrative, mylonitic shear fabric (S_{WRSZ} ; Table 4.2; Plate 4.1) and a shallowly plunging, north-northwest to south-southeast trending mineral-elongation lineation (L_{WRSZ} ; Table 4.2; Currie, 1991; Fig. 4.3; Plate 4.1b). Rocks in the structurally highest ~2 km of the shear zone preserve amphibolite facies stable mineral assemblages, whereas the remainder of the shear zone comprises rocks that preserve greenschist facies assemblages. All of the rocks in the shear zone contain only one planar fabric (S_{WRSZ}) and one mineral-elongation lineation (L_{WRSZ} ; Table 4.2; Plate 4.1), in contrast to the overlying Florence Range suite and underlying Boundary Ranges suite which preserve evidence for multiple phases of deformation (see below).

The Wann River shear zone clearly involves the lowermost ~1 km of Tracy Arm terrane and the uppermost ~3 km of Stikinia, but has not affected the unconformably overlying Mt. Switzer and Sloko Group volcanic rocks, or plutons younger than 127 Ma old (Appendix A). The upper part of the Wann River shear zone is recognised as the region where tight to isoclinal folds in the Florence Range suite

are only partly transposed into the shear zone; structural fabrics above the shear zone contrast markedly with the straight gneisses of the shear zone (cross-sections U-U', V-V' and W-W'; Fig. 4.5). Near the base of the shear zone, strain has been inhomogeneously distributed within intrusive rocks of the Tagish Lake suite, making it difficult to define an absolute base for the Wann River shear zone (see below).

Structures and Metamorphic Fabrics in the Wann River Shear Zone:

Description

The Wann River shear zone involves the lowermost ~1 km of the Tracy Arm terrane (Florence Range suite), the upper ~3 km of Stikinia (Boundary Ranges and Tagish Lake suites) and a biotite-bearing, strongly lineated granitic gneiss that crops out in the bottom of the Wann River valley, and is probably 275 to 175 Ma old (sample LC-91-437; Table 3.1). The structurally highest rocks in the shear zone are pelitic and semi-pelitic straight gneisses and minor marble layers of the Florence Range suite. The gneissic fabric in the pelitic and semipelitic gneisses (S_{WRSZ}) is defined by compositional layering, and the preferred orientation of biotite and muscovite. Elongate quartz and feldspar grains lie in the plane of S_{WRSZ} , but most are nearly equant, exhibit only minor undulose extinction, do not contain subgrains, and have curved grain boundaries. Quartz and feldspar grains are primarily medium-grained; minor fine-grained quartz and feldspar lie along discrete planes that are concordant with S_{WRSZ} , or as inclusions in larger grains. The lack of strong microscopic solid state deformation fabrics in these rocks is considered the result of

annealing. Marbles also exhibit an annealed texture, with an isotropic texture that is defined by equant, medium-grained calcite grains that have curved grain boundaries, minor embayments and straight extinction. The marble layers also commonly contain a centimetre-scale platy parting that is concordant with S_{WRSZ} . In some samples medium-grained calcite are locally surrounded by mantle aggregates of fine-grained calcite. Mineral-elongation lineations in the Florence Range suite (L_{WRSZ} ; Table 4.2) are defined by quartz- and calcite-rodding (Plate 4.1b), generally have shallow plunges, and trend between north-south and northwest-southeast (Fig. 4.3; domains 3 and 9 in Fig. 4.8).

On the western limb of the Willison synform (on White Moose Mountain and in the Wann River valley) and on the eastern limb of the synform, south of the Caplice Fault, the Florence Range suite structurally overlies a Permian hornblende-plagioclase straight gneiss, referred to as the Wann River gneiss. Within the Wann River shear zone, the Wann River gneiss and Florence Range suite appear to be imbricated in two, and possibly three places. On White Moose Mountain, pelitic rocks bounded above and below by the Wann River gneiss are interpreted as a tectonic sliver of the Florence Range suite. The overlying layer of Wann River gneiss is also interpreted as a tectonic sliver because it is bounded above and below by the Florence Range suite rocks (cross-section U-U'; Fig. 4.5). Similarly, the Wann River gneiss layer exposed south of Wann River, is imbricated with pelitic rocks considered part of the Florence Range suite and is interpreted as a tectonic sliver (Fig. 4.5a). West of Hoboe Glacier, pelite and marble layers alternate with hornblende-plagioclase gneiss

along the western margin of the Wann River gneiss. The pelite and marble layers could be infolds of the Florence Range suite, imbricate slices of the Florence Range suite that have been intercalated with the Wann River gneiss, or metasedimentary rocks of the Boundary Ranges suite that were deposited with the Wann River gneiss (Fig. 4.7c). Note that within the Wann River gneiss rare layer-parallel calcareous layers occur (Plate 4.6a).

The Wann River gneiss has a generally annealed texture; quartz exhibits only minor undulose extinction and embayments, both quartz and plagioclase have curved grain boundaries, and neither mineral contains sub-grains (Fig. 4.23). It has a foliation, S_{WRSZ} , which is defined by the preferred orientation of hornblende, biotite and recrystallized, flattened quartz and feldspar grains (Plate 4.6b). S_{WRSZ} is thought to have formed under amphibolite facies conditions on the basis of the coexistence of stable hornblende and andesine plagioclase. Locally preserved mm- to cm-scale, graded compositional layering in the Wann River gneiss is concordant with S_{WRSZ} and is regarded as the transposed bedding of a metamorphosed volcanoclastic rock (Chapter 1). L_{WRSZ} in the Wann River gneiss is defined by the alignment of hornblende crystals and parallels mineral elongation lineations observed throughout the shear zone (Plate 4.6b; Fig. 4.3).

Both the Florence Range suite rocks and the adjacent amphibolite facies Boundary Ranges and Tagish Lake suite rocks lack abundant well preserved shear sense indicators and display an annealed appearance in thin section (Plate 4.6b). However, locally along the Florence Range suite-Wann River gneiss contacts the

annealed texture has been sheared (Plate 4.1f), low angle truncations of lithologic layers exist (Plate 4.6c), and there are mm-scale chloritized shear surfaces. The low angle truncations are exposed on the eastern limb of the Willison synform, west of Hoboe glacier, and are indicative of sinistral shear (Fig. 4.24), as are rotated blocks in the Florence Range suite (Plate 4.6d).

The Wann River gneiss does not exist north of the Caplice Fault and south of the mid-Cretaceous granite. There, the interpreted contact between Tracy Arm terrane and Stikinia is the inferred contact between the Florence Range suite and the Mt. Caplice quartz monzodiorite. Both the Mt. Caplice quartz monzodiorite and the structurally lower Hoboe quartz syenite are in general uniformly medium-grained, hornblende-bearing plutons that exhibit local compositional banding that parallels S_{WRSZ} in the adjacent rocks, and contain a generally shallow and approximately north-south trending lineation that is defined by the preferred orientation of hornblende crystals and quartz-rodging (L_{WRSZ} ; Fig. 4.3). Matrix quartz and feldspar grains may exhibit undulose extinction and comprise subgrains, and generally have curved grain boundaries, although subgrains within recrystallized quartz ribbons may have serrate grain boundaries. Plagioclase augen commonly contain bent twins and myrmekite, and are surrounded by mantle aggregates of plagioclase, or wrapped by recrystallized ribbon quartz (Plate 4.7a). Epidote and chlorite lack a preferred orientation and are less common in these plutons than in the most of the Hale Mountain granodiorite (see below).

The basal ~10 m of the main body of the Wann River gneiss alternates with cm- to m-scale layers of feldspar augen-bearing hornblende gneiss that are interpreted as ductily deformed layers of the Hale Mountain granodiorite (Tagish Lake suite; Plates 4.1d; 4.7a). The contacts between these layers are concordant with S_{WRSZ} and the Florence Range suite-Wann River gneiss contact, but may involve low angle truncations of compositional layering (Plate 4.1e). The Hale Mountain granodiorite is exposed on Hale Mountain, White Moose Mountain, at the base of Mt. Switzer, and west of Hoboe Glacier (Fig. 4.2). It is an Early Jurassic, medium- to coarse-grained hornblende granodiorite with feldspar porphyroclasts (augen). Near its contact with the Wann River gneiss (ie. near the Wann River and west of the Hoboe Glacier) it primarily contains stable hornblende ($\pm$ biotite), with only minor epidote and chlorite. A sample of Hale Mountain granodiorite from the north side of Mt. Lawson (Sample G in Fig. 4.2) contains both textural and compositional evidence for deformation under amphibolite to upper greenschist facies conditions. The medium-grained plagioclase augen in this sample have sericitized cores of uncertain composition, and clear, unaltered oligoclase rims (An_{28} ; Appendix 4.3) that are wrapped by recrystallized quartz ribbons (Plate 4.7a). The plagioclase rims are considered the result of dynamic recrystallization under conditions of amphibolite facies metamorphism, during displacement on the Wann River shear zone. Fine-grained plagioclase grains in the matrix are albitic in composition (An_2 ; Appendix 4.3) and are thought to have formed during subsequent greenschist facies conditions.

The metamorphic grade of the Hale Mountain granodiorite decreases from amphibolite facies to greenschist facies within 500 m of its contact with the Wann River gneiss. Where it preserves primarily greenschist facies minerals, the Hale Mountain granodiorite generally comprises a fine-grained matrix, which is green due to the presence of epidote and actinolite, and medium- to coarse-grained feldspar porphyroclasts (Plates 3.2a and 4.7b; ; Table 3.1; Chapter 3). Throughout the pluton, plagioclase porphyroclasts are commonly mantled by aggregates of quartz and feldspar (oligoclase to albite), wrapped by recrystallized quartz ribbons that are concordant with S_{WRSZ} , or both. The quartz ribbons comprise quartz sub-grains with serrate grain boundaries and trails of fluid inclusions interpreted as having formed along annealed microfractures. Quartz and feldspar grains exhibit undulose extinction and have curved grain boundaries. Feldspar augen may have bent twins, narrow intragranular recrystallized bands, and/or subgrains, and may be fractured. Rotated feldspar augen, extensional shears, C/S fabrics and rare folds observed on the eastern limb of the Willison synform (in the Wann River valley and on top of White Moose Mountain) are indicative of top-to-the-south (to southeast) relative sense of shear (Fig. 4.3; Plates 4.1; 4.6d), whereas those on the eastern limb of the Willison synform are indicative of sinistral displacement (Plates 4.7c, d).

Within ~2 km of the base of the Florence Range suite the Hale Mountain granodiorite is penetratively deformed, but >2 km away deformation fabrics in the Hale Mountain granodiorite are inhomogeneous. Discrete m- to 10 m-scale mylonitic layers are separated by massive layers of the granodiorite (compare Plates 4.1 and

3.1d). At its base, 10 cm- to 1 m-thick layers of the Hale Mountain granodiorite alternate with layers of actinolite schist of the Boundary Ranges suite of similar thickness (Fig. 4.5a; Plate 4.8a). The base of the Hale Mountain granodiorite is considered the approximate base of the Wann River shear zone in this area because structures in the underlying Boundary Ranges suite have not been completely transposed.

Structures and Metamorphic Fabrics in Stikinia From the Wann River Shear Zone Footwall: Description

The Boundary Ranges suite rocks that lie structurally below the Wann River shear zone are exposed at low elevations near the south end of Tagish Lake and west of Hoboe Glacier, and are dominated by actinolite schist. The actinolite schist rarely contains chlorite, but may contain plagioclase, pyroxene and/or hornblende megacrysts. The megacrysts commonly have quartz-filled tails, appear fragmented (Plate 4.8b), and are wrapped by the foliation, which is defined by the preferred orientation of actinolite and flattened feldspar grains (Plate 4.8b). Actinolite schist layers commonly alternate with thin, mm-scale feldspar-rich layers (Chapter 1) that parallel the predominant foliation. They are locally associated with felsic gneiss, marble and metapelite. Felsic gneiss layers (metamorphosed flows, sills or transposed dykes) associated with the actinolite schist also preserve a well-developed foliation, which is defined by the preferred orientation of micaceous minerals, and flattened quartz and feldspar grains.

The discontinuous nature of lithologic layers within the Boundary Ranges suite has in most cases prevented rigorous structural analysis, and most structural interpretations are based on outcrop-scale observations. In rare cases the foliation is axial planar to minor isoclinal folds (outcrop scale or smaller), but in general foliation is concordant with compositional layering and has been folded, along with compositional layering by crenulations, and small- to medium-scale close to isoclinal folds. North of the Wann River, near the base of the Wann River shear zone, compositional layering (including the contact between the Hale Mountain granodiorite and the underlying Boundary Ranges suite), foliation and the axial planes of folds in the Boundary Ranges suite have moderate to horizontal dips and are concordant with S_{WRSZ} in the overlying Wann River shear zone; these structures are considered older than the Wann River shear zone. Fold hingelines generally parallel mineral-elongation lineations defined by the preferred orientation of actinolite in actinolite schist, calcite rodding in marble, and L_{WRSZ} in adjacent rocks. West of Hoboe Glacier compositional layering and foliation in the actinolite schist-dominated assemblage is folded by rare, outcrop scale folds with shallow fold axes that trend approximately north-south. There, the Boundary Ranges suite is bounded to the east by the Llewellyn fault zone and comprises the structurally lowest rocks exposed in that area.

In contrast to the shallow to moderately dipping structures preserved immediately below the Hale Mountain granodiorite, compositional layering, foliation and axial planes of folds in the structurally deeper Boundary Ranges suite rocks exposed northwest of Nelson Lake, along the western shore of Tagish Lake and north

of Fantail Lake are steep to vertical (Fig. 4.2; Mihalynuk et al., 1989b). Locally, the foliation outlines refolded mesoscopic folds (outcrop-scale; Plates 4.8c, d), and north of Fantail Lake a map-scale interference fold involves a marble layer and the foliation, indicating at least three phases of deformation occurred locally (see Figure 1-32-6 in Mihalynuk et al., 1989a).

Rocks of the Boundary Ranges suite that lie structurally below the actinolite schist-dominated sequence are exposed in the Bighorn Creek valley and on Mt. Lawson. They are predominantly felsic metavolcanic rocks and metapelitic rocks, with minor marble layers. All rock-types have been folded by open, upright crenulations and mesoscopic folds (see cross-section V-V' in Fig. 3.3). The pelitic rocks contain biotite $\pm$ quartz, plagioclase, muscovite, andalusite, kyanite, garnet and/or staurolite, and are commonly graphitic (Table 1.1). In thin section, biotite and muscovite define the foliation and exhibit undulose extinction where they are crenulated. Kyanite and andalusite are also folded (Plates 4.8e and 4.9a). Garnet porphyroblasts are wrapped by the foliation and have been rotated relative to the foliation. Their rims are commonly replaced by quartz + biotite $\pm$ plagioclase $\pm$ magnetite $\pm$ ilmenite. Undulose extinction and subgrain development are ubiquitous in quartz, and common in feldspar. The relations between staurolite and textural elements are unclear.

The pelitic rocks are interlayered with Mississippian felsic orthogneisses that contain a well defined foliation defined by micaceous minerals and flattened quartz and feldspar grains. They commonly contain mica fish, rotated feldspar augen wrapped by ribbon quartz, and an elongation mineral lineation defined by quartz and/or

plagioclase (Chapter 3). One exception is the Mississippian Bighorn Creek granite, which lacks a strong penetrative foliation, but contains ample evidence of subsolidus deformation, including mortar texture and exsolution textures in feldspar (Chapter 3).

Interpretation of Structures and Metamorphic Fabrics in Stikinia

From Below the Wann River Shear Zone

Below the Wann River shear zone there is evidence for multiple phases of deformation within the Boundary Ranges suite rocks that underlie the Tagish Lake suite plutons. The folds that are preserved below the Wann River shear zone are older than the shear zone because they do not affect the shear zone fabric and appear to have been transposed by it. As in Tracy Arm terrane, the style of deformation in Stikinia varies with proximity to the Florence Range suite-Wann River gneiss (Tracy Arm-Stikinia) contact; on the northern limb of the Willison synform the orientation of structures in the Boundary Ranges suite vary from shallowly dipping near the Wann River shear zone, to steep at structurally deeper levels (cross-section U-U'; Fig. 4.5). Whether this change in orientation reflects rotation of structures into the shear zone cannot be demonstrated unequivocally: the folds with steeply dipping axial planes either occur close to the Llewellyn fault zone, or far from exposures of the Wann River shear zone. For example, the upward change from minor folds with steep axial planes at the deepest structural levels in the Tagish Lake area (on Mt. Lawson and north of Fantail Lake) to minor folds with shallowly dipping axial planes that are concordant with S_{WRSZ} near the base of the Wann River shear zone (on Hale

Mountain) cannot be shown to be the result of rotation during D_{WRSZ} because the orientation of the Wann River shear zone is not known north of Mt. Lawson.

Shallowly plunging south-southeast to north-northwest trending mineral-elongation lineations observed throughout the Boundary Ranges and Tagish Lake suite rocks, both above and below the Wann River shear zone are thought to have been rotated into the Wann River shear zone, or interpreted as having formed during displacement on the shear zone.

The First Phase of Deformation (D_{1S}) and Metamorphism (M_{1S}) in Stikinia

Contacts between and within metavolcanic and metasedimentary layers are interpreted as transposed bedding and are denoted S_{0S} (Table 4.2; the subscript "S" is used to distinguish fabrics found in Stikinia from those of Tracy Arm terrane). The foliation in the Boundary Ranges suite rocks below the Wann River shear zone is defined by the alignment of micaceous or acicular metamorphic minerals (e.g. biotite, muscovite, actinolite); it is referred to as S_{1S} because it is the oldest preserved structural fabric, and S_{0+1S} because it is commonly concordant with S_0 . The minerals that define S_{0+1S} are thought to have formed during the oldest recognised metamorphic event in Stikinia because locally S_{0+1} is axial planar to isoclinal folds (M_{1S} ; Table 4.2). Kyanite and andalusite are also considered M_{1S} minerals because they lie in the plane of S_{0+1S} and have undulose extinction where they are folded (Plates 4.8e; 4.9a). Because folded kyanite and andalusite have been observed in a single thin section

(Table 3.1), rocks at this structural level must have reached amphibolite facies conditions before or during the first preserved phase of deformation. At the lowest exposed structural levels (ie. on Mt. Lawson) garnet is considered an M_{1S} mineral because it is wrapped by S_{0+1S} . It is assumed garnet and biotite grew in equilibrium, although this relationship has not been established due to the almost ubiquitous alteration of the garnet rims.

The Second and Third Phases of Deformation in Stikinia (D_{2+3S})

Complex folding of S_{0+1S} is common below the Wann River shear zone and is considered indicative of polyphase deformation that preceded the formation of the Wann River shear zone. Locally, at least two generations of structures have folded the S_{1S} , including mesoscopic structures (Plates 4.8c, d) and a map-scale interference fold (see Figure 1-32-6 in Mihalynuk et al., 1989a). The two post- D_{1S} phases of deformation are referred to as D_{2+3S} . The timing of these structures and the rotation of porphyroclasts (e.g. garnet, pyroxene, hornblende, and plagioclase) and fracturing of hornblende and plagioclase porphyroclasts in actinolite schist relative to the formation of other structural elements are uncertain.

Penetrative deformation fabrics in Mississippian orthogneisses could have formed as early as D_{1S} , or as late as D_{3S} . The orthogneisses have not been observed cross-cutting D_{2+3S} structures, and in the Swanson River headwaters the foliation in an undated felsic orthogneiss considered part of Stikinia is folded about D_{2S} and possibly D_{3S} folds. The presence of only a weak penetrative foliation in one of these

metaplutonic rocks, the Bighorn Creek granite, suggests that this pluton acted as a rigid block during D_{WRSZ} (Chapter 3).

The temperatures and pressures that existed during D_{2+3S} and the ages of D_{2+3S} are not known, but the ages of these events are considered older than 186 Ma, because D_{2+3S} structures do not appear to affect even the oldest dated Tagish Lake suite pluton, the Hoboe quartz syenite (189.5 ± 3 Ma; Chapter 3).

The Second Metamorphic Episode in Stikinia (M_{2S}) and the Inverted Metamorphic Gradient in the Uppermost Stikinia Rocks

M_{2S} minerals are only found in the structurally highest ~1 km of the Stikinia rocks, which includes the Wann River gneiss and the structurally highest Tagish Lake suite plutons. The Wann River gneiss has a relatively annealed appearance in thin section (Plate 4.6b), despite the fact that it lies within the Wann River shear zone, has been tectonically juxtaposed against and imbricated with Tracy Arm terrane rocks, and retains a foliation and mineral-elongation lineation considered to have formed during D_{WRSZ} . The lack of dynamic recrystallization textures in these rocks suggests that M_{2S} outlasted displacement on the Wann River shear zone. Amphibolite facies conditions are inferred for M_{2S} on the basis of the hornblende-andesine (actinolite-absent) stable mineral assemblage in the Wann River gneiss and uppermost Tagish Lake suite rocks. The decrease in metamorphic grade with increasing structural depth, from M_{2S} amphibolite facies mineral assemblages to greenschist facies in the underlying Tagish Lake and Boundary Ranges suite rocks is interpreted as a locally inverted

metamorphic sequence, which is discussed in more detail below (cross-section U-U'; Fig. 4.5). It is not considered the result of post-metamorphic thrusting because U-Pb age data indicate that the Boundary Ranges suite is in general right-way-up (Chapter 3).

The structures described above cannot be followed across the Wann River shear zone into the Florence Range suite, but M_{2s} is considered the same age as the youngest stable metamorphic mineral assemblages in the Florence Range suite (see below).

Structures in Tracy Arm Terrane From Above the Wann River Shear Zone: Description

Within the Florence Range suite, above the Wann River shear zone, there is evidence for at least two phases of deformation and two or more metamorphic episodes that preceded the formation of the Wann River shear zone. Despite disruption by younger structures, it has been possible to elucidate this complex structural history by following contacts between contrasting rock-types and unravelling metamorphic fabrics. Isogradic surfaces have not been mapped in the Florence Range suite because diagnostic mineral assemblages are scarce, and metamorphic grade is largely uniform.

Folds with an axial planar foliation are rare in the Florence Range suite; they are generally close to isoclinal and mesoscopic (Plate 4.9b). Their axial planar foliation is the predominant foliation in schist and semipelitic rocks of the Florence Range suite, and except in the cores of these folds, is concordant with lithologic

contacts. It is primarily defined by the preferred orientation of biotite and muscovite and commonly parallels mm- to cm-scale compositional layering defined by variable concentrations of quartz + feldspar. At least two generations of micaceous minerals may be observed in a single thin section; the first generation of muscovite and biotite grains can be distinguished from the second because they tend to be $<100\text{ }\mu\text{m}$ in diameter and have wavy (undulose) extinction where they have been folded (Plate 4.9c), whereas the younger generation(s?) are polygonalized and $>100\text{ }\mu\text{m}$ to 5 mm in diameter (Plate 4.9d). The second generation of micaceous minerals commonly pseudomorphs the first generation, but some are randomly oriented and overprint older fabrics (Plate 4.10a). Other minerals that lie in the plane of the foliation include andalusite, kyanite, staurolite and some fibrolite. Where kyanite is present, it is preserved as anhedral fragments surrounded by randomly oriented, second generation muscovite (Plate 4.10a). The relationships between kyanite and fabric elements are ambiguous because the kyanite fragments are isolated from such foliation, crenulations and other structural elements by the surrounding muscovite, but in most cases the long axes of the kyanite grains appear to parallel foliation. In most kyanite-bearing samples, fibrolite is also present (Plate 4.10b; Table 4.1). Prismatic sillimanite and fibrolite are commonly associated with biotite, with the long axes of the fibrolite either parallel to the long axes of the biotite grains, or randomly oriented (Plate 4.10b).

Locally, fibrolite inclusions are preserved in quartz grains that exhibit undulose extinction and comprise subgrains. Undulose extinction in quartz and plagioclase is ubiquitous, quartz subgrains are extremely common and plagioclase subgrains may

also be present. Both quartz and plagioclase grains are flattened or elongate, with curved or serrate grain boundaries; they commonly define a mineral-elongation lineation in the lowermost Florence Range suite rocks. There is widespread sericitization of plagioclase and porphyroclasts of unknown original composition.

Textural relations between garnet and metamorphic fabrics are difficult to interpret because alteration of garnet rims is almost ubiquitous and has resulted in 10% to 100% replacement of individual garnet crystals by biotite $\pm$ ilmenite $\pm$ quartz $\pm$ plagioclase. Phantom euhedral garnet rims can still be recognised because fine-grained opaque minerals that were adsorbed onto the garnet crystal facies remain and outline the pseudo-hexagonal outline of the original garnet (Plate 4.10c). In some samples garnet is wrapped by the foliation (Plate 4.10d), and in others it has overprinted the foliation and crenulations that involve the foliation (Plate 4.10c). The garnet crystals commonly contain inclusions of quartz and opaque minerals, and have straight inclusion trails that are at an angle to foliation (Plate 4.10c). Pressure shadows and tails outside the second generation garnet crystals contain quartz $\pm$ biotite $\pm$ fibrolite (Plate 4.10d).

Staurolite has been observed in only one rock, a schist from a fault-bounded block within the Willison Creek fault zone. Although the relations between staurolite and metamorphic fabrics outside the fault zone are uncertain, textural relations between other metamorphic minerals in this same sample are the same as those in samples from outside the fault zone (kyanite fragments are surrounded by randomly oriented muscovite, and fibrolite is randomly oriented and associated with biotite),

suggesting that it experienced a similar, if not the same tectonic history as rocks outside the fault zone. Stauroilite fragments occur as clusters of inclusions in second generation muscovite. The inclusions have maintained a common crystallographic orientation, with their long axes generally paralleling the foliation (Table 4.1).

There are at least two generations of leucosome. The oldest generation exists throughout the Florence Range suite as foliation-parallel, mm- to cm-scale pods, or continuous layers of quartz and feldspar, separated by pelitic layers; they are commonly folded. The second generation of leucosome exists above and within the Wann River shear zone, and is locally developed at the contact between the Florence Range suite and the Wann River gneiss in the Wann River valley. It comprises coarse grained trondhjemitic layers that are characterized by randomly oriented biotite books up to 5 cm in diameter (Plates 4.11a, b). The layers are 1 cm to 1 m thick and generally parallel the foliation in adjacent pelitic or semipelitic layers, but locally cross-cut the foliation and appear to have displaced pelitic layers (Plate 4.11b). In thin section the leucosome generally displays an igneous texture and the only evidence for subsolidus deformation are undulose extinction in quartz and feldspar and subgrain development in quartz.

In the amphibolite layers, the foliation generally parallels lithologic contacts, and is defined by the preferred orientation of hornblende, hornblende fragments (formed by fracturing along cleavage planes), biotite, and locally by mm- to cm-scale compositional banding defined by alternating hornblende-rich and plagioclase-rich layers. However, some amphibolite layers may be massive with randomly oriented,

undeformed hornblende crystals. The amphibolite layers primarily comprise hornblende (30-90%) and plagioclase (5-50%; An_{5-15}), and commonly contain garnet, biotite, titanite, chlorite, quartz, rutile, epidote, calcite, tremolite, hematite, apatite and/or ilmenite. In some samples there are two generations of plagioclase. The first generation has altered, inclusion-free plagioclase fragments with undulose extinction, bent twins and deformation twins; whereas the second generation is characterized by fresh, randomly oriented plagioclase grains that have inclusions of hornblende and opaque minerals, and do not appear to be deformed. Garnet crystals are anhedral and fractured, may contain straight inclusion trails of opaque minerals, and have been replaced (up to 100%) along fractures and rims by biotite $\pm$ plagioclase $\pm$ chlorite $\pm$ opaque minerals. Locally, the garnet crystals have been rotated, as evidenced by quartz-filled pressure shadows and tails. Quartz exhibits undulose extinction and subgrain development, and chlorite and biotite are generally characterized by straight extinction. Ilmenite is commonly rimmed by titanite, and both minerals may be found as inclusions in hornblende, biotite or chlorite.

Marble layers are generally massive, although mm-scale graphitic layers commonly parallel the margins of the marble layers. They primarily comprise coarse-grained, randomly oriented calcite, and in thin section generally have an annealed texture, with the exception of curved grain boundaries. They are commonly associated with calc-silicate rocks, especially near their contacts with amphibolite layers. Calc-silicate rocks also form cm-scale layers within thicker (m-scale) marble layers that are not associated with amphibolite. Like the marble layers, minerals in the calc-silicate

layers are generally randomly oriented, grain boundaries are straight or curved, and in some cases exhibit mortar texture. Compositional banding parallels the foliation in adjacent rocks, and is defined by alternating mm- to cm-scale layers with varying amounts of calcite $\pm$ dolomite $\pm$ diopside $\pm$ grossular garnet $\pm$ vesuvianite $\pm$ titanite. Titanite is common as inclusions in calcite and dolomite.

Structurally above the Wann River shear zone, folds that refold those with an axial planar foliation, and involve the foliation in the pelitic and amphibolitic rocks, vary in size from microscopic crenulations to mountain-side scale folds that control the map-pattern (Plates 4.11c to 4.13). Where the first generation micaceous minerals and andalusite crystals are folded by these crenulations, they exhibit undulose extinction (Plate 4.9c).

In the Wann River headwaters area, spectacular, northeast-verging, reclined folds with fold axes that plunge shallowly toward the south-southeast (domain 1; Fig. 4.8) are outlined by marble, schist, amphibolite and the predominant foliation (cross-section V-V'; Fig. 4.5; Plates 4.11d; 4.12a). They lie in the hangingwall of a fault of uncertain age that also carries the Dolphin synform in its hangingwall (Plate 4.12b). The underlying zone of disrupted carbonate lenses, which has not been mapped in detail, is interpreted as the continuation of this fault (Plate 4.12a). It is not clear whether these folds lie on the upper limb or lower limb of a larger structure. They are separated from folds of similar style exposed to the east (domain 2 in Fig. 4.8), by a steep, west-side-down fault. The folds in domain 2 are carried in the hangingwall of a south-dipping fault, referred to as the Wann fault, which is recognised on the basis of

the truncation of the lower limbs of a northeast-verging fold ("WF" in Figs. 4.3; 4.4; 4.5c; 4.6; Plates 4.12c, d). Above the Wann fault, a recumbent, isoclinal synform and its parasitic folds are outlined by schist, marble, amphibolite (labelled " S_{2T}^* " in cross-section W-W'; Fig. 4.5; Plates 4.12c, d). The vergence of this synform is not known because its upper limb has been removed by erosion and its lower limb is truncated by the Wann fault (Fig. 4.5c). The axial plane of the synform is "warped" by D_{3T} structures (labelled " S_{3T} " in cross-section W-W'; Fig. 4.5c) that are also thought to fold S_{WRSZ} ; folding about the D_{3T} synform is reflected by the spread of poles to foliation about a great circle (Fig. 4.6).

The marble layer that defines the upper limb of the synform is folded by northeast-verging folds that resemble folds exposed to the west in domain 1 (cross-section V-V' in Fig. 4.5; compare folds in Plates 4.11d and 4.12a with those in Plate 4.13a). The upper limb marble layer can be followed almost continuously to the southeast to where it intersects cross-section X-X' (Figs. 4.4; 4.6). Projection of the underlying D_{2T} synform shown in cross-section V-V' into the plane of cross-section X-X' (Fig. 4.7) indicates that the synform is probably truncated by the Willison Creek fault zone. The generally flat-lying limbs of the D_{2T} synform are separated from the steep marble layers to the south (exposed east of cross-section X-X' and west of the Willison Creek fault zone) by a fault that is also truncated by the upper décollement of the Willison Creek fault zone (Fig. 4.4; Plate 4.3).

The structures described above are carried in the hangingwall of the upper décollement of the Willison Creek fault zone (Fig. 4.4). Below the fault zone,

compositional layering and foliation are also folded by mesoscopic folds and a major fold referred to as the Hoboe antiform ("HA" in Figs. 4.3; 4.9; see cross-sections Y-Y' and Z-Z' in Fig. 4.7). The Hoboe antiform is located west of the Willison Creek fault zone, on the eastern limb of the Willison synform. It has rotated marble, schist, quartzite and amphibolite layers and foliation in the Florence Range suite from steep and roughly north-south striking south of Mt. Caplice (domain 9 in Fig. 4.8), to moderately west-dipping farther west (Fig. 4.9; Plates 4.13b, c).

Interpretation of Structures and Metamorphic Fabrics in the Florence Range Suite

Above the Wann River shear zone, the Florence Range suite preserves at least two generations of structures and three metamorphic mineral assemblages that formed before displacement on the Wann River shear zone. The deformational events are termed D_{1T} and D_{2T} , and periods of metamorphic mineral growth are referred to as metamorphic events and are denoted M_{1T} , M_{2T} and M_{3T} (the subscript "T" denotes events or structures in Tracy Arm terrane rocks; Table 4.2). Although different deformational events and metamorphic episodes are treated as separate entities, in some cases they probably form a continuum and reflect progressive deformation or progressive metamorphism, respectively.

The First Phase of Deformation (D_{1T}) and First Metamorphic Episode (M_{1T}) in Tracy Arm Terrane

Compositional layering in the Florence Range suite is interpreted as transposed bedding (referred to as S_{0T} ; Table 4.1), with the exception of contacts between metasedimentary layers and amphibolite layers, which could represent transposed intrusive contacts (Chapter 2). S_{0T} is almost everywhere parallel with the predominant foliation in these rocks (S_{1T}); where they are concordant, S_{0T} and S_{1T} are referred to as S_{0+1T} (Table 4.2). S_{1T} is interpreted as forming during the oldest documented phase of deformation that affected the Florence Range suite (designated D_{1T} ; Table 4.2) on the basis of rare folds with an axial planar foliation that has been refolded. No major D_{1T} fold closures have been identified in the study area.

In many cases S_{0+1T} is defined by biotite and muscovite grains that are interpreted as M_{1T} minerals (Table 4.2). On the basis of their structural relationship between M_{1T} minerals and D_{1T} folds, M_{1T} is considered roughly coeval with D_{1T} . Neither the age of D_{1T} and M_{1T} , nor the vergence of D_{1T} are known. Elsewhere, S_{0+1T} is defined by a younger generation of micaceous minerals that have retained the original orientation of the M_{1T} minerals (Plate 4.9d). Andalusite crystals are considered M_{1T} minerals because they lie in the S_{0+1T} plane and have been folded by crenulations that also fold S_{0+1T} (Plate 4.9c). Garnet crystals that are wrapped by S_{0+1T} and have straight inclusion trails are also regarded as M_{1T} minerals because they have overgrown S_{0T} , but show no evidence to suggest that they overgrew any younger structures or M_{1T} minerals (Plate 4.10c).

M_{1T} is considered an upper greenschist to lower amphibolite facies metamorphic event because the coexistence of M_{1T} garnet and biotite has not been demonstrated texturally (most garnet rims are altered, and no biotite inclusions have been found in garnets with straight inclusion trails). It is not clear whether staurolite formed during M_{1T} or M_{2T} , but because it is replaced by randomly oriented muscovite, it most likely formed before M_{3T} (see below). If M_{1T} reached amphibolite facies conditions, hornblende crystals in amphibolitic rocks may have grown during M_{1T} .

M_{1T} andalusite is indicative of medium-temperature - low-pressure conditions, which may occur (1) in contact metamorphic aureoles, (2) if medium to high temperature rocks experience a decrease in pressure, or (3) in an area of high heat flow, such as a continental rift setting (Wickham and Oxburgh, 1985). Andalusite growth in the Florence Range suite is not attributed to contact metamorphism because the oldest known plutons that have intruded these rocks are younger than the peak metamorphic mineral assemblage (M_{3T}), and therefore must be younger than the M_{1T} andalusite crystals (Table 4.2). Rapid denudation leading to andalusite growth is also considered unlikely because there is no evidence for medium- to high-temperature and pressure conditions preceding andalusite growth. Speculatively, if the andalusite in the Florence Range suite could have formed in a continental rift setting, perhaps as Tracy Arm terrane was rifted away from the continental margin that was the source of its detritus, and D_{1T} could have been an extensional event that occurred long before the accretion of Tracy Arm terrane to Stikinia.

The temporal relationship between kyanite and andalusite, and between kyanite and phases of deformation is uncertain because kyanite is isolated from the fabric elements by the polygonalized muscovite that surrounds them (Plate 4.10a).

The Second Phase of Deformation in Tracy Arm Terrane (D_{2T})

Mountain side-scale structures (Plates 4.11c, d; 4.12; 4.13) and mesoscopic parasitic folds and microscopic crenulations (Plates 4.9c, d) that fold S_{0+1T} are denoted D_{2T} structures (Table 4.2). They are preserved in the hangingwall of the Wann River shear zone, and are exposed in both the hangingwall and footwall of the Willison Creek fault zone, but are less common in the footwall. The D_{2T} folds in the hangingwall of the Willison Creek fault zone are primarily northeast-verging, but the overall vergence of D_{2T} structures cannot be indisputably determined because the observed folds could have formed on the lower limb of a larger southwest-verging structure. Alternatively, the D_{2T} structures could be northeast-verging, in which case, if they formed during initial accretion of Tracy Arm terrane, their opposing senses of vergence, relative to that of the Wann River shear zone (southwest-verging) could indicate that the D_{2T} structures reflect a change in vergence during accretion. Alternatively the D_{2T} structures could be older than, and unrelated to the accretionary event.

The Hoboe antiform is preserved in footwall of the Willison Creek fault zone (Fig. 4.9). It is not possible to tell whether Stikinia rocks are folded by the Hoboe antiform on the eastern limb of the Willison synform, but north of the Wann River, on

the northern limb of the Willison synform, the lack of a Hoboe antiform-equivalent structure that folds S_{WRSZ} and lithologic contacts in Stikinia suggests that the Hoboe antiform is older than the Wann River shear zone; it is therefore interpreted as a D_{2T} structure.

Other evidence for post- D_{1T} deformation comes from quartz tails around garnet porphyroblasts and inclusion trails in garnet that are at an angle to S_{0+1T} , reflecting the rotation of either S_{0+1T} or the garnet. Northeast-verging, sheared leucosome pods are interpreted as having been deformed during D_{2T} (Plate 4.14a), and therefore indicate that these rocks were probably at amphibolite facies conditions before or at the same time as D_{2T} . The bending and disaggregation of hornblende crystals from above the Wann River shear zone is also considered younger than D_{1T} , because hornblende grew during or after M_{1T} .

The Second Metamorphic Episode in Tracy Arm Terrane (M_{2T})

Textural relations of garnet provide evidence for the second of at least three metamorphic episodes. The second generation of garnet crystals are interpreted as having formed after M_{1T} and D_{2T} because, unlike the M_{1T} garnets, the M_{2T} garnets have overgrown S_{1T} and crenulations that fold S_{1T} (Plate 4.10c). The M_{2T} garnets are considered younger than M_{3T} because they are bounded by pressure shadows that contain biotite that has been replaced by M_{3T} fibrolite; see below; Table 4.2).

M_{2T} is regarded as an amphibolite facies event that occurred between M_{1T} and M_{3T} on the basis of metamorphic minerals and textural relationships (Table 4.2; see

below). It may have been contemporaneous with D_{2T} . Because "normal" static continental geotherms do not pass from the andalusite P-T stability field to that of kyanite (Kerrick, 1990), the change from andalusite-stable conditions during M_{1T} to kyanite-stable conditions during M_{2T} is interpreted as reflecting an increase in pressure, possibly due to crustal thickening during D_{2T} .

The Third Metamorphic Episode in Tracy Arm Terrane (M_{3T})

Minerals that formed during the third metamorphic episode (M_{3T}) are in general randomly oriented, but may also pseudomorph older minerals that have a preferred orientation. M_{3T} minerals can be distinguished from M_{1T} or M_{2T} minerals if they replace the older minerals. Metamorphic mineral assemblages and textures in the pelitic rocks indicate that during M_{3T} , fibrolite, prismatic sillimanite and leucosome formed, and muscovite, biotite, quartz and plagioclase were recrystallized, presumably at the expense of minerals that no longer appear to be in textural equilibrium, including kyanite, some biotite and possibly staurolite. Fibrolite, prismatic sillimanite, the muscovite grains that surround relic kyanite (Plate 4.10a), and undeformed plagioclase, biotite and quartz grains, are considered M_{3T} minerals that formed via the ionic reactions outlined by Carmichael (1969):

The sum of these equilibria is:

(abbreviations for all mineral phases and components are taken from Kretz, 1983; Table 4.1). On the basis of this interpretation, it is not surprising that plagioclase grains truncate D_{ZT} crenulation fold hinges, or that muscovite, biotite and quartz have been recrystallized (they preserve only minor subsolidus deformation fabrics), because these minerals are products of the reactions outlined above. Kerrick (1990) has questioned the validity of Carmichael (1969)'s assumption that there is a constant Al reference frame, and has suggested that sillimanite- and fibrolite-forming reactions that do not involve kyanite should also be considered. However, other sillimanite-forming reactions that have been derived for rocks from other areas are not supported by textures in the Florence Range suite pelites. The following sillimanite-forming reaction

involves the breakdown of staurolite (McLellan, 1985), and could apply to the growth of sillimanite in the Florence Range suite rocks, but only one of the Florence Range suite samples that was collected contains staurolite, indicating that either all of the staurolite in the other samples has broken down, or staurolite was never present and the growth of sillimanite was at the expense of reactants other than staurolite. This reaction is also considered inappropriate for the Florence Range suite because the

staurolite that has been preserved is surrounded by younger muscovite grains, which suggests that a reaction for the breakdown of staurolite probably involved muscovite as a product, rather than a reactant.

Because garnet is strongly altered (up to 100%) in almost all garnet-bearing Florence Range suite rocks, and fibrolite has replaced biotite in pressure shadows surrounding the relict garnets, the following equilibrium for the breakdown of garnet and the growth of sillimanite is attractive:

(McLellan, 1985). If this reaction occurred, other reactions must have resulted in the breakdown of M_{2T} kyanite and biotite, and the formation of M_{3T} muscovite and leucosome.

The leucosome layers that are characterized by randomly oriented biotite books are interpreted as having formed during M_{3T} because they do not exhibit evidence of penetrative deformation (Plates 4.11a, b). Where the leucosome has cross-cut S_{0+1T} and appears to have displaced pelitic layers (Plate 4.11b), it is regarded as a melt. It is believed to have formed in situ and is not interpreted as a dyke related to the Coast Belt magmatism because it is generally confined to between the pelitic layers, and no rocks of the same age (see below) are exposed near-by. The formation of leucosome and the change from kyanite-stability field conditions to sillimanite-stability field conditions reflects an increase in temperature, a decrease in pressure, and/or an

increase in fluid activity. Presumably some fluid activity and/or solid state diffusion continued after the M_{3T} fibrolite crystallized, because some quartz grains contain fibrolite inclusions.

Stable mineral assemblages in amphibolitic and calc-silicate rocks are consistent with amphibolite facies conditions indicated by the pelitic stable mineral assemblages (hornblende-oligoclase stable in amphibolitic rocks, and diopside and grossular in the calc-silicate rocks), but on the basis of textures it is difficult to determine whether recrystallization and/or neocrystallization of hornblende or calc-silicate minerals has occurred. The widespread spatial association between calc-silicate layers and marble-amphibolite contacts indicates that the calc-silicate minerals probably formed during a regional metamorphic event, but which event is not obvious. Minerals in the calc-silicate layers may have formed in an earlier event and then recrystallized during later metamorphic episodes. Their random orientation, straight grain boundaries and lack of microscopic deformation fabrics suggest that the calc-silicate minerals grew (or were recrystallized) after deformation ceased, and therefore during M_{3T} . Alternatively, less competent minerals such as calcite may have deformed around more competent M_{1T} or M_{2T} calc-silicate minerals, such as diopside, garnet and vesuvianite, and then recrystallized, as indicated by titanite inclusions in undeformed calcite crystals.

Some amphibolite rocks from outside of the Wann River shear zone contain hornblende crystals that are massive, either because they escaped or resisted deformation, or because they have been recrystallized since D_{2T} , possibly during D_{3T} .

The second generation plagioclase grains (ie. those grains that lack deformation fabrics and contain hornblende inclusions) in amphibolitic rocks are considered M_{3T} minerals.

Temperature Reached During M_{3T} in Tracy Arm Terrane

M_{3T} stable mineral assemblages and leucosome provide the best indication of temperatures reached during M_{3T} . Assuming the leucosome formed via the following reaction:

its presence indicates that these rocks probably reached temperatures of at least 650°C, assuming they were at pressures greater than 5 kbar (e.g. Vielzeuf and Holloway, 1988; 1 kbar=100 MPa). Pelitic rock of the Florence Range suite are on the low-temperature side of the "muscovite-out isograd" reaction:

due to the lack of metamorphic potassium feldspar and widespread coexistence of muscovite and quartz. This implies that at ~6 kbar, temperatures were lower than ~725°C (e.g. Spear and Cheney, 1989). Assuming the transition from the kyanite-stable to sillimanite-stable conditions was coeval with leucosome formation, these temperatures indicate that the Florence Range suite probably experienced pressures of

at least 6-7 kbar (e.g. Vielzeuf and Holloway, 1988). Assuming a pressure of ~6 kbar, these temperatures are consistent with the presence of stable diopside and the absence of tremolite (in the presence of calcite + quartz) in the calc-silicate layers (Carmichael, 1991), and garnet, hornblende and andesine in amphibolite (Bégin, 1992).

Geothermometry of the Florence Range suite has been difficult because almost all garnet rims have been replaced by biotite $\pm$ quartz $\pm$ plagioclase, and most biotite has been partially altered to chlorite, as indicated by its anomalously low K concentration (Appendix 4.5). The only sample that proved suitable for garnet-biotite Fe-Mg exchange geothermometry returned temperatures of ~650°C, using the calibration of Ferry and Spear (1978; Appendix 4.4).

DEFORMATION AND METAMORPHISM IN TRACY ARM TERRANE AND STIKINIA, AND THEIR RELATION TO THE WANN RIVER SHEAR ZONE

The structures and metamorphic fabrics observed in the metamorphic rocks of the Tagish Lake area are best explained by a post-184 Ma thrusting event that placed relatively hot (amphibolite facies) Tracy Arm terrane rocks over sub-greenschist to greenschist facies Stikinia rocks, and transposed all older structures into parallelism along the Tracy Arm-Stikinia terrane boundary. Outside the Wann River shear zone, both the Florence Range and the Boundary Ranges suites preserve evidence for multiple phases of deformation (D_{1T} to D_{2T} in Tracy Arm terrane and D_{1S} to D_{3S} in Stikinia), but within the shear zone only one planar fabric (S_{WRSZ}) and one lineation (L_{WRSZ}) are preserved. S_{WRSZ} and L_{WRSZ} are interpreted as a transposition foliation and a

stretching lineation, respectively, that formed during top-to-the-south-southeast displacement on the Wann River shear zone. Within the shear zone, the older structures are thought to have been transposed during D_{WRSZ} on the basis of: (1) absence of evidence for multiple phases of deformation within the shear zone, in contrast to the abundance of polydeformed rocks outside the shear zone (2) the equivalent orientations of S_{WRSZ} in both Tracy Arm terrane and Stikinia, and the Stikinia-Tracy Arm terrane boundary, and (3) the coincidence of L_{WRSZ} in both terranes. Structures adjacent to the shear zone have axial planes that are concordant with S_{WRSZ} , suggesting that they were rotated into parallelism with S_{WRSZ} during D_{WRSZ} . The imbricate contacts between the Boundary Ranges suite and Tagish Lake suite rocks parallel S_{WRSZ} and are interpreted as intrusive contacts that were transposed during displacement on the Wann River shear zone (Chapter 3; Plates 4.1d, e). The coincidence of D_{2T} fold axes and L_{WRSZ} (all are shallow and approximately north-south to north-northwest--south-southeast trending; compare lineations plotted in Figs. 4.3 and 4.8) may be original, or may reflect the rotation of preexisting structures during D_{WRSZ} .

The Willison synform is considered a post-accretionary structure because it folds lithologic contacts in both its footwall and hanging wall. It is not considered a flexural slip fold because although it plunges shallowly toward the south-southwest, pre-existing Wann River shear zone lineations do not plot on the small circle of a stereonet with an axis that also plunges shallowly toward the southwest. Instead, the lineations associated with the Wann River shear zone (L_{WRSZ}) change from generally

south-southeast-plunging on the northern limb of the synform, to north- or south-plunging on the eastern limb of the synform (Figs. 4.3; 4.4; 4.8). The Willison synform is more likely a flexural flow fold that formed due to shearing along the eastern limb of the synform. This interpretation allows for the inferred rotation of the lineations and also explains the reduction in the thickness of lithologic units between the northern limb and the eastern limb.

The sense of shear in the Wann River shear zone is consistent around the Willison synform (Fig. 4.3); most shear sense indicators are indicative of top-to-the-south sense of shear on the western limb of the synform and sinistral motion on its eastern limb (Fig. 4.3). Most shear sense indicators and subsolidus deformation fabrics are preserved in the structurally lower, greenschist facies Hale Mountain granodiorite and Boundary Ranges suite rocks. Deformation fabrics in these rocks are interpreted as forming under greenschist facies conditions because they involve greenschist facies minerals, including actinolite, epidote, albite, and chlorite. They include extensional shears (Plate 4.1a), rotated blocks, C-S fabrics, quartz ribbons (Plates 4.7a, b), rotated, dynamically recrystallized feldspar porphyroclasts (Plates 4.7a, b), mortar texture around feldspar porphyroclasts, bent twins and deformation twins in plagioclase, C-S fabrics, myrmekite and rare isoclinal folds, all of which may form under temperatures consistent with the upper greenschist facies conditions (Hanmer, 1982; Fitz-Gerald and Stünitz, 1993), although the texture of myrmekite suggests that it probably formed after deformation ceased. The axial planes of rare isoclinal folds in the Tagish Lake suite plutons are concordant with S_{WRSZ} , and are not considered the result of more than

one phase of deformation; they most likely nucleated on inhomogeneities within these plutons during D_{WRSZ} .

Despite the penetrative nature of S_{WRSZ} , amphibolite facies Florence Range and Boundary Ranges suite rocks within the Wann River shear zone have a relatively annealed appearance in thin section. With the exception of minor undulose extinction and embayments in quartz, the near absence of subsolidus deformation fabrics and well preserved shear sense indicators in the amphibolite facies rocks is interpreted as indicating that static amphibolite facies metamorphism outlasted displacement on the Wann River shear zone.

M_{2T} and M_{3T} may represent a prograde metamorphic sequence, rather than two separate metamorphic episodes. The breakdown of M_{2T} kyanite, and the formation of M_{3T} sillimanite, fibrolite, leucosome and randomly oriented micaceous minerals after D_{WRSZ} , is considered to be the result of an increase in temperature, a decrease in pressure, and/or an increase in fluid activity that post-dated displacement on the Wann River shear zone. An increase in temperature may be due to the relaxation of isotherms associated with structural thickening caused by the accretion of Tracy Arm terrane. An increase in fluid activity may be related to the thrusting event, and the decrease in pressure could be the result of subsequent unroofing caused by erosion and/or tectonic denudation. The M_{3T} leucosome is interpreted as having formed soon after displacement on the Wann River shear zone ceased, and during peak metamorphic temperatures. Accretion and leucosome formation were probably also associated with an increase in fluid activity.

The Inverted Metamorphic Gradient

Clearly, the metamorphic minerals that overprint S_{WRSZ} formed after Tracy Arm terrane and Stikinia were together, and the metamorphic events that formed these minerals, M_{3T} and M_{2S} , are the same metamorphic event (M_{2S+3T}). It follows that the inverted M_{2S} metamorphic sequence in the uppermost rocks of Stikinia formed during thermal equilibration after amphibolite facies Tracy Arm terrane rocks were thrust over sub-greenschist to greenschist facies Stikinia rocks that were part of a normal metamorphic sequence of unknown age (M_{1S}). The amphibolite facies rocks in structurally higher areas of the Stikinia footwall (ie. the Wann River gneiss) and at the base of the Tracy Arm terrane hangingwall remained at amphibolite facies temperatures for a sufficient length of time to allow shear fabrics to anneal. In contrast, the structurally lower rocks of the footwall (ie. the Tagish Lake suite plutons and some of the underlying Boundary Ranges suite rocks) either reached amphibolite facies conditions, but did not remain at those conditions for long enough for shear fabrics to anneal, or remained at greenschist facies conditions and retained the shear fabrics that formed during displacement on the Wann River shear zone.

The age of the Mt. Caplice quartz monzodiorite (180.9 ± 3.1 Ma; Chapter 1) provides a maximum age limit for the Wann River shear zone of <184 Ma because the intrusive contains a foliation and mineral-elongation lineation that parallel S_{WRSZ} and L_{WRSZ} preserved throughout the shear zone (Fig. 4.3). Because Tracy Arm terrane must have been in contact with the uppermost rocks of Stikinia by the time rocks from both terranes were metamorphosed under static, amphibolite facies conditions (M_{3T+2S}), the

minimum age of the thrusting event, can be constrained by the ages of metamorphism and cooling in both terranes determined from U-Pb chronometry (presented below).

POST- D_{WRSZ} DEFORMATION

Because deformation, plutonism and volcanism have affected this area since M_{3T+2S} , minor deformation of M_{3T+2S} minerals and post- M_{3T+2S} retrogression of minerals are to be expected. Undulose extinction and subgrain development in quartz with fibrolite inclusions is attributed to post- M_{3T+2S} deformation, and by analogy, some of the undulose extinction in quartz and plagioclase, and subgrains in quartz in other samples may also be considered the result of post- D_{WRSZ} , or may be due to late movement on the Wann River shear zone. However, the age of the strain that resulted in the formation of subgrains and undulose extinction in quartz and feldspar is difficult to determine and may vary from place to place. their formation may have been synchronous with late displacement on the Stikinia-Tracy Arm terrane boundary, where localized chloritized shear surfaces probably reflect shearing that occurred after the metamorphic rocks cooled to greenschist facies conditions; shearing may have been localized along the terrane boundary due to the different rheologic properties of the Tracy Arm terrane and Stikinia rocks. Above and below the terrane boundary subsolidus deformation structures in quartz and feldspar may be older than the Wann River shear zone. In some cases the alteration of minerals, such as the replacement of biotite by chlorite in both pelitic and amphibolitic rocks, can be attributed to the return to greenschist facies conditions after the amphibolite facies conditions of M_{3T+2S} .

THE AGES OF METAMORPHISM AND COOLING IN TRACY ARM TERRANE AND STIKINIA OF THE TAGISH LAKE AREA

Previously, the ages of metamorphism and deformation in the metamorphic rocks of Stikinia west of the Llewellyn fault zone were inferred from observations east of the Llewellyn fault zone, where metamorphism is considered pre-Triassic on the basis of metamorphic clasts in a relatively unmetamorphosed and undeformed Triassic conglomerate in Stikinia (Souther, 1971; Bultman, 1979). This interpretation is consistent with the relatively unmetamorphosed state of the Late Triassic Stuhini Group exposed east of the Llewellyn fault zone, but contrasts with observations made west of the Llewellyn fault zone, where greenschist to lower amphibolite facies metamorphic fabrics affect Tagish Lake suite plutons as young as 180.9 ± 3.1 Ma (U-Pb zircon age for the Mt. Caplice quartz monzodiorite; Chapter 3). The difference in ages of metamorphism across the Llewellyn Fault zone is not surprising because Stikinia was once an active arc (e.g. Wheeler et al., 1991; Monger et al., 1991 and references therein) within which metamorphism and deformation was probably diachronous, varying both laterally and with structural level.

Recently, the age of metamorphism in Tracy Arm terrane of the Tagish Lake area was erroneously assumed to be pre-Triassic, primarily because all metamorphic rocks in the Tagish Lake area were assumed to belong to one terrane. The pre-Triassic age for metamorphism was derived from an interpretation of Rb/Sr data for a granodiorite that was thought to intrude the metamorphic rocks of Tracy Arm terrane (Werner, 1978; Monger et al., 1991), and the incorrect assumption that Tracy Arm

terrane rocks at the south end of Atlin Lake are intruded by a relatively undeformed Late Triassic granitic body (Late Triassic hornblende granodiorite and hornblende gabbro in Fig. 4.2; Brew et al., 1992). However, the ages of these plutons provide no insight into the age of metamorphism in Tracy Arm terrane because they intruded Stikinia and not Tracy Arm terrane. The "granodiorite" analysed by Rb/Sr methods is now referred to as the Mt. Caplice quartz monzodiorite, and is known to be Early Jurassic in age (Chapter 3) and separated from Tracy Arm terrane by the Caplice fault (Fig. 4.2). The Late Triassic granitic body cited by Brew et al. (1992) is separated from Tracy Arm terrane rocks by the Llewellyn fault zone (Fig. 4.2).

Previously published data indicate that metamorphism in the Florence Range suite is definitely older than a large, undeformed, shallow-level, mid-Cretaceous granodiorite that intruded both the Florence Range and Tagish Lake suites northeast of Mt. Caplice (sample K in Fig. 4.2). K-Ar hornblende and biotite cooling ages of 97.5 ± 3.4 Ma and 96.0 ± 3.4 Ma, respectively, for this granodiorite (R.L. Armstrong, pers. comm., 1991) are only slightly younger than its U-Pb zircon crystallization age of 103 ± 0.7 Ma (Appendix A). Cooling of this pluton through the closure temperature for diffusion of Ar in biotite ($280 \pm 40^\circ\text{C}$; Harrison et al., 1985) within ~ 7 Ma of its crystallization, implies that the country rocks it intruded (ie. Florence Range and Tagish Lake suites) were relatively cool by ~ 100 Ma. Al-in-hornblende geobarometry for the same pluton also indicates it crystallized at ~ 7 km depth (Appendix 4.1; Table 4.3).

K-Ar ages for hornblende from amphibolite in the Florence Range suite provide no clear information regarding the age of this important period of Tracy Arm terrane metamorphism because they are inconsistent (144 ± 5 Ma, 214 ± 8 Ma, and 244 ± 16 Ma; unpublished data, R.L. Armstrong, pers. comm., 1991); their isotopic systems may have been variably affected by excess argon, or they may have lost Ar during subsequent metamorphic episodes related to local heating by intrusions.

The Ages of Metamorphism and Cooling in the Florence Range and Boundary Ranges Suites Determined from U-Pb Analyses

Zircon, monazite, titanite and rutile were analysed using U-Pb isotopic methods to determine ages of metamorphism and cooling of the Florence Range and Boundary Ranges suites. Dates on zircon from M_{3T} leucosome, and monazite from a Florence Range suite schist addressed the age of M_{3T} and therefore the minimum age of the Wann River shear zone and the Stikinia-Tracy Arm terrane boundary. The age of the leucosome is interpreted as having formed when the Florence Range suite reached maximum temperatures. The closure temperature for diffusion of Pb in monazite ($\sim 725 \pm 25^\circ\text{C}$; Parrish, 1990) is nominally higher than temperatures experienced by the Florence Range suite during most of its history (Appendix 4.4), making it unlikely that the U-Pb systematics in monazite have been significantly reset.

Titanite and rutile were analysed because they are common to rocks from both Tracy Arm terrane and Stikinia exposed near the terrane boundary. The ages therefore provide a test for the proposed model, which predicts that after amphibolite facies

Tracy Arm terrane rocks were thrust over cooler Stikinia rocks, thermal equilibration would result in rocks near the terrane boundary from both terranes cooling through the same temperatures at roughly the same time. Because the closure temperature for diffusion of Pb in titanite ($\sim 600 \pm 50^\circ\text{C}$; Tucker et al., 1987; Heaman and Parrish, 1991) is probably lower than the maximum temperature experienced by the Florence Range suite, the U-Pb systematics in titanite may have been reset during the most recent $\sim 650^\circ\text{C}$ metamorphic episode (M_{3T}). The growth of titanite at the expense of ilmenite in the Wann River gneiss may have occurred during M_{1S} or M_{2S} , and its age could reflect the age of one or both of the metamorphic episodes. Alternatively, if titanite grew during M_{1S} and temperatures $>550^\circ\text{C}$ were maintained in the uppermost Stikinia rocks for a sufficient period of time during M_{2S} , the U-Pb systematics in titanite from this suite may have been completely reset, and could reflect the approximate age of M_{2S} . The U-Pb ages for rutile from these two suites are expected to be the same; they should reflect cooling after Tracy Arm terrane was thrust over Stikinia because temperatures reached during both M_{3T} and M_{2S} exceeded the closure temperature for diffusion of U and Pb in rutile ($420 \pm 30^\circ\text{C}$, assuming a grain size of 90 to 210 μm and cooling rate of 0.5 to 1°C ; Mezger et al., 1989).

Analytical Methods

U-Pb analyses of zircon, monazite, titanite and rutile fractions were performed at the Geological Survey of Canada in Ottawa using analytical procedures summarized by Parrish et al. (1987; 1991). Blanks for U were 5 pg or less, and blanks for Pb were

7-20 pg for zircon analyses, from 17 to 30 pg for monazite analyses, from 20-300 pg for titanite analyses and were ~26 picograms for rutile analyses. Analytical results are presented in Table 4.5.

The Age of Metamorphism in the Florence Range Suite

(A) Leucosome from the Florence Range suite (LC-86-84b)

A sample of coarse-grained trondhjemitic leucosome with large, randomly oriented biotite crystals (>5 cm in diameter; Plates 4.11a, b) was collected from a 30 cm thick leucosome layer near the Wann River headwaters to determine the age of M_{3T} (sample A in Fig. 4.2). At that locality, leucosome layers are from 1 to 30 cm thick and separated by garnet- and biotite-bearing, quartzofeldspathic semipelitic layers (paleosome) of similar width. The leucosome is considered a product of M_{3T} because it lacks a penetrative metamorphic fabric. Undulose extinction in quartz and plagioclase and subgrain development in quartz are the only evidence of subsolidus deformation and are likely the result post- M_{3T} deformation.

All analyses of zircon from the leucosome are discordant (Fig. 4.10a); the most discordant fraction (D) comprised clear, subhedral to anhedral, equant to slightly elongate zircon crystals which are interpreted as inherited xenocrysts, whereas the other four fractions (A, B, C, and E) comprised clear zircon shards (length:width ~2:1) with only minor inclusions, that are thought to have crystallized at the same time as the leucosome. Despite the lack of visible cores in fractions A, B, C and E, inheritance of an older component of zircon is considered the primary cause of discordance in all

fractions. Some of the inherited zircon is probably be older than 1850 Ma, on the basis of the regression of data from all five fractions (Fig. 4.10a).

The youngest and oldest limits for the age of zircons from the M_{3T} leucosome are based on the lower intercepts determined from the regression of the four youngest fractions ($173 \pm 2.4/-2.9$ Ma) and the regression of all five fractions (179.7 ± 4 Ma); the best estimate of the age of the leucosome is therefore interpreted as 177 ± 7 Ma. This estimate does not include the insignificant correction for discordance due to ^{206}Pb deficiency, which is caused by exclusion of ^{230}Th from zircon during crystallization (Mattinson, 1973). Pb-loss is not considered significant in this interpretation, because only fracture-free, clear crystals were analysed, all crystals were strongly abraded prior to dissolution (Krogh, 1982), and U-Pb ages for metamorphic monazite (see below) indicate that the age of metamorphism was probably <178.3 Ma, and younger than the maximum assigned age limit for the zircon fractions (184 Ma). The interpreted age of 177 ± 7 Ma is also considered reasonable because the fraction closest to concordia, Fraction A, has Pb/U ages that are well within the assigned uncertainty range ($^{207}\text{Pb}/^{235}\text{U}$ age = 182 ± 0.4 Ma; $^{206}\text{Pb}/^{238}\text{U}$ age = 179.0 ± 0.3 Ma).

The 177 ± 7 Ma U-Pb age for zircon from metamorphic leucosome (Fig. 4.10a) is interpreted as the age of the M_{3T} leucosome, and therefore the age of M_{3T} . Because the leucosome is interpreted as a local melt, its age is thought to reflect the time that the Florence Range suite reached peak metamorphic temperatures.

(B) Monazite from schist of the Florence Range suite (LC-89-59)

Monazite was extracted from a garnet-muscovite-biotite schist exposed south of the Wann River (sample B in Fig. 4.2; Table 4.1). The monazite occurs in the matrix and as inclusions in M_{3T} biotite grains. All fractions were multigrain fractions and comprised clear, yellow, slightly elongate, anhedral crystals with minor dark inclusions and no visible cores. The analyses spread above and below concordia between ~180 Ma and 174 Ma, when calculated using common Pb isotope ratio corrections calculated for 175 Ma using the Stacey and Kramers (1975) Pb isotope growth curve (Fig. 4.10b). The two youngest analyses (M2 and M3) plot above concordia, probably in part the result of excess ^{206}Pb from ^{230}Th decay. Assuming Th/U of the metamorphic fluid is 3, corrected ages using common Pb isotope ratio corrections calculated for 175 Ma based on the Stacey and Kramers (1975) Pb isotope growth curve, fraction M2 has a $^{207}\text{Pb}/^{235}\text{U}$ age of 177.4 ± 0.9 Ma and $^{206}\text{Pb}/^{238}\text{U}$ age of 177.2 ± 0.9 Ma, and fraction M3 has a $^{207}\text{Pb}/^{235}\text{U}$ age of 174.2 ± 1.5 Ma and $^{206}\text{Pb}/^{238}\text{U}$ age of 174.8 ± 0.5 Ma ($^{206}\text{Pb}/^{238}\text{U}$ ages have been corrected for excess ^{206}Pb , but do not include uncertainty due to correction). The $^{206}\text{Pb}/^{238}\text{U}$ ages and $^{207}\text{Pb}/^{235}\text{U}$ ages for fractions M2 and M3 are similar, but continue to plot above concordia, possibly in part because an incorrect Th/U and/or initial Pb isotopic ratios for the fluids from which monazite crystallized were used (Fig. 4.10b).

The analyses have also been recalculated using initial common Pb ratios determined from $^{206}\text{Pb}/^{204}\text{Pb}$ vs $^{238}\text{U}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{235}\text{U}/^{204}\text{Pb}$ isochron plots for fractions M2, M5 and M6 (Table 4.5). Regression of Pb/Pb ratios involving

fraction M3 do not converge, possibly indicating that monazite crystals in this fraction formed from a different fluid than those of the other fractions, or experienced Pb-loss. The calculated Pb isotopic ratios are interpreted as reflecting the initial Pb isotopic composition of the fluid from which the monazite crystallized. They appear to be compare favorably with Pb isotope correlation diagrams for rocks with a North American continental margin assemblage protolith (Sevigny and Parrish, in press). The recalculated analyses for fractions M2, M5 and M6 cluster about 177.5 Ma, and fraction M3 remains above ~173 Ma (Fig. 4.10b inset). The agreement between fractions M2, M5 and M6 argues strongly for a monazite crystallization age of 177.5 ± 1.5 Ma. This age is considered the age of M_{3T} , and by inference the age of M_{2S} , and is supported by the 177 ± 7 Ma U-Pb zircon age for the M_{3T} leucosome.

(C) Titanite from a calc-silicate rock of the Florence Range suite (LC-89-86)

Titanite was separated from a Florence Range suite garnet-diopside-vesuvianite calc-silicate exposed south of the Wann River headwaters (sample C in Fig. 4.2), and was analysed to constrain the cooling history of Tracy Arm terrane. The titanite grains are anhedral and commonly occur as inclusions in calcite crystals. Titanite analyses shown in Figure 4.10c were determined using common Pb isotope ratio corrections calculated for 180 Ma using the Stacey and Kramers (1975) Pb isotope growth curve. Despite their large error ellipses, which allow them to fortuitously intersect concordia between 212 and 192 Ma, they do not overlap one another. The discrepancy between these analyses could be caused by the use of inappropriate initial common Pb isotopic

ratios in the common Pb correction, more than one generation of titanite in each fraction, partial resetting of the U-Pb systematics of M_{1T} or M_{2T} titanite crystals during M_{3T} , or a combination of these causes.

The possibility that the discrepancy between these analyses was caused by the use of inappropriate initial common Pb isotopic ratios in the common Pb correction was tested by recalculating the titanite analyses using initial Pb isotope compositions determined from $^{206}\text{Pb}/^{204}\text{Pb}$ vs $^{238}\text{U}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{235}\text{U}/^{204}\text{Pb}$ isochron plots (Figure 4.10d; Table 4.5), which result in the titanite analyses plotting on top of one another, with the most precise analysis located directly over concordia near 190 Ma. The calculated Pb isotopic ratios are slightly lower than those calculated for the monazite fractions described above.

If only the $^{206}\text{Pb}/^{238}\text{U}$ ages for titanite are considered here, the age of the titanite analysis with the smallest uncertainty (S1; 189.6 ± 1.8 Ma) is older than the age of M_{3T} determined from both the zircon from U-Pb analyses of zircon from metamorphic leucosome and monazite from the Florence Range suite. This age is therefore interpreted as the result of partially reset U-Pb systematics in M_{1T} or M_{2T} titanite, or a mixture between M_{3T} titanite and M_{1T} and/or M_{2T} titanite.

(D) Rutile from semipelite of the Florence Range suite (LC-91-469)

Rutile from a medium-grained sillimanite-garnet-biotite-muscovite semipelite collected near the Willison Creek headwaters was analysed to constrain the cooling history of the Florence Range suite (sample D in Fig. 4.2). Rutile is anhedral, clear,

contains minor inclusions and occurs along relic garnet crystal faces (now replaced by biotite + quartz + fibrolite + fine-grained white mica) and fractures. U-Pb analyses of rutile fragments (>149 μm diameter) plot to the far right of concordia with large uncertainties when calculated using initial Pb isotope ratios based on the Stacey and Kramers (1975) Pb isotope growth curve (Fig. 4.10c). Their discordance is interpreted as indicating that these initial Pb isotope ratios are inappropriate for these samples. Initial Pb ratios determined from Pb isochron diagrams (Table 4.5) compare favorably with those determined for western North American rocks with a continental margin assemblage protolith (Sevigny and Parrish, in press), are similar to those calculated for the Florence Range suite titanite, and slightly lower than those calculated for the Florence Range suite monazite. Results calculated using the initial Pb ratios determined from Pb isochron diagrams are shown in Figure 4.10d. The $^{206}\text{Pb}/^{238}\text{U}$ are considered most reasonable because they are least affected by the common Pb correction. They indicate that this sample cooled through $420 \pm 30^\circ\text{C}$, the closure temperature for U and Pb in rutile (assuming a grain size of 90 to 210 μm and a cooling rate of $0.5\text{--}1^\circ\text{C}/\text{Ma}$; Mezger et al., 1989) 142.5 ± 2.5 Ma ago.

The Age of Metamorphism in the Boundary Ranges Suite

Titanite and rutile from two samples of Permian Wann River gneiss were analysed to determine the ages of M_{2S} and cooling in the Boundary Ranges suite. The titanite sample is from the main body of the Wann River gneiss, which is exposed north of the Wann River and lies below the Tracy Arm-Stikinia contact, within the

Wann River shear zone (sample E in Fig. 4.2). The rutile sample is from a ~30 m thick sliver of Wann River gneiss that is imbricated with Tracy Arm Terrane rocks, ~475 m above the main body of the Wann River gneiss (sample F in Fig. 4.2).

(E) Titanite from the Wann River gneiss (LC-91-458b)

Titanite grains in the Wann River gneiss are anhedral, are generally associated with hornblende, and commonly occur as inclusions in hornblende crystals. The analyses for two fractions of anhedral titanite from the Wann River gneiss are difficult to interpret because one error ellipse plots just below concordia near 183 Ma (S1) and the other plots near 196 Ma (S2; Fig. 4.10c). The discordance and discrepancy between these analyses may reflect partial resetting, mixing between components of titanite that formed at different times and/or Pb-loss. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages are least affected by Pb-loss, and are >205 Ma (with uncertainties of ~30 Ma; Table 4.5), consistent with the interpretation that the Boundary Ranges suite experienced a pre-Early Jurassic metamorphic event that was followed by partial resetting of the U-Pb systematics after 181 Ma. The U-Pb systematics may have been reset during M_{2S} or due to increased temperature caused by the intrusion of the Early Jurassic Hale Mountain granodiorite (186 ± 1.5 Ma; Chapter 3). Alternatively, a second titanite component may have formed during M_{2S} and mixing between M_{1S} titanite and M_{3S} titanite may be the cause of the discrepancy between these analyses and their discordance. All interpretations assume that some titanite grew before M_{2S} , which is reasonable because in this sample titanite occurs as inclusions in hornblende crystals.

This data is best interpreted as indicating that the Wann River gneiss cooled below the closure temperature for U and Pb in titanite ≤ 183 Ma ago.

(F) Rutile from the Wann River gneiss (LC-89-91b)

Two fractions of clear, pale red, $>149\mu\text{m}$ rutile fragments and crystals from the Wann River gneiss have error ellipses that lie on concordia between 144.9 and 146.1 Ma (Table 4.5; Figs. 4.10c; 4.10d). The $^{206}\text{Pb}/^{238}\text{U}$ ages are 145.4 ± 0.6 and 145.5 ± 0.6 , and the $^{207}\text{Pb}/^{235}\text{U}$ ages are 146.1 ± 1.6 and 146.0 ± 1.8 for fractions R1 and R2, respectively. On the basis of the Pb/U age with the largest uncertainty, which includes all of the other Pb/U ages and their uncertainties, the Wann River gneiss cooled through the closure temperature for U and Pb in rutile ($420 \pm 30^\circ\text{C}$; Mezger et al., 1989), 146.0 ± 1.8 Ma ago, at a time in agreement with rutile ages for a sample from the Florence Range suite (Sample "D"; LC-91-469).

The Age of the Stikinia-Tracy Arm Terrane Boundary

The maximum age of the Stikinia-Tracy Arm terrane boundary and the Wann River shear zone is limited by the age of the youngest Tagish Lake suite pluton, the Mt. Caplice quartz monzodiorite (180.9 ± 3.1 Ma; Chapter 3) for two reasons. First, the absence of plutons and dykes of the Early Jurassic Tagish Lake suite in Tracy Arm terrane, in contrast to their abundance in Stikinia indicates that Tracy Arm terrane and Stikinia were not amalgamated during the emplacement of the 190-184 Tagish Lake suite plutons (Fig. 4.2; Chapter 3). Secondly, the 180.9 ± 3.1 Ma Mt. Caplice quartz

monzodiorite has been affected by D_{WRSZ} and predates amalgamation. Because the transposition fabric of the Wann River shear zone is the oldest documented link between Stikinia and Tracy Arm terrane, its mid-Jurassic age does not support the supposition that Tracy Arm terrane (Nisling) and Stikinia were together before Triassic time.

The 177.5 ± 1.5 Ma U-Pb age for monazite from Tracy Arm terrane schist is interpreted as the age of M_{3T+2S} metamorphic minerals that overprinted the Wann River shear zone transposition fabric, and therefore provides a minimum age of 176 Ma for displacement along the Wann River shear zone and the amalgamation of Tracy Arm terrane and Stikinia. This age is supported by the 177 ± 7 Ma U-Pb zircon age for the M_{3T} leucosome, which dates maximum metamorphic temperatures. Thus, the Stikinia-Tracy Arm terrane boundary and the Wann River shear zone formed between 184 and 176 Ma, not long after the Mt. Caplice quartz monzodiorite was emplaced at ~21 km depth (Table 4.3). Assuming this pluton was not rapidly unroofed before Stikinia and Tracy Arm terrane were juxtaposed, the currently exposed Wann River shear zone probably formed at similar depths, consistent with the syn- to post- D_{WRSZ} upper greenschist to amphibolite facies stable mineral assemblages preserved in the shear zone.

THE AGES OF STRUCTURES PRESERVED OUTSIDE THE WANN RIVER SHEAR ZONE AND POST-ACCRETION DEFORMATION, COOLING, AND DENUDATION OF STIKINIA AND TRACY ARM TERRANE

Structures and metamorphic mineral assemblages preserved above the Wann River shear zone in Tracy Arm terrane (D_{IT} , D_{2T} , M_{IT}) and below the shear zone in Stikinia (D_{IS} , D_{2+3S} , M_{IS}) are older than the ca. 180 Ma shear zone. Their relation to the amalgamation of Tracy Arm terrane and Stikinia is not clear; they could have formed during initial convergence, or much earlier. They are not correlated between Tracy Arm terrane and Stikinia because they cannot be followed across the Wann River shear zone.

Structures and metamorphic minerals in Stikinia that are older than Wann River shear zone may be related to possible Lower Permian to Middle Triassic deformation and metamorphism of Stikinia rocks in the "type-Stikinia" area near the Stikine River to the south (Fig. 4.1; Read et al., 1983; Anderson, 1989; 1993; Greenwood et al., 1991) and from east of the Llewellyn fault zone (e.g. Souther, 1971), but due to the possible diachronous nature of metamorphism and deformation such a correlation is speculative. Andalusite in these rocks may have formed in a region of high heat flow within the arc.

After the metamorphic rocks of Stikinia and Tracy Arm terrane reached peak temperatures, they began to cool, resulting in the formation of greenschist facies minerals such as chlorite and the alteration of garnet. Bultman (1979) reported a 165 ± 3 Ma K-Ar hornblende cooling age for the Early Jurassic Hale Mountain granodiorite

(Table 4.5); this age can be interpreted to indicate that this sample, and presumably the adjacent metamorphic rocks cooled through the closure temperature for diffusion of Ar in hornblende ($500 \pm 50^\circ\text{C}$; Harrison, 1981) 165 ± 3 Ma ago. Not surprisingly, cooling ages for rutile are slightly younger. The rutile ages in both Tracy Arm terrane and Stikinia are similar, indicating that this area cooled through $420 \pm 30^\circ\text{C}$, the closure temperature for U and Pb in rutile (Mezger et al., 1989), ~ 147 - 142 Ma ago. (Figs. 4.10c; 4.10d). The change from $\sim 700^\circ\text{C}$ ca. 175 Ma ago, to $\sim 420^\circ\text{C}$ ca. 145 Ma ago is indicative of a time-averaged cooling rate of $\sim 9^\circ\text{C}/\text{Ma}$.

Indications that the metamorphic rocks in the Tagish Lake area were deformed and experienced varying petrogenetic conditions after most of the displacement on the Wann River shear zone, and probably after M_{3T} , include (1) folding of S_{WRSZ} by the Willison synform and broad, open, upright folds, (2) truncation of the Wann River shear zone by the Llewellyn fault zone, Willison Creek fault zone and steep faults, (3) the unconformable relationships between Florence Range suite rocks containing S_{WRSZ} and the overlying Mt. Switzer and Sloko Group volcanic rocks, and (4) the intrusive relationship between Tertiary plutons and the Florence Range suite rocks (Appendix A). The mid-Cretaceous depth of both Florence Range and Boundary Ranges suite rocks on the west side of the Llewellyn fault zone can be constrained to less than 13 km, and possibly less than 6 km, by Al-in-hornblende geobarometry and K-Ar cooling ages for the mid-Cretaceous pluton that intruded them (Table 4.3; Appendix 4.1). By ~ 58 Ma volcanic rocks were deposited on a surface that is approximately equivalent to that exposed today.

THE GEOMETRY OF THE STIKINIA-TRACY ARM TERRANE BOUNDARY DURING ACCRETION

Two models for the accretion of Tracy Arm terrane to Stikinia are consistent with the known constraints for the geometry of the Stikinia-Tracy Arm terrane boundary: (1) Tracy Arm terrane was obducted over Stikinia, or (2) Tracy Arm terrane was inserted into Stikinia at mid-crustal levels as a wedge. The constraints for the geometry of the Stikinia-Tracy Arm terrane boundary during accretion include (1) amphibolite facies Tracy Arm terrane metasedimentary rocks sit directly atop a generally right-way-up section of Stikinia rocks that were probably at a ~21 km depth ca. 180 Ma ago; (2) shear sense indicators demonstrate that Tracy Arm terrane was thrust from the north-northwest over Stikinia (Fig. 4.3); (3) emplacement of Tracy Arm terrane over Stikinia between 184 and 176 Ma ago resulted in a change from kyanite-stable (M_{2T}) to sillimanite-stable conditions (M_{3T}) in Tracy Arm terrane, and the formation of an inverted metamorphic gradient in the uppermost Stikinia rocks; (4) no detritus in a Middle Jurassic or older sedimentary rock can be unequivocally linked to a source in Tracy Arm terrane; and (5) north of the Tagish Lake area, in the Aishihik Lake area of southwestern Yukon, metamorphic rocks considered part of Stikinia (Aishihik assemblage; Chapter 5) overlie part of Tracy Arm terrane (Nisling assemblage; Chapter 2). Purely strike-slip models are not considered here because the formation of an inverted metamorphic gradient cannot easily be attributed to a steep strike-slip fault. Furthermore, the near horizontal to moderately dipping orientation of the Stikinia-Tracy Arm terrane boundary and Wann River shear zone observed on the

northern limb of the Willison synform is considered to be closer to the original orientation of the shear zone and terrane boundary than the near-vertical orientation of fabrics preserved on the western limb of the Willison synform. However, displacement along the Stikinia-Tracy Arm terrane boundary may have had a component of strike slip motion. The implications of the top-to-the-south sense of shear observed in the Tagish Lake area are discussed in Chapter 5.

From the constraints described above, it follows that Tracy Arm terrane was detached from its original lower crustal and lithospheric basement, either during or before it was thrust over Stikinia. Presumably, the basement to the middle and upper crust of Tracy Arm terrane either underplated the Stikinia arc, or was subducted into the mantle. The remaining Tracy Arm terrane middle and upper crust may have been sufficiently buoyant to have been delaminated from its middle and lower crustal basement and either thrust over the arc rocks of Stikinia, or inserted into Stikinia as a wedge.

Both the obduction and the insertion models allow for the creation of an inverted metamorphic sequence. To preserve this sequence, the upper Stikinia rocks must have cooled soon after the hotter Tracy Arm terrane rocks were emplaced because equilibration of the isotherms in the absence of erosion would have eventually resulted in a "normal" metamorphic sequence that was continuous from the uppermost Tracy Arm terrane rocks, across the terrane boundary, to the lowest Stikinia rocks. A number of authors have pointed out the rapid cooling and fault displacement rates required to preserve an inverted metamorphic gradient, especially if the footwall rocks

were at near surface temperatures when thrusting began (e.g. England and Molnar, 1993; Ruppel and Hodges, 1994). In this case, slightly lower rates are required because amphibolite facies rocks were thrust over greenschist facies rocks. The displacement rates required to preserve an inverted metamorphic gradient are 4 cm/yr (Ruppel and Hodges, 1994), which is not considered unreasonable for subduction of Tracy Arm terrane below a volcanic arc, because the Australian continental lithosphere has been subducted under the Indonesian arc at rates between 3 and 7 cm/yr (Hamilton, 1979). The higher radiogenic heat production in Tracy Arm terrane, relative to Stikinia may have contributed to the preservation of the inverted metamorphic gradient.

The obduction model requires that at least 15 km of Tracy Arm terrane were erosionally or tectonically removed, but neither has the upper crustal component of Tracy Arm terrane, nor a significant volume of sedimentary rocks derived from Tracy Arm terrane have been recognised. Gehrels and Greig (1991) have suggested that some detrital zircons in the Late Jurassic sediments of the Gravina Belt (Fig. 4.1) were derived from "Yukon-Tanana terrane" (in part equivalent to Tracy Arm terrane; Chapter 2), but Tracy Arm terrane cannot be unequivocally identified as the provenance of Middle Jurassic or older sediments. Rare metamorphic clasts found in Late Triassic Stuhini Group or Early Jurassic Laberge Group conglomerates of Stikinia were likely derived from within Stikinia (Chapters 1 and 2).

The strongest argument for the insertion model is the Stikinia-over-Tracy Arm terrane relationship observed in the Aishihik Lake area. The upper member-to-the-west

sense of shear determined from shear bands in the Aishihik Lake batholith (Johnston, 1993), which is lithologically and isotopically similar to, and coeval with the Tagish Lake suite, is consistent with the insertion model. If Tracy Arm terrane formed a wedge that was inserted into Stikinia at mid-crustal levels, it could have caused significant crustal thickening, without contributing detritus to the Middle Jurassic sedimentary record in Stikinia or causing an increase pressure experienced by the rocks that lay above the base of the Tracy Arm terrane wedge. Furthermore, if concomitant rapid erosion and/or normal faulting occurred in the overlying Stikinia arc these rocks may have experienced a decrease in pressure. As mentioned earlier, the change from kyanite-stable (M_{2T}) to sillimanite-stable conditions (M_{3T}) in Tracy Arm terrane after accretion may, at least in part, reflect a decrease in pressure.

The Lower to Middle Jurassic (Hettangian to Bajocian) conglomerates of the Laberge Group are indicative of significant Early to mid-Jurassic erosion. Increasing proportions of plutonic clasts and decreasing input from volcanic and sedimentary sources with time, are interpreted as reflecting progressive erosion of the Stikinia arc (Dickie, 1989), but the age of the Laberge Group is primarily limited by fossil ages, and the uncertainty associated with correlating absolute ages with fossil ages makes it difficult to directly relate the accretion of Tracy Arm terrane to the deposition of sedimentary sequences in Stikinia.

Normal faults that may be at least as old as Middle Jurassic have been recognised in Stikinia in the Stikine River area (eg. Souther and Symons, 1974; Brown and Gunning, 1989; Logan et al., 1992). A possible post-Late Triassic normal fault has

also been recognised in the Tagish Lake area, near Hoboe Glacier, on the east side of the Llewellyn fault zone, where the Mt. Mussen shear zone juxtaposes relatively unmetamorphosed (sub-greenschist facies) Triassic volcanic and sedimentary rocks of Stikinia (Stuhini Group) against foliated and sheared greenschist facies metavolcanic rocks that resemble the Boundary Ranges suite rocks exposed on the west side of the Llewellyn fault zone (Fig. 4.2; Plate 4.14b). The relationship between the accretion of Tracy Arm terrane and the Mt. Mussen shear zone is not clear because the age of the shear zone can only be limited to Late Triassic or younger..

THE AMALGAMATION OF TRACY ARM TERRANE AND STIKINIA AND ITS RELATION TO COEVAL TECTONISM IN THE CORDILLERA

The accretion of Tracy Arm terrane to Stikinia was coeval with a period of terrane accretion and intense deformation in the Canadian Cordillera that began in latest Early Jurassic to early Middle Jurassic time, probably due to the opening of the northern Atlantic ocean and increased normal convergence between North America and subducting crust to the west (e.g. Coney and Evenchick, in press). The amalgamation of Tracy Arm terrane to the western margin of Stikinia occurred either at the same time as, or soon after Stikinia was accreted to the composite western margin of North America (Cache Creek-Quesnellia-Slide Mountain-Kootenay-North America). Although the terranes of the Intermontane Superterrane were probably close to one another by Early Triassic time (e.g. Monger et al., 1982), the age of the juxtaposition of Stikinia and Cache Creek terrane along the King Salmon fault can be

constrained to Aalenian time (Mortimer, et al., 1990; Ricketts et al., 1992). However, the age of the Stikinia-Tracy Arm terrane boundary cannot be directly compared with the Aalenian accretion of Stikinia (Ricketts et al., 1992; Fig. 4.1), or with the ages of other major structures of similar age in the Cordillera that have been constrained by fossil ages, because the uncertainties of the absolute ages that have been applied to paleontologically defined epochs are large. For example, on the basis of Harland et al.'s (1990) time scale, the 184 to 176 Ma Stikinia-Tracy Arm terrane boundary formed in Toarcian to Aalenian time, possibly at the same time as the King Salmon fault, but if the Decade of North American Geology 1983 geologic time scale (Palmer, 1983) is used, the accretion of Tracy Arm terrane occurred during the Bajocian to Bathonian epochs, after the formation of the King Salmon fault. The 184 to 176 Ma age for the Stikinia-Tracy Arm terrane boundary can not be used to determine whether the Insular superterrane was accreted to the western margin of the Intermontane superterrane during Middle Jurassic or mid-Cretaceous time, but it does indicate that these superterranes could not have been together before 184 Ma ago.

In the Stikine River area, there is a widespread upper Pliensbachian to Toarcian hiatus in Stikinia (Anderson, 1993, and references therein). This hiatus followed the intrusion of plutons of the Early Jurassic Texas Creek plutonic suite, many of which, like the coeval Tagish Lake suite plutons, are ductily deformed (e.g. McClelland and Mattinson, 1991; Brown et al., 1992; Porter, 1992). It has been suggested that this deformation may correlate with a Middle Jurassic contractional event along the King Salmon Fault (Brown et al., 1992). The hiatus may also reflect the initial underplating

by buoyant continental or transitional crust. As pointed out by Thorkelson et al. (in review), inherited Precambrian zircons appear to be more common in Middle Jurassic and younger Stikinia plutons than in the older Stikinia plutons. Tracy Arm terrane rocks presently occur west of the Skeena and Stikine arches and farther north, but have not been recognised west of the Bowser Basin or south of the Skeena arch. It is tempting to relate the mid-Jurassic evolution of the arches to the accretion of Tracy Arm terrane, but the present distribution of Tracy Arm terrane, relative to Stikinia may not reflect its original relative location because structures with unknown offset, such as the Llewellyn Fault zone, may have been the loci of significant strike-slip displacement since Tracy Arm terrane was accreted.

Earliest Middle Jurassic metamorphism and deformation are not unique to the Tagish Lake area, but are common throughout the northern and central Cordillera. Middle Jurassic (187-137 Ma) K-Ar cooling ages for muscovite and biotite from metamorphic rocks in the northeastern Cordillera (previously referred to as Yukon-Tanana terrane, and now considered part of the Nisutlin subterrane of Kootenay terrane and Nisling terrane; Wheeler et al., 1991; Fig. 4.1) are considered indicative of Early Jurassic, and possibly older metamorphism (Tempelman-Kluit, 1976; Tempelman-Kluit and Wanless, 1975; Wilson et al., 1985). In addition, a U-Pb titanite cooling age from the Aishihik Lake area indicates that metamorphic rocks in that area were metamorphosed by Middle Jurassic time (184 ± 2 Ma; Johnston and Timmerman, 1994; Fig. 4.1). Deformation farther north and west may have been slightly older. In Alaska, Early Jurassic (~204-185 Ma) $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende, biotite and white mica

cooling ages for amphibolite gneisses of "Yukon-Tanana terrane" are interpreted as the result of thrusting of terranes over parautochthonous North American strata (Hansen et al., 1991). West of the Tracy Arm terrane, in southeast Alaska, there is evidence for Early to Middle Jurassic deformation; in the Alexander terrane, where dextral shear along the Duncan Canal shear zone is interpreted as recording either juxtaposition of Alexander terrane and Stikinia, or northward translation of Alexander terrane during Early to Middle Jurassic time (McClelland and Gehrels, 1990).

Pressures determined from equilibrium metamorphic mineral assemblages and isotopic ages from parautochthonous and autochthonous rocks of ancestral North America in the southern and central Omineca Belt indicate that crustal thickening resulted in doubling of the stratigraphic thickness of sediments on the western margin of ancestral North America during Middle Jurassic time (Pigage, 1978; Archibald et al., 1983; Mortensen et al., 1987; Calderwood et al., 1990). Restoration of the proposed cumulative 500 to 1000 km displacement on Cretaceous and younger dextral strike-slip faults from east of Stikinia in the northern Rocky Mountain trench (Eisbacher, 1985; Gabrielse, 1985) indicates that the greater part of Stikinia (and therefore presumably Tracy Arm terrane) was located west of what is now southeastern British Columbia. If this is the case, the coeval onset of orogeny that result in the formation of the Omineca Belt may be linked to deformation in Stikinia.

Latest-Early Jurassic to earliest-Middle Jurassic deformation and metamorphism have been reported as far south as the Canada-United States border, but farther south most Jurassic tectonism appears to be younger (~163 to 155 Ma Nevadan Orogeny;

e.g. Schweickert et al., 1984; Graymer and Jones, 1994), with the exception of deformation that can be linked to Mesozoic terrane accretion in the Northern Sierra and Eastern Klamath terranes, two terranes that, like Stikinia, have been included in the "McCloud belt terranes" (e.g. Miller, 1987). Deformation in Northern Sierra terrane could be as old as 179 Ma (Snoke et al., 1982), but farther west in the northern Sierra Nevada metamorphic belt deformation and metamorphism occurred <174 Ma ago (Hacker, 1992). The oldest accretionary event in the terranes of the northern Klamath Mountains (Eastern Klamath terrane) occurred between 172 and 162 Ma (Hill, 1985).

Thus, the 184-176 Ma accretion of Tracy Arm terrane to the western margin of Stikinia is roughly coeval with the accretion of terranes along the northern and central portions of the western composite margin of North America. Younger Mesozoic events are related to continued contraction of the orogen and the accretion of terranes to the west.

APPENDIX 4.1: ALUMINUM-IN-HORNBLENDE GEOBAROMETRY IN THE EARLY JURASSIC TAGISH LAKE SUITE AND A MID-CRETACEOUS, UNDEFORMED GRANODIORITE.

Hornblende crystals from three Early Jurassic Tagish Lake suite plutons and an undeformed mid-Cretaceous granodiorite were analysed to determine their depth and presumably the depth of the adjacent layered metamorphic rocks at the time of magma crystallization. Of the five samples analysed, two are from the Hale Mountain granodiorite (one from near the top of White Moose Mountain and one from west of Hoboe glacier), one is from the Mt. Caplice quartz monzodiorite, one is from the Hoboe quartz syenite, and one is from the mid-Cretaceous granodiorite (103 ± 0.7 Ma; Appendix A; Fig. 4.2) that is exposed north of Mt. Caplice and west of the Llewellyn fault zone. The Tagish Lake suite plutons intruded only Stikinia rocks (Chapter 3 ; Fig 2), whereas the undeformed granodiorite is in intrusive contact with metamorphic rocks belonging to both Stikinia and Tracy Arm terrane (Fig. 4.2).

It has been shown empirically (Hammarstrom and Zen, 1986; Hollister et al., 1987) and experimentally (Johnson and Rutherford, 1989; Rutter et al., 1989; Thomas and Ernst, 1990; Schmidt, 1991) that the total aluminum content (Al_T) in hornblende varies linearly with pressure between at least ~2 and 10 kbar, if the following conditions are met: (1) the sample contains the assemblage plagioclase (An_{25-35} , for the calibration of Hollister et al., 1987), quartz, potassium feldspar, biotite, hornblende, titanite, apatite and magnetite or ilmenite and (2) the hornblende formed in a SiO_2 -rich magma. The latter condition is thought to be fulfilled if hornblende shows textural

evidence of being in equilibrium with quartz. If the Al_T content is within a 95% confidence limit, the uncertainty associated with this geobarometer is ± 1 kbar for the calibration Hollister et al. (1987), and ± 2 kbar for the calibration of Hammarstrom and Zen (1986). For the calibration of Johnson and Rutherford (1989) the uncertainty varies with the Al_T content of the hornblende, and the maximum estimated uncertainty is ± 0.5 kbar at pressures less than 8 kbar and slightly more at higher pressures (± 0.6 kbar).

All of the samples that were analysed have the equilibrium mineral assemblage required for this geobarometer (see Chapter 3 for rock descriptions of the Tagish Lake suite samples and Appendix A for a description of the mid-Cretaceous pluton), and the hornblende grains exhibited textural evidence of being in equilibrium with quartz. It is assumed that the hornblende crystal compositions have not changed since they crystallized from a magma, despite subsolidus deformation of adjacent quartz grains (undulose extinction and curved grain boundaries) in some samples (G-J). The mid-Cretaceous pluton (Sample K) has a granitic texture, indicating that it has not been deformed. Hornblende crystals in all samples, except the Hale Mountain granodiorite sample from west of Hoboe glacier (Sample H; Fig. 4.2; Plate 4.7a), showed no evidence of deformation or recrystallization.

Analytical Methods and Results of geobarometry

The hornblende samples were analysed using the same methods as those used for microprobe analyses biotite and garnet (this paper), except that the Kakanui

hornblende standard (U.S.N.M. Standard 143965) was used for calibration. Hornblende was not analysed for H₂O, F or Cl, which accounts for the weight% totals <100 (Appendix 4.2). Structural formulæ for hornblende were calculated on the basis of 23 oxygens. Analytical uncertainty for the relative total Al content is 0.1%.

The hornblende grains that were analysed are regarded as crystals that formed during primary magmatic crystallization because the measured Si contents are ≥ 7.5 , the compositional limit for hornblende proposed by Leake (1971), and because their Ca contents are ≥ 1.6 . Despite zoning in plagioclase grains from some samples, the plagioclase compositions in all samples were between 24% and 40% anorthite (Appendix 4.3; Table 4.3), within or close to the limits of the geobarometer. It is assumed that the plagioclase grains have retained their original compositions, although the crystallization composition of plagioclase grains in sample H is not well known because the cores of the plagioclase augen are sericitized and the oligoclase (An₂₈) rims are interpreted as having formed during dynamic recrystallization of the feldspars, under amphibolite facies conditions.

Pressures determined using four different calibrations (Hammarstrom and Zen, 1986; Hollister et al., 1987; Johnson and Rutherford, 1989; Schmidt, 1991) varied by as much as 2 kbar (Appendix 4.2). Because the variation is larger than the proposed uncertainties for most of the calibrations, the assigned uncertainty for each sample of ± 2 kbar is intended to reflect the accuracy of the results (Table 4.3 and Appendix 4.2). However, the precision of a single calibration is probably less than ± 2 kb, and differences greater than the uncertainty of the calibration probably reflect genuine

relative changes in pressure. The calibration Hollister et al. (1987) is preferred because it is an empirical calibration, and has a low uncertainty due to the large dataset used to define the best linear fit. The results of Hammarstrom and Zen's (1986) empirical investigation are similar (Appendix 4.2), but are not used because they have a larger uncertainty. Pressures determined using the experimental calibrations of Johnson and Rutherford (1989) and Schmidt (1991) are also presented in Appendix 4.2 for comparison, but pressures calculated using the calibration by Thomas and Ernst (1990) have not been included in Appendix 4.2 because the rocks experienced pressures lower than those used to derive this calibration.

Pressures determined from compositions of hornblende crystal rims from plutons of the Tagish Lake suite calculated using the calibration of Hammarstrom and Zen (1986) are from 5.6 to 7.2 kbar (excluding uncertainty; Table 4.3), indicating that these plutons were intruded at mid-crustal levels. Although the penetrative deformation in sample H (Plate 4.7a) renders the results from this sample suspect, the pressures determined for this sample are considered reasonable because they are consistent with those derived from other Tagish Lake suite plutons. Pressures derived from hornblende cores are expected to be slightly higher than those determined from rim compositions because the ascent of magma during intrusion is assumed to be linked to a concomitant decrease in pressure. Alternatively, a decrease in pressure could also reflect rapid denudation coeval with crystallization. Lower rim pressures were determined for one sample of the Hale Mountain granodiorite (sample G) and the mid-Cretaceous granodiorite (sample K), as expected, but all of the other samples have

core pressures that are lower than the rim pressures (Table 4.3), suggesting that either the resolution of the barometer is not sufficient to distinguish between small changes in pressure, as indicated by the prescribed uncertainty of ± 1 kbar for the barometer (Hollister et al., 1987), or pressure increased during hornblende crystallization. A possible explanations for an increase in pressure within an arc environment are loading caused by the growth of a volcanic edifice or by magma emplacement (Brown and Walker, 1993).

The difference between the pressures determined for the Hale Mountain granodiorite sample from White Moose Mountain (G; 7.2 kbar) and those for the sample from west of Hoboe Creek (H; 6.5 kbar; Table 4.3), excluding uncertainty, is less than the resolution of this geobarometer (± 1 kbar), but if it is real, it could be interpreted as indicating that these samples formed at different structural levels within the same pluton. However, it should be noted that the Hale Mountain granodiorite pluton exposed on White Moose Mountain is not contiguous with the Hale Mountain granodiorite exposed west of Hoboe Creek, and these two samples could be from two different, but lithologically identical plutons that intruded the Boundary Ranges suite at about the same stratigraphic level (Fig. 4.2). The only significantly different pressure determined for the Tagish Lake suite is for the youngest pluton, the Mt. Caplice quartz monzodiorite (J; 5.6 ± 2 kbar; Table 4.3), which returns the lowest pressure, suggesting that either this area was denuded during the latest Early Jurassic, or the Mt. Caplice quartz monzodiorite crystallized at a higher structural level than the other Tagish Lake suite plutons.

Al-in-hornblende analyses for the mid-Cretaceous granodiorite indicate that this pluton crystallized under much lower pressures than the Tagish Lake suite plutons (~1.8 kbar; Table 4.3), and that by ~100 Ma the adjacent metamorphic rocks were at depths less than 13 km (assuming a specific gravity of 2.65 g/cm³ for the rock column and an uncertainty of ± 2 kbar) and possibly less than 6 km (excluding uncertainty). Shallow crystallization of this pluton is in agreement with the rapid cooling inferred from the similar K-Ar cooling ages and U-Pb zircon crystallization age for this pluton discussed previously.

Appendix 4.2. Electron microprobe analyses of hornblende and pressures calculated for plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous pluton from the Tagish Lake area, based on the Al-in-hornblende barometer¹ (page 1 of 5).

SAMPLE LC-90-226 (Hale Mountain granodiorite)²
UTM 8V, 536680E 6593470N

	grain 1			grain 2			grain 3		
	core	rim	rim	core	rim	rim	core	rim	rim
SiO ₂	41.76	43.94	43.70	42.23	41.57	41.55	39.99	42.31	42.85
Al ₂ O ₃	12.08	11.85	10.82	11.06	12.64	12.78	13.11	12.33	11.43
TiO ₂	0.85	0.42	0.38	1.13	0.63	0.67	0.66	0.72	0.54
FeO	18.66	17.66	17.99	18.73	18.77	18.83	19.88	18.68	18.24
MnO	0.25	0.21	0.24	0.31	0.31	0.25	0.25	0.20	0.22
Cr ₂ O ₃	0.02	n.d. ³	n.d.	n.d.	0.05	n.d.	n.d.	n.d.	n.d.
MgO	9.34	10.04	10.54	9.73	9.62	9.50	10.20	9.80	9.70
CaO	11.81	11.58	11.98	11.80	11.83	11.79	10.28	11.83	11.96
K ₂ O	1.38	1.34	1.08	1.33	1.39	1.48	1.21	1.35	1.27
Na ₂ O	2.02	1.36	1.54	1.97	1.96	1.92	2.01	1.65	1.48
Cl	0.03	0.01	n.d.	0.05	0.02	0.03	0.01	n.d.	n.d.
TOTAL	98.20	98.41	98.27	98.34	98.79	98.80	97.60	98.87	97.69

Cations per 23 oxygen, Total Fe as FeO

Si	6.34	6.56	6.57	6.41	6.28	6.28	6.14	6.36	6.50
Al ^{IV}	1.66	1.44	1.43	1.59	1.72	1.72	1.86	1.64	1.50
Al ^{VI}	0.51	0.65	0.48	0.39	0.53	0.55	0.51	0.54	0.54
Ti	0.10	0.05	0.04	0.13	0.07	0.08	0.08	0.08	0.06
Fe	2.37	2.21	2.26	2.38	2.37	2.38	2.55	2.35	2.31
Mn	0.03	0.03	0.03	0.04	0.04	0.03	0.03	0.03	0.03
Cr	0.00	—	—	—	0.01	—	—	—	—
Mg	2.12	2.24	2.36	2.20	2.17	2.14	2.33	2.20	2.19
Ca	1.92	1.85	1.93	1.92	1.91	1.91	1.69	1.90	1.94
K	0.27	0.26	0.21	0.26	0.27	0.28	0.24	0.26	0.24
Na	0.59	0.40	0.45	0.58	0.57	0.56	0.60	0.48	0.44
Cl	0.01	—	—	0.01	—	0.01	—	—	—

Calculated pressures^{1,4}

1	7.0	6.6	5.7	6.0	7.4	7.5	8.0	7.1	6.4
2	7.4	7.0	6.0	6.4	7.9	8.1	8.6	7.6	6.8
3	5.7	5.4	4.6	4.9	6.1	6.2	6.6	5.8	5.2
4	7.3	6.9	6.1	6.4	7.7	7.8	8.3	7.4	6.7

Notes: ¹Calibrations for the Al in hornblende barometer are from: 1 - Hammarstrom and Zen (1986); 2 - Hollister et al. (1987); 3 - Johnson and Rutherford (1989); and 4 - Schmidt (1991).
²Sample "G" in Fig. 2. ³n.d.=not detected. ⁴Pressure is measured in kbar (1 kbar=100 MPa). All analyses are representative individual analyses.

Appendix 4.2 (continued). Electron microprobe analyses of hornblende and pressures calculated for plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous pluton from the Tagish Lake area, based on the Al-in-hornblende barometer¹ (page 2 of 5).

SAMPLE LC-90-272 (Hale Mountain granodiorite)²
UTM 8V, 549440E 6552800N

	grain 1			grain 2			grain 3		
	core	rim	rim	core	rim	rim	core	rim	rim
SiO ₂	44.01	44.01	43.90	44.06	44.34	43.46	43.31	43.79	44.39
Al ₂ O ₃	10.85	11.94	11.08	10.86	10.84	11.81	11.27	11.01	11.20
TiO ₂	0.95	0.74	0.84	1.21	0.69	0.80	0.94	0.91	0.70
FeO	16.77	16.12	16.02	16.24	15.68	16.63	16.21	16.35	17.02
MnO	0.30	0.32	0.33	0.40	0.31	0.32	0.25	0.32	0.30
Cr ₂ O ₃	0.02	n.d. ³	0.02	0.07	0.12	n.d.	0.02	0.05	n.d.
MgO	10.07	10.17	10.47	10.25	10.91	9.65	9.99	10.54	10.11
CaO	11.83	12.13	11.97	11.83	11.97	11.86	11.72	11.72	12.00
K ₂ O	1.15	1.19	1.11	1.12	1.00	1.16	1.05	1.10	1.05
Na ₂ O	1.34	1.28	1.53	1.43	1.30	1.26	1.44	1.48	1.38
Cl	0.04	0.02	0.09	0.10	0.03	0.02	n.d.	0.04	n.d.
TOTAL	97.33	97.92	97.36	97.57	97.19	96.97	96.20	97.31	98.15
Cations per 23 oxygen, Total Fe as FeO.									
Si	6.63	6.57	6.60	6.62	6.65	6.57	6.58	6.59	6.63
Al ^{IV}	1.37	1.43	1.40	1.38	1.35	1.43	1.42	1.41	1.37
Al ^{VI}	0.56	0.67	0.57	0.54	0.57	0.67	0.60	0.54	0.60
Ti	0.11	0.08	0.10	0.14	0.08	0.09	0.11	0.10	0.08
Fe	2.11	2.01	2.02	2.04	1.97	2.10	2.06	2.06	2.12
Mn	0.04	0.04	0.04	0.05	0.04	0.04	0.03	0.04	0.04
Cr	—	—	0.00	0.01	0.01	—	0.00	0.01	—
Mg	2.26	2.26	2.35	2.30	2.44	2.17	2.26	2.37	2.25
Ca	1.91	1.94	1.93	1.90	1.92	1.92	1.91	1.89	1.92
K	0.22	0.23	0.21	0.21	0.19	0.22	0.20	0.21	0.20
Na	0.39	0.37	0.45	0.42	0.38	0.37	0.42	0.43	0.40
Cl	0.01	—	0.02	0.02	0.01	—	—	0.01	—
Calculated pressures ^{1,4}									
1	5.8	6.6	6.0	5.8	5.7	6.7	6.2	5.9	6.0
2	6.1	7.1	6.3	6.1	6.0	7.1	6.6	6.2	6.4
3	4.7	5.4	4.8	4.7	4.6	5.4	5.1	4.8	4.9
4	6.2	7.0	6.3	6.1	6.1	7.0	6.6	6.3	6.4

Notes: ¹Calibrations for the Al in hornblende barometer are from: 1 - Hammarstrom and Zen (1986); 2 - Hollister et al. (1987); 3 - Johnson and Rutherford (1989); and 4 - Schmidt (1991). ²Sample "H" in Fig. 2. ³n.d.=not detected. ⁴Pressure is measured in kbar (1 kbar=100 MPa). All analyses are representative individual analyses.

Appendix 4.2 (continued). Electron microprobe analyses of hornblende and pressures calculated for plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous pluton from the Tagish Lake area, based on the Al-in-hornblende barometer¹ (page 3 of 5).

SAMPLE LC-90-314 (Hoboe quartz syenite)²
UTM 8v 547850E 6556325N

	grain 1			grain 2			grain 3		
	core	rim	rim	core	rim	rim	core	rim	rim
SiO ₂	41.86	42.02	42.88	42.04	41.93	42.45	41.97	41.56	41.14
Al ₂ O ₃	10.55	11.66	10.49	11.46	11.87	11.18	10.68	11.36	11.40
TiO ₂	1.29	0.80	0.86	0.74	0.74	0.80	1.36	0.64	1.07
FeO	19.36	19.54	18.80	19.29	19.10	19.15	19.06	19.28	19.34
MnO	0.38	0.34	0.40	0.33	0.33	0.30	0.35	0.25	0.34
Cr ₂ O ₃	n.d. ³	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	9.29	9.16	9.80	9.50	9.36	9.67	9.51	9.12	9.23
CaO	11.56	11.67	11.66	11.84	11.70	11.69	11.61	11.57	11.69
K ₂ O	1.25	1.28	1.12	1.19	1.26	1.24	1.26	1.19	1.28
Na ₂ O	1.38	1.76	1.72	1.50	1.68	1.48	1.75	1.63	1.51
Cl	n.d.	0.04	0.04	0.02	0.06	0.06	0.06	0.02	0.01
TOTAL	96.92	98.27	97.77	97.91	98.03	98.02	97.61	96.62	97.01

Cations per 23 oxygen, Total Fe as FeO.

Si	6.45	6.40	6.53	6.41	6.38	6.46	6.43	6.42	6.35
Al ^{IV}	1.55	1.60	1.47	1.59	1.62	1.54	1.57	1.58	1.65
Al ^{VI}	0.37	0.49	0.41	0.47	0.51	0.46	0.36	0.49	0.42
Ti	0.15	0.09	0.10	0.08	0.08	0.09	0.16	0.07	0.12
Fe	2.50	2.49	2.39	2.46	2.43	2.44	2.44	2.49	2.50
Mn	0.05	0.04	0.05	0.04	0.04	0.04	0.04	0.03	0.04
Cr	—	—	—	—	—	—	—	—	—
Mg	2.14	2.08	2.22	2.16	2.12	2.19	2.17	2.10	2.12
Ca	1.91	1.90	1.90	1.93	1.91	1.90	1.90	1.92	1.93
K	0.24	0.25	0.22	0.23	0.25	0.24	0.25	0.24	0.25
Na	0.41	0.52	0.51	0.44	0.50	0.44	0.52	0.49	0.45
Cl	—	0.01	0.01	0.01	0.02	0.02	0.02	0.01	—

Calculated pressures^{1,4}

1	5.7	6.6	5.6	6.4	6.8	6.2	5.8	6.5	6.5
2	6.0	7.0	5.8	6.8	7.2	6.5	6.1	6.9	6.9
3	4.6	5.4	4.5	5.2	5.6	5.0	4.7	5.3	5.3
4	6.1	6.9	6.0	6.8	7.1	6.5	6.2	6.8	6.9

Notes: ¹Calibrations for the Al in hornblende barometer are from: 1 - Hammarstrom and Zen (1986); 2 - Hollister et al. (1987); 3 - Johnson and Rutherford (1989); and 4 - Schmidt (1991).
²Sample "I" in Fig. 2. ³n.d.=not detected. ⁴Pressure is measured in kbar (1 kbar=100 MPa). All analyses are representative individual analyses.

Appendix 4.2 (continued). Electron microprobe analyses of hornblende and pressures calculated for plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous pluton from the Tagish Lake area, based on the Al-in-hornblende barometer¹ (page 4 of 5).

SAMPLE LC-90-343 (Mt. Caplice quartz monzodiorite)²
UTM 8v 564825E 6557750N.

	grain 1			grain 2			grain 3		
	core	rim	rim	core	rim	rim	core	rim	rim
SiO ₂	43.70	41.84	43.47	43.82	44.02	43.64	43.91	42.82	43.01
Al ₂ O ₃	9.58	11.25	9.96	9.63	9.73	9.78	9.11	10.67	10.65
TiO ₂	1.25	0.91	1.08	1.31	1.11	1.02	1.16	1.03	1.07
FeO	18.84	19.91	18.52	18.79	18.96	19.19	18.64	19.24	19.20
MnO	0.26	0.29	0.26	0.27	0.25	0.28	0.29	0.27	0.33
Cr ₂ O ₃	n.d. ³	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	10.84	9.38	10.48	10.64	10.36	10.28	10.74	10.22	10.08
CaO	11.63	11.62	11.51	11.63	11.87	11.70	11.32	11.72	11.96
K ₂ O	1.06	1.19	1.09	1.09	1.07	1.07	0.90	1.24	1.25
Na ₂ O	1.57	1.52	1.46	1.96	1.65	1.63	1.61	1.39	1.69
Cl	0.16	0.04	0.08	0.13	0.09	0.13	0.15	0.02	0.05
TOTAL	98.89	97.95	97.91	99.27	99.11	98.72	97.83	98.62	99.29

Cations per 23 oxygen, Total Fe as FeO.

Si	6.58	6.40	6.58	6.57	6.60	6.59	6.66	6.47	6.46
Al ^{IV}	1.42	1.60	1.42	1.43	1.40	1.41	1.34	1.53	1.54
Al ^{VI}	0.27	0.42	0.36	0.27	0.32	0.33	0.29	0.37	0.35
Ti	0.14	0.10	0.12	0.15	0.12	0.12	0.13	0.12	0.12
Fe	2.37	2.55	2.35	2.36	2.38	2.42	2.36	2.43	2.41
Mn	0.03	0.04	0.03	0.03	0.03	0.04	0.04	0.03	0.04
Cr	—	—	—	—	—	—	—	—	—
Mg	2.43	2.14	2.36	2.38	2.32	2.31	2.43	2.30	2.26
Ca	1.88	1.90	1.87	1.87	1.91	1.89	1.84	1.90	1.93
K	0.20	0.23	0.21	0.21	0.20	0.21	0.17	0.24	0.24
Na	0.46	0.45	0.43	0.57	0.48	0.48	0.47	0.41	0.49
Cl	0.04	0.01	0.02	0.03	0.02	0.03	0.04	—	0.01

Calculated pressures^{1,4}

1	4.6	6.3	5.0	4.7	4.7	4.8	4.3	5.6	5.6
2	4.8	6.7	5.3	4.8	4.9	5.0	4.4	6.0	5.9
3	3.7	5.1	4.1	3.7	3.8	3.9	3.4	4.6	4.5
4	5.1	6.6	5.4	5.1	5.2	5.3	4.7	6.0	6.0

Notes: ¹Calibrations for the Al in hornblende barometer are from: 1 - Hammarstrom and Zen (1986); 2 - Hollister et al. (1987); 3 - Johnson and Rutherford (1989); and 4 - Schmidt (1991). ²Sample "J" in Fig. 2. ³n.d.=not detected. ⁴Pressure is measured in kbar (1 kbar=100 MPa). All analyses are representative individual analyses.

Appendix 4.2 (continued). Electron microprobe analyses of hornblende and pressures calculated for plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous pluton from the Tagish Lake area, based on the Al-in-hornblende barometer¹ (page 5 of 5).

SAMPLE LC-90-359 (mid-Cretaceous pluton)²
UTM 8v, 544950E 6564725N

	grain 1			grain 2		
	core	rim	rim	core	rim	rim
SiO ₂	46.90	47.05	47.59	47.53	48.15	47.72
Al ₂ O ₃	7.79	7.33	7.07	6.98	6.11	6.48
TiO ₂	1.73	1.54	1.41	1.53	1.10	0.93
FeO	12.24	12.74	13.16	13.03	13.59	13.76
MnO	0.37	0.41	0.42	0.46	0.43	0.40
Cr ₂ O ₃	0.02	0.02	0.02	n.d. ³	0.01	n.d.
MgO	14.67	14.32	14.44	14.35	14.16	14.11
CaO	11.68	11.51	11.95	11.81	11.97	12.08
K ₂ O	0.74	0.61	0.66	0.58	0.56	0.62
Na ₂ O	1.69	1.45	1.41	1.29	1.06	0.88
Cl	0.07	0.08	0.07	0.05	n.d.	0.01
TOTAL	97.90	97.06	98.20	97.61	97.14	96.99

Cations per 23 oxygen, Total Fe as FeO.

Si	6.86	6.94	6.96	6.98	7.11	7.06
Al ^{IV}	1.14	1.06	1.04	1.02	0.89	0.94
Al ^{VI}	0.21	0.22	0.18	0.19	0.17	0.20
Ti	0.19	0.17	0.16	0.17	0.12	0.10
Fe	1.50	1.57	1.61	1.60	1.68	1.70
Mn	0.05	0.05	0.05	0.06	0.05	0.05
Cr	—	0.00	0.00	—	—	—
Mg	3.20	3.15	3.15	3.14	3.12	3.11
Ca	1.83	1.82	1.87	1.86	1.89	1.92
K	0.14	0.12	0.12	0.11	0.10	0.12
Na	0.48	0.42	0.40	0.37	0.30	0.25
Cl	0.02	0.02	0.02	0.01	—	0.00

Calculated pressures^{1,4}

1	2.8	2.5	2.2	2.2	1.4	1.8
2	2.8	2.4	2.1	2.1	1.2	1.6
3	2.2	1.9	1.7	1.6	1.0	1.3
4	3.4	3.1	2.8	2.7	2.0	2.4

Notes: ¹Calibrations for the Al in hornblende barometer are from: 1 - Hammarstrom and Zen (1986); 2 - Hollister et al. (1987); 3 - Johnson and Rutherford (1989); and 4 - Schmidt (1991).
²Sample "K" in Fig. 2. ³n.d.=not detected. ⁴Pressure is measured in kbar (1 kbar=100 MPa). All analyses are representative individual analyses.

Appendix 4.3. Electron microprobe analyses of plagioclase from plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous granodiorite from the Tagish Lake area (page 1 of 3).

SAMPLE LC-90-226¹
(Hale Mountain granodiorite)

SAMPLE LC-90-272²
(Hale Mountain granodiorite)

	<u>grain 1</u>	<u>grain 2</u>	<u>grain 3</u>	<u>grain 4</u>	<u>grain 1</u>	<u>grain 2</u>		
			(rim)	(rim)				
SiO ₂	68.82	67.07	61.41	61.55	58.25	57.85	57.67	57.90
Al ₂ O ₃	19.76	20.49	23.96	23.94	25.03	25.15	24.91	24.56
CaO	0.35	0.34	5.68	5.58	7.35	7.45	7.55	7.32
Na ₂ O	11.18	10.44	7.63	7.75	7.17	7.29	7.37	7.33
K ₂ O	0.14	0.85	0.38	0.32	0.63	0.40	0.43	0.60
FeO	<u>0.18</u>	<u>0.16</u>	<u>0.11</u>	<u>0.11</u>	<u>0.07</u>	<u>0.14</u>	<u>0.14</u>	<u>0.01</u>
TOTAL	100.43	99.35	99.17	99.25	98.50	98.28	98.07	97.72
<u>Cations per 16 oxygen.</u>								
Si	5.98	5.91	5.49	5.50	5.29	5.27	5.27	5.31
Al	2.02	2.13	2.52	2.52	2.68	2.70	2.68	2.65
Ca	0.08	0.03	0.54	0.53	0.72	0.73	0.74	0.72
Na	1.80	1.78	1.32	1.34	1.26	1.29	1.31	1.30
K	0.02	0.10	0.04	0.04	0.07	0.05	0.05	0.07
Fe	0.01	0.01	0.01	0.01	—	0.01	0.01	—
An ³	2	2	29	28	36	36	36	36

Notes: ¹Sample "G" in Fig. 2. ²Sample "H" in Fig. 2. ³An = % Anorthite content. All analyses are representative individual analyses.

Appendix 4.3 (continued). Electron microprobe analyses of plagioclase from plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous granodiorite from the Tagish Lake area (page 2 of 3).

SAMPLE LC-90-314 ¹ (Hoboe quartz syenite) UTM 8v 547850E 6556325N				SAMPLE LC-90-343 ² (Mt. Caplice quartz monzodiorite) UTM 8v 564825E 6557750N			
	grain 1	grain 2	grain 3	grain 1		grain 2	
SiO ₂	61.15	59.83	61.62	61.12	62.06	62.30	61.96
Al ₂ O ₃	25.13	25.17	24.43	23.05	23.36	23.50	23.49
CaO	5.62	6.20	4.92	4.98	5.05	5.12	5.26
Na ₂ O	8.33	7.73	8.45	8.24	8.44	8.44	8.33
K ₂ O	0.42	0.41	0.44	0.81	0.52	0.48	0.54
FeO	0.10	0.07	0.10	0.19	0.08	0.13	0.06
TOTAL	100.75	99.41	99.96	98.39	99.51	99.97	99.64
<u>Cations per 16 oxygen.</u>							
Si	5.40	5.36	5.47	5.52	5.54	5.53	5.521
Al	2.62	2.66	2.56	2.46	2.46	2.46	2.467
Ca	0.53	0.60	0.47	0.48	0.48	0.49	0.503
Na	1.43	1.34	1.45	1.44	1.46	1.45	1.439
K	0.05	0.05	0.05	0.09	0.06	0.05	0.062
Fe	0.01	—	0.01	0.01	0.01	0.01	0.004
An ³	27	31	24	25	25	25	26

Notes: ¹Sample "I" in Fig. 2. ²Sample "J" in Fig. 2. ³An = % Anorthite content. All analyses are representative individual analyses.

Appendix 4.3 (continued). Electron microprobe analyses of plagioclase from plutons of the Early Jurassic Tagish Lake suite and a mid-Cretaceous granodiorite from the Tagish Lake area (page 3 of 3).

SAMPLE LC-90-359¹
(mid-Cretaceous granodiorite)
UTM 8v, 544950E 6564725N

	grain 1		grain 2	
SiO ₂	59.64	59.55	62.24	61.70
Al ₂ O ₃	26.09	26.16	24.88	24.96
CaO	8.11	8.19	6.35	6.48
Na ₂ O	7.28	6.88	8.29	7.89
K ₂ O	0.57	0.52	0.38	0.51
<u>FeO</u>	<u>0.32</u>	<u>0.29</u>	<u>0.13</u>	<u>0.15</u>
TOTAL	102.01	101.59	102.27	101.69
<u>Cations per 16 oxygen.</u>				
Si	5.25	5.25	5.42	5.41
Al	2.71	2.72	2.55	2.58
Ca	0.76	0.77	0.59	0.61
Na	1.24	1.18	1.40	1.34
K	0.06	0.06	0.04	0.06
Fe	0.02	0.02	0.01	0.01
An ²	38	40	30	31

Notes: ¹Sample "K" in Fig. 2. ²An = % Anorthite content. All analyses are representative individual analyses.

APPENDIX 4.4: GARNET-BIOTITE FE-MG GEOTHERMOMETRY OF SEMI-PELITIC ROCKS OF THE FLORENCE RANGE SUITE

Although most Florence Range suite samples lack mineral assemblages in apparent textural equilibrium and appropriate for geothermometry, two pelite samples appeared to be reasonable candidates for garnet-biotite Fe-Mg exchange geothermometry. One is from the core of the D_{2T} synform, and the other is from the Mt. Caplice area (locations are shown in Fig. 4.2, and sample descriptions are presented in Table 4.1). Because the biotite in these samples is not deformed (Plate 4.14c), it was assumed that the minerals in these samples formed during M_{3T} and retain compositions that reflect the temperature associated with M_{3T} .

Florence Range suite sample from the core of the D_{2T} synform (Sample LC-89-148)

The Florence Range suite sample from the core of the D_{2T} synform is a lineated garnet-biotite semipelite (Sample L in Fig. 4.2). The biotite grains are polygonalized and appear to be in textural equilibrium with garnet. They are randomly oriented in the plane perpendicular to the lineation, and appear to be controlled by a mineral-elongation lineation defined by aligned quartz and feldspar grains in the planes that parallel the lineation. Some biotite grains have been partially altered to chlorite. Garnet is poikiloblastic and may be fractured.

Florence Range suite sample from the Mt. Caplice area (Sample LC-89-315)

The sample from the Mt. Caplice area is from the eastern limb of the Hoboe antiform (Sample M in Fig. 4.2). It is a foliated garnet-sillimanite-muscovite-biotite semipelitic rock (Plate 4.10b). Garnet is wrapped by the foliation, which is defined by the preferred orientation of biotite, muscovite and recrystallized quartz ribbons. There are two generations of biotite. The first parallels the foliation, and the second is randomly oriented, and commonly occurs adjacent to garnet, or in pressure shadows around the garnet. Sillimanite is associated with biotite and also occurs in the garnet pressure shadows. The second generation biotite appears to be in textural equilibrium with the garnet and is interpreted as forming during M_{3T} , as is the sillimanite. Plagioclase has been partially replaced by fine-grained white mica.

Analytical Methods

Microprobe analyses were obtained by energy dispersive spectrometry (EDS) using a Cambridge Microscan 5 instrument at Carleton University, operated at 20 kV accelerating potential, with a specimen current of 9.0 nanoamperes. Microprobe analyses of garnet and biotite are presented in Appendices 4.4 and 4.5, respectively. The cores and rims of the garnets, biotite grains adjacent to the garnet rims, and matrix biotite were analysed. Only biotite grains that were not associated with chlorite were analysed. Totals for biotite are <100% because H_2O , Cl and F were not analysed;

it is assumed that the difference between the calculated totals and 100% is equal to the analytical uncertainty + the H₂O + Cl + F content.

Results of Geothermometry

Of the numerous calibrations for the garnet-biotite Fe-Mg exchange geothermometer, both empirical and experimental, the calibration by Ferry and Spear (1978) is used here because the composition of the garnets in this sample is within the recommended limits (Table 4.4) and corrections for non-ideal mixing of Ca (e.g. Hodges and Spear, 1978) were not considered necessary. Because the garnets have relatively high Mn contents, the calibration of Ganguly and Saxena (1984) has also been applied because it corrects for non-ideal mixing of Mn and Fe-Mg solid solution, as well as non-ideal mixing of Ca (Table 4.4). In addition, temperature was estimated using the PTAXSS computer program (Berman et al., 1987; Brown et al., 1988), which uses an internally consistent thermodynamic data set and also accounts for non-ideal mixing using the calibrations of Berman (1988; 1990) for garnet and of McMullin et al. (1991) for biotite. The calibration by Perchuk and Lavrent'eva (1983) was not used because it was experimentally calibrated using a system with Fe/(Fe+Mg) held at 0.6, whereas garnets in this sample have Fe/(Fe+Mg) ratios of ~0.8 (Appendix 4.5).

Florence Range suite sample from the core of the D_{2T} synform
(Sample LC-89-148)

With the exception of titanium, the composition of biotite in the matrix is similar that of biotite grains adjacent to garnet crystals (Appendix 4.4). Analyses of the rims and cores of four garnet grains (Appendix 4.5) indicate that two garnet grains are normally zoned, with Ca- and Mn-rich cores and Fe- and Mg-rich rims; suggesting the garnets grew during conditions of increasing pressure and/or temperature (e.g. Tracy et al., 1976); the lack of normal zoning in the other two garnets may reflect re-equilibration, homogenization or rapid growth of the garnet rims.

The normally zoned garnet crystals in Sample L return the most consistent results, but all analyses are well within the prescribed uncertainty limits. Assuming the garnet and biotite were in equilibrium at pressures of ~6 kbar, and on the basis of an inherent uncertainty of $\pm 50^\circ\text{C}$, the calibrations applied indicate that this sample experienced temperatures of $\sim 797 \pm 76^\circ\text{C}$ (calibration of Ferry and Spear, 1978), $777 \pm 87^\circ\text{C}$ (calibration of Ganguly and Saxena, 1984), or $841 \pm 73^\circ\text{C}$ (determined using the PTAXSS program of Berman et al., 1987; Table 4.4).

Comparison with the stable mineral assemblages observed in the Florence Range suite, which indicate they were on the low temperature side of the muscovite-out isograd (ie. they contain coexisting muscovite and quartz and lack metamorphic potassium feldspar), indicating that the results for Sample L are at higher than the published temperatures for the muscovite-out isograd at 6 kbar (e.g. $\sim 700^\circ\text{C}$; Spear and Cheney, 1989), and are therefore unrealistically high. This implies that at least one

of the assumptions required by the geothermometer were not met; the garnet may not have grown in equilibrium with the biotite, and/or the biotite composition may have been altered, as indicated by the presence of chlorite.

Florence Range suite sample from the Mt. Caplice area (Sample LC-89-315)

Both biotite and garnet in Sample M are generally uniform in composition, and garnets in are not zoned. They return temperatures that are lower than those determined for Sample L. Assuming pressures of ~6 kbar, the calibrations applied indicate that this sample experienced temperatures of $\sim 655 \pm 92^\circ\text{C}$ (calibration of Ferry and Spear, 1978), $672 \pm 70^\circ\text{C}$ (calibration of Ganguly and Saxena, 1984), and $620 \pm 70^\circ\text{C}$ (determined using the PTAXSS program of Berman et al., 1987; Table 4.4). These temperatures are in agreement with the temperature conditions inferred from the M_{3T} stable mineral assemblage and are considered reasonable.

Appendix 4.5. Electron microprobe analyses of garnet from schist of the Florence Range suite (page 1 of 2).

SAMPLE LC-89-148¹ (UTM 8V, 536750E 6570100N).

	grain 1		grain 2		grain 3		grain 4	
	(rim)	(core)	(rim)	(core)	(rim)	(core)	(rim)	(core)
SiO ₂	37.42	37.85	37.52	37.64	36.99	36.82	37.46	37.14
Al ₂ O ₃	22.26	22.46	22.39	21.98	21.44	21.05	21.41	21.00
TiO ₂	n.d. ²	n.d.	n.d.	n.d.	n.d.	0.04	n.d.	0.09
FeO	31.08	32.16	31.54	32.09	30.35	27.96	31.34	28.36
MgO	4.09	4.08	4.11	4.27	3.65	2.60	3.92	3.34
MnO	1.88	1.96	2.33	2.07	3.54	6.07	2.69	5.10
CaO	4.06	2.96	3.28	3.00	4.15	5.12	3.67	4.63
TOTAL	100.79	101.47	101.17	101.05	100.12	99.66	100.49	99.66
Cations per 24 oxygen. Total Fe as FeO.								
Si	5.90	5.93	5.90	5.93	5.91	5.94	5.95	5.96
Al ^{IV}	0.10	0.07	0.10	0.07	0.09	0.06	0.05	0.04
Al ^{VI}	4.04	4.08	4.05	4.01	3.95	3.94	3.96	3.93
Ti	—	—	—	—	—	0.01	—	0.01
Fe	4.10	4.21	4.15	4.23	4.06	3.77	4.16	3.81
Mg	0.96	0.95	0.96	1.00	0.87	0.62	0.93	0.80
Mn	0.25	0.26	0.31	0.28	0.48	0.83	0.36	0.69
Ca	0.69	0.50	0.55	0.51	0.71	0.88	0.62	0.80
Alm ³	68	71	69	70	66	62	69	62
Pyp	16	16	16	17	14	10	15	13
Spes	04	04	05	05	08	14	06	11
Gros	11	08	09	08	12	14	10	13

Notes: ¹Sample "L" in Fig. 2. See Table 4.1 for sample description. ²n.d.=not detected.

³Alm=Fe/Fe+Ca+Mg+Mn in garnet in %. All analyses are representative individual analyses.

Appendix 4.5 (continued). Electron microprobe analyses of garnet from schist of the Florence Range suite (page 2 of 2).

SAMPLE LC-89-315¹ (UTM 8V, 546425E 6555900N).

	grain 1		grain 2		grain 3		grain 4	
	(rim)	(core)	(rim)	(core)	(rim)	(core)	(rim)	(core)
SiO ₂	37.50	37.36	37.83	37.78	37.50	37.44	38.02	38.18
Al ₂ O ₃	21.13	21.03	21.28	21.04	21.38	21.06	21.59	21.49
TiO ₂	n.d.	0.01	n.d.	0.02	0.02	n.d.	n.d.	0.01
Cr ₂ O ₃	n.d.	0.01	n.d.	0.03	n.d.	n.d.	n.d.	0.04
FeO	24.56	26.43	26.51	24.99	25.37	26.26	28.63	28.47
MgO	3.09	3.56	3.64	3.22	3.33	3.47	4.41	4.40
MnO	11.89	10.00	8.81	11.77	11.15	9.70	6.17	6.58
CaO	1.78	1.72	2.33	1.88	1.93	1.90	1.99	2.09
TOTAL	99.95	100.12	100.40	100.73	100.68	99.83	100.81	101.26
Cations per 24 oxygen. Total Fe as FeO.								
Si	6.02	5.99	6.02	6.02	5.98	6.01	6.00	6.00
Al ^{IV}	—	0.01	—	—	0.02	—	—	—
Al ^{VI}	4.00	3.97	3.99	3.96	4.00	3.98	4.02	3.98
Ti	—	—	—	—	—	—	—	—
Cr	—	—	—	—	—	—	—	—
Fe	3.30	3.54	3.53	3.33	3.38	3.53	3.78	3.74
Mg	0.74	0.85	0.86	0.76	0.79	0.83	1.04	1.03
Mn	1.62	1.36	1.19	1.59	1.51	1.32	0.82	0.88
Ca	0.31	0.30	0.40	0.32	0.33	0.33	0.34	0.35
Alm ³	0.55	0.59	0.59	0.55	0.56	0.59	0.63	0.62
Pyp	0.12	0.14	0.14	0.13	0.13	0.14	0.17	0.17
Spes	0.27	0.22	0.20	0.26	0.25	0.22	0.14	0.15
Gros	0.05	0.05	0.07	0.05	0.05	0.05	0.06	0.06

Notes: ¹Sample "M" in Fig. 2. See Table 4.1 for sample description. ²n.d.=not detected.
³Alm=Fe/(Fe+Ca+Mg+Mn) in garnet in %. All analyses are representative individual analyses.

Appendix 4.6. Electron microprobe analyses of biotite from schist of the Florence Range suite (page 1 of 2).

SAMPLE LC-89-148¹ (UTM 8V, 536750E 6570100N).

	<u>matrix</u>	<u>grain 1</u>	<u>grain 2</u>	<u>grain 3</u>	<u>grain 4</u>
SiO ₂	35.53	34.75	34.77	34.43	34.54
Al ₂ O ₃	19.11	19.96	19.29	19.79	19.63
TiO ₂	2.64	1.78	2.00	2.05	2.60
FeO	19.88	20.25	19.83	20.64	20.42
MgO	8.89	9.15	8.67	8.46	8.39
MnO	0.11	0.06	0.06	0.13	0.04
CaO	n.d. ²	n.d.	n.d.	0.01	n.d.
Na ₂ O	n.d.	n.d.	n.d.	0.28	0.35
K ₂ O	<u>9.47</u>	<u>9.44</u>	<u>9.40</u>	<u>9.02</u>	<u>9.18</u>
TOTAL	95.63	95.39	94.02	94.81	95.15
<u>Cations per 23 oxygen. Total Fe as FeO.</u>					
Si	5.40	5.30	5.38	5.30	5.30
Al ^{IV}	2.60	2.70	2.62	2.70	2.70
Al ^{VI}	0.82	0.90	0.90	0.89	0.84
Ti	0.30	0.20	0.23	0.24	0.30
Fe	2.53	2.58	2.57	2.66	2.62
Mg	2.01	2.08	2.00	1.94	1.92
Mn	0.01	0.01	0.01	0.02	0.01
Ca	—	—	—	—	—
Na	—	—	—	0.08	0.10
K	1.84	1.84	1.86	1.77	1.80

Notes: ¹Sample "L" in Fig. 2. See Table 4.1 for sample description. ²n.d.=not detected.
All analyses are representative individual analyses.

Appendix 4.6 (continued). Electron microprobe analyses of biotite from schist of the Florence Range suite (page 2 of 2).

SAMPLE LC-89-315¹ (UTM 8V, 546425E 6555900N).

	<u>matrix</u>	<u>grain 1</u>	<u>grain 2</u>	<u>grain 3</u>	<u>grain 4</u>
SiO ₂	35.84	36.25	35.43	35.84	36.16
Al ₂ O ₃	21.10	21.07	21.00	20.65	21.65
TiO ₂	3.49	3.27	2.85	2.92	1.20
Cr ₂ O ₃	0.06	n.d. ²	0.01	0.02	n.d.
FeO	16.31	16.68	17.16	16.97	17.18
MgO	10.34	10.80	10.60	10.13	11.54
MnO	0.38	0.40	0.39	0.38	0.23
CaO	0.03	0.09	0.07	0.11	0.07
Na ₂ O	0.21	n.d.	0.20	n.d.	n.d.
K ₂ O	8.67	8.60	8.49	8.47	8.41
Cl	n.d.	n.d.	0.02	n.d.	n.d.
TOTAL	96.43	97.16	96.22	95.49	96.44

Cations per 23 oxygen. Total Fe as FeO.

Si	5.28	5.29	5.25	5.34	5.32
Al ^{IV}	2.72	2.71	2.75	2.66	2.68
Al ^{VI}	0.94	0.92	0.92	0.96	1.07
Ti	0.39	0.36	0.32	0.33	0.13
Cr	0.01	—	—	—	—
Fe	2.01	2.04	2.13	2.11	2.11
Mg	2.27	2.35	2.34	2.25	2.53
Mn	0.05	0.05	0.05	0.05	0.03
Ca	—	0.01	0.01	0.02	0.01
Na	0.06	—	0.06	—	—
K	1.63	1.60	1.60	1.61	1.58

Notes: ¹Sample "M" in Fig. 2. See Table 4.1 for sample description. ²n.d.=not detected.
All analyses are representative individual analyses.

Appendix 4.7

Geometric construction for the up-plunge projection of D_{2T} folds in the Florence Range suite exposed west of the southern Wann River headwaters. Projection is based on a mean lineation trend of 126° and plunge of 24° (Fig. 4.6). Interpretation of relative ages of structures is shown in Fig. 4.5 b.

Figure 4.1. Terrane map of northwest British Columbia, southwest Yukon and southeast Alaska (modified from Wheeler and McFeely, 1991 and Wheeler et al., 1991). KSF = King Salmon Fault; LFZ = Llewellyn fault zone. Inset: AK = Alaska; BC = British Columbia; YT = Yukon Territory.

Figure 4.2. Geologic map of the Tagish Lake area (modified from Mihalynuk et al, 1989a; 1990a; Currie 1990a; 1991; 1992a), showing the locations of U-Pb geochronology samples (A to F; Tables 4.5; 4.6; Fig. 4.10), Al-in-hornblende geobarometry samples (G to K; Table 4.2), garnet-biotite geothermometry samples (L and M; Table 4.4), and cross sections shown in Figures 4.5 and 4.7 (U to Z). CCF = Chicken Creek fault; CF = Caplice Fault; STTB = Stikinia-Tracy Arm terrane boundary.

Figure 4.3. Structural map of the Tagish Lake area showing the approximate extent of the Wann River shear zone and the locations of observed shear sense indicators, and equal area stereonet projections of stretching lineations (L_{WRSZ}) for metamorphic rocks from the Tagish Lake area. Shaded areas represent metamorphic rocks, patterned areas represent unmetamorphosed rocks ($\leq$ zeolite facies); patterns and structural symbols correspond to those used in Figure 4.2. The orientation of the axial trace of the Willison synform has been determined from the intersection of the mean foliations from the eastern and northern limbs of the synform and the map-pattern. The diagonally lined area has not been mapped in detail and lacks structural data. CF = Caplice fault; CCF = Chicken Creek fault; DS - Dolphin synform; HA - Hoboe antiform; STTB = Stikinia-Tracy Arm terrane boundary; WCFZ - Willison Creek fault zone; WS- Willison synform; WF - Wann fault.

Figure 4.4. Geology map of the southern Tagish Lake area showing marble layers in the Florence Range suite that outline medium-scale structures. CF = Caplice fault; DS - Dolphin synform; HA - Hoboe antiform; WCFZ - Willison Creek fault zone; WS - Willison synform; WF - Wann fault.

Figure 4.4. Geologic map of the southern Tagish Lake area. CF = Caplice fault. WCFZ = Willison Creek fault zone. WF = Wann fault.

Figure 4.5. Cross-sections that cross the western limb of the Willison synform: (a) cross-section from south of the Wann River to north of White Moose Mountain (U-U'; note imbrication of Tracy Arm terrane-Stikinia contact; STTB = Stikinia-Tracy Arm terrane boundary); (b) the western Wann River headwaters (V-V'; Plates 4.11d; 4.12a); (c) downplunge projection of the Wann thrust fault and D_{2T} folds (W-W'; see Fig. 4.6 for the geometric construction for this downplunge projection, and Plates 4.12c, d). Cross-section locations are shown in Figure 4.2. * denotes D_{2T} synform described in text. AMPH = amphibolite facies mineral assemblages; GS = greenschist facies mineral assemblages.

(a) South of Wann River to North of White Moose Mountain

(b) Wann River headwaters (west)

(c) Wann River headwaters (east)

CRETACEOUS AND TERTIARY IGNEOUS ROCKS

- Mt. Swiss volcanic rocks
- intrusive rocks

STIKINIA

- JURASSIC
 - Tagish Lake suite
- TRIASSIC
 - Stuhini Group volcanic rocks
- PALEOZOIC (?)
 - Boundary Ranges suite
 - (undivided)
- PERMIAN
 - Wann River gneiss

TRACY ARM TERRANE

- Age uncertain
- Florence Range suite
 - undivided
 - marble

- Geologic contact
- Fault
- Wann River shear zone
- transposition fabric (S_{WRSZ})
- S_{2T} ★
- D_{2T} synform described in text

Figure 4.6. Geologic map and up-plunge projection of D_{2T} and D_{3T} folds and the Wann fault, in the Florence Range suite exposed above the Wann River shear zone (see Plates 4.12c, d). Squares indicate the location of points used to constrain the contacts in this projection. The interpreted relative ages of structures are shown in Fig. 4.5c. Projection is based on a mean lineation trend of 126° and plunge of 24° . See Appendix 4.7 for the geometric construction of an up-plunge projection

Figure 4.7. Cross-sections showing medium-scale structures in the Florence Range suite: (a) cross-section from the south end of Atlin Lake, north of Willison Creek (W-W'); note that folds from cross-section W-W' have been projected into this cross-section); (b) cross-section from the Mt. Caplice area (Y-Y'); (c) cross-section from the Hoboe Glacier (Z-Z'). All cross-sections are the same scale and their locations are shown in Figure 4.2. All cross sections were constructed perpendicular to D_{2T} fold axes, with the exception of cross-section V-V', which was drawn at an angle of 70° to the fold axes so that it would incorporate as much geology as possible. STTB = Stikinia-Tracy Arm terrane boundary.

(a) Willison Creek

(b) Mt. Caplice

(c) Hoboe Glacier

- CRETACEOUS AND TERTIARY**
- plutonic rocks
 - Sloko Group volcanic rocks
- STIKINIA**
- EARLY TRIASSIC**
- Stuhini Group volcanic rocks
- EARLY JURASSIC**
- Tagish Lake suite
 - Hoboe quartz syenite
 - Hale Mountain granodiorite
 - Boundary Ranges metamorphic suite
- PERMIAN**
- Wann River gneiss
- PALEOZOIC AND YOUNGER (?)**
- undivided
 - leucogabbro
- TRACY ARM TERRANE**
- pre-Middle Jurassic
- marble
 - a
 - amphibolite
 - biotite schist
- Llewellyn fault zone**

Figure 4.8. Structural domain map for Florence Range suite rocks, with associated equal area stereonet projections of poles to S_{0+1T} (the predominant foliation + compositional layering; shown as circles) and D_{2T} fold hingeline lineations (squares). Open circles represent the mean poles to foliation and mean lineations. Grey areas on the location map represent different structural domains discussed in text. The diagonally lined area has not been mapped in detail and lacks structural data. CF = Caplice fault; CCF = Chicken Creek fault; STTB = Stikinia-Tracy Arm terrane boundary.

Cretaceous to Tertiary

intrusive rocks

TRACY ARM TERRANE

Florence Range suite

undivided

marble

calcsilicate

STIKINIA

Jurassic

Tagish Lake suite

Mt. Caplice quartz monzodiorite

Permian

Boundary Ranges suite

Wann River gneiss

89 } Foliation S_{0-1T} (dipint, vertical)

23 } Fold axis L_{2T}

Trace of Hoboe antiform

glacier

Figure 4.9. Geologic map of the Mt. Caplice area showing the Hoboe antiform (Plates 4.52; 4.53, and Willison synform (Plate 4.15). See cross section Y-Y' in Figure 4.7b.

Figure 4.10. U-Pb concordia diagrams for: (a) zircon from leucosome of the Florence Range suite shown in Plate 4.11b; inset shows lower intercepts derived from the regression of the four youngest fractions ($173 \pm 2.4/-2.9$ Ma) and all fractions (179 ± 4 Ma); (b) monazite from schist of the Florence Range suite (dashed ellipses have been corrected for excess ^{206}Pb); inset shows results calculated using initial Pb ratios determined from Pb isochron diagrams ($^{206}\text{Pb}/^{204}\text{Pb}=22.66 \pm 2.21$ and $^{207}\text{Pb}/^{204}\text{Pb}=16.39 \pm 0.42$; Table 4.5); (c) titanite and rutile fractions from the Florence Range suite (LC-89-86 and LC-91-469; C and F in Figure 4.1) and from the Wann River gneiss of the Boundary Ranges suite (LC-89-91b and LC-91-469; D and E in Figure 4.1); and (d) analyses for titanite and rutile from the Florence Range suite; inset shows analyses for rutile from the Wann River gneiss (LC-89-91b) for comparison. Analyses shown in Figures 4.10a, 4.10b, 4.10c and the inset in Figure 4.10d were calculated using initial Pb ratios from Stacey and Kramers (1975) Pb isotope growth curve for 180 Ma), whereas those in Figure 4.10d use initial Pb isotope ratios considered more suitable for an evolved continental margin - North America (see text for discussion). Data are presented in Table 4.5.

- Plate 4.1a. View east of an upper-member-to-the-south extensional shear in the Early Jurassic Hale Mountain granodiorite, from near the top of White Moose Mountain. This rock lies within the Wann River shear zone, below the contact between Stikinia and Tracy Arm terrane (labelled "STTB" in Figs. 4.2; 4.3; hammer head is 16 cm long).
- Plate 4.1b. The stretching lineation (L_{WRSZ}) exposed on the outcrop surface of a Florence Range suite marble from south of the Wann River plunges shallowly toward the south-southeast. It is defined by calcite rodding and lies in the plane of the Wann River shear zone transposition foliation (S_{WRSZ}). Both the L_{WRSZ} and S_{WRSZ} are interpreted as forming during displacement on the Wann River shear zone (pen is 14 cm long).
- Plate 4.1c. View west of characteristic straight gneisses of the Wann River shear zone exposed south of the Wann River. The contact between the biotite-bearing semipelitic rocks (B; Florence Range suite?) and the underlying hornblende-plagioclase gneiss of the Boundary Ranges suite (Wann River gneiss (W); Stikinia) is interpreted as the Tracy Arm Terrane-Stikinia contact.
- Plate 4.1d. Sheared contact between a Permian hornblende-plagioclase gneiss of the Boundary Ranges suite (Wann River gneiss; W) and the Early Jurassic Hale Mountain granodiorite of the Tagish Lake suite (H) exposed south of the Wann River. The foliation in both the Wann River gneiss and the Hale Mountain granodiorite parallels their imbricate contact and is interpreted as forming during displacement on the Wann River shear zone.
- Plate 4.1e. View south of truncated layering within the Wann River shear zone, on White Moose Mountain. This contact occurs in a zone of alternating layers of Hale Mountain granodiorite and Wann River gneiss, and reflects the imbricate nature of contacts within this zone.
- Plate 4.1f. View southwest of the contact between hornblende-plagioclase straight gneiss (W; Wann River gneiss; Stikinia) and overlying biotite semipelite (B; Florence Range suite?; Tracy Arm terrane) exposed south of the Wann River, that is interpreted as the Stikinia-Tracy Arm terrane boundary (STTB in Fig. 4.2). Localized shearing along this contact is attributed to deformation that post-dates annealing of the gneissic fabric in the adjacent rocks.

Plate 4.2a. View northeast across the Wann River of marble layers that are folded about the Willison synform (WS). The western limb dips moderately, approximately toward the south and the eastern limb is sub-vertical. D = diorite; M = marble; Q = quartz monzonite; R = rhyolitic felspar porphyry.

Plate 4.2b. View south of a steep, approximately north-south striking, west-side-up fault that offsets the Wann River shear zone. The fault juxtaposes hornblende-plagioclase gneiss, marble and felsic gneiss of the Boundary Ranges suite (Stikinia; west) against marble and biotite schist of the Florence Range suite that have been folded about the Dolphin synform on the east side of the fault (see Plate 4.2c and Figs. 4.3; 4.4).

Plate 4.2c. View southwest of marble and biotite schist layers in the Florence Range suite that outline the Dolphin synform (DS; Figs. 4.3; 4.4). It is not clear whether or not the Dolphin synform involves the Wann River shear zone or rocks belonging to Stikinia.

Plate 4.2d. View north of the upper décollement of Willison Creek Fault showing top-to-the-east sense of shear. The upper décollement juxtaposes hangingwall marble against graphitic quartzite and marble layers at the top of the fault zone. It dips 52° toward 270° , and is not lineated. The strike and dip of the steepest quartzite-marble contacts is 160/76 (hammer is 32 cm long).

Plate 4.3a. View northwest of Willison Creek Fault zone where it is exposed on the ridge between the Wann River and Willison Creek (cross-section X-X' in Fig. 4.7). U = upper décollement; L = lower décollement. Lithologic layers above and below the fault zone are reasonably coherent.

Plate 4.3b. Line drawing of Plate 4.3a.

Plate 4.4a. View south of the Willison Creek fault zone, where it is exposed at the toe of Willison Glacier. Faults within the fault zone truncate S_{0+1T} and mafic dykes. It is not clear whether or not the marble layers represent one layer that has been structurally repeated or if they represent more than one layer.

Plate 4.4b. Line drawing of Plate 4.4a.

Plate 4.4c. View southwest to the base of the Willison Creek fault zone (lower décollement; LD) exposed ~1 km south of Mt. Caplice (Figs. 4.4; 4.9). The lower décollement lies above the Hoboe antiform (Fig. 4.9).

Plate 4.4d. View east of a pegmatite dyke with top-to-the-south offsets exposed immediately above (west of) the upper décollement of the Willison Creek fault zone hangingwall, near the base of Willison Glacier (pen is 16.5 cm long).

Plate 4.5a. A xenolith of Florence Range suite pelite and the pegmatite dyke that intruded it are preserved in a quartz monzonite fault block in the Willison Creek fault zone (hammer is 32 cm long).

Plate 4.5b. Rotated garnets in a schist of the Florence Range suite are indicative of dextral shear (pen is 16.5 cm long).

Plate 4.5c. A Florence Range suite marble xenolith is preserved in a quartz monzonite fault block in the Willison Creek fault zone, and intruded by a mafic dyke. Approximately 0.5 m above the base of the outcrop, both the dyke and the quartz monzonite are truncated by a fault. Lineations on the surface of the outcrop are glacial stria.

Plate 4.5d. View south of a fault in the Willison Creek fault zone that is exposed on the ridge north of Willison Creek. The fault truncates compositional layering and a metamorphic foliation (S_{0+1T}). The fault is labelled "F" in Plate 4.14.

Plate 4.5e. View south of a marble block that contains a folded metamorphic foliation, which is truncated by a fault within the Willison Creek fault zone. D = mafic dyke, B = biotite schist, M = marble.

Plate 4.6a. Carbonate layer (C) within the Wann River gneiss (W) from the Wann River shear zone in the Wann River valley. The carbonate is thought to have been deposited at the same time as the metavolcanic Wann River gneiss. The lithologic contacts are interpreted as sheared and transposed bedding surfaces (hammer head is 16 cm long).

Plate 4.6b. Photomicrograph of the Wann River gneiss. The alignment of hornblende crystals, and flattened quartz and feldspar crystals within the transposition foliation (S_{WRSZ}) are interpreted as indications of deformation and syntectonic growth during displacement on the Wann River shear zone. With the exception of minor undulose extinction and embayments in quartz, the near absence of subsolidus deformation fabrics in the Wann River gneiss suggests that metamorphism outlasted displacement on the Wann River shear zone (field of view is 3 mm wide; Sample LC-89-91B).

Plate 4.6c. Low angle truncation of compositional layering in biotite semipelitic rocks of the Florence Range suite (B) along their contact with the Wann River gneiss (W) from west of Hoboe Glacier. This truncation is indicative of sinistral shear along the contact. N = north; S = south.

Plate 4.6d. View southwest of a sheared felsic dyke that is bounded by biotite schist of the Florence Range suite from within the Wann River shear zone, north of the Wann River. The mineral elongation lineation in the biotite schist plunges 27° toward 144° . Rotated blocks of the dyke are indicative of top-to-the-southeast displacement.

- Plate 4.7a. Photomicrograph of Hale Mountain granodiorite from the north side of Mt. Lawson (Sample G in Fig. 4.2) with medium-grained plagioclase augen (P), deformed hornblende (H), and recrystallized quartz ribbons (field of view is 3 mm wide; Sample LC-90-226).
- Plate 4.7b. Photomicrograph of a rotated plagioclase porphyroclast (P) in the Hale Mountain granodiorite that is associated with epidote (E) and wrapped by quartz ribbons (Q). Hornblende crystals (H) are deformed, and lie in the plane of the foliation (field of view is 3 mm wide; Sample LC-88-35-1).
- Plate 4.7c. C\S fabrics in a calc-silicate rock exposed on the eastern limb of the Hoboe antiform, near Mt. Caplice. (NE = northeast; SW = southwest; coin is 1.7 cm in diameter).
- Plate 4.7d. Folds with sinistral shear sense in a hornblende leucogabbro of the Boundary Ranges suite exposed west of Hoboe Glacier.

- Plate 4.8a. Contact between relatively undeformed Hale Mountain granodiorite (H) and folded actinolite schist of the Boundary Ranges suite (B) from the north side of Mt. Switzer (south of Tagish Lake). Although the Hale Mountain granodiorite is in general structurally higher than the actinolite schist of the Boundary Ranges suite, locally layers of the hornblende granodiorite are overlain by actinolite schist because along this contact 10 cm- to m-scale layers of actinolite schist alternate with layers of hornblende granodiorite over ~10 m.
- Plate 4.8b. Photomicrograph of Boundary Ranges actinolite schist with deformed hornblende (H) and plagioclase (P) porphyroclasts. The foliation is defined by the preferred orientation of actinolite and flattened plagioclase grains (field of view is 3 mm wide; Sample LC-89-14-1).
- Plate 4.8c. Refolded fold in the actinolite schist of the Boundary Ranges suite. S_{1S} is folded by a D_{2S} fold. Field of view is 34 cm wide.
- Plate 4.8d. Outcrop-scale refolded fold in a Boundary Ranges suite actinolite schist from the south shore of Tagish Lake. S_{1S} is folded by a close, upright D_{2S} fold.
- Plate 4.8e. Photomicrograph of andalusite (A) in a schist from the structurally deepest Boundary Ranges suite rocks exposed in the Tagish Lake area (from Mt. Lawson), parallels S_{0+1} defined by biotite (B) and muscovite (M). The micaceous minerals and the andalusite are kink-folded and exhibit undulose extinction (field of view is 3 mm wide; Sample LC-90-286d).

Plate 4.9a. Photomicrograph of staurolite (S) and folded kyanite (K) in a Boundary Ranges suite schist (field of view is 3 mm wide; Sample LC-90-286d).

Plate 4.9b. D_{1T} fold in the biotite schist and marble of the Florence Range suite. Axial planar micas define the predominant foliation in these rocks (S_{1T} ; pen is 16.5 cm long).

Plate 4.9c. Photomicrograph of folded andalusite (A) in schist of the Florence Range suite. In this thin section the muscovite (M) and biotite (B) that define S_{0+1T} have not been recrystallized, and like the andalusite, they exhibit undulose (wavy) extinction due to folding about D_{2T} or D_{3T} crenulations (field of view is 3 mm wide; Sample LC-91-483).

Plate 4.9d. Photomicrograph of polygonalized muscovite and biotite that outline crenulations in a Florence Range suite schist. The crenulations formed during D_{2T} or D_{3T} and folded M_{1T} micaceous minerals, which have since been pseudomorphed by M_{3T} muscovite and biotite. Some of the M_{3T} micas are randomly oriented and overprint the earlier S_{0+1T} and S_{2T} fabrics (field of view is 3 mm wide; Sample LC-89-127b).

Plate 4.10a. Photomicrograph of M_{2T} kyanite (K) replaced by M_{3T} muscovite (M) in a Florence Range suite schist (field of view is 3 mm wide; Sample LC-90-264).

Note that some, but not all muscovite parallels S_{0+1T} .

Plate 4.10b. Photomicrograph of M_{3T} fibrolite (F) replacing M_{3T} biotite (B) in a Florence Range suite schist. Note that the fibrolite both pseudomorphs the biotite (ie. long axes are parallel) and is randomly oriented. Fibrolite is also associated with prismatic sillimanite (S). Results of garnet-biotite Fe-Mg exchange geothermometry for this sample indicate that it reached temperatures of $\sim 650^{\circ}\text{C}$ (Appendix 4.4; field of view is 3 mm wide; Sample LC-90-315).

Plate 4.10c. Photomicrograph of garnet from a Florence Range suite schist. Garnet is 70% replaced by biotite, opaque minerals, quartz. The phantom edge of the garnet is marked "G". The core of the garnet contains straight quartz inclusion trails and its outer rim has overgrown at least one generation of crenulations (field of view is 3 mm wide; Sample LC-90-483).

Plate 4.10d. Photomicrograph of garnet (G) from a Florence Range suite schist that is wrapped by the foliation. One relict garnet face is bounded by a recrystallized quartz ribbon (Q). Biotite (B) in the altered garnet rim and in the tail (T) has been partially replaced by sillimanite (S; field of view is 3 mm wide; LC-91-469).

Plate 4.11a. Middle Jurassic (177 ± 7 Ma) leucosome from the Florence Range suite that formed after the accretion of Tracy Arm terrane to Stikinia (Sample LC-89-84b; A in Fig. 4.1).

Plate 4.11b. Close-up of Middle Jurassic (177 ± 7 Ma) leucosome from the Florence Range suite shown in Plate 4.35 (Sample LC-89-84b; A in Fig. 4.1).

Plate 4.11c. Refolded fold in the Florence Range metamorphic suite that involves the axial planar foliation of the refolded fold (S_{IT}).

Plate 4.11d. View southeast of northeast-verging folds outlined by marble and schist of the Florence Range suite shown in cross-section V-V' (Fig. 4.5c).

SIT

Plate 4.12a. View northwest of northeast-verging folds shown in Plate 4.11d and cross-section V-V' (Fig. 4.5c). Below the folds there are disrupted and folded carbonate lenses that are considered the continuation of the layer of fault blocks exposed above the fault shown in Plate 4.48.

Plate 4.12b. View southwest of the fault that carries the marbles shown in Plates 4.11d and 4.12a and cross-section V-V' (Fig. 4.5c) in its hangingwall.

Plate 4.12c. View southeast of the Wann thrust fault and D_{2T} and D_{3T} folds in the Florence Range suite. This view corresponds with the downplunge projection of cross-section W-W' in Fig. 4.5b.

Plate 4.12d. Line drawing of Plate 4.12c.

Plate 4.13a. View southwest of northeast-verging folds in the Florence Range suite, from above the Wann thrust fault. These folds lie in the upper limb of a D_{2T} synform shown in cross-section W-W' and labelled " D_{2T}^* " (Fig. 4.5b).

Plate 4.13b. View northwest of the Hoboe antiform. Marble layers on the east side of the antiform are steep and approximately north-south striking and those on west side dip moderately toward the west (see also Plate 4.43 and Fig. 4.9).

Plate 4.13c. View north of the Hoboe antiform. Marble layers are steep and approximately north striking on the east side of the antiform and moderately dipping on the west (see also Plate 4.52 and Fig. 4.9).

Plate 4.14a. View *northwest* of sheared, moderately dipping M_{2T} leucosome lenses in Florence Range suite schist. The orientation of these lenses is considered indicative of D_{3T} top-to-the-northeast sense of shear. Leucosome at the top of the photograph is lithologically similar to, and may be the same age as the Early to Middle Jurassic leucosome in Plate 4.35 (pen is 16.5 cm long).

Plate 4.14b. View east of the Mt. Mussen shear zone (MMSZ), which juxtaposes relatively unmetamorphosed volcanic rocks of the Triassic Stuhini Group of Stikinia (left), against possible Boundary Ranges suite metamorphosed actinolite schist and minor marble that are also considered part of Stikinia (right). The shear zone dips $\sim 45^\circ$ toward the northeast (Fig. 4.2).

Plate 4.14c. Photomicrograph of a Florence Range suite semipelite sample that was analysed for garnet-biotite Fe-Mg exchange geothermometry. The garnet is in apparent textural equilibrium with biotite, but the biotite is probably partially altered to chlorite, and the sample returns unrealistically high temperatures (Sample LC-89-148)

LITHOLOGY¹ MINERALOGY²

CHARACTERISTICS

BIOITITE AND MUSCOVITE SCHIST

LC-89-57c 40% Qtz, 30% Pl, 20% Bt, 10% Ms, 3% Ky, 2% Sil (fibrolite), acc. Zr, Mz, Tml

LC-89-59 38% Qtz, 35% Ms, 15% Bt, 5% Pl, 5% Grt, acc Mz, Op.

LC-89-127b 35% Ms, 30% Bt, 30% Qtz, 5% Plg, 3% Sil (fibrolite), 1% Chl, 1% Op, acc. Zr or Mz, Ap, Tml

LC-89-142 35% Ms, 30% Qtz, 15% Bt, 10% Plg, 8% Chl, 1% Grt, 1% Op?, acc. ?

LC-89-148 50% Qtz, 35% Plg, 20% Bt, 5% Grt, acc. Ap (geothermometry sample)

LC-89-157 25% Qtz, 20% Ms, 20% Bt, 20% Plg, 10% Sil (fibrolite), 3% Ky, 2% St, acc. Zr, Ap

LC-90-264 25% Ms, 25% Qtz, 15% Bt, 10% Plg, 10% Ky, 5% Grt, acc. Zr

LC-89-315 40% Qtz, 15% Bt, 15% Ms, 15% Pl, 10% Sil (Sil and fibrolite), 10% Plg, 10% Ky, 3% Grt, 2% Op acc. Tour (geothermometry sample)

LC-90-483 40% Ms, 30% Qtz, 20% Bt, 5% And, 5% Tour, 2% Grt, 1% Sil, acc. Op, Mz or Zr.

Medium-grained, grey schist with planar foliation defined by preferred orientation of Bt, Ms, and flattened Qtz and Pl grains. Relict porphyroblasts of unknown original composition have been replaced by Sil+Bt. Matrix Bt has also been replaced by Sil. There are two generations of Ms: one is coplanar with foliation; the second generation Ms is randomly oriented and surrounds relic Ky. Undulose extinction, curved grain boundaries and subgrains are exhibited by Qtz grains. Recrystallized Qtz has inclusions of Sil.

Medium-grained, grey schist. Foliation ($S_{0.17}$) defined by preferred orientation of Ms and Bt. $S_{0.17}$ is folded around open folds, and wraps around Grt. A second generation of biotite is randomly oriented. Relict porphyroblasts have been replaced by Qtz+Bt+Pl. Grt has inclusions of Qtz and Ms. There is subgrain development in Qtz and deformation twins in Pl.

Medium-grained, grey schist. Foliation ($S_{0.17}$) defined by preferred orientation of polygonalized Ms and Bt and flattened Qtz and Pl grains. Relict porphyroblasts of unknown original composition have been replaced by Qtz+Bt+Pl. Qtz exhibits subgrain development and there are deformation twins in Pl. Biotite is ~40% altered to Chl.

Medium-grained, grey schist. Foliation defined by preferred orientation of Ms and Bt. Micaceous minerals are folded around open folds, and wrap around relict porphyroblasts (Grt). Micas have been recrystallized since folding, as indicated by their lack of undulose extinction and the presence of blocky Ms grains that overprint foliation. Bt is intergrown (replaced by?) Chl. ~95% of Grt has been replaced by Chl and Ser. Grt has inclusions of Bt. Subgrain development and undulose extinction in Qtz and deformation twins in Pl.

Medium-grained, ineated Grt-Bt semipelite. Lineation is defined by the preferred orientation of recrystallized, elongate Qtz and Plg grains.

Medium grained grey schist with foliation defined by alignment of Ms and Bt. Ms and Bt are folded by crenulations that are truncated by Qtz, Plg, and Bt grains. There are two generations of biotite, one parallel to $S_{0.17}$ and the other is blocky and randomly oriented. Both are associated with fibrolite. Ky is coarse-grained and altered. Sil is randomly oriented. Staurolite fragments maintain the same crystallographic orientation, but are surrounded (replaced?) by second generation biotite. Qtz exhibits undulose extinction, has serrate grain boundaries, and contains subgrains and sillimanite inclusions. Plg is blocky and has overgrown possible D_T or D_T crenulations.

Medium-grained, grey schist. Foliation (defined by preferred orientation of Ms and Bt) wraps around relict Grt (<7 mm diameter) and parallels cm-scale compositional layering between Qtz-rich and mica-rich layers ($S_{0.7}$). Micas have been recrystallized since folding, as indicated by their lack of undulose extinction and the presence of blocky Ms grains that overprint foliation. Ky is replaced by late Ms. ~30 to 100% of Grt porphyroblasts have been replaced by intergrown Bt +Sil. Grt has ~straight inclusions of Qtz. Qtz grains have serrate boundaries and undulose extinction. Pl has undulose extinction.

Medium-grained, grey schist. Foliation (defined by preferred orientation of Ms and Bt) wraps around relict Grt (<7 mm diameter). Up to 80% of Grt porphyroblasts has been replaced by Bt + Qtz + Op + Sil + Pl. Grt has Qtz inclusions. Qtz ribbons have been recrystallized; subgrains have serrate boundaries ? and undulose extinction?. Sil both parallels $S_{0.17}$ and is randomly oriented.

Medium-grained, grey schist. Foliation (defined by preferred orientation of Ms and Bt) and And are folded around open crenulations. Grt: cores have straight inclusions of Qtz. Grt rims have overgrown D2 folds and have been replaced by Bt + Qtz.

1. Capital letters are the same as those shown in Figure 1 and in Tables 2 and 3; 2. acc. - accessory mineral (comprises <1% of rock). Where possible, the abbreviations of Kretz (1983) have been used. Act - actinolite, Ap - apatite, Bt - biotite, Cal - Calcite, Chl - chlorite, Ep - epidote, Hbl - hornblende, Grt - garnet, Ms - muscovite, Pl - plagioclase, Py - pyrite, Op - opaque minerals, Or - orthoclase, Qtz - quartz, Rt - rutile, St - staurolite, Ti - titanite, Zrn - zircon.

Table 4.1 (continued). Lithologic descriptions of Florence Range suite rocks with diagnostic mineral assemblages and metamorphic fabrics.

LITHOLOGY ¹	MINERALOGY ²	CHARACTERISTICS
AMPHIBOLITE		
LC-89-159 Hanging valley north of Willison Creek	70% Hbl, 10% Grt, 10% Plg, 10% Bt, 1% Qtz, 1% Op, acc?	Massive, fine-grained, garnet amphibolite. Grt ~4 mm diameter. >60 % of Grt has been replaced by Pl + Bt + Qtz. Pl and Qtz are interstitial. Hbl grains were probably once medium-grain size, but have been broken along cleavage surfaces.
LC-89-248 Base of Willison Glacier	60% Hbl, 30% Plg, 5% Bt, 2% Op, 3% Grt, acc. Ap	Massive, fine-grained, garnet amphibolite. Hbl is fragmented and folded. Grt crystals were originally ~0.5mm diameter and comprised ~10% of rock, but have been replaced (~60%) by Bt + Plg + Chl. Relic Grt crystals have straight inclusion trails of Op minerals, is fractured, and has been replaced by Plg + Bt + Chl. Plg exhibits undulose extinction and is fragmented. Hbl grains were probably once medium-grain size, but have been broken along cleavage surfaces.
LC-89-306 Hoboe Creek area	55% Hbl, 25% Plg, 10% Qtz, 5% Op, 3% Grt, 2% Bt acc. Ap	Fine- to medium-grained, garnet amphibolite with weak compositional layering. Hbl is fractured. Grt is anhedral, with only fragments remaining, and has been replaced (up to 90%) by Bt + Plg. Relic Plg exhibits undulose extinction.
LC-89-318 North side of Caplice Glacier	60% Hbl, 20% Plg, 10% Grt, 5% Qtz, 3% Chl, 2% Bt, acc. Ep, Op	Fine- to medium-grained, garnet amphibolite with cm-scale compositional layering and a foliation defined by alignment of Hbl and Hbl fragments. Grt is wrapped by the foliation, contains inclusions of Op minerals, and is replaced (up to 30%) by Chl + Plg. Qtz fills Grt tails. There are 2 generations of Plg. One is altered, bent, fragmented and sericitized. The other exhibits straight extinction and is not visibly altered. Chl and Bt exhibit straight extinction and are randomly oriented.
LC-90-320 Hanging valley north of Caplice Glacier	40% Hbl, 30% Plg, 15% Bt, 10% Qtz, 3% Op, 2% Gar, acc. Ap	Fine-grained, garnet amphibolite with mm-scale compositional layering. Hbl fragments are aligned parallel to layering. Bt is blocky, exhibits straight extinction, and most, but not all grains parallel layering. Plg and Qtz exhibit undulose extinction and Qtz comprises subgrains. Relic Grt is anhedral, and replaced by Plg+Op + Bt + Qtz.
QUARZITE		
LC-89-3C-2 South end of Atlin Lake	95% Qtz, 3% Ms, 2% Chl	Fine grained, pale grey quartzite. Foliation defined by alignment of elongate Qtz grains and Ms. Qtz grains wrapped by thin Ms grains. Subgrain development and undulose extinction in Qtz.
LC-90-449 Wann River valley	95% Qtz, 3% Op, 2% Bt	Fine grained, dark grey quartzite. Foliation defined by the preferred orientation of Bt and the variable grain-size of Qtz, which increases with decreasing Op (graphite?) content. Qtz grains have serrate grain boundaries and Ms inclusions, and exhibit subgrain development and undulose extinction.
CALCSILICATE		
LC-89-189b Wann River headwaters	45% Cal, 25% Grt, 15% Di, 10% Tr, 5% Ti	Fine- to medium-grained calc-silicate with cm-scale white-, green-, and pink-weathering compositional layering.
LC-89-85 Wann River headwaters	44% Cal, 40% Di, 20% Qtz	Fine- to medium-grained calc-silicate with mm-scale alternating Qtz- and Ce-rich layers. Qtz exhibits undulose extinction and has curved grain boundaries.
LC-89-86 Wann River headwaters	50% Cal, 25% Ves, 15% Di, 5% Tr, 4% Ti, 1% Ep	Fine- to medium-grained calc-silicate with well developed calcite-, vesuvianite, and diopside-rich compositional layers. Calcite crystals contain titanite inclusions.
FELSIC GNEISS LC-89-92b Wann Iver headwaters	50% Plg, 35% Qtz, 10% Bt, 3% Ms, 1% Chl, acc. Ap, Zr	Fine- to medium-grained tonalite with igneous texture. Foliation is defined by the alignment of flattened Qtz and Plg grains and the preferred orientation of Ms. Grain boundaries are curved. Qtz and Plg exhibit undulose extinction.
LEUCOSOME		
LC-89-84b Wann River headwaters	leucosome: 60% Plg, 35% Qtz, 5% Bt, acc. paleosome: 30% Bt, 34% Plg, 34% Qtz, 1% Grt, 1% Op, acc. Ap, Zr, Op	Coarse-grained tonalite with igneous texture. Moderate alignment of Bt and Ms. Grain boundaries are curved. Qtz exhibits undulose extinction and subgrains. Plg exhibits minor alteration to sericite and undulose extinction. Medium-grained schist with fine-grained anhedral garnet porphyroblasts. Foliation defined by Bt and elongate Qtz grains.

1. Capital letters are the same as those shown in Figure 1 and in Tables 2 and 3; 2. acc. - accessory mineral (comprises <1% of rock), Act - actinolite, Ap - apatite, Bt - biotite, Cal - Calcite, Chl - chlorite, Ep - epidote, Hbl - hornblende, Grt - garnet, Ms - muscovite, Pl - plagioclase, Py - pyrite, Op - opaque minerals, Or - orthoclase, Qtz - quartz, Rt - rutile, St - staurolite, Ti - titanite, Zm - zircon.

Table 4.2. Structural events and periods of metamorphic mineral assemblage growth in Tracy Arm terrane and Stikinia.

<u>TRACY ARM</u>			<u>STIKINIA</u>		
	<u>CHARACTERISTICS</u>	<u>AGE</u>		<u>CHARACTERISTICS</u>	<u>AGE</u>
D _{1T}	Compositional layering (S _{0T}) interpreted as bedding		D _{1S}	Compositional layering (S _{0S}) interpreted as bedding	
D _{1T}	Layer-parallel foliation (S _{0+1T}), outcrop scale isoclinal folds with axial planar foliation	>176 Ma ~syn-M _{1T}	D _{1S}	Layer-parallel foliation (S _{0+1S}), out-crop scale isoclinal folds with axial planar foliation	~syn-M _{1S}
M _{1T}	Biotite-grade; upper greenschist to lower amphibolite facies	~syn-D _{1T} >176 Ma	M _{1S}	Andalusite- to kyanite-grade; amphibolite facies	~syn-D _{1S} >176 Ma
D _{2T}	Outcrop- and mountain side-scale close folds that fold S _{0+1T} and leucosome		D _{2+3S}	Outcrop-scale close to isoclinal folds that fold S _{0+1S}	>176 Ma
M _{2T}	Kyanite-grade	>176 Ma			
STRUCTURES AND METAMORPHIC FABRICS THAT AFFECT BOTH TRACY ARM TERRANE AND STIKINIA ROCKS					
D _{WRSZ}	Wann River shear zone - characterized by a transposition foliation (S _{WRSZ}) that is coplanar with the Stikinia-Nisling terrane boundary, and a mineral-elongation lineation (L _{WRSZ}). Interpreted as forming during displacement on the Stikinia-Tracy Arm terrane boundary.				184-176 Ma
M _{3T}	Sillimanite-grade, with no preferred orientation of M _{3T} minerals.	~177 Ma	M _{2ST}	Amphibolite facies in structurally highest levels and greenschist facies in underlying structural levels	~177 Ma
Caplice fault					
Mt. Switzer volcanism (Tertiary ~ 62 Ma) ¹					
Steep faults (shown as normal faults on Figs. 4.2 and 4.4)					
Llewellyn fault zone?					
Sloko Group volcanism (Tertiary ~ 56 Ma) ¹					
Willison Synform and Willison Creek Fault zone					
Llewellyn fault zone?					
Notes: D=deformational event, M=metamorphic episode, S=surface, L=lineation, subscript "T" denotes events that affected Tracy Arm terrane, subscript "S" denotes events that affected Stikinia.					
¹ Pers. Comm. M. G. Mihalynuk, 1994.					

Table 4.3. Pressures determined for Early Jurassic Tagish Lake suite plutons and a mid-Cretaceous pluton from the Tagish Lake area, on the basis of the Al-in-hornblende geobarometer¹.

Sample number	Sample description and location ²	Pressure ^{3,4}		Depth ⁵ (km)
		core	rim	
LC-90-226 "G" in Fig. 2	Hale Mountain granodiorite (An ₂₈ ; Z ⁷ =186 ±1.5 Ma; Ti ⁸ =184.5 ±0.9 Ma) UTM 8V, 536680E 6593470N	7.5 +1.2/-1.0 ⁶	7.2 +0.9/-1.2	27 +3/-5
LC-90-272 "H" in Fig. 2	Hale Mountain granodiorite (An ₃₆ ; Z ⁷ =186 ±1.5 Ma; Ti ⁸ =184.5 ±0.9 Ma) UTM 8V, 549440E 6552800N	6.3 +0.3/-0.2	6.5 +0.6/0.5	24 ±2
LC-90-314 "I" in Fig. 2	Hoboe quartz syenite (An ₂₄₋₃₁ ; Z ⁷ =189.5 ±3 Ma; Ti ⁸ = 186.5 ±0.4 Ma) UTM 8V, 547850E 6556325N	6.3 +0.5/-0.3	6.7 +0.5/-0.7	25 +2/-3
LC-90-343 "J" in Fig. 2	Mt. Caplice quartz monzonite (An ₂₅₋₂₆ ; Z ⁷ =180.9 ±3.1 Ma; Ti ⁸ = 177.0 ±0.6 Ma) UTM 8V, 564825E 6557750N.	4.7 +0.1/-0.3	5.6 +1.1/-0.7	21 +4/-3
LC-90-359 "K" in Fig. 2	Massive, medium-grained granodiorite (An ₃₀₋₄₀ ; Z ⁷ =103 ±0.7 Ma) UTM 8V, 544950E 6564725N	2.4 +0.4/-0.3	1.8 ±0.6	6.8 ±2

¹Calibrations for the Al in hornblende geobarometer are from Hollister et al. (1987).

²Sample locations are shown in Figure 2. ³Pressure is measured in kbar (1 kbar = 100 MPa). ⁴Errors shown here reflect the variation between analyses of different grains but do not include the proposed error for the geobarometer of ±1 kbar (Hollister et al., 1987).

⁵Depth calculated assuming specific gravity of rock column is 2.65 g/cm³. ⁶Pressures determined using other calibrations of this barometer may be as much as 2 kbar lower (Appendix 2.2); error assignment of ±2 kbar is therefore recommended for all analyses.

⁷U-Pb zircon age (Chapter 3). ⁸U-Pb titanite age (Chapter 3).

Table 4.4. Temperatures determined from the garnet-biotite Fe-Mg geothermometer for samples from the Florence Range suite¹.

Garnet-Biotite Pair	Ferry and Spear (1978)	Ganguly and Saxena (1984)	Berman (1990)
Sample LC-89-148			
1	771	740	818
2	803	768	841
3	796	800	847
4	817	801	859
Average	797±76 ²	777±87	841±73
Sample LC-89-315			
1	613	652	595
2	666	686	640
3	661	691	639
4	679	660	608
Average	655±92 ²	672±70	620±70

Notes: ¹ Temperatures measured in °C. Assumed pressure = 6 kbar. ² Assigned error based on the variation in temperatures determined for different garnet-biotite pairs and an inherent error of ±50°C for each geobarometer. ³ Sample description and location given in Table 4.1.

Table 4.5. U-Pb data for zircon and monazite fractions from the Florence Range suite.

Fraction size ¹	Wt. μ g	U ppm	Pb ² ppm	206Pb ³ 204Pb	Pbc ⁴ pg	208Pb %	206Pb ⁵ 238U	207Pb ⁵ 235U	Corr. Coeff.	207Pb ⁶ 206Pb	207Pb ⁷ 206Pb	207/206 Age ⁶ age (Ma)	207/206 Age ⁷ age (Ma)
FLORENCE RANGE SUITE													
(A) Leucosome (LC-89-84b) leucosome from southern headwaters of the Wann River. UTM 8v 6570400N 536000E.													
zircon													
A (74-105)	17	2012	52.09	6056	10	1.8	0.02815 $\pm$ 0.10%	0.1964 $\pm$ 0.11%	0.89	0.05061 $\pm$ 0.05%	0.05058	223.1 $\pm$ 2.2	221.87
B (74-105)	9	1670	48.35	3346	9	2.4	0.03117 $\pm$ 0.10%	0.2333 $\pm$ 0.12%	0.81	0.05429 $\pm$ 0.07%	0.05427	383.3 $\pm$ 3.1	382.1
C (74-105)	3	3792	119.1	78	417	2.9	0.03354 $\pm$ 0.87%	0.2661 $\pm$ 2.5%	0.69	0.05753 $\pm$ 2.0%	0.05750	512 $\pm$ 86	510.9
D (74-105)	6	865	63.50	3169	8	8.7	0.07143 $\pm$ 0.12%	0.9038 $\pm$ 0.13%	0.94	0.09176 $\pm$ 0.04%	0.09174	1462.5 $\pm$ 1.6	1462.1
E (74-105)	10	1975	54.64	1711	22	3.2	0.02958 $\pm$ 0.10%	0.2156 $\pm$ 0.13%	0.83	0.05285 $\pm$ 0.08%	0.05282	322.3 $\pm$ 3.4	321.0
(B) schist (LC-89-59) from north of Willison Creek. UTM 8v 6569510N 539390E.													
monazite													
M2 (74-105)	19	2350	237.6	2458	32	75	0.02791 $\pm$ 0.25%	0.1899 $\pm$ 0.28%	0.53	0.04935 $\pm$ 0.25%	0.04942	164.4 $\pm$ 11.9	167.6
M3 (74-105)	19	1760	161.5	388	159	73	0.02752 $\pm$ 0.16%	0.1872 $\pm$ 0.45%	0.69	0.04933 $\pm$ 0.36%	0.04939	163.5 $\pm$ 17.0	166.3
M5 (74-105)	9	1021	146.2	219	77	82	0.02820 $\pm$ 0.22%	0.1946 $\pm$ 0.88%	0.65	0.05005 $\pm$ 0.76%	0.05017	197.1 $\pm$ 35.3	202.7
M6 (74-105)	14	1180	131.0	502	59	77	0.02800 $\pm$ 0.13%	0.1928 $\pm$ 0.43%	0.60	0.04994 $\pm$ 0.37%	0.05002	192.3 $\pm$ 17.2	196.0
M2 ⁸	19	2350	238	2458	33	75	0.02788 $\pm$ 0.25%	0.1893 $\pm$ 0.31%	0.53	0.04924 $\pm$ 0.28%		159.3 $\pm$ 13.1	
M3 ⁸	19	1760	161	388	167	73	0.02724 $\pm$ 0.16%	0.1801 $\pm$ 0.45%	0.72	0.04795 $\pm$ 0.35%		96.9 $\pm$ 16.6	
M5 ⁸	9	1021	146	219	79	82	0.02785 $\pm$ 0.36%	0.1856 $\pm$ 1.5%	0.89	0.04833 $\pm$ 1.2%		115.6 $\pm$ 58.1	
M6 ⁸	14	1180	131	502	61	77	0.02788 $\pm$ 0.20%	0.1898 $\pm$ 0.79%	0.85	0.04937 $\pm$ 0.62%		165.6 $\pm$ 29.1	

Notes: (1) sizes (i.e. +105) refer to length aspect of crystals in microns; (2) radiogenic Pb; (3) measured ratio, corrected for spike and fractionation; (4) total common Pb in analysis corrected for spike and fractionation; (5) corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; (6) corrected for blank and common Pb, errors are two sigma in Ma; decay constants are those of Steiger and Jager (1977); for analytical details see Parrish et al. (1987); (7) corrected for initial Th exclusion, assuming whole rock Th/U=3/1 (see Scharer, 1984); (8) assumed initial common Pb composition determined from a Pb isochron diagram is $^{206}\text{Pb}/^{204}\text{Pb}=22.66 \pm 2.21$, $^{207}\text{Pb}/^{206}\text{Pb}=16.39 \pm 0.42$ - see text for discussion.

Table 4.6. U-Pb data for titanite and rutile fractions from the Florence Range and Boundary Ranges suites.

Fraction size ¹	Wt. μ g	U ppm	Pb ² ppm	$^{206}\text{Pb}^3$ / ^{204}Pb pg	Pbc ⁴ %	$^{206}\text{Pb}^5$ / ^{238}U	$^{207}\text{Pb}^5$ / ^{235}U	Corr. Coeff.	$^{207}\text{Pb}^6$ / ^{206}Pb	$^{207}\text{Pb}^6$ / ^{206}Pb age (Ma)
FLORENCE RANGE SUITE (continued)										
(C) calcisilicate (LC-89-86), from the southern Wann River headwaters. UTM 8v 536330E 6571460N.										
titanite										
S1 (+149)	100	58	1.961	103	131	18	0.03066 $\pm$ 0.55%	0.2157 $\pm$ 1.8%	0.69	0.05103 $\pm$ 1.4%
S1'	100	58	1.961	103	135	19	0.02985 $\pm$ 0.59%	0.2057 $\pm$ 1.6%	0.71	0.04998 $\pm$ 1.2%
(D) biotite schist (LC-91-469), from south of the upper Wann River, UTM 8v 539325E 6573700N.										
rutile										
R1 (+149)	691	20	1.919	40	989	16	0.02405 $\pm$ 2.5%	0.1901 $\pm$ 6.7%	0.70	0.05733 $\pm$ 5.3%
R2 (+149)	777	18	1.334	50	685	17	0.02352 $\pm$ 1.7%	0.1906 $\pm$ 4.5%	0.69	0.05879 $\pm$ 3.6%
R1 ⁸	691	20	1.919	40	956	19	0.02530 $\pm$ 1.9%	0.2023 $\pm$ 7.9%	0.70	0.05798 $\pm$ 6.7%
R2 ⁸	777	18	1.334	50	664	19	0.02427 $\pm$ 1.3%	0.1978 $\pm$ 5.3%	0.70	0.05910 $\pm$ 4.5%
BOUNDARY RANGES SUITE										
(E) Wann River gneiss (LC-91-458b), compositionally layered hornblende-plagioclase gneiss from north of the Wann River. UTM 8v 539350E 6569275N.										
titanite										
S1 (+149)	132	145	4.058	241	155	7.7	0.02860 $\pm$ 0.23%	0.1980 $\pm$ 0.74%	0.68	0.05023 $\pm$ 0.61%
S2 (+149)	116	162	4.913	244	162	7.9	0.03100 $\pm$ 0.23%	0.2153 $\pm$ 0.90%	0.63	0.05037 $\pm$ 0.78%
(F) Wann River Gneiss (LC-89-91B), compositionally layered hornblende-plagioclase gneiss from south of the upper Wann River, UTM 8v 536750E 6573525N.										
rutile										
R1 (+149)	1805	39	0.75	291	372	1.1	0.02283 $\pm$ 0.21%	0.1551 $\pm$ 0.60%	0.70	0.04926 $\pm$ 0.48%
R2 (+149)	2352	39	0.75	281	503	1.1	0.02285 $\pm$ 0.22%	0.1549 $\pm$ 0.64%	0.69	0.04916 $\pm$ 0.52%

Notes: (1) sizes (i.e. +105) refer to length aspect of crystals in microns; (2) radiogenic Pb; (3) measured ratio, corrected for spike and fractionation; (4) total common Pb in analysis corrected for spike and fractionation; (5) corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; (6) corrected for blank and common Pb, errors are two sigma in Ma; decay constants are those of Steiger and Jager (1977); for analytical details see Parrish et al. (1987); (7) assumed initial common Pb composition is $^{206}\text{Pb}/^{204}\text{Pb}=21.04\pm0.52$, and $^{207}\text{Pb}/^{204}\text{Pb}=15.85\pm0.34$ - see text for discussion. (8) assumed initial common Pb composition is $^{206}\text{Pb}/^{204}\text{Pb}=19.89\pm0.24$, $^{207}\text{Pb}/^{204}\text{Pb}=15.85\pm0.57$ - see text for discussion.

CHAPTER 5: AN EVALUATION OF TECTONIC MODELS FOR THE ORIGIN AND TECTONIC HISTORY OF TRACY ARM TERRANE

ABSTRACT

Tracy Arm terrane is a narrow sliver of metamorphosed continental margin-type rocks exposed in the northwestern Cordillera. It is in part derived from a Precambrian source, and is tectonically bounded by arc-like terranes that record little continental influence (Stikinia to the east and Wrangellia and Alexander terranes to the west), though other studies have interpreted some contacts with adjacent assemblages as depositional. Tracy Arm terrane has been correlated with lithologically similar parautochthonous assemblages in eastern Yukon and mainland Alaska on the basis of the apparent association between the quartz-rich metasedimentary assemblages and Paleozoic volcanic assemblages. This correlation implies that Tracy Arm terrane formed next to the western margin of ancestral North America, and supports tectonic models involving "obduction" and "tectonic enclosure". However, critical evaluation of the association between the quartz-rich metasedimentary assemblages and volcanic assemblages in the Coast Mountains shows that in many cases the inferred association between these assemblages could be tectonic, rather than depositional, and that their correlation with parautochthonous assemblages is equivocal. It is therefore proposed that Tracy Arm terrane was separated from continental margin rocks to the east by Paleozoic to Mesozoic arc terranes prior to mid-Jurassic time. Three additional models for the origin and accretion of Tracy Arm terrane are also possible. The "exotic" model interprets Tracy Arm terrane as a continental ribbon that was rifted from the

trailing edge of the land mass (Australia?) that formed the rifted complement to the western margin of Pangea during, or soon after the Neoproterozoic break-up of Laurentia and formation of Panthalassa. In the "dextral strike-slip" model Tracy Arm terrane originated along the southwestern margin of Pangea and was transported along dextral strike-slip faults to its present position, west and outboard of arc-like and oceanic terranes. The "sinistral" model envisages that Tracy Arm terrane either originated along the eastern margin of Eurasia (probably the Siberian Platform), or adjacent to northern North America (Arctic Alaska terrane, before the Canada Basin formed). Sinistral motion is consistent with the large-scale sense of shear inferred from other areas along the western margin of Pangea for the mid-Mesozoic.

INTRODUCTION

Tracy Arm terrane has been referred to as the principal conundrum of the Canadian Cordillera (Coney and Evenchick, in press) because it is interpreted as a passive margin sequence of Paleozoic and/or Precambrian age, but is separated from the nearest continental margin rocks (ancestral North America) by arc-like terranes that have little continental affinity (Stikinia and Cache Creek terranes; Fig. 5.1). It was juxtaposed against the western margin of Stikinia between 184 and 173 Ma, during or slightly before the accretion of Stikinia to the composite western margin of North America (Cache Creek-Quesnellia-Slide Mountain-Kootenay-North America; Chapter 4). After mid-Cretaceous time, Tracy Arm terrane was transported primarily northward, along with terranes to the east, by large dextral transcurrent fault systems

to its present position in the northern Coast Mountains of British Columbia, Yukon and southeast Alaska (Gabrielse, 1985). Although the mid-Mesozoic and younger tectonic history of Tracy Arm terrane is reasonably well known, it is not clear how, when or where Tracy Arm terrane was detached from its source of continental clastic detritus or transported to a position next to Stikinia. Even the depositional age is poorly known.

The purpose of this chapter is to evaluate the tectonic models that attempt to explain why the Tracy Arm terrane is presently orphaned from its continental parentage and account for its early history (Fig. 5.2; Table 5.1). Despite the numerous models suggested for its tectonic history, proposed correlations for Tracy Arm terrane have not been rigorously tested. Furthermore, the implications of a terrane with strong continental affinity sandwiched between terranes composed primarily of relatively unevolved arc material are not always considered in Cordilleran tectonic syntheses (eg. Wernicke and Klepacki, 1988; Umhoefer, 1987; Plafker, 1990; van der Heyden, 1992).

The proposed models may be separated into two groups: those that assume that rocks of the Tracy Arm terrane are correlative with sequences that are parautochthonous to North America and those that do not require such correlations, and assume that tracts of oceanic and island arc crust separated the Tracy Arm terrane from ancestral North America (Table. 5.1; Fig. 5.2). Hence in this chapter, a discussion of the validity of lithostratigraphic correlations between the Tracy Arm terrane and sequences that are parautochthonous with respect to North America precedes evaluation of tectonic models proposed for this terrane.

TRACY ARM TERRANE

Tracy Arm terrane consists of discontinuous fragments of metamorphosed continental margin-type sediments exposed in the northern Coast Mountains from near Bella Coola (~53.5°N latitude; Wheeler et al., 1991; Gehrels et al., 1992a), through southeastern Alaska and northwestern British Columbia, to northern Yukon and Alaska (Figs. 5.1; 5.3; Coney et al., 1980; note that Tracy Arm terrane has also been included as part of Nisling terrane; see Chapter 2). In general, these fragments are exposed west of Stikinia and east of Wrangellia and Alexander terrane, which together comprise the Insular Superterrane (Monger, 1989). Tracy Arm terrane primarily comprises rocks of sedimentary protolith, including thick, continuous layers of carbonate rock and quartzite and pelitic to semipelitic rock. Locally, it includes amphibolite, or lower grade equivalents (basites), and possible metachert. The quartz-rich clastic rocks commonly contain detrital Precambrian zircons, reflecting at least partial derivation from a Precambrian source. However, on their own, metasedimentary rocks with a component of inherited Precambrian detritus cannot be used as a hallmark for Tracy Arm terrane because similar rocks exist in other terranes within the Coast Belt. For example, a component of inherited Precambrian zircon has been documented in plutons of Stikinia (Bevier and Anderson, 1991; Thorkelson, et al., in review), and the structurally lowest (oldest) Stikinia rocks may include a component of quartz-rich strata (McClelland, 1992; Chapter 2).

Where they have been studied, the fragments that make up Tracy Arm terrane have been given informal names, such as the Tracy Arm assemblage and Kah Shakes

sequence in southeastern Alaska, the Nisling assemblage ("lower unit" of the Aishihik metamorphic suite of Johnston and Timmerman, 1994a) in Yukon, and the Florence Range suite in northwestern British Columbia (Fig. 5.3; Chapter 2). Tracy Arm terrane rocks may also be preserved as pendants in the Coast Belt that have not been studied in detail and are referred to as "undivided metamorphic rocks" (Fig. 5.2; Gehrels and Berg, 1984; Wheeler et al., 1991). Tracy Arm terrane is tentatively correlated with Dillinger, Nixon Fork, Kilbuck, and Idono terranes of mainland Alaska (Chapter 2). It has also been correlated by others with Yukon-Tanana terrane, in part on the basis of its apparent association with Paleozoic volcanic rocks (e.g. Gehrels et al., 1990; 1992a; Mortensen, 1992).

CONTINENTAL MARGIN-LIKE ASSEMBLAGES AND THEIR RELATION TO ADJACENT ARC-LIKE ASSEMBLAGES IN BRITISH COLUMBIA, YUKON, AND ALASKA

The inferred correlation of Tracy Arm terrane with parautochthonous North American rocks has largely been based on a spatial association of continental margin-like assemblages with mid-Paleozoic volcanic assemblages in British Columbia, Yukon, and Alaska (e.g. Gehrels et al., 1990; 1992a). Mid-Paleozoic volcanic assemblages are reasonably common in Cordilleran terranes; they occur in Eastern Klamath terrane, Northern Sierra terrane, Stikinia, Chilliwack terrane, Slide Mountain terrane, Kootenay terrane, the southern Brooks Range, Ruby terrane, Wrangellia and possibly Quesnellia (Rubin et al., 1990; Wheeler et al., 1991; Monger et al., 1991). In

some places Paleozoic arc-like assemblages are in depositional contact with continental margin-like assemblages (e.g. the Lower Paleozoic Milford Group unconformably overlies Lower Paleozoic and older assemblages considered part of ancestral North America; Klepacki and Wheeler, 1985; Klepacki, 1989), but in the Coast Mountains the contact relations between the arc-like and continental margin-like assemblages are uncertain, often because the original contact is metamorphosed, tectonically reworked, not preserved and/or not exposed. The arc-like and continental margin-like assemblages of the Coast Mountains, and their mutual boundaries are described below from north to south.

Yukon-Tanana Terrane and the Nisutlin Subterrane of Kootenay Terrane, Yukon and Eastern Alaska

As it was originally defined, Yukon-Tanana terrane comprised heterogeneous gneiss and schist of "quartzo-feldspathic and silicic volcanic-rich protolith of probable Precambrian to known Paleozoic age, and of unknown continental affinity" (Coney et al., 1980; also referred to as Yukon Crystalline terrane, e.g. Churkin et al., 1982). It was originally thought to include nappes of upper Paleozoic oceanic assemblages that had been thrust over these rocks (e.g. Coney et al., 1980), but these were later differentiated and designated the Seventy Mile terrane (Jones et al., 1981; Fig. 5.1), Salcha terrane (Churkin et al., 1982), Innoko, Tozitna and/or Angayucham terranes (Silberling et al., 1984); these oceanic terranes have been correlated with Slide Mountain terrane of the Canadian Cordillera (Silberling et al., 1984; Fig. 5.1). Yukon-

Tanana terrane was distinguished from other terranes in Yukon and Alaska primarily on the basis of metamorphic grade, although Coney et al. (1980) noted some distinctive lithologic units, and Churkin et al. (1982) attempted to subdivide it into subterrane (Y₁₋₄).

Mortensen (1992) included rocks previously considered part of Tracy Arm terrane in his definition of Yukon-Tanana terrane and used lithologic, isotopic and U-Pb age criteria to divide Yukon-Tanana terrane into "lower, middle and upper units". The "lower unit" mainly comprises quartzofeldspathic metasediments and minor pelitic schist with marble, calc-silicate and calcareous schist bands and has been correlated with continental margin-like rocks of Tracy Arm terrane (Mortensen, 1992). The "middle unit" comprises carbonaceous phyllite, schist, and quartzite with rare marble and pebble conglomerate, all interlayered with abundant mafic and felsic metavolcanic rocks that are at least in part Devonian-Mississippian in age; it was considered equivalent to Wheeler et al.'s (1991) Nasina assemblage of Nisling terrane (Mortensen, 1992). For the purposes of this chapter the name "Nasina assemblage" is used for the "middle unit" of northern and eastern Yukon (the "middle unit" of southwestern Yukon is discussed below). In southeast Yukon, the "upper unit" includes quartzite and massive, Carboniferous to Early Permian, fossiliferous carbonate rocks, whereas in northernmost Yukon it includes mid-Permian felsic metaplutonic and metavolcanic, and minor mafic metavolcanic rocks that are interlayered with impure quartzite and rare carbonaceous schist (Mortensen, 1992). On the basis of lithologic similarities Mortensen and Jilson (1985), and later Wheeler et al. (1991) considered the "upper

unit" to be correlative with coeval rocks to the south that belong to the Barkerville subterrane of Kootenay terrane. Wheeler et al. (1991) referred to them as the Nisutlin subterrane of Kootenay terrane. The "upper unit" is herein referred to as the "Nisutlin assemblage". Because Permian dykes in northernmost Yukon form a "definite link between the Nasina assemblage and rocks of the Nisutlin subterrane" (Mortensen, 1992), both the Nasina and Nisutlin assemblages are regarded as part of the Nisutlin subterrane of Kootenay terrane (Fig. 5.3). Although Kootenay terrane remains "suspect" because it records deformational and magmatic events not recognised in the autochthonous miogeoclinal sequence to the east (e.g. Wheeler et al, 1991; Coney and Evenchick, in press), rocks in both the Barkerville and Nisutlin subterrane are generally considered distal equivalents of North American strata that have been telescoped along thrust faults to their present positions adjacent to ancestral North America (Tempelman-Kluit, 1979; Monger, 1989).

As it has been defined by Mortensen (1992), Yukon-Tanana terrane includes both continental margin-like and arc-like assemblages, but a geologic link between these assemblages has yet to be demonstrated. Some of the metasedimentary rocks in the Yukon-Tanana terrane have been intruded by 340 to 361 Ma orthogneisses, but none of these orthogneisses form a "stitching pluton" between the "lower unit" and the Nisutlin subterrane. In fact, there are no known pre-Cretaceous geologic links between the "lower unit" and the "middle and upper units" of Yukon-Tanana terrane (ie. the Nasina and Nisutlin assemblages; J.K. Mortensen, pers. comm., 1992). Consequently, the terrane affiliation of the "lower unit" is not clear; it has been correlated with the

Nisling assemblage of southwestern Yukon (see below; Mortensen, 1992), but could also represent an autochthonous North American miogeoclinal sequence, onto which the Nisutlin subterrane was tectonically emplaced, possibly during the Mesozoic. A North American connection is supported by similar U-Pb ages for single detrital zircon crystals from both "Yukon-Tanana rocks" and the North American miogeocline (Mortensen, 1990; J.K. Mortensen pers. comm. in Gehrels et al., 1991), but the original relations between "lower unit" and ancestral North America are uncertain because they are separated by either the Northern Rocky Mountain trench or the Finlayson Lake fault zone (Mortensen and Jilson, 1985). The "lower unit" is more reasonably assigned a name that implies no terrane affinity, and is herein referred to as the "Yukon assemblage" (Fig. 5.3); the term "Yukon-Tanana terrane", which has led to much confusion, should be abandoned.

Although a pre-Cretaceous link between continental margin-like rocks in Yukon-Tanana terrane (Yukon assemblage) and Paleozoic arc-like rocks (Nisutlin assemblage) from east of Stikinia has yet to be demonstrated, this apparent association has been used to correlate these rocks with continental margin-like rocks from west of Stikinia. The latter have been included in Yukon-Tanana terrane (e.g. Mortensen, 1992; Gehrels et al., 1990; 1992a) and Tracy Arm terrane (Coney et al, 1980; Chapter 2). They have been studied in the most detail in the Aishihik Lake area of southwestern Yukon.

Aishihik Lake area, Southwestern Yukon

Most recently the metamorphic rocks in the Aishihik Lake area have been referred to as the Aishihik metamorphic suite and divided into "lower and upper packages" (Johnston and Timmerman, 1994a, b, c), which correspond with the "lower and middle units" of Yukon-Tanana terrane described by Mortensen (1992), respectively; no rocks corresponding to the "upper unit" of Yukon-Tanana terrane have been recognised in Aishihik Lake area (Mortensen, 1992). The "upper package" includes metabasite, metagabbro, metafelsite, hornblende diorite gneiss and an Early Mississippian two-mica granitic gneiss. It has been correlated with the Nasina assemblage by Mortensen (1992) and Johnston and Timmerman (1994a), and has also been interpreted as the northern extension of Stikinia and correlated with the Paleozoic rocks of the Stikinia type-area, which are referred to as the Stikine assemblage (see below; Fig. 5.3; Currie, 1992; Chapter 3). The "upper package" lies structurally below the Early Jurassic Aishihik Batholith of the Aishihik plutonic suite (187 $\pm$ 9.7/-0.9 Ma; Johnston, 1993; Johnston and Timmerman, 1994b and 1994c; Johnston and Erdmer, in press). The contact between the Aishihik batholith and the "upper package" is interpreted as a sheared intrusive contact and within the batholith there are gneissic banding, annealed mylonite and discontinuous shear bands with an upper member-to-the-west sense of shear. The foliated Aishihik suite and Aishihik batholith have been intruded by undeformed plutons and dyke swarms of the Early Jurassic Long Lake suite (198.7 Ma to 186.4 Ma).

The "lower package" of the Aishihik metamorphic suite is exposed west of, and structurally below the "upper package" and comprises quartzofeldspathic-micaschist, marble, quartzite, metachert and rare metabasite. Unlike the "upper package", these rocks have not been intruded by plutons of the Aishihik plutonic suite. The "lower package" has been correlated with the Nisling assemblage of Wheeler et al. (1991) by Mortensen (1992) and Johnston and Timmerman (1994a). The contact between the "upper and lower packages" is sheared and lithologic layers near the contact may be discontinuous (S.T. Johnston, pers. comm., 1994).

Johnston and Erdmer (in press) suggested that the "lower package" could have formed the depositional basement of the "upper package". Because they considered both "packages" part of a single terrane (Nisling terrane; herein considered part of Tracy Arm terrane), they concluded that the Aishihik suite plutons provided a link between Nisling terrane (Tracy Arm terrane) and Stikinia. Johnston and Timmerman (1993a) have argued that isogradic surfaces that parallel the late Early Jurassic Aishihik batholith form an Early Jurassic link between the "upper and lower packages" of the Aishihik suite on the basis of metamorphic mineral assemblages interpreted as mid-Jurassic in age on the basis of a 184 ± 2 Ma U-Pb date for titanite from the Aishihik batholith "metamorphic aureole". Alternatively, the observed normal isogradic sequence could have been produced if greenschist facies "upper package" rocks were thrust over lower amphibolite facies rocks of the "lower package". Thus, because the nature of the contact between the two packages is ambiguous, and because none of the dykes or plutons of Early Jurassic age that intruded the Aishihik plutonic suite and the

"upper package" also intruded the "lower package", the "lower and upper packages" are herein referred to as separate assemblages. The term "Aishihik assemblage" is retained for the "upper package", and "Nisling assemblage" is considered a reasonable term for the "lower package" because it is exposed in the Nisling Range and conforms to the original definition of Nisling assemblage (Wheeler et al., 1991; Fig. 5.3).

The continental margin-like rocks of Nisling assemblage are herein included in Tracy Arm terrane (Chapter 2) and the largely igneous (arc-like?) Aishihik assemblage is included in Stikinia (Chapter 3), making the sheared contact between them the Stikinia-Tracy Arm terrane boundary, and possibly the upper décollement of a proposed Tracy Arm terrane wedge that was inserted into Stikinia during accretion (Chapter 4). The proposed lower décollement is exposed to the south, in the Tagish Lake area (Chapter 4; Fig. 5.3).

Tagish Lake Area, Northwestern British Columbia

The geology of the Tagish lake area is strikingly similar to that of the Aishihik Lake area. In both areas an arc-like assemblage of metavolcanic and metasedimentary rocks intruded by Early Mississippian felsic orthogneisses and variably deformed Early Jurassic calc-alkaline plutons lies to the east of, and is in sheared contact with, a continental margin-like assemblage. The continental margin-like assemblage in the Tagish Lake area is referred to as the Florence Range suite and is considered part of Tracy Arm terrane (Chapter 2). The Florence Range suite primarily comprises thick, continuous marble and pelitic to semipelitic layers with minor amphibolite and

quartzite layers. It is imbricated with, and lies structurally above a multiply deformed, generally right-way-up sequence of metapelitic and metavolcanic rocks with minor marble termed the Boundary Ranges suite (Chapter 3). The structurally lowest, and presumably oldest rocks in the Tagish Lake area belonging to this suite are Devonian-Mississippian felsic gneisses, Mississippian or older felsic gneisses, locally graphitic metapelitic rocks and marble. They are overlain by Mississippian or older to Permian intermediate to mafic metavolcanic rocks (Currie 1992; Chapter 2). Both the Boundary Ranges suite and the foliated ca. 185 Ma granitic rocks of the Tagish Lake suite that intruded the Boundary Ranges suite are considered part of Stikinia (see Chapter 3, for a detailed correlation of metamorphosed Stikinia rocks from the Tagish Lake area with coeval rocks from Stikinia exposed to the south). The spatial association of the Tracy Arm terrane rocks with the mid-Paleozoic to mid-Mesozoic assemblages of arc volcanic, plutonic and sedimentary rocks of Stikinia is the result of tectonic juxtaposition that occurred 184-176 Ma ago (Chapter 4). Shear sense indicators from near the terrane boundary indicate that Tracy Arm terrane was thrust from the north-northwest (present coordinates) over Stikinia.

The Florence Range suite is considered equivalent to the Nisling assemblage on the basis of lithologic similarity, and the Boundary Ranges suite is correlated with the Aishihik assemblage. The Boundary Ranges suite and the Aishihik assemblage are considered equivalent to one another because they are lithologically similar; both assemblages lie within the same semi-continuous belt of metamorphosed arc-like rocks that is exposed east of Tracy Arm terrane (Chapter 3; Fig. 5.3), and both assemblages

have been intruded by Early Jurassic plutons that are lithologically, isotopically and texturally similar, and essentially coeval (Chapter 3). This belt continues to the south to the Taku River area and the Stikinia type-area near the Stikine River (Chapter 3). There, the Paleozoic metamorphic rocks are referred to as the Stikine assemblage and are exposed along the western margin of Stikinia (Monger, 1977; Brown et al., 1991; Asitka Group of Wheeler and McFeely, 1991, Chapter 3). The present geometry of the western margin of Stikinia may have formed in a manner analogous to that of Wrangellia, in which underplating by accreting terranes may have contributed to the uplift and eventual exposure of the structurally lowest rocks along its western margin.

Taku River area

Metamorphic rocks in the Taku River area that are exposed east of the Llewellyn fault have been included in the Stikine assemblage, and those exposed west of the Llewellyn Fault have been divided into three assemblages; from west to east they are the Whitewater, Boundary Ranges, and Mount Stapler suites (Mihalynuk et al., 1994). The Boundary Ranges suite of the Taku River area comprises metamorphosed sedimentary and volcanic rocks that are regarded as an extension of the Boundary Ranges suite rocks exposed in the Tagish Lake area; like the Boundary Ranges suite to the north, it is herein correlated with the Stikine assemblage and considered part of Stikinia. The Whitewater suite comprises amphibolite facies quartz-rich graphitic schist, quartzite, metabasite and ultramafite, with minor carbonate rock layers that rarely exceed 3 m in thickness and comprise less than 1% of the section,

and felsic orthogneiss bodies. Mihalynuk et al. (1994) have suggested that this suite may be correlative with parts of Yukon-Tanana terrane, but it could also be included in Stikinia because, with the exception of the ultramafite, equivalent rock-types also occur in the structurally lowest rocks of the Boundary Ranges suite in the Tagish Lake area (described above). The Mt. Stapler suite includes rock-types characteristic of the Whitewater and Boundary Ranges suites and is gradational with both assemblages (Mihalynuk et al., 1994) and is herein also considered correlative with the Stikine assemblage.

As suggested by Mihalynuk et al. (1994), the metamorphic suites exposed west of the Llewellyn fault in the Taku River area could be facies equivalents of, or the basement to the lower part of the Stikine assemblage. The proposed correlation between the Whitewater suite and Yukon-Tanana terrane is based only on lithologic similarity, and is equivocal. Tracy Arm terrane does not appear to be present in the Taku River area because all three suites (Whitewater, Boundary Ranges, and Mount Stapler suites) lack the thick carbonate rock layers of Tracy Arm terrane exposed farther north (Florence Range suite and Nisling assemblage) and include significant igneous components that are not found in the Tracy Arm terrane.

Stikine River Area

East of the Stikine River and north of the Iskut River there are layered silicious rocks with quartzites that contain Precambrian and Early to Middle Paleozoic detrital zircons (McClelland and Mattinson, 1991; Carboniferous, W.C. McClelland pers.

comm., 1993). They are associated with Devonian orthogneisses (369 ± 5.1 Ma and 370 ± 6.6 Ma; W.C. McClelland pers. comm., 1993), and are overlain by Early Permian or older metamorphic rocks of the Stikine assemblage of Stikinia (McClelland, 1992). McClelland (1992) interpreted their metamorphosed contact with the overlying volcanic rocks as a stratigraphic contact, and on the basis of this interpretation included the silicious rocks in the Paleozoic Stikine assemblage. He also correlated these rocks with the Paleozoic rocks of Yukon-Tanana terrane on the basis of the Devonian orthogneisses and the Precambrian zircons in the quartzites (Chapter 2).

Geologic relations between layered metamorphic rocks and an Early Jurassic quartz monzodiorite pluton (Bevier and Anderson, 1991) exposed on the west side of the Stikine River (near its confluence with the Iskut River), are very similar to relations between the Aishihik assemblage and the Early Jurassic Aishihik batholith in the Aishihik Lake area; the quartz monzodiorite intrudes felsic gneiss and lesser metapelite, mafic metavolcanic, metaconglomerate of possible volcanic origin, marble, calc-silicate and quartzite of the Elbow Mountain crystalline complex, and crosscuts an earlier foliation in the metamorphic rocks (Porter, 1992). As in the Aishihik Lake area, the most recent metamorphism is considered the result of Early Jurassic plutonism because the garnet-in isograd in the layered rocks is sub-parallel to the pluton's margin (Porter, 1992).

Both the Elbow Mountain crystalline complex and the Early Jurassic plutons from west of the Stikine River are considered part of Stikinia (Porter, 1992; Fig. 5.3).

The terrane affiliation of the quartz-rich metasedimentary rocks from east of the Stikine River is less clear because pre-Early Jurassic ties between the continentally derived sediments and the overlying Stikinia rocks have remained elusive (McClelland and Mattinson, 1991). Their inferred depositional contact with the overlying metavolcanic rocks and the apparent lack of Paleozoic detrital zircons in Tracy Arm terrane quartzites exposed elsewhere argue against including them in Tracy Arm terrane; they are more likely some of the stratigraphically lowest rocks presently exposed in Stikinia (Fig. 5.3).

No bone fide Tracy Arm terrane rocks have been identified in the Stikine River area, but due to the close spatial association of Stikinia and Tracy Arm terrane some confusion has arisen. For example, Smith (1992) included the volcanogenic massive sulfide deposits at Anyox (west-central British Columbia) in Tracy Arm terrane, but considered the geochemistry of the volcanics at Anyox consistent with those of the volcanic-arc assemblages in Stikinia and Alexander terrane. Clearly, the massive sulfide deposit did not form within Tracy Arm terrane.

In summary, there is a semi-continuous, north-northeast trending belt of Paleozoic arc-like rocks of probable Stikinia affinity that extends along the eastern margin of the Coast Mountains from the Aishihik Lake area, to south of the Stikine River. It is in general exposed east of Tracy Arm terrane, but has not been shown to be in stratigraphic contact with Tracy Arm terrane (Fig. 5.3).

Juneau

On the west side of the Coast belt, south of Juneau, Tracy Arm terrane is primarily composed of quartz-rich metasedimentary rocks of the Tracy Arm assemblage (Chapter 2), and may also be preserved as pendants in, and along the western margin of Late Cretaceous to Paleocene granitoids of the Coast belt (Fig. 5.3). The Tracy Arm assemblage is bounded on the west by metamorphic rocks of the Endicott Arm assemblage, which together with the Port Houghton assemblage, forms the hangingwall of the moderately to steeply east-dipping, mid-Cretaceous (?) Sumdum fault (Gehrels et al., 1990). The footwall of the Sumdum fault consists of Taku terrane volcanic, sedimentary and plutonic rocks (Gehrels et al., 1990). The Tracy Arm assemblage comprises amphibolite facies quartz-rich metasedimentary rocks and subordinate metapelite, metabasalt, metarhyolite, and marble that are characterized by an evolved isotopic signature ($\epsilon_{\text{Nd}(0)}$ values are from -26 to 0, and T_{DM} values are from 850 to 2450 Ma; Samson et al., 1991). U-Pb ages for detrital zircons are similar to detrital zircon ages from the western North American miogeocline, the western Canadian shield and Yukon-Tanana terrane in eastern Alaska, and Yukon (Nisutlin subterrane, Gehrels et al., 1991).

The Tracy Arm assemblage strongly resembles the Florence Range suite; it has been included in Tracy Arm terrane (Coney et al., 1980; Chapter 2) and correlated with Yukon-Tanana terrane (presumably the "lower unit" of Mortensen, 1992; Gehrels et al., 1990, 1992a). The Endicott Arm and Port Houghton assemblages more closely resemble rocks of the Nisutlin subterrane and Paleozoic Stikinia and Wrangellia

(Chapter 2). Their relation to the Tracy Arm assemblage is described as gradational and interpreted as depositional by Gehrels et al. (1992a), but could also be tectonic.

Ketchikan, Alaska

South of Juneau, near Ketchikan, lower Paleozoic meta-silicic tuff, quartzite, marble, metabasalt, calc-silicate, pelitic schist, siliceous quartzofeldspathic schist and orthogneiss that were included in both Tracy Arm terrane and Taku terrane by Berg et al. (1988) are referred to as the Kah Shakes sequence (Rubin and Saleeby, 1992; Fig. 5.3). They are lithologically similar to the Florence Range suite and are herein considered part of Tracy Arm terrane (Chapter 2), although they were correlated with Yukon-Tanana terrane by Rubin and Saleeby (1992). The Kah Shakes sequence is structurally interleaved with Alexander terrane, Taku terrane, and upper Jurassic to lower Cretaceous Gravina sequence rocks by mid-Cretaceous thrusts (Rubin and Saleeby, 1992). The Taku terrane rocks include the Alava sequence, which comprises Late Triassic volcanic and carbonate rocks that could represent the westernmost vestige of Stikinia (Rubin and Saleeby, 1991) or part of Wrangellia (Brew and Ford, 1994; Gehrels and Barker, 1994), and upper Pennsylvanian and lower Permian metavolcanic and metasedimentary rocks, which are correlated with Yukon-Tanana terrane by Rubin and Saleeby (1992), but could also be part of Stikinia (Chapter 2).

Skeena River area

South of the Stikine River, the Scotia-Quaal metamorphic belt of the Skeena River area is exposed along the western margin of Stikinia. It is composed of north-northwest trending bands of metavolcanic, metaplutonic and metasedimentary rocks, including a quartzite unit and Devonian orthogneiss units (Gareau, 1991a; 1991b). Polyphase metamorphism and multiple phases of deformation have made it impossible to tell whether the contacts between these lithologic units are depositional or faulted. A 191.7 ± 0.6 Ma orthogneiss has intruded the contacts between all of the major lithologic units, except the quartzite, indicating that they were all together before Middle Jurassic time. Although the Scotia-Quaal metamorphic belt has been correlated with Tracy Arm terrane (Nisling terrane; Gareau, 1991a; 1991b), its metavolcanic and metaplutonic components, especially those of Devonian age, strongly suggest that these metamorphic rocks instead represent the southernmost exposures of the Stikine assemblage (Chapter 2). The minor quartzites may be analogous to those observed in the Stikine River area by McClelland (1992), or could be metacherts, consistent with the slightly negative ϵ_{Nd} values obtained for two quartzite samples (present values are -2.9 ± 0.6 and -4.5 ± 0.6 , and calculated $\epsilon_{Nd(0)}$ values for 385 Ma are 0 and -1.8; Gareau, 1991a).

Bella Coola area

Schist, marble, metapelite and quartzite in the Bella Coola area are separated from fossiliferous Mesozoic volcanic and sedimentary rocks of Stikinia to the east by the Menzies Point shear zone (Rusmore et al., 1992). ϵ_{Nd} values for the metamorphic rocks are variable (-10.3 to +6.6, and $^{87}Sr/^{86}Sr$ ratios are from 0.7042 to 0.7193), and preliminary single grain detrital zircon U-Pb ages are Late Archean and Early Proterozoic (Boghossian et al., 1993). It is not clear whether these rocks are related to the Paleozoic Stikine assemblage or Tracy Arm terrane; they are included in Tracy Arm terrane and correlated with the "lower" Yukon-Tanana terrane by Gehrels et al. (1992b) and considered part of the lower Stikine assemblage by Boghossian et al. (1993). They are bounded along their western margin by a ~83 Ma granodiorite, and west of the granodiorite are lower grade rocks, including a granitoid clast conglomerate with late Silurian detrital zircons, siliciclastics, marble, and metapelite that are considered part of Alexander terrane. Still farther west, a second sliver of quartz-rich metasedimentary rocks is preserved between the Alexander terrane and Wrangellia (Boghossian et al., 1993). Boghossian et al. (1993) have suggested that the western sliver of supracrustal rocks could be either part of the quartz-rich assemblage exposed to the east that has been imbricated with the adjacent Alexander and Wrangellia terrane rocks by strike-slip faults (Stikinia or Tracy Arm terrane?), or an exotic crustal fragment (Tracy Arm terrane?). If the latter interpretation is correct, Tracy Arm terrane could extend as far south as Bella Coola.

SUMMARY: TRACY ARM TERRANE AND ITS RELATION TO ADJACENT TERRANES

As defined here, Tracy Arm terrane is a distinct terrane that has pre-mid-Jurassic stratigraphic and tectonic histories that are different from those of adjacent terranes to the west (Taku, Wrangellia and Alexander terranes) and east (Stikina). Tracy Arm terrane is separated from these terranes by faults, and lacks their characteristic, and volumetrically significant igneous components (Chapters 1 and 2). It is exposed in the western Canadian Cordillera between Bella Coola and Aishihik Lake and includes the western sliver of continental margin-like sediments exposed near Bella Coola, the Kah Shakes sequence, the Tracy Arm assemblage, the Florence Range suite, the Nisling assemblage, and possibly some of the intervening undivided metamorphic rocks. Where these assemblages are spatially associated with Paleozoic volcanic rocks (including Devonian orthogneisses) their contacts with the volcanic rocks have not been demonstrated to be primary. In the Tagish Lake area, the contact between the Tracy Arm terrane rocks and a Paleozoic volcanic assemblage is demonstrably tectonic and formed during the mid-Jurassic. In the western Coast Mountains the contacts between the Tracy Arm terrane and adjacent Paleozoic volcanic rocks could be the result of considerable tectonic mixing caused by middle Cretaceous and/or Paleogene tectonic events. Such deformation may have redeformed what was probably a previously imbricated and metamorphosed composite terrane of Tracy Arm terrane and Stikinia fragments. Consequently, the association between

continental margin-like sequences and Paleozoic volcanic sequences is not considered a primary association, and does not support correlations with Yukon-Tanana terrane.

In contrast, the quartz-rich metasedimentary assemblages in the Stikine and Skeena river areas could be conformably, or unconformably overlain by Paleozoic arc-like assemblages (see Rubin et al., 1990). Quartz-rich metasediments appear to comprise the lowermost, and presumably oldest exposed rocks of Stikinia and are in possible conformable contact with Paleozoic and younger volcanic rocks also belonging to Stikinia (McClelland, 1992). If a mid-Paleozoic arc fringed the western margin of North America (Pangea), as suggested by Miller (1987) and Rubin et al. (1990), such a relationship would not be surprising. Nor would it be surprising if the Yukon assemblage was part of distal North America, over which the Nisutlin subterrane was deposited or thrust. The relation between Yukon assemblage and Nisling assemblage is not clear; they could be proximal equivalents or unrelated. Until this ambiguity is resolved it will be difficult to ascertain whether strata from either side of Stikinia are correlative and possibly contiguous, and which of the tectonic models described in Table 5.1 is correct. In general, however, Tracy Arm terrane lies to the west of the main exposures of Stikinia, with the possible exception of areas where the Stikinia-Tracy Arm terrane boundary has been tectonically reworked.

EVALUATION OF PROPOSED MODELS FOR THE TECTONIC HISTORY OF TRACY ARM TERRANE

The origin and tectonic history of Tracy Arm terrane remains unclear because (1) the origin and age of Tracy Arm terrane strata are unknown, (2) the ages of detrital zircons do not uniquely tie Tracy Arm terrane to a particular craton, (3) assuming Tracy Arm terrane was rifted from a continental margin, the timing of rifting is not known, and (4) correlations between assemblages considered part of Tracy Arm terrane are equivocal because they are separated by mid-Jurassic and younger plutons and structures with unknown or poorly constrained displacements. Despite these limitations, several tectonic models have been proposed for the early tectonic history of Tracy Arm terrane.

In addition to the models presented in Figure 5.2 and Table 5.1, a model which assumes that at least part of the Stikinia arc was built on top of Tracy Arm terrane has also been proposed (e.g. Jackson et al., 1991; McClelland and Mattinson, 1991; Mortensen, 1992). It is not considered here because there is currently no evidence supporting a pre-mid Jurassic link between these two terranes, and three lines of evidence based on observations made in the Tagish Lake area contradict this assumption (Chapters 1 and 4; Currie and Parrish, 1994). First, the absence of Early Jurassic plutons in Tracy Arm terrane (Florence Range suite), in contrast with their abundance in Stikinia, argues strongly against a pre-Early Jurassic juxtaposition of these two terranes (Fig. 5.3). Second, Tracy Arm terrane structurally overlies Stikinia. Finally, the earliest geologic feature shared by both terranes is a transposition fabric

that is interpreted as forming when Tracy Arm terrane was thrust over Stikinia; the transposition fabric affects rocks as young as 180.9 ± 3.1 Ma, indicating that Tracy Arm terrane was accreted to Stikinia after 184 Ma ago (Chapter 4).

All of the models discussed below are considered possible, but on the basis of the available data, all have limitations and there is no compelling evidence that favours one model to the extent that it discounts all other models. These models are evaluated below with regard to the age and geometry of the Stikinia-Tracy Arm terrane boundary (Chapter 4), the viability of inter-terrane correlations (discussed above), and available structural and plate motion data.

Obduction Model

The "orogenic float" model (Oldow et al., 1990; Brown et al., 1993) predicts that terranes will be stripped from their original lower crust and mantle and tectonically emplaced onto the overriding plate as they are accreted, possibly causing more inboard terranes and supracrustal rocks to be transported in the same direction along a lower décollement. The present superficial configuration of terranes in the Cordillera could imply that Stikinia, Cache Creek, Quesnellia and Slide Mountain terranes were thrust over a broad western North American margin, forming a klippe that appears to separate Tracy Arm terrane from contiguous rocks to the east (Table 5.1; Fig. 5.2a), but three lines of evidence argue against this interpretation. First, Tracy Arm terrane rocks cannot be directly linked to ancestral North American strata. At best they can be tenuously correlated with the potentially parautochthonous Yukon

assemblage. Second, Tracy Arm terrane overlay Stikinia in the Tagish lake area. Finally, the inferred continental margin platform depositional setting for Tracy Arm terrane suggests it was deposited on transitional or continental crust (Chapter 2), but the isotopic signatures of most Middle Jurassic and younger (post accretionary) plutons that intrude terranes to the east (Cache Creek and Stikinia) do not reflect the presence of significant amounts of underlying older radiogenic crust. Instead, $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios for Mesozoic and Cenozoic plutons that intrude Stikinia, Cache Creek, Quesnellia and Slide Mountain terranes are in general isotopically unevolved relative to those that intrude terranes to the east (Kootenay and autochthonous North America) and west (Tracy Arm terrane), which tend to have $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios >0.704 (eg. Armstrong, 1988; Mihalynuk et al., 1992; Samson et al., 1991; inset of Fig. 5.1).

The presence of Precambrian zircons in Stikinia rocks has been used to suggest that Tracy Arm terrane (Nisling terrane) underlies Stikinia (Thorkelson et al., in review). Indeed, some Mesozoic plutons and volcanic rocks in Stikinia do contain a component of inherited Precambrian zircon, including plutons as young as Triassic age (the Triassic Zippa Mountain syenite, M.L. Bevier, pers. comm. 1992; the Early Jurassic Spatsizi River Stock and Cold Fish Volcanics, Thorkelson, 1992), but these zircons were not necessarily derived from Tracy Arm terrane; they could have been derived from the quartz-rich metasediments of the lowermost Stikine assemblage, as shown by the detrital zircon study by McClelland and Mattinson (1991).

Enclosure Model

The "enclosure model" (Table 5.1; Fig. 5.2b) of Nelson and Mihalynuk (1993) and Mihalynuk et al. (1994b) was developed primarily to explain the presence of the exotic Tethyan fauna-bearing oceanic rocks of Cache Creek terrane, between Quesnellia and Stikinia, both of which have McCloud-type fauna. Oroclinal folding of a Stikinia-Quesnellia arc about the captured Cache Creek ocean is plausible, but the concomitant counter-clockwise rotation of Tracy Arm terrane envisioned by Nelson and Mihalynuk (1993) is questionable because it considers Tracy Arm terrane an extension of the Yukon assemblage rocks (both are referred to as Yukon-Tanana terrane by Nelson and Mihalynuk, 1993 and Mihalynuk et al., 1994b), and regards these quartz rich metasedimentary rocks as having been adjacent to the Stikinia-Quesnellia arc by Late Triassic time. Tracy Arm terrane rocks cannot be followed continuously and connected with Yukon assemblage rocks, and evidence from the Tagish Lake area indicates that the Tracy Arm terrane was not adjacent to Stikinia until after Early Jurassic time. Alternatively, if east-dipping subduction was initiated along the new western margin of Stikinia at some point during the closure of an oroclinal bend in the Stikinia-Quesnellia arc, convergence (possibly oblique) between Stikinia and a plate to the west could have resulted in the juxtaposition of Tracy Arm terrane and Stikinia during mid-Jurassic time (Fig. 5.2b).

Exotic Model

The absence of a bordering continent, the apparent lack of continuity between Tracy Arm terrane rocks and North American rocks to the east, the fault-bounded nature of Tracy Arm terrane, and its present location between two arc-like terranes, suggests that Tracy Arm terrane could have separated from an adjacent continent and formed a continental ribbon before it was accreted to the western margin of Stikinia (Table 5.1; Fig. 5.2c). Other examples of rifted continental margin fragments that have formed independent entities, such as the Lomonsov Ridge, which rifted from Barents Shelf, the Blake Plateau, and Lord Howe Ridge, which rifted from Australia, are thought to have formed as a result of crustal-scale simple shear during continental separation (Lister et al., 1986). The "exotic model" requires a continental margin assemblage <1.1 Ga old and ultimate sources of zircon that have ages equivalent to detrital zircons found in Tracy Arm terrane (ie. 1.05-1.40 Ga, 1.75-2.00 Ga, ~2.3 Ga, 2.5-2.7 Ga, and ~3 Ga; Gehrels et al., 1991; Appendix 2.1).

Tracy Arm terrane could have been rifted from the western margin of Pangea (North America; see the following discussion of the minimum displacement model), or the land mass that formed the rifted complimentary margin to the western margin of Laurentia during the Neoproterozoic formation of Panthalassa (Table 5.1; Fig. 5.2c). Recent reconstructions for the Neoproterozoic supercontinent place Australia (Bell and Jefferson, 1987), or a composite continental block of East Antarctica and Australia (e.g. Dalziel, 1991; Hoffman, 1991; Moores, 1991; Ross et al., 1992) along the western margin of Laurentia (western North America), and Siberia adjacent to its

northern margin (Hoffman, 1991; Ross et al., 1992; Condie and Rosen, 1994). Tracy Arm terrane could have been rifted from the eastern margin of the Late Proterozoic Central Australian Platform Cover, which has been interpreted as a rifted continental margin because it lacks miogeoclinal strata (Scheibner, 1985). Rifting probably occurred during or soon after the break-up of Laurentia. Post-Cambrian rifting is unlikely because the autochthonous platformal sediments were isolated from the adjacent oceanic crust by Middle Cambrian terranes composed of sediments of deep marine aspect that were thrust over the platform-like margin during Cambrian time (Berry and Crawford, 1988). By mid-Paleozoic time the Central Australian Platform Cover was bounded by a belt of lower Paleozoic terranes of the Tasman orogenic system (Scheibner, 1985; Coney et al., 1990).

U-Pb ages for most detrital zircons from Tracy Arm terrane can be found in the Precambrian crystalline crust of the eastern Australian craton, which is unconformably overlain by the Central Australian Platform Cover. Grenvillian age zircons could have been derived from the Musgrave Inlier and Albany-Fraser Belts (Maboko et al., 1991; Pidgeon, 1990; Fletcher et al., 1991). The ultimate sources of 1.75-1.90 Ga zircons could have been areas affected by the Barramundi (1880-1840 Ma; Page and Williams, 1988; Wyborn, R.W. et al., 1988) and Kimban (1843-1709 Ma; Fanning et al., 1988; Mortimer, et al., 1988) orogenies, including the Mt. Isa, Tennant Creek and Arunta inliers and the Willyama Block. There are crystalline rocks in Australia with ~2.00 Ga and 2.5-2.7 Ga T_{DM} model ages (McCulloch, 1987). The only exposed Archean crust in eastern Australia, from which the ~3 Ga detrital zircons

could be derived is found in the Gawler craton (Fanning et al., 1988; Mortimer, et al., 1988).

Minimum Displacement Model

Rifting of continental margin rocks from the western margin of Pangea or Laurentia could have occurred via crustal-scale simple shear during continental separation, a jump in a spreading center (e.g. Sheridan et al., 1981), or subduction of a spreading center (Murchey, 1994). The time of rifting in the minimum displacement model is not well defined, but is generally considered Late Devonian to Early Mississippian (Tempelman-Kluit, 1979; Morrison et al., 1979; Jackson, 1992). Murchey (1994) has suggested that the Devonian-Mississippian oceanic Golconda, Slide Mountain and Angayucham terranes could have been parts of a single wedge-shaped basin that formed along the western margin of North America due to the subduction of a spreading center, analogous to the formation of the Gulf of California (rifting due to back-arc spreading is considered unlikely because there is no evidence for subduction of old crust). If initial rifting occurred "inboard" of the continental margin, it could have resulted in the formation of the Tracy Arm terrane (Table 5.1; Fig. 5.2d). Alternatively, a fragment of North America may have formed a continental ribbon at another time, possibly during the latest Proterozoic formation of Panthalassa (in contrast to the "exotic model" this model assumes that Tracy Arm terrane is a fragment of North America).

Gehrels et al. (1991) noted the similarity in U-Pb single grain ages for detrital zircons from Tracy Arm terrane and those found in North American strata. Potential source areas for all but the 1.05-1.4 Ga detrital zircons from the Tracy Arm assemblage and Florence Range suite can be found near what was once the western margin of Pangea. The ultimate source of the 1.75-2.00 Ga zircons could have been the Talston magmatic zone, Thelon tectonic zone, Great Slave Lake shear zone (~2.0-1.9 Ga; Hanmer et al., 1992, and references there in), Yavapai-Mazatzal craton (1.67-1.78 Ga), Rimbey Magmatic Arc (1.78-1.85 Ga), and/or Fort Simpson magmatic arc (1.84 to 1.86 Ga); the ~2.3 Ga zircons may be from western Rae province (Bostock and van Breemen, in press), and the 2.5-2.7 Ga and ~3 Ga zircons could have originated in the Wyoming Craton and/or Medicine Hat Block (Ross et al., 1991). The 1.05-1.4 Ga detrital zircons are more problematic; they could have been derived from (1) the Pearya terrane (Grenvillian age rocks exposed on northern Ellesmere island that are thought to have been translated westward along a sinistral strike-slip fault from the Baltic shield to their present position during the Permian; Trettin, 1987), (2) a Grenvillian age Laurentian source, which lay west of the Belt basin, but which has been obscured by post-Belt sedimentation, (3) from Grenvillian plutonic rocks, or (4) sedimentary rocks containing ca. 1.1 Ga detrital zircons. Evidence suggesting that a Grenvillian age orogen may underlie the miogeoclinal strata of western North America includes a fold belt of possible Grenvillian age in the subsurface, east of the Cordillera (Cook, 1988), a 1.2 Ga granitic boulder from the Late Proterozoic Windermere Supergroup was probably derived from a nearby granitic source within Laurentia (Ross

et al., 1992), and a ~1.1 Ga granite clast (U-Pb zircon age) from a diatreme in the northern Cordillera may have been derived from the crystalline basement underlying the sedimentary cover (Jefferson and Parrish, 1989). Approximately 1.1 Ga xenocrystic zircons in diatremes exposed in the Cordillera may be derived directly from Grenvillian age crust, but could also be from reworked detritus from the Early and Late Proterozoic sedimentary rocks (Windermere or Belt-Purcell supergroups; Parrish and Reichenbach; 1991). If ca. 1.1 Ga detrital zircons were derived from the upper part of the Belt-Purcell Supergroup, their ultimate source could have been the Albany-Fraser and Musgrave belts of Australia (Ross et al., 1992). Alternatively, as suggested by Rainbird et al., (1992) for Grenvillian age detrital zircons from a Neoproterozoic sedimentary inlier on Victoria Island in northern Canada (Fig. 5.5), detrital zircons of similar age could have been transported over thousands of km from the Grenville orogen to the western margin of Pangea, by a large river system.

Although a North American origin for Tracy Arm terrane is possible, the "minimum displacement model" is considered unrealistic because it does not allow for the large body of evidence for large, margin-parallel (latitudinal) transportation of terranes both before and after their accretion to the western margin of North America. Plate reconstruction studies suggest that there were sinistral relative plate motions along the western margin of North America at about 170, 140 and 120 Ma, and that relative plate motions were dextral at other times (Engelbreton et al., 1985; D.C. Engelbreton and K. Kelley pers. comm. in Monger et al., 1994), paleomagnetic data from mid-Cretaceous rocks from the Canadian Cordillera east of the Coast Belt

indicate northward displacement of up to 1500 km (e.g. Irving and Wynne, 1990), paleontological studies reflect at least 1800 km northward movement for Stikinia since the Late Pliensbachian (Taylor et al., 1984), and structural studies provide abundant evidence for significant post-mid Cretaceous dextral strike-slip displacement on numerous faults separating the terranes and autochthonous strata (>1000 km; Gabrielse, 1985) and sinistral and dextral movement before mid-Cretaceous time (Avé Lallemant and Oldow, 1988). Of special note, but not accounted for in Fig. 5.2d, are the faunal differences between Cache Creek and adjacent terranes that reflect the juxtaposition of coeval sedimentary rocks that were deposited at different latitudes (Monger and Ross, 1971).

Dextral Strike-Slip Model

Given the right plate configuration, it is possible for continental fragments to travel across oceans as passengers on oceanic plates. Tracy Arm terrane need not have crossed the open ocean, but could have been transported along a continental margin, inboard of a subduction zone. If the oceanic plate that lay to the west of Pangea was subducted at an oblique angle, the convergence vector between the subducting crust and the overlying lithospheric plate may be partitioned into (1) a margin-perpendicular component that is taken up by orthogonal subduction of crust and possibly shortening normal to the subduction zone, and (2) a margin-parallel component that is accommodated by a strike-slip fault that bounds the "inboard" side of the crustal sliver (e.g. Fitch, 1972; Beck, 1983).

Before general acceptance of terrane theory, Jones et al. (1972) suggested that truncated facies trends in Paleozoic rocks of western North America in California represented the locus of rifting of an anomalous continental fragment that was transported northward on dextral strike-slip faults to its present position in southeast Alaska. Although the continental fragment they were referring to was Alexander terrane, which is now considered an arc-like terrane that originated next to Gondwana (Gehrels and Saleeby, 1987), a southwestern Pangea origin for Tracy Arm terrane cannot be ruled out (Table 5.1; Fig. 5.2e). The western North American crystalline provinces described above as possible sources of the Tracy Arm terrane detrital zircons could have also provided detritus to Tracy Arm terrane if it originated south of its present latitude, with the possible exception of the volumetrically minor ~2.3 Ga zircons. The obvious ultimate source of the ca. 1.1 Ga (1.05-1.40 Ga) zircons for detritus deposited next to southern North America are the Grenvillian age rocks now exposed in Mexico (Fig. 5.5), and which border Laurentia on the southeast.

The dextral strike-slip model is supported by the abundant evidence for post-mid-Cretaceous northward movement of terranes that lie to the east of Tracy Arm terrane referred to above (eg. Gabrielse, 1985). However, dextral motion may not apply to the entire tectonic histories of all terranes. Paleomagnetic studies of Permian to Jurassic rocks are indicative of smaller net displacements than those for the younger rocks, leading some workers to suggest that terranes of the Pacific northwest sector of the Cordillera first moved south "en bloc" by ~1000 km or more and then north by as much as 2000 km (Irving and Wynne 1990, and references therein).

Sinistral Strike-Slip Model

Plate reconstruction studies (Engebretson et al., 1985), paleomagnetic data (Irving and Wynne, 1990), the inferred top-to-the-south-southeast sense of shear for the Tracy Arm-Stikinia terrane boundary in the Tagish Lake area and structural studies for other parts of the Cordillera (Avé Lallemant and Oldow, 1988 and references therein) are consistent with clockwise rotation of the oceanic plate (Kula, Izanagi or Farallon plate) that lay to the west of Pangea during earliest Middle Jurassic time, and perhaps earlier (e.g. Avé Lallemant and Oldow, 1993). From these observations it follows that clockwise rotation could have resulted in the formation and sinistral translation (relative to Pangea) of a continental ribbon that lay to the north or west of Tracy Arm terrane's present location (Table 5.1; Fig. 5.2f).

Before pre-mid-Jurassic sinistral translation can be considered, dextral strike-slip displacements on the Tintina-Northern Rocky Mountain trench (~450 km) and Denali fault systems (~350 km; Lanphere, 1977; Eisbacher, 1976) must be restored and Cretaceous basinal deposits results removed, as shown in the pre-100 Ma terrane map of Figure 5.4. Continued restoration of possible terrane trajectories results in the post-sinistral displacement Jurassic view of northern Pangea and Panthalassa shown in Figure 5.5. The Ruby terrane and adjacent Angayucham terrane are restored to positions south of the Arctic Alaska terrane (Fig. 5.4), in accord with Jones et al.'s (1986) conclusion that Ruby terrane has moved substantially with respect to the terrane to the north. Other authors have shown continuity between the Ruby terrane and Arctic Alaska terrane, but Jones et al. (1986) noted that lithological packages of

quite different aspect appear to be juxtaposed along a major terrane boundary between the Arctic Alaska and Ruby terranes. Whether Arctic Alaska terrane was rotated, or translated along a strike-slip fault is still a matter of conjecture (Lawver and Scotese, 1990 and references therein), but in either case, this view suggests that the northern margin of Panthalassa could have fallen on a great or small circle, and sinistral translation of a Eurasian terrane sliver along a transcurrent or oblique convergent margin may have been possible.

The Jurassic restoration shown in Figure 5.5 has a relatively simple terrane configuration along the western margin of Pangea. Parautochthonous terranes (Kootenay, Nisutlin, Coldfoot/Hammond, Ruby) either occur east of, or are tectonically overlain by klippen of oceanic terranes (Slide Mountain, Angayucham, Seventy Mile terranes). Arc-like terranes including Quesnellia, Stikinia, and the Togiak and Koyakuk terranes, and the oceanic Cache Creek terrane are west of these oceanic and parautochthonous terranes. Tracy Arm terrane lies along the western boundary of Stikinia, and is separated from Pangea (North America) and the parautochthonous terranes by oceanic and arc terranes, suggesting that Tracy Arm terrane originated either to the north or northwest of its mid-Jurassic location.

It will be difficult to determine whether Tracy Arm terrane originated in the northwest quadrant of Panthalassa, possibly as part of the Siberian platformal sequence (trajectory #1 in Fig. 5.2f), or the northeast quadrant of Panthalassa, adjacent to Ancestral North America (trajectory #2 in Fig. 5.2f). In fact, division of the Panthalassan margin into northwestern and northeastern quadrants may be academic,

depending on the age of the Tracy Arm terrane sediments, because some strata exposed in northern Alaska (i.e. the Arctic Alaska terrane) are nearly continuous with rocks to the west in Russia (Churkin et al., 1985), and because the paleogeography of Proterozoic to Paleozoic river systems in these areas is unknown. Without paleogeographic constraints, a conservative approach is to assume that the sources of detritus for the Siberian platformal sequence were the crystalline rocks exposed in Eurasia, and the provenance of sediments deposited on the North American margin was the North American craton. Even so, distinguishing between a North American and a Siberian provenance may be formidable because the ages of crystalline basement rocks in both cratons are so similar that a number of authors have discussed possible Siberian and Canadian paleogeographic connections that are predicated on similarities between the ages of crystalline basement rocks in these cratons (Churkin and Whitney, 1983; Sears and Price, 1978; Condie and Rosen, 1994).

Possible North American sources for the Precambrian zircons in Tracy Arm terrane are discussed above with respect to the "minimum displacement model" and possible Eurasian sources are discussed below. The Aldan Shield of Eurasia consists of crystalline rocks of appropriate ages that could have been the provenances of all of the dated detrital zircons from clastic rocks of the Tracy Arm terrane, with the exception of Grenvillian age zircons; they include rocks with ages of 1870-1942 Ma (Nutman et al., 1992), 2.1-2.45 (Glebovitsky and Drugova, 1993 and references therein), 2.5-2.8 Ga (Bibikova and Williams, 1990; Nutman et al., 1992), ~3 Ga (Nutman et al., 1992). To date no evidence of ~1 Ga igneous activity or metamorphism has been identified in

the Siberian craton, but analogous to the North American craton, crust of Grenvillian age may exist beneath the Proterozoic to Mesozoic cover of Siberia. Condie and Rosen (1994) have suggested that the Grenville Orogen may continue from eastern North America into the Angara fold belt of Siberia.

The large magnitude strike-slip displacements required by a Eurasian origin for Tracy Arm terrane are not considered unreasonable because ~3000 km of oceanic crust have been subducted beneath the western margin of North America in the last 180 Ma (Engelbreton et al., 1992). As Tracy Arm terrane migrated southward over this long distance, it may have been attenuated and dispersed along the western margin of Pangea, analogous to the proposed dispersal of Wrangellia (Hillhouse, 1977; Irving and Yule, 1987). Consequently, it would not be surprising to find Tracy Arm-like terranes north of the Tagish Lake area, and possibly north of the Denali fault. Terranes that resemble Tracy Arm terrane include the Nixon Fork, Kilbuck, and Idono terranes of Alaska. The Nixon Fork terrane is composed of stratigraphic sequence of Paleozoic reefal carbonate rocks (Ordovician to Devonian) deposited on Precambrian basement and overlain by Permian, Triassic and Cretaceous fossiliferous sedimentary rocks (Silberling et al., 1984). The basement rocks are mainly pelitic and calc-schists with minor marble, quartzite, felsic metavolcanic rocks. Patton and Dutro (1979) interpret this terrane as a rifted continental margin sequence derived from Western North America (Canadian Cordillera) because the transgressions and regressions are coeval with those of the North American stratigraphic record. On the basis of paleomagnetic data that suggests the Nixon Fork terrane came from the south during Permian time,

and was rotated in a clockwise manor (Plumley and Coe, 1983; Coe et al., 1985), Plafker (1990) argued against strike-slip faults causing northward translation of the Nixon Fork terrane; the same data could point toward a more southerly Eurasian origin for the Nixon Fork terrane.

The Kilbuck terrane comprises gneiss and schist exposed in western Alaska (Fig. 5.1), including biotite-hornblende gneiss, garnet amphibolite, quartz-mica schist, marble and ca. 2.0 Ga orthogneisses (Box et al., 1990; Kanektok metamorphic complex of Hoare and Coonrad, 1979). Radiogenic ages for these rocks are indicative of Precambrian metamorphism (Hoare and Coonrad, 1979). A possible outlier of the Kilbuck terrane is the Idono complex of Miller and Bundtzen (1985), which is exposed to the west of the Nixon Fork terrane and has yielded zircons with U-Pb ages of 2.03 Ma. Box et al., (1990) did not consider the ca. 2.0 Ga provinces in western North America probable sources of detritus for the Kilbuck terrane because unlike the Kilbuck terrane rocks, the North American rocks have negative ϵ_{Nd} values and/or are associated with 1.8 Ga and younger plutonism. Instead they suggested that the possible provenances of appropriate age for the ~2.0 Ga orthogneisses of the Kilbuck terrane are Australia, the Baltic Shield of Finland, the Guiana and west African shields flanking the southern Atlantic Ocean, and the "poorly dated Siberian craton". The ca. 2.0 Ga provinces that have since been documented in the Aldan shield of the Siberian craton (e.g. Nutman et al., 1992; Glebovitsky and Drugova, 1993 and references therein) provide a plausible alternative source for the basement rocks of the Kilbuck terrane.

CONCLUSIONS

On the basis of available data it is not possible to unequivocally determine Tracy Arm terrane's origin and pre-accretionary (pre-mid Jurassic) history. Potential origins include western North America (Pangea or Laurentia), either north or south of its present location, the Siberian craton, or Australia (Gondwana). Because Tracy Arm terrane cannot be unequivocally correlated with parautochthonous North American rocks, proposed correlations between Tracy Arm terrane and the Yukon assemblage (Yukon-Tanana terrane) cannot support the "obduction and enclosure" models. Neither the "exotic" nor the "dextral strike-slip" model can be dismissed, but evidence for pre-Middle Jurassic sinistral motion points to the "sinistral strike-slip" model. If the "sinistral strike-slip" model is correct, it will still be difficult to determine whether Tracy Arm terrane originated next to Eurasia or North America without better characterization of the source areas (most notably the possibility of Grenvillian age rocks in the crystalline basements of both areas), and the age of Tracy Arm terrane.

Figure 5.1. Terrane map of British Columbia, Yukon and Alaska (modified from Plafker, 1990) showing the location of the Tagish Lake area. Van = Vancouver; Anc = Anchorage. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio contours are shown in the inset (modified from Armstrong, 1988).

(a) OBDUCTION MODEL

(b) ENCLOSURE MODEL

Early Jurassic (208-187 Ma)

(c) EXOTIC MODEL

(d) MINIMUM DISPLACEMENT MODEL

(e) DEXTRAL STRIKE-SLIP MODEL

(f) SINISTRAL STRIKE-SLIP MODEL

Figure 5.2. Cartoons for the six proposed models for the origin of Tracy Arm terrane. Q = Quesnellia; S.L.= Sea Level; ST = Stikinia. Models are modified from (a) Coney, 1989; (b) Nelson and Mihalynuk, 1993; (d) Tempelman-Kluit, 1979; (e) Jones et al, 1972. See Table 5.1 a descriptions of each model. ① and ② in Fig. 5.2f are possible trajectories for Tracy Arm terrane

Figure 5.3. Terrane map of northern British Columbia and Yukon (modified from Wheeler and McFeely, 1991) with schematic sections showing proposed terminology and terrane affiliation for metamorphic rocks. Terrane and assemblage names used in this paper are shown in bold. (1)Johnston and Timmerman (1994a, b, c); (2)Mortensen (1992); (3)Wheeler et al. (1991); (4)Coney et al., 1980; (5)Mihalynuk et al., 1994; (6)Gehrels et al., (1992); (7)Brown et al. (1991); Anderson (1993); Porter (1993); (8)Rubin and Saleeby (1991; 1992); (9)Mortensen (1990a); (10)Chapter 2. *from the Cleary Sequence, north of Fairbanks, Alaska; from the Nasina quartzite from west of Dawson, Yukon. AMS = Aishihik metamorphic suite; B.C. = British Columbia; FFZ = Finlayson Fault Zone; LFZ = Llewellyn fault zone; SF = Sumdum fault; YT = Yukon; YTT = Yukon-Tanana terrane.

Figure 5.4. Terrane map of the northern Cordillera for mid-Cretaceous time. Cretaceous basinal deposits have been removed and displacements on the Tintina-Northern Rocky Mountain Trench (~450 km) and Denali (up to 350 km) fault systems have been resored, but Arctic Alaska terrane has not been restored to its original position.

Figure 5.5. Jurassic view of northern Pangea and Panthalassa (present continental margins are shown for ease of recognition). Restoration of assumed trajectories of allochthonous terranes places these terranes farther south than shown in Fig. 5.4. The geology of Precambrian North America is modified from Hoffman (1989) and Ross et al. (1991), and that of Eurasia is modified from Zonenshain et al. (1990). Present continental margins are shown for ease of recognition.

TABLE 5.1. MODELS PROPOSED FOR THE ORIGIN AND TECTONIC HISTORY OF TRACY ARM TERRANE

MODELS THAT REQUIRE CORRELATION BETWEEN TRACY ARM TERRANE AND PARAUTOCHTHONOUS NORTH AMERICAN TERRANES

- (A) *Obduction Model* Arc-like terranes that formed along the western margin of Pangea were thrust over a broad stable continental margin of present North American affinity and subsequently folded to form klippen that in map-view appear to separate Tracy Arm terrane from parautochthonous rocks to the east (Coney, 1989; Gehrels et al., 1990; 1992a; Hansen, 1990; Fig. 5.2a).
- (B) *Enclosure Model* Quesnellia and Stikinia formed an essentially continuous arc west of terranes that are parautochthonous to North America, until anticlockwise rotation of Stikinia resulted in oroclinal enclosure of the exotic Cache Creek ocean (Fig. 5.2b). Concomitant rifting inboard of some of the continentally derived sediments deposited along the northern margin of western North America resulted in the formation of parautochthonous terranes (Yukon-Tanana terrane), and large-scale anticlockwise rotation placed the northwestern-most of the parautochthonous rocks (Tracy Arm terrane) on the west side of the oceanic and arc-like terranes (Cache Creek terrane, Quesnellia and Stikinia; Nelson and Mihalynuk, 1993; Fig. 5.2b).

MODELS THAT DO NOT REQUIRE CORRELATION BETWEEN TRACY ARM TERRANE AND PARAUTOCHTHONOUS NORTH AMERICAN TERRANES

- (C) *Exotic Model* During, or soon after the Neoproterozoic break-up of Laurentia and formation of Panthalassa, a continental ribbon (Tracy Arm terrane) was rifted from the trailing edge of the land mass (Australia?) that formed the rifted complimentary margin to the western margin of Pangea (this paper; Fig. 5.2c).
- (D) *Minimum Displacement Model* After a crustal fragment was rifted from the western margin of ancestral North America, possibly during the Late Devonian or Early Mississippian, ocean basins and island arcs formed between the newly formed continental ribbon (Tracy Arm terrane) and the continental margin. In Late Triassic to Middle Jurassic time, Tracy Arm terrane and the arc and oceanic rocks were amalgamated and accreted to the western margin of North America (Tempelman-Kluit, 1979; Morrison et al., 1979; Jackson, 1992; Fig. 5.2d).
- (E) *Dextral Strike-Slip Model* Tracy Arm terrane is a fragment of the central-western or southwestern margin of Pangea that was rifted from the margin and then transported along dextral strike-slip faults to its present position, west of arc-like and oceanic terranes (e.g. Jones et al., 1972; Fig. 5.2e).
- (F) *Sinistral Strike-Slip Model* Tracy Arm terrane originated next to either Eurasia (the Siberian Platform) or the northern margin of western Pangea (possibly Arctic Alaska terrane), and was transported to a location west of Stikinia by Jurassic or older sinistral transcurrent faults (this paper; Fig. 5.2f).

APPENDIX A: U-Pb GEOCHRONOLOGY OF CRETACEOUS AND TERTIARY PLUTONIC ROCKS OF THE TAGISH LAKE AREA, NORTHEASTERN COAST MOUNTAINS, BRITISH COLUMBIA

ABSTRACT

U-Pb age determinations for Cretaceous and Tertiary plutons that intrude metamorphic rocks of Tracy Arm terrane and Stikinia along the eastern margin of the Coast Plutonic Complex in northwestern British Columbia provide age constraints for the structural history of these terranes. The Stikinia-Tracy Arm terrane boundary was broadly folded and offset by steep north-trending faults after 185 Ma and before the intrusion of a 127.2 ± 0.6 Ma granite. Deformation fabrics in a 101.6 ± 0.6 Ma granite are interpreted as the result of mid-Cretaceous or younger motion on the Llewellyn fault, which juxtaposes the metamorphic rocks of Stikinia and Tracy Arm terrane against unmetamorphosed rocks of Stikinia to the east. To the northwest a hornblende granodiorite, thought to be part of the same plutonic complex, is 103 ± 0.7 Ma. A 55.9 ± 0.2 Ma age for an massive diorite tentatively constrains the age of the Mt. Switzer volcanic suite to Eocene or older. Coeval with the diorite is a quartz monzonite (55.7 ± 0.2 Ma) that is exposed 7 km to the east.

GEOLOGICAL FRAMEWORK

Cretaceous and Tertiary intrusive rocks in the Tagish Lake area lie on the eastern margin of the Coast Plutonic Complex of northeastern British Columbia, where they intrude metamorphic rocks of Tracy Arm terrane and Stikinia (Currie, 1991,

1992b; Chapter 1; Fig. A.1). The Tracy Arm terrane is a narrow terrane interpreted by Wheeler et al. (1991) as representing a continental margin setting. It is separated from continental margin rocks of ancestral North America by oceanic rocks of Slide Mountain and Cache Creek terranes, and by primitive arc rocks of Stikinia and Quesnellia south of latitude 61°N (Fig. A.1). How continental margin rocks of Tracy Arm terrane came to be located outboard of oceanic and arc-like terranes remains a mystery. However, U-Pb age determinations for plutons from the Tagish Lake area constrain the structural history of Tracy Arm terrane, making it possible to begin to compare the complex deformational history of this terrane with tectonic events that affected other parts of the Cordillera.

In the Tagish Lake area metamorphic rocks from both Stikinia and Tracy Arm terrane are polydeformed. Deformation fabrics that formed prior to the juxtaposition of the two terranes were transposed during the amalgamation of the two terranes along a sinistral transpressive ductile shear zone between 185 Ma and 170 Ma. This occurred during or before the accretion of Stikinia to the ancestral margin of North America (Chapter 1). The resulting terrane boundary, referred to as the Wann River shear zone, and the associated penetrative ductile fabrics have been folded into broad open folds that are reflected in the map pattern (Fig. A.1). Later faulting offset the terrane boundary, and was followed by intrusion of Cretaceous to Tertiary plutons. By some time in the Early Tertiary the metamorphic rocks were exposed and unconformably overlain by Mt. Switzer and Sloko Group volcanic rocks (Christie, 1957; Mihalynuk et al., 1989; 1990; Currie, 1990a; 1991; 1992b; Fig. A.1).

The metamorphic rocks of Stikinia and Tracy Arm terrane are presently juxtaposed against unmetamorphosed volcanic and sedimentary rocks of Stikinia along the Llewellyn Fault (Fig. A.1), for which the magnitude, age and sense of movement are poorly constrained. Some west-side-down motion occurred on the fault after the deposition of Early Tertiary Sloko Group volcanic rocks; they are preserved at elevations of 1200m west of the fault, but are not preserved east of the fault where peaks up to 1981m are underlain by Triassic volcanic rocks (Fig. A.1). However, the magnitude of offset may be less than the difference between these elevations (780m), since the volcanics may have been deposited in an area of uneven topography (see Souther, 1971). In contrast, based on the juxtaposition of amphibolite facies metamorphic rocks west of the fault against unmetamorphosed rocks on the east, the net offset on the Llewellyn Fault has a component of west-side-up movement. The contrasting senses of offset suggest that movement on the Llewellyn Fault has been variable and long-lived. It is likely that the Llewellyn fault was also the locus of some strike-slip motion, but there are presently no reliable data to constrain either the magnitude or sense of this offset.

Five U-Pb age determinations for plutons from the Tagish Lake area are presented. Three massive plutons (labeled A to C in Fig. A.1) were collected to obtain a youngest possible age for deformation that affected the metamorphosed rocks, and samples from massive and deformed parts of a fourth plutonic complex were collected to determine the oldest age limit for at least some of the deformation associated with movement on the Llewellyn Fault (labeled D and E, respectively in Fig. A.1).

ANALYTICAL METHODS

U-Pb analyses of zircon fractions were performed at the Geological Survey of Canada in Ottawa using analytical procedures summarized by Parrish et al. (1987). Blanks for U are 1 picogram or less, and blanks for Pb range from 7-14 picograms. The clearest crystals were analyzed and all zircon fractions were strongly abraded prior to dissolution (Krogh, 1982). Analytical results are presented in Table A.1.

Some of the analyses give discordant results. Three possible causes of discordance are: Pb-loss; ^{206}Pb deficiency; and inheritance of a component of older zircon. It is unlikely that Pb-loss is significant in these samples because they were all strongly abraded, and the U concentration of all samples except LC-89-35c is low to moderate (Table A.1). The ^{206}Pb deficiency is caused by the exclusion of ^{230}Th from zircon during crystallization (Mattinson, 1973). Since ^{230}Th is an intermediate daughter product of the ^{238}U - ^{206}Pb decay chain (half life = $7.52 \times 10^4\text{Ma}$), the exclusion of ^{230}Th will result in a deficit of ^{206}Pb , and the $^{206}\text{Pb}/^{238}\text{U}$ age will be slightly less than the $^{207}\text{Pb}/^{235}\text{U}$ age. ^{206}Pb deficiency has been corrected in all analyses presented here using the method outlined in Schärer (1984); this method has been successfully applied to zircon analyses (Coleman and Parrish, 1991). Inheritance of an older component of zircon has affected some of the analyses, and is discussed below where appropriate.

RESULTS AND INTERPRETATIONS

Granite intruding fabrics in metamorphic rocks of Stikinia (A).

The oldest dated pluton that is apparently unaffected by deformation that folded the Stikinia-Tracy Arm terrane boundary (Wann River shear zone) is a pink granite with chilled margins (labelled A in Fig. A.1); it also intrudes the metamorphic rocks of Tracy Arm terrane and Stikinia (Currie, 1992b), steep north-trending faults, and the shear fabric associated with the Wann River shear zone. Although this granite is shown in Fig. A.1 as a continuous body, it may contain phases of other lithologies and/or different ages (eg. see Mihalynuk et al, 1990).

U-Pb isotopic data for two zircon fractions (A and B) from the granite are concordant after correction for ^{206}Pb deficiency (Table A.1, Fig. A.2a; Schärer, 1984), and overlap concordia between 126.4 and 127.6 Ma. It is therefore inferred that 127 ± 0.6 Ma is the crystallization age of the granite. Two additional fractions are discordant, primarily due to a component of inherited zircon. Some of the inherited zircon likely comes from the Early Jurassic Hale Mountain granodiorite (Currie, 1991) which is intruded by the granite (included in Stikinia in Fig A.1); the error ellipse for fraction E plots near a mixing line with a lower intercept of 127 Ma (the age of the granite) and an upper intercept of 185 Ma (the age of the Hale Mountain granodiorite). However, some of the inherited zircon must be older than Jurassic since both fractions C and E plot to the right of this mixing line.

The Early Cretaceous age for the granite provides a youngest possible age for the steep north-trending faults and the deformation that folded the Stikinia-Tracy Arm terrane boundary and control the map pattern in the Tagish Lake area.

Hornblende diorite on the north side of Mt. Switzer (B)

The hornblende diorite body on the north side of Mt. Switzer (labelled B in Fig A.1) intrudes previously deformed Jurassic plutonic rocks of Stikinia and the 127 ± 0.6 Ma granite (Fig. A.1). Although the contact relationship between the diorite and the volcanic rocks was not observed, it is inferred that the diorite also intruded the Mt. Switzer volcanic suite because adjacent to the diorite body, conglomerates in the Mt. Switzer volcanic suite contain clasts of amphibolite, biotite schist, quartzite, granite and volcanic rocks, but lack diorite clasts. All of the lithologies represented in the conglomerate are exposed near the conglomerate; the granite clasts are likely derived from the 127 ± 0.6 Ma granite that crops out to the west, the source of the metamorphic rocks is thought to be the Tracy Arm terrane which also lies to the west, and the volcanic rocks are assumed to be derived from the Mt. Switzer volcanic suite. The lack of diorite clasts suggests that the diorite was not present at the time the conglomerate was deposited, and is therefore younger than the Mt. Switzer volcanic suite.

U-Pb analyses for two fractions of large, clear, euhedral zircon with minor inclusions from the diorite body are concordant, overlapping concordia at 55.9 ± 0.2 Ma (Table A.1; Fig. A.2b). The correction for ^{206}Pb deficiency (Table A.1, Fig. A.2b;

Schärer, 1984) increases the $^{206}\text{Pb}/^{238}\text{U}$ age of these analyses by about 0.1 Ma. The interpreted age of crystallization for the diorite is 55.9 ± 0.2 Ma.

The Eocene age for the diorite provides a probable minimum age for the Mt. Switzer volcanic rocks.

Pink biotite quartz monzonite (C)

A small pink biotite quartz monzonite body (labelled C in Fig. A.1) intrudes the Wann River shear zone on the north side of the upper Wann River, where Permian gneiss of Stikinia (Currie, 1992b) and metasedimentary rocks of Tracy Arm terrane (Fig. A.1) are juxtaposed. This shear zone is thought to have been active between 185 and 170 Ma (Chapter 1).

Zircons from the quartz monzonite are clear, euhedral and contain inclusions. U-Pb isotopic data for three zircon fractions (A, B, and F) are nearly concordant to concordant and have $^{206}\text{Pb}/^{238}\text{U}$ ages of 55.7 ± 0.3 Ma, before correction for ^{206}Pb deficiency and from 55.7 to 55.8 Ma after the correction is made (Table A.1; Schärer, 1984; Fig. A.2c). Fraction (A) also has a $^{207}\text{Pb}/^{235}\text{U}$ age of 55.7 ± 0.2 Ma and its error ellipse overlaps concordia at 55.7 ± 0.2 Ma. The consistency of the results is evidence for a crystallization age of 55.7 ± 0.2 Ma. Three additional fractions plot to the right of concordia, primarily due to a component of older inherited zircon. Error ellipses for these fractions overlap mixing lines with lower intercepts of 55.7 Ma (the age of the quartz monzonite) and upper intercepts of 129 Ma (the age of a pluton that crops out west of the quartz monzonite) and 185 Ma (the age of the Hale Mountain granodiorite

which crops out north of the quartz monzonite and is included in Stikinia in Fig. A.1). The inherited zircon component is therefore probably Jurassic to Cretaceous in age.

The age of the quartz monzonite is within error of that of the 55.9 ± 0.2 Ma diorite exposed 5 km to the west, suggesting that these plutons may be genetically related.

Massive granodiorite from north of Mt. Caplice (D)

An massive biotite-hornblende granodiorite (labelled D in Fig. A.1) from north of Mt. Caplice and 6km west of the Llewellyn Fault intrudes metamorphic rocks of the Tracy Arm terrane, Stikinia, and the tectonic boundary between them (Wann River shear zone; Fig. A.1).

Pale yellow, euhedral zircons with inclusions have concordant U-Pb analyses that overlap concordia between 103.1 and 103.7 Ma, after correction for ^{206}Pb deficiency (Table A.1; Fig. A.3a). The error ellipse for a third fraction (C) is discordant, likely due to a component of inherited zircon, of probable Cretaceous age. If fractions A and B also contain a component of inherited zircon, the crystallization age of the granodiorite might be slightly younger than 103.1 Ma, and therefore the age of the granodiorite is interpreted as 103 ± 0.7 Ma.

Deformed granite from the west side of the Llewellyn Fault (F)

A moderately altered granite adjacent to and west of the Llewellyn Fault (labelled E in Fig. A.1) is poorly indurated with brittle fabrics that dip steeply toward

the west. Unlike the massive plutons, in which quartz grains have straight extinction and lack subgrains, quartz grains from this granite sample have well developed serrated subgrain boundaries. It is inferred that these deformation fabrics post-date crystallization of the granite and are due to movement on the Llewellyn Fault.

Two of three analyses for zircon fractions for this sample are nearly concordant (A and B; Fig. A.3b), and after correction for ^{206}Pb deficiency one fraction (B) touches concordia at 102.2 Ma. The third fraction is highly discordant, primarily due to a component of inheritance, some of which may be older than 1.1 Ga. The data for these three zircon fractions yield a regression line with a lower intercept of 101.6 ± 0.3 Ma (MSWD = 1.63), younger than 102.2 Ma, the age at which fraction B touches concordia. It is not possible to determine whether the younger age for the lower intercept is due to Pb-loss or whether fraction B contains a component of inherited zircon and the contact between its error ellipse and concordia is fortuitous. Consequently, the interpreted age for the granite includes both ages, and is 101.6 ± 0.6 Ma.

The age of the granite (101.6 ± 0.6 Ma) and is close to the age of the granodiorite to the west (103 ± 0.7 Ma), suggesting that these rocks may be phases of the same pluton. The Early Cretaceous age for this deformed granite indicates that some of the movement on the Llewellyn Fault occurred after 101.6 ± 0.6 Ma.

CONCLUSIONS

The main conclusions, based on the U-Pb age determinations presented here are:

1. Broad folds which control the pattern of map units and fold the Stikinia-Tracy Arm terrane boundary in the Tagish Lake area formed after 185 Ma and before 128 Ma.
2. At least some of the deformation associated with movement on the Llewellyn fault occurred after 102 Ma.
3. The age of the Mt. Switzer volcanic suite is tentatively inferred as 56 Ma or older.

Figure A.1. Geologic map of the Tagish Lake area showing locations of samples discussed in text: A, LC-89-35c; B, LC-89-75c; C, LC-89-175c; D, LC-90-359; E, LC-89-261c. Geology is based on work by Mihalynuk et al. (1989a, b, 1990) and Currie (1990a, 1991, 1992a, b). The Wann River shear zone (WRSZ) forms the boundary between Tracy Arm terrane and Stikinia (Chapters 1 and 4). Inset shows location of Tracy Arm terrane and adjacent terranes (modified from Wheeler et al., 1991). Location of Tracy Arm terrane in southeast Alaska is based on presence of quartz-rich metaclastic rocks, some of which are known to contain detrital zircon of Proterozoic age. Ak - Alaska, BC - British Columbia, YT - Yukon Territory.

Figure A.2a. U-Pb concordia diagram for granite (Sample LC-89-35c) from the south end of Tagish Lake. Data have been corrected for ^{206}Pb deficiency (Table A.1).

Figure A.2b. U-Pb concordia diagram for diorite (Sample LC-89-75c) from southwest of Mt. Switzer. Data have been corrected for ^{206}Pb deficiency (Table A.1).

Figure A.2c. U-Pb concordia diagram for quartz monzonite from north of the upper Wann River (Sample LC-89-175c). Data have been corrected for ^{206}Pb deficiency (Table A.1).

Figure A.3a. U-Pb concordia diagram for massive granodiorite from north of Mt. Caplice (Sample LC-90-359). Data have been corrected for ^{206}Pb deficiency (Table A.1).

Figure A.3b. U-Pb concordia diagram for deformed granite from adjacent to the Llewellyn Fault (Sample LC-90-261c). Data have been corrected for ^{206}Pb deficiency (Table A.1).

Table A.1. U-Pb data for zircon fractions from massive plutons in the Tagish Lake area.

Fraction size ¹	Wt. μ g	U ppm	Pb ² ppm	206Pb/204Pb	Pbc ⁴ pg	208Pb %	206Pb ⁵ 238U	207Pb ⁵ 235U	Corr. Coeff.	207Pb ⁶ 206Pb	207Pb ⁷ 206Pb	207/206 Age ⁶ age (Ma)	207/206 Age ⁷ age (Ma)
(A) LC-89-31 Undeformed, pink, medium-grained muscovite-biotite ($\pm$ garnet) granite; locally contains orthoclase phenocrysts up to 2 cm in diameter. UTM 8v, 538130E 6586250N.													
A +149	137	2216	42.3	4558	84	6.1	0.01988 $\pm$ 0.21%	0.1333 $\pm$ 0.22%	0.98	0.04864 $\pm$ 0.04%	0.04860	130.4 $\pm$ 2.1	128.9
B +149	89	2313	43.8	2118	124	3.0	0.01993 $\pm$ 0.12%	0.1338 $\pm$ 0.15%	0.88	0.04868 $\pm$ 0.07%	0.04864	132.7 $\pm$ 3.4	131.0
C +149	151	2389	42.8	7948	59	3.2	0.02059 $\pm$ 0.16%	0.1432 $\pm$ 0.17%	0.98	0.05044 $\pm$ 0.04%	0.05040	215.2 $\pm$ 1.7	213.6
E +149	22	991	20.6	859	38	6.3	0.02163 $\pm$ 0.10%	0.1466 $\pm$ 0.23%	0.67	0.04916 $\pm$ 0.18%	0.04914	155.8 $\pm$ 8.5	154.5
(B) LC-89-175c Undeformed fine- to medium-grained pink quartz monzonite from southeast of Mt. Switzer. UTM 8v, 534260E 6575550N.													
A +149	28	302	2.66	664	7	11.2	0.00868 $\pm$ 0.14%	0.0564 $\pm$ 0.50%	0.58	0.04711 $\pm$ 0.43%	0.04707	55.0 $\pm$ 20	53.1
B +149	35	374	3.59	374	20	11.7	0.00869 $\pm$ 0.15%	0.0572 $\pm$ 0.69%	0.64	0.04772 $\pm$ 0.61%	0.04767	85.4 $\pm$ 27	83.3
C +149	76	324	3.70	508	33	11.4	0.01038 $\pm$ 0.15%	0.0683 $\pm$ 0.40%	0.59	0.04772 $\pm$ 0.34%	0.04765	85.4 $\pm$ 16	82.0
D +149	161	345	3.10	725	44	11.8	0.00879 $\pm$ 0.11%	0.0574 $\pm$ 0.27%	0.67	0.04734 $\pm$ 0.21%	0.04730	66.6 $\pm$ 10	64.6
E +149	165	332	3.08	1214	26	12.1	0.00908 $\pm$ 0.17%	0.0594 $\pm$ 0.27%	0.62	0.04751 $\pm$ 0.21%	0.04734	74.7 $\pm$ 10	70.9
F +149	110	337	2.99	523	40	11.6	0.00869 $\pm$ 0.14%	0.0569 $\pm$ 0.38%	0.63	0.04749 $\pm$ 0.31%	0.04745	73.8 $\pm$ 15	71.8
(C) LC-89-75c Undeformed medium-grained hornblende diorite. UTM 8v, 537250E 657890N.													
A +149	406	311	2.91	511	142	16.3	0.00871 $\pm$ 0.16%	0.0566 $\pm$ 0.36%	0.70	0.04714 $\pm$ 0.27%	0.04712	56.2 $\pm$ 13.0	55.5
B +149	352	291	2.72	656	89	16.3	0.00870 $\pm$ 0.13%	0.0566 $\pm$ 0.28%	0.71	0.04724 $\pm$ 0.21%	0.04723	61.5 $\pm$ 9.9	60.8
(D) LC-90-359 Undeformed medium-grained biotite-hornblende granodiorite from Mt. Caplice. UTM 8v, 544950E 6564725N.													
A +149	52	310	5.40	1006	16	16.3	0.01615 $\pm$ 0.14%	0.1074 $\pm$ 0.32%	0.57	0.04823 $\pm$ 0.26%	0.04820	110.7 $\pm$ 12.5	108.9
B +149	148	301	5.23	1584	29	16.1	0.01619 $\pm$ 0.10%	0.1077 $\pm$ 0.19%	0.68	0.04824 $\pm$ 0.14%	0.04820	111.1 $\pm$ 6.8	109.2
C +149	162	279	4.92	1186	40	16.6	0.01632 $\pm$ 0.10%	0.1087 $\pm$ 0.19%	0.72	0.04828 $\pm$ 0.13%	0.04824	113.1 $\pm$ 6.3	111.3
(E) LC-90-261C Deformed medium- to coarse-grained granite from adjacent to the Llewellyn fault. UTM 8v, 551075E 6561325N.													
A 105-149	161	562	9.76	2340	39	17.2	0.01592 $\pm$ 0.10%	0.1060 $\pm$ 0.14%	0.82	0.04826 $\pm$ 0.08%	0.04822	112.2 $\pm$ 3.9	110.3
B 105-149	123	427	7.41	1242	43	16.9	0.01596 $\pm$ 0.10%	0.1062 $\pm$ 0.20%	0.67	0.04825 $\pm$ 0.15%	0.04821	111.6 $\pm$ 7.2	109.6
C 105-149	182	353	6.41	948	73	16.6	0.01676 $\pm$ 0.11%	0.1161 $\pm$ 0.21%	0.69	0.05025 $\pm$ 0.16%	0.05021	206.7 $\pm$ 7.3	204.0

Notes: ¹sizes (i.e. +105) refer to length aspect of crystals in microns; ²radiogenic Pb; ³measured ratio, corrected for spike and fractionation; ⁴total common Pb in analysis corrected for spike and fractionation; ⁵corrected for blank Pb and U, common Pb, errors quoted are one sigma in percent; ⁶corrected for blank and common Pb, errors are two sigma in Ma; decay constants are those of Steiger and Jäger (1977); for analytical details see Parrish et al. (1987); ⁷corrected for initial Th exclusion, assuming whole rock Th/U=3/1 (see Schärer, 1984).

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LEGEND
LAYERED ROCKS

Tertiary
TSG SLOKO GROUP: basaltic through rhyolitic flows, tuffs, and epiclastic rocks

STIKINIA
Triassic
URSG STUHINI GROUP: argillites, greywackes, volcanic lith-arenites, conglomerates, limestone, pyroxene-phyric flows, tuffs

Probable Paleozoic
BOUNDARY RANGES SUITE
IPBR mainly actinolite schist and associated felsic and marble layers; some felsic orthogneisses are Mississippian in age; may also include leucogabbro; actinolite schist may contain hornblende-, pyroxene-, or feldspar porphyroclasts, and felsic layers may contain feldspar porphyroclasts; layered rocks are at least in part Mississippian or older (Chapter 3)

Permian
PBRW WANN RIVER GNEISS: hornblende-plagioclase gneiss; 270 ±5 Ma (Chapter 3)

TRACY ARM TERRANE
Age uncertain (>176 Ma)
PPFR FLORENCE RANGE SUITE: undivided; biotite schist, muscovite-biotite schist, semi-pelitic rocks, quartzite, marble, calc-silicate, and hornblende amphibolite, with rare felsic gneiss layers and possible metachert
or c marble
q quartzite

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Notes: ¹ All contacts between the Florence Range suite and the Boundary Ranges suite are interpreted as tectonic contacts. CF = Caplice Fault. WCFZ = Williston Creek fault zone.

MASSIVE INTRUSIVE ROCKS
Probable Eocene
Ehd hornblende diorite
Egm quartz monzonite
Probable Mesozoic or Tertiary
MTg biotite granite and granitic to tonalitic dykes

Cretaceous
mKg hornblende-biotite granite, locally granodioritic in composition and brittily fractured near the Llewellyn fault zone (101.6 ±0.6 Ma; Appendix A)

FOLIATED INTRUSIVE ROCKS
STIKINIA
Jurassic
TAGISH LAKE SUITE
LJRMcm Mt. Caplice quartz monzodiorite: lineated and foliated hornblende quartz monzodiorite; 180.9 ±3.1 Ma (Chapter 3)
LJRMg Hale Mountain granodiorite: variably foliated and lineated hornblende granodiorite; locally mylonitic; 186 ±1.5 Ma (Chapter 3)
LJHqs Hoboe quartz syenite: variably foliated and lineated hornblende quartz syenite; 189.5 ±3.0 Ma (Chapter 3)

Probable Mesozoic
Mhd variably foliated hornblende diorite to gabbro