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**Regional-Scale Transposition and Late Large-Scale Folding in the Teslin
Zone,
Pelly Mountains, Yukon**

**by
CHRIS C. GALLAGHER, B. Sc.**

**A thesis submitted to the Faculty of
Graduate Studies and research in partial fulfillment
of the requirements for the degree of
Master of Science, Department of Earth Sciences**

**Carleton University
Ottawa, Ontario
April, 1999**

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The undersign hereby recommend to the Faculty of Graduate
Studies and Research, acceptance of this thesis
"Regional-Scale Transposition and Late Large-Scale Folding in the Teslin Zone, Pelly Mountains,
Yukon"
submitted by Chris C. Gallagher, B.Sc.,
in partial fulfillment of the requirements for
the degree of Master of Science.

Chairman, Department of Earth Sciences

Thesis supervisor

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Carleton University
April, 1999

ABSTRACT

The d'Abbadie fault zone was originally interpreted to represent a major structural boundary that separated flat lying rocks of the Cassiar Platform to the east from steeply dipping mylonitic rocks of the Teslin Zone to the west. Detailed geologic mapping of the Dycer Creek area and to the north at Last Peak has revealed several major findings: i) rocks of the Teslin Zone and Cassiar Platform are documented on both sides of the fault; ii) regional transposition S_{T2} fabrics observed on both sides of the fault are correlative with one another; iii) late km-scale F_3 folding resulted in the modification of a near horizontal regional foliation on both sides of the fault into a steeply dipping orientation. These findings lead to several conclusions. The d'Abbadie fault zone is not a major tectonic boundary separating terranes of contrasting tectonic history, and the entire Dycer Creek area has undergone a common deformation history.

The dominant regional fabric documented in the area consists of a variably dipping, southeast-striking S_{T2} transposition fabric that is observed on both sides of the fault. S_{T2} is interpreted to have developed coevally in both domains during northeast-vergent isoclinal F_2 folding. The original orientation of S_{T2} , based on regional arguments and field observations, is interpreted to have been near horizontal or gently west dipping. S_{T2} has been rotated, by large-scale F_3 open folds, from a gently west dipping to a steeply dipping orientation in the eastern domain. F_3 minor open folds, observed in both domains, are correlated with zones of steeply dipping S_{T2} . These field observations led to the identification of late regional scale F_3 folds (e.g. Two-Cub antiform) as having been responsible for the steeply dipping fabrics of the Dycer Creek area. These findings are consistent with those to the north at Last Peak, although the geometric relationships between the two late folds requires further study.

The final phase of deformation involved two Cretaceous dextral strike-slip events. A local northwest-striking, ductile, dextral strike-slip zone, the Dycer Creek shear zone, and the d'Abbadie fault zone, a brittle, north-south striking, vertical fault zone. In the Dycer Creek area, kinematic indicators observed in the d'Abbadie fault zone are consistent with dextral strike-slip motion (Gallagher *et al.*, 1998).

The timing of regional ductile deformation is poorly constrained. Hansen (1989, 1992a,b) suggests that, based on regional cooling ages and a lack of recovery microstructures in Teslin Zone rocks, regional deformation occurred in Early Jurassic time. In the hope of better constraining the timing of regional ductile D_2 and D_3 deformation, geochronology samples of four plutonic bodies have been collected during the 1995, 1996 and 1997 field seasons. Initial results of the study are promising as U-Pb dating of zircons from the Mendocina Orthogneiss within the Teslin Zone has confirmed a Devonian age, and dating of monazites from the Dycer Creek Stock and Last Peak Granite has bracketed the age of regional F_3 folding between 113 and 98 Ma. These findings suggest that at least part of the regional deformation may be younger than previously proposed.

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ORIGINAL CONTRIBUTION

The Teslin Zone and eastern Cassiar Platform was originally mapped by Tempelman-Kluit (1977 and 1984) at a scale of 1:250000. More detailed mapping, at an undisclosed scale, was conducted by Vicki Hansen (1989). Previous isotopic studies of the area were restricted to U-Pb zircon dating of the Mendocina Orthogneiss and Ar/Ar cooling ages of muscovite, biotite and hornblende.

During the summer of 1997 I mapped the Dycer Creek area, located on the Livingstone Creek map sheet (105 E/12) at a scale of 1:25000. Particular attention was paid to structural observations on both sides of the d'Abbadie fault zone. Geologic and structural maps were digitized in AutoCAD and then imported into CorelDRAW 7.0 for refinement. Structural and other geologic data was entered into FieldLog for analysis. I completed structural and petrographic analysis of ~60 thin sections of samples collected in the field. U-Pb analyses of four samples, collected from the Dycer Creek area, were conducted at the Carleton University geochronology laboratory under the supervision of Dr. Sharon Carr and Dr. J. Blenkinsop. I completed sample crushing, mineral separation, mineral selection, fraction preparation, data reduction and interpretation. Sample dissolution, column chemistry and mass spectrometry on the TIMS were conducted by Dr. Sharon Carr.

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CHAPTER 1: INTRODUCTION

FIGURES:

Figure 1-1: Regional geology of the Teslin Zone and Cassiar Platform

Subduction zones manifest themselves at destructive plate margins where two tectonic plates converge. At such locations the tectonic plate with the greatest density (oceanic crust) will 'sink' underneath the overriding plate (continental crust). Within the subduction zone channel the majority of oceanic crust and pelagic sediments are dragged underneath another tectonic plate, exposed to high pressures and low to moderate temperatures and are finally recycled somewhere in the mantle. Accretion of oceanic and pelagic material to the continental land mass occurs in two possible ways: delamination and subsequent obduction of material onto the craton (thin skinned tectonics); or delamination and subsequent underplating of subducted material onto the accretionary wedge. Between two approaching crustal terranes, be they continents, micro-continents, island arcs or seamounts, at least one subduction zone must be present in order to remove the oceanic crust that separates them. When collision occurs, the oceanic crust and related sediments that once separated the two bodies have been consumed by subduction processes. The zone along which this occurred, by definition, is a suture zone "zones along which oceanic lithosphere has been totally subducted" (Seyfert, 1987, page 775).

Suture zones are characterized by a tectonic melange consisting of chaotic deformation and knocker geometry, high pressure - low temperature (eclogite and blueschist) metamorphism, and most importantly different tectonic plates on either side of the zone. These are major crustal-scale structures that propagate all the way to the Moho and separate two distinct tectonic plates from one another. Identifying such high strain zones in the field is difficult as most major suturing events are preceded by accretion of micro-continents and island arcs, and overprinted by younger orogenic activity. Many

terrane boundaries are identified as sutures, but they do not mark a zone of oceanic crustal recycling and many do not penetrate the lithosphere.

The Teslin Zone, located approximately 150 km northeast of Whitehorse (inset, Fig. 1-1), Yukon, has been interpreted, by some workers, to be a suture zone (Tempelman-Kluit, 1979; Hansen, 1989, 1992 a,b). It is characterized by a steeply dipping regional foliation that is believed to be terrane specific with respect to the Yukon-Tanana Terrane. Regional foliations were originally thought to have developed as steeply dipping mylonites in a west-dipping Permo-Triassic subduction zone complex. It was therefore proposed that the Teslin Zone is a suture zone that separates rocks of the North American craton from allochthonous rocks to the west. According to the model, the zone is crustal in scale such that it propagates all the way to the moho, and marks the western most extent of North American rocks in the northern Cordillera. Recent work, completed by Stevens *et al.* (1996, 1997), de Keijzer and Williams (1996, 1997a, 1998) and de Keijzer *et al.* (in press) is not consistent with previous interpretations of the Teslin Zone. They both question the role of subduction in the evolution of the Teslin Zone and do not consider it a suture zone, although the authors' separate interpretations of the tectonics of the zone contrast significantly (See Chapter 2).

An understanding of the tectonic evolution of the Teslin Zone will shed considerable light onto the early accretionary history of the Northern Cordillera. A widely accepted model pertaining to the evolution of the Teslin Zone and the Yukon-Tanana Terrane as a whole remains to be presented, as considerable work remains before the zone can be truly understood. The primary objective of this thesis is to further the

understanding of the structural and tectonic evolution of the Teslin Zone by analysis of structural and geochronologic data collected in the Dycer Creek area. The Dycer Creek area was chosen as a field area to obtain this goal as it provides an opportunity to examine rocks of the Teslin Zone, d'Abbadie fault zone and eastern Cassiar Platform. Detailed geologic mapping and U-Pb geochronologic sampling was conducted during the summer of 1997 and these data have been compiled with unpublished data collected by Harvey during the 1995 and 1996 field seasons. Objectives of the study were as follows: i) to determine the structural geometry of the Dycer Creek area to gain insight into the tectonic evolution of the Teslin Zone and neighbouring eastern Cassiar Platform; ii) to compare and contrast structural, lithologic and U-Pb geochronologic data on both sides of the d'Abbadie fault zone to determine if it is a major tectonic boundary; and iii) conduct U-Pb isotope analysis on geochronology samples to better constrain the timing of regional ductile deformation.

Figure 1-1: Regional geology of the Teslin Zone from South Big Salmon River to Last Peak (Modified after Hansen, 1989). Inset shows the location of the Dycer Creek area with respect to the various tectonic terranes of Wheeler and McFeely, 1991.

Terranes:
 NA = North America
 YT = Yukon Tanana
 TSZ = Teslin Zone
 NS = Nisling
 DY = Dorsey
 QN = Quesnellia
 CC = Cache Creek
 ST = Stikinia
 WM = Windy McKinley
 AX = Alexander
 WR = Wrangellia

m = undivided
 metamorphic rocks
 DFZ = D'Abbadie fault

Km

0 5 10 15

CHAPTER 2: REGIONAL GEOLOGY AND PREVIOUS WORK

FIGURES:

Figure 2-1: Structural geometry of the Teslin Suture Zone and neighbouring Cassiar Platform (after Hansen, 1989)

Figure 2-2: Tectonic evolution of the Teslin Tectonic Zone (after Stevens and Erdmer, 1996)

Figure 2-3: Structural geometry of the Teslin Zone (after de Keijzer and Williams, in press)

2.1 Introduction

The geology of the Yukon territory consists of numerous tectonostratigraphic terranes that form the main building blocks of the northern Canadian Cordillera (Galbrielse and Yorath, 1992). Eastern and central Yukon is composed of in-place or slightly displaced autochthonous terranes of Proterozoic and younger North American strata (Cassiar, Selwyn and McKenzie terranes; inset Fig. 1-1). West of these lie demonstrably allochthonous terranes of late Proterozoic and Mesozoic oceanic and island-arc affinity (Quesnel, Cache Creek, Wrangellia; inset Fig. 1-1; Coney, 1989). These allochthonous terranes are a product of subduction-related arc volcanism and sedimentation outboard of the western North American cratonal margin. Accretion of these terranes onto the North American plate began as early as the Devonian and initial assembly ended in the late Jurassic or early Cretaceous. Paleogeography of these terranes remains uncertain and some may have travelled great distances prior to accretion onto North America (Coney, 1989). This collage of terranes has been dissected by numerous dextral strike-slip faults that formed as a result of partitioning of oblique subduction and a change from dominantly oblique convergence to nearly margin parallel translation.

There is some controversy over the style, boundary conditions and polarity of the subduction zone or zones related to the formation and accretion of allochthonous terranes to the North American craton. Separating North American rocks from allochthonous terranes to the west lies the Yukon-Tanana Terrane, the easternmost of the Yukon suspect terranes. The size, age and location of the Yukon-Tanana Terrane makes it an important key to understanding the early accretionary history of the Northern Cordillera (Hansen,

1989). -

2.2 Yukon-Tanana Terrane Geology

Introduction

The Yukon-Tanana Terrane extends north-westward from the southern Yukon to central Alaska and forms the largest tectonostratigraphic terrane recognised in North America (inset, Fig. 1-1). It was originally defined by Coney *et al.* (1980) as an assemblage of rocks containing a strong L-S tectonite fabric in contrast with adjacent rocks. The Yukon-Tanana's size and location between autochthonous North America to the east and allochthonous terranes to the west make it the focus of a number of controversial issues. Firstly, does the Yukon-Tanana represent one single terrane or is it a composite of several terranes? Several authors have treated it as a single terrane (Gehrels *et al.*, 1990, 1991; Mortensen, 1992; Nelson and Mihalynuk, 1993; Mihalynuk *et al.*, 1994) but others have suggested that it comprises several terranes of varying tectonic affinity (Churken *et al.*, 1982; Foster *et al.*, 1985; Aleinikoff *et al.*, 1987; Hansen, 1990a, 1998). Secondly, does the Yukon-Tanana represent autochthonous strata of North American affinity, allochthonous strata or a combination of both?

Rock type and assemblages recognised

The paleosetting during deposition of highly metamorphosed and deformed protoliths of the Yukon-Tanana Terrane is not known. There is a growing consensus that the terrane represents the distal portion of the North American miogeocline. Coney (1989) suggests that it represents the detrital equivalent of the miogeocline and Mortensen (1992) interprets protolith deposition to be consistent with passive continental sedimentation.

Stevens *et al.* (1997) use chemical and isotopic signatures of Yukon-Tanana Terrane metasedimentary rocks to identify the setting of deposition. They interpret this data to be consistent with deposition along the outermost margin of ancestral North America where it received Proterozoic cratonic sedimentation and detritus from a chemically and isotopically juvenile signature (basalt-andesite) that is not recognised in the miogeocline.

Rocks at the lowest structural level are relatively homogeneous, consisting of metamorphosed quartzofeldspathic-mica schists and locally abundant metapelite and marble (Coney, 1989; Mortensen, 1992). These rocks have poorly constrained depositional ages but are known to be pre-Devonian (Mortensen, 1992). At that time, the Yukon-Tanana Terrane was apparently tectonically inactive as deposition is consistent with passive continental margin sedimentation (Mortensen, 1992). Higher structural levels are lithologically much more heterogeneous, consisting of metamorphosed mafic and felsic volcanic rocks, orthogneiss, chlorite schist, marbles and quartzite (Coney, 1989). Ultramafic and related sedimentary rocks, possibly representative of dismembered ophiolite, are closely related to Yukon-Tanana Terrane rocks (Mortensen, 1992).

Terrane nomenclature in the Yukon-Tanana Terrane is extremely complex. This confusion is primarily a result of two schools of thought, one dominantly in the Yukon region and the other in Alaska. This complexity may also reflect a general poor understanding of the terrane itself. The Yukon-Tanana Terrane (Coney *et al.*, 1980) has been sub-divided into a myriad of sub-terrane and assemblages, the simplest of which is that of Wheeler *et al.* (1988). They correlate a large portion of the terrane with Nisling Terrane of Wheeler and McFeeley (1987); the remainder is correlated with Kooteny

Terrane and named the Nisutlin sub-terrane. The Nisutlin sub-terrane is characterized by metamorphosed white-mica phyllitic quartzite, chlorite-quartz gritstone, calcareous chloritic phyllite to quartzite, graphitic quartzite and phyllite (Hansen, 1989). Ultramafic and mafic rocks, consisting of weakly to un-metamorphosed slices of oceanic layer 1 strata and associated sedimentary rocks, occur as klippen and imbricate thrust slices within the Yukon-Tanana Terrane. These were originally grouped as the Anvil allochthon or assemblage (Tempelman-Kluit, 1979, 1981), and subsequently renamed Slide Mountain Terrane (Monger, 1984; Monger and Berg, 1987). There remains debate as to the true affinity of these rocks and their tectonic significance in the northern Cordillera. For a comprehensive review of terrane nomenclature and subterrane comparison see Mortensen (1992).

Magmatism

Three major phases of magmatism and associated volcanism, distinguished on the basis of lithology, crystallization age, and isotope systematics, have been documented (Mortensen, 1992). The first major phase of magmatism had begun by the mid to late-Devonian and resulted in Devonian to Mississippian granitoid plutons and related mafic and felsic volcanic rocks. Deformed 364-343 Ma peraluminous orthogneiss (S-type), with a strong inherited Proterozoic Pb component, is recognised in Alaska, western and southeastern Yukon and the Teslin Suture Zone (Aleinikoff *et al.*, 1986; Mortensen, 1983, 1990a; Mortensen and Jilson, 1985; Hansen *et al.*, 1989). Deformed 365-348 Ma hornblende-bearing (I-type) quartz diorite, granodiorite and quartz monzonite, with minor Proterozoic inheritance, is recognised in south-eastern and western Yukon and central

Alaska (Mortensen, 1983, 1986; Mortensen and Jilson, 1985; Aleinikoff and Nokleberg, 1985; Aleinikoff *et al.*, 1987). These two assemblages of plutonic rocks have been attributed to continental-arc magmatism associated with a west-dipping Devonian subduction zone on the western margin of North America (Gilbert and Bundtzen, 1979; Dusel-Bacon and Aleinikoff, 1985; Nokleberg and Aleinikoff, 1985; Mortensen and Jilson, 1985; Nokleberg *et al.*, 1989).

The second major phase of Yukon-Tanana Terrane magmatism occurred in the Permian. It consists of well documented, weakly to strongly foliated plutons representing a 263-256 Ma calc-alkaline magmatic suite (Mortensen, 1996). The calc-alkaline chemistry of the suite and its temporal equivalence with Permian eclogite facies metamorphism is interpreted as evidence of a Permian subduction zone and related arc volcanism (Mortensen, 1996; Creaser *et al.*, 1997).

Abundant late-Triassic to early-Jurassic calc-alkaline plutonism is documented throughout the Yukon-Tanana Terrane and western terranes (Mortensen, 1996; Creaser *et al.*, 1997). These massive to rarely foliated, 212-185 Ma hornblende, granodiorite and quartz monzonite suites are thought to post-date regional D₁ deformation in the Yukon-Tanana Terrane and possibly mark the final assembly of the Intermontane Belt (Mortensen, 1992; Monger and Berg, 1987).

Regional deformation

The Yukon-Tanana Terrane is "characterized by a widespread subhorizontal mylonitic and gneissic fabrics superimposed upon a complex lithologic succession" (Coney, 1997, p. 111) that is interpreted to be a result of regional D₁ deformation

(Mortensen, 1992). The tectonic significance of this deformation event remains unresolved and continues to fuel the controversy over the tectonic evolution of the Yukon-Tanana Terrane. D_1 deformation was originally attributed to subduction zone related processes by Tempelman-Kluit (1979). Hansen (1989, 1990b) suggested that Cretaceous extensional processes are responsible, while others (Mortensen, 1992) favour a continent-continent collision model.

Most of the regional D_1 deformation is thought to have occurred in the early to mid-Triassic (Mortensen, 1992). Mid-Permian calc-alkaline suite rocks consistently contain weakly to strongly developed foliations correlative with the regional D_1 fabrics. These rocks place a lower age bracket of mid-Permian on the timing of D_1 deformation. An upper age bracket of late Triassic is provided by massive and undeformed late-Triassic plutons observed cross-cutting the regional D_1 fabric (Foster, 1970; Cushing, 1984; Foster *et al.*, 1987).

Metamorphism and cooling

Regional metamorphism, thought to have peaked in the late Permian to mid-Triassic, is primarily middle to upper greenschist-facies; however, amphibolite-facies rocks have been documented locally (Dusel-Bacon and Foster, 1983; Read, 1988; Dusel-Bacon *et al.*, 1989). The timing of peak metamorphism is interpreted to be syn to post- D_1 because in some areas strain fabrics are well preserved while in others they have been totally recrystallized (Mortensen, 1992).

Permian eclogite and blueschist-facies metamorphic rocks have been documented in the Yukon-Tanana Terrane throughout the Yukon and Alaska. Creaser *et al.* (1997, p.

709) suggest that they “constitute one of the best-preserved belts of Paleozoic high-pressure rocks”. The presence of these rocks is the best direct evidence that subduction related processes are involved in the evolution of the Yukon-Tanana Terrane. Pods of eclogite found near Last Peak preserve prograde eclogite-facies assemblages that have been overprinted by younger, greenschist-facies metamorphism (Creaser *et al.*, 1997). P-T work completed by Erdmer *et al.* (1996) reveals peak metamorphic conditions of 625°C and minimum pressures of 1.4 GPa. The pods of eclogite contain a penetrative ductile foliation that is correlated with the regional D₁ foliation. A concordant U-Pb age of metamorphic zircon in the Last Peak eclogite is 269 ± 2 Ma (Creaser *et al.*, 1997). ⁴⁰Ar/³⁹Ar muscovite ages obtained by Erdmer *et al.* (1996) are consistent with eclogites cooling through 350°C between 270 and 260 Ma. Further to the east, eclogite preserved in the St. Cyr klippe yielded metamorphic zircon crystallization ages that are equivalent within error to those of the Last Peak eclogite (Erdmer *et al.*, 1996).

Widespread Jurassic to mid-Cretaceous muscovite cooling ages are documented throughout the Yukon-Tanana Terrane. Since regional metamorphism is interpreted to have reached the thermal peak in the late Jurassic to mid-Triassic, the cause for such young cooling ages must be investigated. Mortensen (1992) suggested several interpretations: i) prolonged post-D₁ cooling from a late-Permian to Jurassic event; ii) a younger dynamothermal event; iii) a combination of both. In Cretaceous time, most of the deformation was in the form of dextral strike-slip faults and localized extensional faults.

2.3 Previous Work in the Teslin [Suture, Tectonic] Zone

The Teslin Zone is the narrowest portion of the Yukon-Tanana Terrane, measuring

15-20 km wide by 200 km long, and is distinguished from the rest of the Yukon-Tanana Terrane by an anomalous northwest-striking, steeply dipping regional foliation. Rocks of this zone, correlated with rocks of the Yukon-Tanana Terrane that contain a steeply dipping foliation, have been given several names: i) the Teslin Suture Zone (Tempelman-Kluit, 1979; Hansen, 1989); the Teslin Tectonic Zone (Stevens and Erdmer, 1996); the Teslin Zone (de Keijzer and Williams, 1998). For the remainder of the thesis it will be referred to as the Teslin Zone. In the central Yukon, (inset, Fig. 1-1) it is bounded to the west by the Cretaceous Big Salmon Fault and to the east by the north-south trending d'Abbadie fault zone. The nature of the transition from the steeply dipping foliation in the Teslin Zone to shallowly dipping in the north remains unresolved, primarily as a result of poor exposure. The stratigraphy of the Teslin Zone is generally poorly understood as a result of complex deformation and transposition, although metasedimentary and metavolcanic rocks of the Nisutlin and Anvil assemblages have been recognised (Hansen, 1989; Stevens and Erdmer, 1996; de Keijzer and Williams, 1997; Harvey *et al.*, 1997). Devonian peraluminous orthogneiss and Cretaceous leucogranite have been identified in the Teslin Zone (Hansen, 1989; Harvey *et al.*, 1997a; Gallagher *et al.*, 1998).

The Teslin Zone was originally modelled after a west-dipping Permo-Triassic subduction zone that marked the western extent of the North American craton (Tempelman-Kluit, 1979). The model suggests that the rocks of the Teslin Zone formed as a tectonic melange, with steeply dipping mylonitic fabrics resulting from chaotic deformation within the subduction channel in Permian or Triassic time. The melange was then thrust onto North American strata and preserved as tectonic klippen as far as 400 km

to the east of the zone. These interpretations were based on the anomalously steep-dipping regional fabrics, a general increase in metamorphic grade to the west and the presence of eclogite-facies metamorphic rocks within the Teslin Zone.

Hansen *et al.* (1987a, 1988, 1989, 1990, 1992a, 1992b) presented temporal and kinematic data that are interpreted to further constrain the evolution of the subduction complex. Hansen (1989) interpreted the Teslin Zone as a km-scale shear zone that consists of a steeply dipping mylonitic fabric containing two differently oriented stretching lineations. According to this model, the Teslin Zone's geometry and regional fabric are a direct result of the polarity and local sense of shear of the subduction zone. Hansen (1989) identified zones of dip-slip and strike-slip motion on the basis of stretching lineation orientations. Shear sense of these zones was interpreted from quartz c-axis orientations, mylonitic S-C fabrics, mica fish and asymmetric pressure shadows on pyrite porphyroblasts. Dominant zones of normal and reverse dip-slip shear were found to be cross-cut by distinct younger zones of dextral strike-slip shear (Fig. 2-1).

Hansen (1992a, 1992b) also provided P-T and geochronologic data that constrain the timing of deformation within the Teslin Zone. Quantitative pressure-temperature analysis of Teslin Zone rocks indicate that they have undergone moderate to high P-T conditions. Zones of dip-slip shear have undergone the highest P-T conditions (575-675°C and 0.8-1.6 GPa) and zones of strike-slip shear were active at lower P-T conditions (425-550°C and 0.5-0.8 kbar; Hansen 1987a). $^{40}\text{Ar}/^{39}\text{Ar}$ methods revealed zones with different cooling histories, and place an upper limit on regional ductile deformation. Rocks of the Teslin Zone were interpreted as having cooled rapidly

(50°C/Ma) in the early Jurassic (~195 Ma). Cassiar Platform rocks cooled below 350°C between 150 and 110 Ma (Hansen, 1989). The absence of a strongly annealed fabric led Hansen to suggest that the rocks of the Teslin Zone were uplifted to cool upper crustal levels shortly after deformation. Hansen (1989) interpreted these data as evidence that regional ductile deformation of the Teslin Zone had ended by the mid-Jurassic.

Hansen (1989, 1992a, 1992b) modelled the tectonic evolution of the Teslin Zone as a west-dipping Permo-Triassic subduction zone in which oceanic crust (Anvil assemblage) and pelagic sediments (Nisutlin assemblage) were subducted under an outboard magmatic arc. Subducted rocks were underplated and ductily deformed under high P-T conditions during both downflow and return flow within the accretionary wedge. This dip-slip motion is interpreted to be preserved as zones of reverse and normal dip-slip shear within Teslin rocks (Fig. 2-1 ①). Dip-slip motion then changed to strike-slip motion within the subduction channel under lower P-T conditions (Fig. 2-1 ②). The subducted complex was subsequently thrust onto North America as a result of the collision between North America and the outboard magmatic arc in early Jurassic time (Fig. 2-1 ③). The uplift of subduction complex rocks (Yukon-Tanana Terrane) resulted in rapid cooling rates in the Jurassic. North American rocks were buried and underwent slower exhumation and cooling in the early Cretaceous. Finally, in the late Cretaceous, rocks of the Teslin Zone were juxtaposed against rocks of the Cassiar Platform along a dextral strike-slip shear zone (Fig. 2-1 ④).

Stevens and Erdmer (1993, 1996, 1997) presented structural and geochronologic data that are not consistent with the previous subduction zone models of Templeman-Kluit

and Hansen. They argue that, although subduction does play a role in the evolution of the Yukon-Tanana Terrane, the fabrics present within the Teslin Zone are a result of complex shear derived from structural divergence. They interpreted the presence of Mississippian-Devonian orthogneiss, that cross-cuts primary sedimentary structures, as evidence that rocks of the Nisutlin assemblage constituted a sedimentary package by early Mississippian time (Stevens and Erdmer, 1996, 1997). In effect, the assemblage did not form in an early Mesozoic subduction complex but most likely developed as an arc-related basinal sequence in the Mississippian.

Based on three transects across the Teslin Zone, south of the South Big Salmon River (Fig. 1-1), Stevens and Erdmer (1996, 1997) identified numerous domains defined by a unique combination of regional fabric orientations, stretching lineations and folding style. Kinematic analysis of the separate domains revealed structural divergence consisting of both east and west-directed shear as well as strike-slip motion.

Stevens and Erdmer (1996) presented a model incorporating the westward subduction of Slide Mountain oceanic crust under an outboard magmatic arc in the Permian. Eclogite and blueschist-facies metamorphism occurred along the basal contact between the Slide Mountain and Nisutlin assemblages (Fig. 2-2 iii). Closing of the Slide Mountain ocean resulted in oblique thrusting of Nisutlin and Anvil assemblage rocks eastward over a west-dipping detachment in late Triassic to mid-Jurassic time (Fig. 2-2 ii). Peraluminous orthogneiss to the west acted as a tectonic wedge that resulted in westward-directed back-thrusting and structural divergence in the Teslin Zone (Fig. 2-2 ii). Diverging east and west-directed shear zones are used to explain the variety of fabrics and

structures documented. Finally, in the Cretaceous, dextral strike-slip faulting offset the western margin of the zone and localized extensional features that are now exposed in the peraluminous orthogneiss (Fig. 2-2 i).

de Keijzer and Williams (1996, 1997a, 1998) and de Keijzer *et al.* (in press) present structural data that is interpreted to be consistent with late, km-scale folding that produced the steeply dipping fabrics of the Teslin Zone. de Keijzer and Williams (1998) conducted detailed geologic mapping within the Teslin Zone, adjacent rocks of the Cassiar Platform and a tectonic klippe that is correlated with rocks of Slide Mountain Terrane. They concluded that regional fabrics consist of transposed compositional layering and variably plunging intersection lineations resulting from F_1 and F_2 folding. A mylonitic fabric, with fold hinges parallel to moderately south to south-east plunging stretching lineations, is documented along the western margin of the Teslin Zone. de Keijzer and Williams (1998) observed both steeply and shallowly dipping fabrics within the Teslin Zone. The km-scale, northeast-verging F_3 'Grizzly Synform' is identified within this fold belt (Fig. 2-3) and is thought to explain the geometry of part of the Teslin Zone. de Keijzer and Williams (1998) correlated anomalously steeply dipping fabrics with the overturned limb of the synform, while shallowly dipping fabrics represent the lower limb.

East of the d'Abbadie fault zone, klippen of ultramafic rocks, correlated with Slide Mountain Terrane, lie on top of Cassiar Platform rocks (Hansen, 1989; de Keijzer, 1998). de Keijzer and Williams documented transposition fabrics that extend up section into the klippen throughout the studied portion of the North American platform. They correlated fabrics of the Cassiar Platform and klippen with those of the Teslin Zone. The contact

between the klippen and rocks of the Cassiar Platform is marked by a mylonitic zone that de Keijzer and Williams (1998) correlated with the mylonitic zone documented in the western Teslin Zone. They interpret this inferred continuous mylonitic zone as a refolded detachment upon which rocks of the Teslin Zone were thrust onto the North American platform (Cassiar Terrane). The detachment is thought to have had an original gently west-dipping orientation.

de Keijzer and Williams (1998, in press) suggested that regional fabrics observed in the Teslin Zone are not the result of subduction related processes, but rather obduction related processes, and that the steep dip of fabrics in the zone are the result of late large-scale folding. The northeast-directed thrusting that produced the mylonitic zone is interpreted to have occurred prior to or during regional transposition. The thrust sequence was then folded by the northeast-verging Grizzly synform resulting in steeply dipping fabrics on the overturned limb of the synform.

Figure 2-1: Block diagram of the Teslin Zone and western Cassiar Platform, modified after Hansen (1989), depicting distinct zones of (1) dip-slip and (2) strike-slip shear that developed within the subduction zone channel in the Jurassic. Also depicted are (3) klippen of Teslin Zone affinity that were thrust on top of Cassiar Platform rocks; and (4) Late Cretaceous plutons (Kqm) and dextral strike-slip faults (DFZ).

Figure 2-2: Tectonic evolution of the Teslin Tectonic Zone from the Late Proterozoic to Early Cretaceous time (after Stevens and Erdmer, 1996).

i) Early - Mid-Cretaceous

- continued dextral strike-slip motion bisects Teslin Zone flower structure
- local extension exposes Devonian orthogneiss

ii) Late Triassic - Early to Mid Jurassic

- collision of crust containing orthogneiss with Teslin Zone resulted in west-directed back thrusting and dextral strike-slip motion
- development of a divergent flower structure
- incorporation of eclogite into Teslin Zone rocks

iii) Permian - Early Triassic

- westward subduction of Slide Mountain oceanic rocks under Yukon-Tanana
- eclogite and blueschist facies metamorphism
- east directed thrusting of Teslin Zone rocks onto North American miogeocline

iv) Mississippian

- deposition of Nisutlin and Anvil assemblages along distal North American shelf
- Nisutlin protoliths derived from North American craton
- Anvil protoliths derived from bimodal source consisting of distal North American and allochthonous sediments

i) Early - Mid - Cretaceous

ii) Late Triassic - Early to Middle Jurassic

iii) Permian - Early Triassic

iv) Mississippian

Figure 2-3: Simplified structural cross-section of the Last Peak Area, approximately 15 km north of the Dycer Creek area. Lithologic contacts represent transposed stratigraphy that developed during F_2 folding, that was rotated into a steep attitude by late large-scale F_3 folding (after de Keijzer and Williams, 1998).

CHAPTER 3: GEOLOGY OF THE DYCKER CREEK AREA

FIGURES:

Figure 3-1: Geologic map of the Dycer Creek area (in pocket)

Figure 3-2a: Western Domain Lithologic Column

Figure 3-3b: Eastern Domain Lithologic Column

PLATES:

Plate 3-1: Marble / biotite-chlorite psammite contact

Plate 3-2: compositional layering in marble

Plate 3-3: WD quartzite compositional layering

Plate 3-4: WD graphitic phyllite compositional layering

Plate 3-5: WD calcareous graphitic schist

Plate 3-6: ED massive carbonate

Plate 3-7: ED graphitic phyllite

TABLES:

Table 3-1: summary of lithologies and mineralogy of DCA map units

3.1 Introduction

The Dycer Creek area is located in the centre of the Teslin Zone approximately 100 km northeast of Whitehorse (inset, Fig. 1-1) and includes rocks of the Teslin Zone and the Cassiar Platform. The 126 km² map area is bounded to the south by Dycer Creek. No other major topographic features delineate the mapping area that is roughly bisected by the north-south trending d'Abbadie fault zone

For comparative purposes the map area will be discussed in terms of two domains (Fig. 1-1). The eastern domain lies east of the d'Abbadie fault zone and consists of Proterozoic greenschist to sub-greenschist facies carbonates and graphitic schists of the Cassiar Platform were intruded by the Little Lake Granite and Dycer Creek Stock. The western domain lies west of the d'Abbadie fault zone and consists of rocks interpreted to be affiliated with the Anvil and Nisutlin assemblages of the Yukon-Tanana Terrane (Hansen, 1989; Stevens and Erdmer, 1996; Tempelman-Kluit, 1979). Rocks of the western Domain are composed of greenschist to lower amphibolite-facies metasedimentary and metavolcanic rocks and two deformed granitoids; the Last Peak Granite and the d'Abbadie Orthogneiss.

Rocks of the Dycer Creek area have undergone penetrative polyphase deformation including at least one generation of transposition. As a result, unit thicknesses observed in the field are structural thicknesses and do not represent primary sedimentary thickness. There may also be unrecognized unconformable and - or structural contacts between units in the field and therefore the lithologic column (Fig. 3-2 a; 3-2 b) may not represent the original stratigraphic succession. Deformation has apparently destroyed all primary tops

indicators within the units. The descriptions of map units are presented in order from structurally lowest to highest levels. Table 3-1 is supplied as a supplement to this chapter and contains detailed descriptions and mineralogy of the map units.

3.2 Geology of the Western Domain

The western Domain has been divided into two assemblages (after Harvey *et al.*, 1996), on the basis of lithology and interpreted depositional environment; the Last Peak assemblage and the Mendocina assemblage (Fig. 1-1). They are parallel, northwest trending assemblages that are bounded by the d'Abbadie fault zone to the southeast and extend northwest to the limits of mapping (Fig. 1-1). Most of the units display recessive weathering characteristics, and topography is characterized by low, rolling, glaciated ridges and valleys. Typically these ridges and valleys run perpendicular to the regional strike, but the eastern portion of the domain is dominated by a large, relatively sharp, ridge that parallels the regional northwest strike. This ridge most likely formed as a result of the presence of the resistant Last Peak Granite and the Mendocina Orthogneiss that outcrop on its eastern flank (Fig. 3-1). The ridges that run perpendicular to the regional strike are typically at an elevation of 1500 metres (just above treeline) and are of low topographic relief. The best exposure in the domain is along the northwest-striking ridge (east), while the remainder of the domain (centre and west) is characterized by poor exposure.

3.2.1 Last Peak Assemblage

Metasedimentary rocks

The Last Peak assemblage is a ~ 7 km thick heterogeneous metasedimentary sequence exposed in the eastern portion of the western Domain (Fig. 1-1). It consists of

calcareous graphitic schist, quartzofeldspathic muscovite schist and biotite chlorite psammite with minor quartzite, graphitic phyllite and marble. Contacts between lithologies are typically gradational although the marble unit is in sharp contact with neighbouring units (Fig. 3-1; Fig. 3-2 a).

Biotite Chlorite Psammite

The psammite is at the lowest structural level within the assemblage and is a ~300 m thick, light greenish grey unit that weathers dark green with rusty red streaks (Fig. 3-2 a). It is well exposed on the eastern and western flanks of the north trending ridge just west of the d'Abbadie fault zone. (Fig. 3-1; Plate 3-1). It is a heterogeneous unit that varies through the section and along strike from biotite to chlorite rich psammite with local occurrences of actinolite. Contacts between the various sub-units are gradational at a scale of 10 m. The psammite is characterized by 0.5 to 1 cm thick quartzofeldspathic layers that consist of fine grained quartz and minor amounts of potassic feldspar. A mm thick fine-grained anastomosing foliation is defined by a chlorite $\pm$ biotite parting.

Marble

The marble unit is in sharp contact with the biotite chlorite psammite that occurs structurally below it (Fig. 3-2 a). It is a 20 to 50 metre thick homogeneous layer that is continuous along strike for several km. The resistant unit is well exposed on the northwest trending ridge and is repeated by F_2 folding (Fig. 3-1; Plate 3-1). It weathers to a light grey rough surface; 0.5 to 1 m thick layers of medium grained calcite and 10 to 20 cm thick dark grey layers composed of muscovite, calcic clinoamphibole and very fine grained graphitic material define the compositional layering (Plate 3-2).

Muscovite Psammite

The resistant quartzofeldspathic muscovite schist is in sharp contact with the underlying marble and is a ~250 m thick homogeneous unit that is variably exposed along the crest of the north-south trending ridge. (Fig. 3-1; Fig. 3-2 a). The unit is highly altered displaying oxidized muscovite, pyrite (?) and sericitized feldspar porphyroclasts resulting in a rusty red to orange weathered surface. The fresh surface of the rock reveals light grey 1-2 cm thick quartzofeldspathic domains bounded by an anastomosing foliation defined by a medium-grained rusty red muscovite parting. Medium-grained quartzofeldspathic domains consist of 80% to 90% quartz; the remainder is 0.5 to 1.5 mm potassic feldspar porphyroclasts.

Quartzite

The contact between the underlying muscovite schist and the quartzite is not exposed and its characteristics are not known (Fig. 3-2 a). The quartzite has a structural thickness of approximately 200 metres, weathers dark grey, and is a resistant well exposed layer. The unit forms a steep knoll along the southern end of the west-trending ridge, resulting in excellent exposure. Preferential weathering occurs along layers of fine grained muscovite and graphite resulting in a rough weathered surface. The rock is medium grained with sharp to mildly diffuse quartz-rich layers and finer grained layers of muscovite and graphite defining compositional layering (Plate 3-3).

Graphitic Phyllite

The 300 metre thick graphitic phyllite is in gradational contact with the underlying

quartzite (Fig. 3-2 a), but is poorly exposed as a result of its recessive nature. It is dark grey to black on both the fresh as well as the weathered surfaces. Variations in quartz and graphite abundances result in compositions ranging from graphitic phyllite to graphite-bearing quartzite. Compositional layering is defined by mm-scale dark layers of graphite or graphitic material and quartz rich layers (Plate 3-4).

Calcareous-Graphitic Schist

The 5 km thick unit is at the highest structural level and comprises approximately 50% of the Last Peak assemblage by area. It is characterized by a reddish-brown to orange weathering colour, with distinct intrafolial pockets of preferentially weathered out carbonate (Plate 3-5). The heterogeneous, poorly exposed unit displays gradational map-scale variations in mineralogy and composition. Sub-units consist of up to 20 metre thick concordant pods of graphitic schist and phyllite and minor graphitic quartzite. Compositional layering in the calcareous graphitic schist is defined by 1-2 cm thick discontinuous coarse layers of quartz and carbonate, and fine-grained chlorite-, muscovite- and graphite-rich layers. Graphitic schist and quartzite are similar in nature to the graphitic phyllite and quartzite map units.

3.2.2 The Mendocina Assemblage

The Mendocina assemblage is an undivided metasedimentary and metavolcanic sequence situated along the western margin of the western domain. Recessive weathering characteristics result in poor exposure along east-west trending ridges and valleys. Its structural thickness is poorly constrained, but is at least 2.5 km thick (Fig. 3-2 a). It is characterized by the presence of quartz-bearing chlorite schists that grade into concordant

resistant pods of quartzite up to 500 metres long and 75 metres wide. No igneous bodies intrude the assemblage in the western domain.

Chlorite psammite

This poorly exposed unit is characterized by a magnetite-zoisite-chlorite psammite that has a dark green weathered surface with preferentially weathered-out, mm-scale pockets of carbonate. The fresh surface of the rock reveals mm-scale compositional layering defined by fine-grained dark green chlorite-rich layers and coarse-grained, light grey to white quartz-rich layers. Calcite occurs as fine- to medium-grained interstitial material throughout the rock. Zoisite is observed throughout the rock in both the chlorite rich and quartz-carbonate rich layers. Accessory minerals consist of euhedral pyrite and magnetite porphyroclasts concentrated in the chlorite-rich layers.

South of the mapping limits, rocks of the Mendocina assemblage vary from dark green, recessive, carbonaceous, epidote-chlorite schist to psammite (Gallagher, unpublished mapping). Lithologic variation from schist to psammite is a reflection of varying amounts of chlorite within the rocks. The light green fresh surface of these rocks reveals mm to ½ cm thick, carbonaceous, quartz-rich layers alternating with phyllosilicate layers of equal or slightly greater thickness. Mineralogy within the quartz- rich layers is highly variable. Typically, quartz is the dominant mineral phase and albite comprises 5% to 15% of the rock. Calcite locally varies from fine-grained interstitial to medium-grained crystals that comprise the majority of the layer. Phyllosilicate layers vary from fine-grained chlorite partings to medium-grained chlorite layers that constitute up to 50% of the rock. Zoisite and epidote, that form up to 30% of the rock, are primarily concentrated in the

chlorite rich layers of the rock. Skeletal garnet, averaging 4 mm in diameter, is locally observed as are anhedral albite porphyroblasts.

2.3 Geology of the Eastern Domain

The eastern domain consists of a single lithologic assemblage, the Dycer Creek assemblage, that is bounded to the west by the d'Abbadie fault zone and extends east to the map boundary (Fig. 3-1). In the eastern part of the domain, resistant intrusive bodies result in topographic highs providing good exposure of the metasedimentary rocks in the east. Along the western margin of the domain, recessive graphitic phyllite and marble underlie low rolling ridges and valleys of variable orientation. The centre of the eastern domain is dominated by a large shallow cirque that provides good exposure of igneous and metasedimentary rocks. Lithologic contacts are typically sharp.

Metasedimentary rocks

The Dycer Creek assemblage is a well exposed heterogeneous assemblage with a minimum thickness of 3.5 km (Fig. 3-2 b). It consists of sub-greenschist to greenschist facies carbonate and pelitic metasedimentary rocks and variably deformed granites (see chapter 6).

Marble

The marble is the structurally lowest unit, has a thickness of approximately 1.5 km and comprises the majority of the metasedimentary rocks within the assemblage (Fig. 3.2b). It is a recessive, homogeneous, massive, light grey unit. Despite outcrop-scale homogeneity (Plate 3-6), the marble unit contains concordant, map-scale lenses of muscovite schist and calcareous quartzite. Compositional layering is defined by 10 to 15

cm thick carbonate layers and 0.5-1 cm thick quartz-bearing carbonate layers.

Discontinuous, 0.5 cm thick, resistant quartz-bearing layers result in a distinct rough weathered surface.

Graphitic phyllite

The graphitic phyllite is in sharp contact with the underlying marble (Fig. 3-2b). It is a homogeneous, dark grey to black, very fine-grained unit of unknown thickness, but is at least 300 metres thick. It is poorly exposed in most areas, except along one east-west trending ridge where it and its contact with the marble unit are well exposed.

Compositional layering is defined by quartz-rich layers intercalated with highly graphitic layers.

Carbonaceous Quartzite

Contact relationships, unit thickness and structural position of the carbonaceous quartzite are poorly understood as a result of poor exposure. Exposure of this unit is limited to one small hill in the southwestern portion of the eastern domain. The rock consists of ~100 % well rounded quartz grains with minor calcite concentrated along grain boundaries. The rock is well indurated but does display preferential weathering along grain boundaries resulting in a very rough, sugary weathered surface.

Figure 3-2a: Lithologic section depicting transposed stratigraphy and distinguishing characteristics of the Last Peak and Mendocina assemblages of the Western Domain.

Figure 3-2b: Lithologic section depicting transposed stratigraphy and distinguishing characteristics of the Dycer Creek assemblage of the Eastern domain.

Plate 3-1: View looking north-west towards Mount Last Peak along the contact between the marble and the biotite chlorite psammite.

CHLORITE

Plate 3-2: View looking south (up plunge) of a mesoscopic F_3 fold within the marble unit. The S_{T2} foliation can be seen rotated from a shallow to steep orientation.

Plate 3-3: Looking downplunge of the Panda antiform (north) at typical quartzite of the western domain. Near horizontal S_1 and S_2 axial plane cleavages are well defined.

Plate 3-4: Typical cm-scale compositional layering within the graphitic phyllite of the western domain. (Pavement surface looking north)

Plate 3-5: Photo looking west towards typical graphitic phyllite of the eastern domain. (Pavement surface looking north)

Plate 3-6: View looking north towards massive unfoliated marble of the eastern domain.

Plate 3-7: Photo looking west towards typical graphitic phyllite of the eastern domain.

Table 3-1 a: Western domain map units lithologic characteristics and possible protoliths

MAP UNIT	LITHOLOGY	UNIT DESCRIPTION	MODAL MINERALOGY	POSSIBLE PROTOLITHS
Chlorite Psammite	Magnetite zoisite chlorite psammite	<ul style="list-style-type: none"> - fine grained, dark green recessive unit - mm-scale compositional layering defined by chlorite rich and quartz rich layers - preferential weathering of interstitial calcite 	30% quartz 30% chlorite 25% zoisite 10% calcite 5% magnetite ± pyrite	<ul style="list-style-type: none"> - immature mafic volcanoclastic sediment
Chlorite Psammite	Epidote zoisite chlorite schist	<ul style="list-style-type: none"> - fine to medium grained, dark to pistachio green recessive unit - mm to ½ cm-scale compositional layering defined by zoisite / chlorite rich and quartzfeldspathic rich layers - skeletal garnet and anhedral albite are locally observed 	60% chlorite 15% epidote - zoisite 10% quartz 10% calcite 5% pyrite / biotite	<ul style="list-style-type: none"> - immature mafic volcanoclastic sediment
Calcareous Graphitic Schist	Calcareous graphitic schist	<ul style="list-style-type: none"> - medium grained, reddish - brown to orange recessive unit - cm-scale compositional layering defined by quartz / carbonate and muscovite / graphite layers - preferentially weathered out intrafolial pockets of calcite are common 	70% graphite 10% muscovite 10% quartz 10% chlorite	<ul style="list-style-type: none"> - pelitic rock with significant input such as a pelagic ooze - deep water basin
Calcareous Graphitic Schist	Graphitic schist	<ul style="list-style-type: none"> - medium grained, dark black recessive unit - mm-scale compositional layering defined by graphite rich layers and quartz rich layers 	80% graphite 20% quartz	<ul style="list-style-type: none"> - significant organic component - pelagic ooze deposited in a deep water basin
Graphitic Phyllite	Quartz graphite phyllite	<ul style="list-style-type: none"> - fine grained, dark black recessive unit - mm-scale compositional layering defined by pure graphite layers with rare interstitial quartz and quartz rich layers 	70% graphite 30% quartz	<ul style="list-style-type: none"> - significant organic component - deep water pelagic ooze

Quartzite	Quartzite	<ul style="list-style-type: none"> - medium to coarse grained, light grey resistant unit - compositional layering defined by cm-scale quartz rich domains alternating with mm-scale fine grained graphitic layers 	95% quartz 5% graphitic	- mature sediment source (quartz arenite) with minor organic component
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Table 3-1 a (cont.): Western domain map units lithologic characteristics and possible protoliths

MAP UNIT	LITHOLOGY	UNIT DESCRIPTION	MODAL MINERALOGY	POSSIBLE PROTOLITHS
Muscovite Psammite	Quartzofeldspathic muscovite psammite	<ul style="list-style-type: none"> - medium grained, rusty-red to orange resistant unit - light grey 1-2 cm thick quartzofeldspathic domains bounded by an anastomosing foliation defined by a medium-grained rusty red muscovite parting - highly oxidized muscovite and pyrite 	75% quartz 10% K-feldspar 10% muscovite 5% pyrite	- immature sediment source (quartz and feldspar) with periodic pelitic input
Marble	Dolomitic marble	<ul style="list-style-type: none"> - medium grained, light grey resistant unit - 1/2 m-scale dolomitic beds alternating with 10 - 20 cm thick dirty marbles containing calcic amphibole, muscovite and graphitic 	90% dolomite 5% tremolite (?) 5% chlorite ± graphitic	- carbonate protolith - episodic continental sediment input
Biotite Chlorite Psammite	Biotite chlorite semipelitic schist	<ul style="list-style-type: none"> - fine grained, dark green recessive unit - cm-scale quartzofeldspathic layers alternating with cm-scale chloritic schist layers 	60% chlorite 25% quartz 10% albite 5% biotite	- immature volcanoclastic sediment
Biotite Chlorite Psammite	Biotite chlorite psammite	<ul style="list-style-type: none"> - fine grained, dark green recessive unit - cm-scale quartzofeldspathic layers bounded by anastomosing chlorite, biotite ± actinolitic parting 	55% quartz 25% chlorite 10% albite 10% biotite 5% actinolite	- immature volcanoclastic sediment

Table 3-1 b: Eastern domain map units lithologic characteristics and possible protoliths

MAP UNIT	LITHOLOGIES	UNIT DESCRIPTION	MODAL MINERALOGY	POSSIBLE PROTOLITHS
Graphitic Phyllite	Graphitic phyllite	<ul style="list-style-type: none"> - fine-grained, homogeneous, dark grey recessive unit - fine grained graphite with minor interstitial quartz 	95% graphite 5% quartz	<ul style="list-style-type: none"> - pelagic ooze - deep water restricted basin
Marble	Marble	<ul style="list-style-type: none"> - medium grained, homogeneous, light grey recessive unit - massive light grey dolomitic marble with cm-scale discontinuous quartz rich layers - weathers to a rough sugary surface 	85% dolomite 10% quartz 5% muscovite $\pm$ calcic amphibole	<ul style="list-style-type: none"> - carbonate protolith - episodic continental sediment input
Marble	Muscovite semipelitic schist	<ul style="list-style-type: none"> - medium grained, dark rusty red resistant unit - in gradational contact with surrounding marble - muscovite schist with moderate quartz and minor k-feldspar abundance 	55% muscovite 30% quartz 10% albite 5% pyrite $\pm$ chalcocopyrite	<ul style="list-style-type: none"> - continental sediment input - quartz rich semi-pelite
Carbonaceous Quartzite	Carbonaceous quartzite	<ul style="list-style-type: none"> - fine to medium-grained, light grey resistant unit - homogeneous quartzite with carbonate cement - weathers to a rough sugary surface 	100% quartz	<ul style="list-style-type: none"> - quartz arenite with carbonate cement

CHAPTER 4: STRUCTURE OF THE DYCKER CREEK AREA

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4.1 INTRODUCTION

For comparative purposes, the structures and geometry within the Dycer Creek area are discussed in terms of the eastern and western domains (Fig. 3-1). Three generations of structures (D_1 , D_2 , D_3) were identified in both domains and each is characterized by a penetrative foliation and fold generation.

4.2 STRUCTURE OF THE WESTERN DOMAIN

4.2.1 Foliations

Fabric hierarchy is complicated when transposition has taken place. This is the case in the western domain where two episodes of transposition have occurred and four foliations are documented, S_0 , S_1 , S_2 , S_3 . Two transposition events have produced two composite transposition foliations: S_{T1} (where S_0 is parallel to S_1) and S_{T2} (where S_0 and S_1 are parallel or nearly parallel to S_2). Figure 4.2 illustrates the complex foliation relationships documented in the domain.

Centimetre-scale S_0 compositional layering, defined by quartz-rich, quartzofeldspathic or carbonaceous layers alternating with phyllosilicate layers (Plate 4-1), is the oldest foliation preserved in the domain. The layering is interpreted to represent primary sedimentary layering that has been attenuated and thinned by transposition during D_1 deformation (Fig. 4-2 i).

S_1 represents a well developed, mm-scale, cleavage that is axial planar to isoclinal F_1 folds. The cleavage is defined by preferentially oriented muscovite and graphite $\pm$

chlorite $\pm$ biotite, concentrated in the phyllosilicate-rich layers of S_0 (Plate 4-1). In the hinges of F_1 folds, S_1 and S_0 are at high angles to one another and are easily distinguishable (Fig. 4-2 i). Along the limbs of F_1 folds, S_1 is parallel to S_0 , resulting in the oldest preserved composite transposition foliation S_{T1} (Fig. 4-2 i). S_{T1} is typically steeply northeast or southwest-dipping (Fig 4.4a; i).

S_2 cleavage is preserved as a penetrative crenulation cleavage defined by the S_1 mineral assemblage and a pressure solution cleavage (Plate 4-1), or a penetrative axial planar cleavage (Plate 4-2). It is defined by preferentially oriented muscovite $\pm$ chlorite $\pm$ biotite that is axial planar to isoclinal F_2 folds (Plate 4-2). Weathered phyllosilicate rocks typically exhibit microlithons with an S_2 intercleavage spacing that varies from 1 to 10 cm. In the hinges of F_2 folds, S_2 sharply cross cuts the earlier transposition foliation S_{T1} (Plates 4-1 and 4-2; Fig. 4-2 ii). In the limbs of F_2 folds, S_2 is parallel to S_{T1} , forming the dominant transposition foliation S_{T2} (Fig. 4-2 ii). S_{T2} is a consistently northwest or southeast striking near vertical foliation that defines the regional foliation of the western domain (Fig. 4-4 a ii).

S_3 is a penetrative cleavage with a typical intercleavage spacing of 1-3 cm on the weathered surface of phyllosilicate-rich rocks. The cleavage is moderately to poorly developed and is best expressed within the calcareous graphitic schist. It is less prominent in other units near the d'Abbadie fault zone. S_3 axial surface cleavage is defined by discrete domains of preferentially oriented graphite, chlorite and muscovite. S_3 is consistently east dipping at $\sim 45^\circ$ and cross cuts the near vertical S_{T2} transposition foliation (Fig. 4-4 a; Plate 4-3 and 4-4). To the north at Last Peak strongly developed cleavage is

axial planar to sinusoidal open F_3 folds (de Keijzer and Williams, in press).

4.2.2 Folds

Minor folds

Rare F_1 folds are preserved as isoclinal, cm-scale hooks and sheared hinges that fold S_0 compositional layering (Plate 4-8). F_1 minor folds are rare throughout the western domain, but are more common in the west where D_2 deformation appears to be less well intense. Continued shortening and folding of F_1 structures during D_2 may have resulted in many F_1 minor folds being 'sheared out' beyond recognition. Outcrop-scale F_1 folds are less common than decimetre-scale folds. A single outcrop-scale F_1 fold, that folds S_0 compositional layering of the marble unit of the Last Peak assemblage, is recognized (Plate 4-5).

The most common folds in the domain are F_2 folds, with amplitudes ranging from 1 to 15 cm, that fold the S_{T1} foliation (Fig. 4-2 ii). These are tight to isoclinal, upright to slightly overturned folds with typically steeply east-dipping axial planes parallel to S_{T2} (Plate 4-6). They are shallowly plunging to the northwest at approximately 20° - 30° , although a small percentage plunge shallowly to the southeast (Fig. 4-4 b i). Outcrop-scale F_2 folds, with an average wavelength of 0.5 m, are identical in style to smaller-scale folds and are common within every lithology.

Minor F_3 folds are open to close, and plunge shallowly to the north-northwest and south-southeast (Fig. 4-4 b ii), with axial surfaces that dip at 45° to the east (4-4 a iii; Plate 4-7). They also form rare, doubly verging, symmetrical, box folds in the western portion of the map area within the calcareous graphitic schist. F_3 folds that do not form

symmetrical box folds have a consistent Z-asymmetry throughout the western domain (Plate 4-7). Outcrop-scale F_3 folds that have rotated originally shallowly west-dipping S_{T2} transposition foliations into a steep-attitude in the marble unit of the western domain have been documented (Plate 3-2). F_2 isoclinal folds that partially define the S_{T2} foliation have been refolded around the hinge of the mesoscopic F_3 fold.

Map-scale folds

Typically, F_1 minor folds are not common enough or well enough preserved to use asymmetry reversals as indicators of map-scale F_1 closures. To the north of the map area at Last Peak, a 0.5 km wavelength isoclinal F_1 fold is mapped in rocks of the Nisutlin assemblage (Harvey *et al.*, 1996; de Keijzer, 1998). Lithologic asymmetry is documented in the western domain along strike ~15 km south of the mapped Last Peak F_1 closure. This asymmetry is best explained by the continuation of the F_1 structure from Last Peak to the Dycer Creek area (Fig. 4-5). Observations are consistent with the presence of a map-scale isoclinal F_1 fold closure, whose axial surface trace is continuous for ~15 km, and supports the conclusion that D_1 deformation may be regional in nature and that the S_{T1} transposition foliation developed as a regionally significant structure.

A macroscopic F_2 fold system, consisting of tight to isoclinal folds with wavelengths of 200-500 m, is identified by numerous F_2 asymmetry reversals and polyphase fabric (S_{T1} and S_2) relationships. These folds, viewed down plunge to the northwest, have a Z-asymmetry with vertical to steeply east-dipping axial planes and shallowly northwest-plunging fold axes (Fig. 4-5). Map-scale F_2 fold style and geometry is consistent across the domain, although the wavelength decreases to the west where D_2

structures are less well developed. Map-scale F_2 isoclinal folds have rotated S_{T1} parallel to S_2 resulting in the regional transposition foliation S_{T2} .

All F_3 folds in the western domain (cm-scale and meter-scale) have a consistent Z-asymmetry looking down plunge. This is consistent with the entire western domain being situated on the same limb of a macroscopic F_3 fold with an antiformal hinge situated to the east. This large-scale folding event is reflected in the distribution of poles to the S_{T2} transposition foliation (Fig 4.4a; ii). S_{T2} is clearly not planar and therefore must have been refolded during a younger deformation event (map-scale F_3 folding).

4.2.3 Lineations

Observed lineations

The oldest lineation preserved, L_{T1-2} , is a well developed intersection lineation between S_{T1} and S_2 that is common throughout the western domain (Plate 4-9). The lineation plunges gently to the northwest or southeast, and parallels minor F_2 fold hinges (Fig. 4-4 b iii). The second lineation, L_{T2-3} , is a variably developed intersection lineation between S_{T2} and S_3 . The intersection lineation parallels F_3 fold hinges and plunges shallowly to moderately to the northwest (Fig. 4-4 b iv).

A penetrative crenulation lineation, L_3 , parallel to mesoscopic F_3 fold hinges on the S_{T2} surface is recognized in the western domain (Fig. 4-4 b v). The lineation is best observed within cleavage domains or partings in graphite-rich units.

There is a localized, shallowly dipping mineral lineation, L_4 , that is developed on the S_{T2} surface. The lineation is defined by actinolite and is observed only along the eastern margin of the domain in the biotite-chlorite psammite unit. Insufficient data have

been collected to confidently correlate the lineation with a specific generation of structure within the domain.

A well developed, predominantly shallowly north to northwest-plunging stretching lineation, L_5 , is documented in the variably deformed granitoids and metasedimentary rocks of the western domain (Fig. 4-1; Fig. 4-4 b v; Plates 4-10 and 4-11). The lineation, defined by stretched quartz and elongate biotite aggregates, is localized within the Dycer Creek shear zone; a 800 meter wide Cretaceous dextral strike-slip shear zone. (Fig. 3-1; See chapter 6 for detailed kinematic analysis of the shear zone and the stretching lineations within it.)

Interpretation of lineations

Regional lineations documented in the western domain are the result of numerous folding events and are not consistent with mylonitic deformation mechanisms as suggested by previous workers (Hansen, 1989; Stevens and Erdmer, 1996). The two intersection lineations, L_{T1-2} and L_{T2-3} , are the result of regional F_2 and F_3 folding respectively; while the L_3 crenulation lineation developed during regional F_3 folding. There are no regional stretching lineations observed in the western domain. This is in stark contrast with assertions of previous authors (Hansen, 1989, page 720) who described rocks of the western domain as having 'two coplanar, yet differently oriented, elongation lineations, L_{ds} and L_{ms} , locally marked by rodded quartz, smeared micas and aligned metamorphic minerals'. Intersection lineations can easily be mistaken for stretching lineations if polyphase foliation relationships are not fully understood (Fig. 4-3). It is possible that Hansen misinterpreted the two generations of intersection lineations as stretching

lineations.

The western domain is not characterized by regional stretching lineations.

Stretching lineations are limited to the Dycer Creek shear zone that post-dates regional deformation and is not interpreted to be regionally significant (Chapter 6). Processes that developed these rare stretching lineations within the domain are not considered to be significant in terms of regional deformation or foliation development.

4.3 Structure of the Eastern Domain

4.3.1 Foliations

Foliation relationships observed in the eastern domain are similar to those of the western domain. Four foliations are documented in the domain (S_0 , S_1 , S_2 , S_3) and are correlated with those of the western domain. At least one transposition foliation, correlative with S_{T2} of the western domain, is observed.

S_0 is preserved as cm-scale compositional layering that is interpreted to represent transposed primary sedimentary layering. S_0 is defined by alternating carbonate and quartz rich layers in the marble unit. In the graphitic phyllite it is defined by layers of graphite and minor muscovite alternating with quartz-rich layers.

S_1 is a cleavage that is well developed within the graphitic phyllite, but is not observed in the marble. S_1 is defined by the preferential orientation of graphite and muscovite that is parallel to S_0 . S_1 cleavage is interpreted to have developed during F_1 regional folding in the eastern domain. The parallel nature of S_0 and S_1 , even in the hinges of F_2 folds, is consistent with a transposition event during F_1 folding - resulting in the S_{T1} transposition foliation.

S_2 is well defined in the graphitic phyllite of the eastern domain as a 2-3 cm spaced cleavage, but is difficult to distinguish in other units. The cleavage is defined by preferentially oriented graphite and muscovite. S_2 is axial planar to isoclinal F_2 folds that fold S_0 and S_1 . Fabric relationships between S_0 , S_1 and S_2 are identical to those of the eastern domain and form a regional transposition foliation S_{T2} of variable orientation (Fig. 4-2 ii). There is a gradual increase in S_{T2} dip from $\sim 20^\circ$ west in the east to near vertical in the western part of the map area (Plates 4-12, 4-13 and 4-14). This transition in dip is reflected in a stereogram of poles to S_{T2} foliations within the eastern domain (Fig. 4-6 a i). The distribution of poles to foliations is consistent with a rotation of the S_{T2} foliation about a shallowly northwest-plunging fold axis. Shallowly dipping foliations observed in the east can be correlated with regional flat lying foliations of the Cassiar Platform and Yukon-Tanana terrane. Therefore the original orientation of the S_{T2} transposition foliation, based on regional arguments, is interpreted to have been near horizontal to gently west dipping. Steeply dipping S_{T2} foliations must therefore be a result of late folding that post-dated regional foliation development.

S_3 developed locally within the phyllites as a crenulation cleavage with 2-5 cm spacing. The cleavage dips $\sim 45^\circ$ to the northeast and cross cuts the variably oriented regional transposition foliation S_{T2} (Fig. 4-6 a ii; Plate 4-17). S_3 also developed locally as a set of conjugate cleavages associated with symmetrical F_3 box folds.

4.3.2 Folds

Minor folds

F_1 fold hinges are rare, but cm-scale isoclinal sheared fold hinges that fold S_0

compositional layering have been identified. Both S_1 cleavage and S_0 compositional layering along the limbs are axial planar to rootless F_1 folds. It is difficult to assign any sense of vergence to these folds as they are commonly sheared or rootless.

Mesoscopic F_2 folds are dominant, occurring as tight to isoclinal folds that fold composite S_0 - S_1 throughout the eastern domain. A variably developed axial planar cleavage, S_2 , has developed during folding and they plunge at $\sim 20^\circ$ to the northwest (Fig. 4-6 b iv). The variation in the dip of the F_2 axial planes is identical to that observed for the S_{T2} foliation. As a result, F_2 folds are reclined in the east (Plate 6-15) and become upright in the west (Plate 6-16).

F_3 folds are open, have interlimb angles of 60° to 120° and plunge moderately northwest (Fig. 4-6 b v; Plate 6-14). Where the S_{T2} regional foliation dips steeply, F_3 folds are observed; however, they are not observed within the shallow lying rocks in the eastern part of the domain (Fig. 4-1). The distribution of F_3 minor folds is therefore correlated with the orientation of the regional transposition foliation S_{T2} . Commonly F_3 folds form doubly verging box folds and kink bands within the graphitic phyllite. Where F_3 folds do not form symmetric box folds they have a consistent Z-asymmetry when looking down plunge to the northwest.

Map-scale folds

Map-scale F_1 folds cannot be identified on the basis of minor fold asymmetry reversals, polyphase fabric relationships, or lithologic asymmetries. Map-scale folds may have developed but were then overprinted and nearly obliterated by younger deformation as observed in the western domain.

A map-scale isoclinal fold system, consisting of northeast-vergent tight to isoclinal F_2 folds, is inferred from minor fold asymmetry reversals and polyphase fabric (S_0 - S_1 and S_2) relationships. Looking down-plunge to the northwest, the folds have wavelengths averaging 300 to 750 meters and are of consistent Z-asymmetry. These folds have rotated S_0 and S_1 parallel to S_2 axial planar cleavage, resulting in the regional transposition foliation S_{T2} .

F_2 axial planes and regional transposition foliations, S_{T2} , of the eastern domain have clearly been deformed by a younger folding event (Plates 4-12; 4-13; 4-14). A stereoplot of poles to S_{T2} foliations shows that these foliations, developed during D_2 deformation, have been rotated about a moderately northwest-plunging fold axis (Fig 4.6a; i). Mapping of S_{T2} foliations throughout the eastern domain has led to the identification of the Two-Cub F_3 antiform whose axial surface trace occurs along the eastern margin of the domain (Fig. 4-1). The Two-Cub antiform rotated originally near horizontal S_{T2} , defined by reclined isoclinal F_2 folds, into a vertical or steeply dipping orientation (Fig. 4-7). It is the northwest plunging Two-Cub antiform, which post-dates regional foliation development, that controls the geometry of the regional foliations of the eastern domain (Fig. 4-7).

4.3.3 Lineations

The oldest regional lineations observed in the eastern domain are an L_{1-2} intersection lineation in the phyllitic units. The variably plunging lineation, formed from the intersection of S_1 with S_2 axial planar cleavage, is not parallel to any generation of structure in the eastern domain (Fig. 4-6 b i). It is possible that the high variability of L_{1-2}

may reflect confusion with a younger $L_{2,3}$ intersection lineation that was not identified in the field. Moderately plunging L_3 crenulation lineations parallel to F_3 fold axes are common within the graphitic phyllite.

Local stretching lineations, consisting of quartz rods and elongate aggregates of biotite, are limited to the Little Lake Granite. The limitation of these lineations to the granite suggests that they are not regionally significant fabric elements.

4.4 The d'Abbadie Fault Zone

The north-south striking fault zone extends from the South Big Salmon River to Last Peak (Fig. 1-1). The zone was originally mapped as a west-dipping thrust fault system by Tempelman-Kluit (1979). Hansen (1989) re-interpreted the zone as a primarily strike-slip system. Based on mapping at Last Peak, de Keijzer *et al.* (1998, in press) concluded that youngest motion along the fault zone is normal.

In the Dycer Creek area, the ~500m thick d'Abbadie fault zone is localized in a forested, north-south striking valley bottom (Plate 4-18). Rocks within the fault zone are primarily dark grey to black graphitic phyllite and light grey marble that are correlated with units of the western domain. The rocks appear unaltered in the Dycer Creek area and are cross-cut by cm-scale brittlely deformed quartz veins; further to the south at Livingstone Creek the fault zone is marked by dark rusty red, oxidized marbles. Fault zone contacts are poorly exposed within the Dycer Creek area.

Structure and kinematic indicators of the d'Abbadie fault zone

The fault zone is characterized by discrete, nearly vertical, roughly north-south striking microfaults separating 5-15 cm thick zones of very fine-grained cataclasite. A

pre-existing foliation, defined by very fine grained graphitic partings, appears to be contained in the cataclasite. Further work is required to determine the exact relationship of the foliation to that of the regional structure. Nearly vertically plunging, mesoscopic “drag” z-folds of micro-faults are common. Rare, nearly horizontal slickenfibres that are also folded by the drag folds are observed on exposed micro-fault surfaces. Quartz veins in graphitic phyllite commonly display apparent dextral offset of several cm (Plate 4-19). Fourth-order drag folds, nearly horizontal slickenfibres and sheared quartz veins are consistent with dextral strike-slip motion. There is no evidence of normal or reverse motion in this part of the fault zone, although localized or an early stage of dip-slip motion cannot be ruled out.

Mendocina Orthogneiss and Last Peak Granite

The Dycer Creek shear zone, a ~750 meter wide, north-northwest striking, dextral strike-slip shear zone (described in chapter 6) is identified in the Mendocina orthogneiss, Last Peak granite and surrounding country rocks. This shear zone was originally interpreted to be associated with regional deformation in the Jurassic (Hansen, 1989). Geochronologic work completed by Harvey *et al.* (1996a) revealed that this shear zone is Early Cretaceous in age. The temporal, kinematic and spatial relationships of the Dycer Creek shear zone with the d’Abbadie fault zone led Harvey *et al.* (1996a) to suggest that foliations within the Dycer Creek shear zone are a result of dextral motion on the d’Abbadie.

Field relationships presented in this paper are not consistent with the interpretations of Harvey *et al.* (1996a). It is generally accepted that shear planes within

shear zones are parallel to the zone boundaries (Hanmer and Paschier, 1991). Shear planes observed in the d'Abbadie fault zone are north-south striking, while shear planes observed in the Last Peak shear zone are northwest to north-northwest striking. The oblique nature of the shear planes within these two domains raises the issue of whether the two foliations are coeval. If the foliations in the two shear zones developed coevally, in other words under the same stress regime, they would have been parallel. The Dycer Creek shear zone cannot reflect a jog in the d'Abbadie fault system as it has the wrong geometry for a dextral system. These two shear zones are therefore most likely separate dextral strike-slip systems that developed at different times in the Cretaceous. Cross-cutting relationships between the shear planes in the granitoids and those in the d'Abbadie fault zone were not observed. However, the map pattern suggests that the Dycer Creek shear zone is truncated by the brittle motion of the d'Abbadie fault.

Constraints on timing and magnitude of displacement on the d'Abbadie fault zone

A km-scale, northwest-trending F₃ fold hinge identified in the eastern domain is a prime candidate for a piercing point. Unfortunately, this antiformal fold hinge is not observed in the western domain. No other potential piercing points could be determined and therefore the exact magnitude of displacement remains unresolved. Therefore, metamorphic, structural and lithologic offset across the fault zone must be used to constrain the magnitude of displacement. Harvey *et al.* (1996) cite an apparent displacement of ~ 4 km based on the offset of lithologic contacts (Harvey, unpublished mapping, 1996). At the latitude of the Dycer Creek area, there is a similarity in structural style on both sides of the fault, and no offset of metamorphic isograds is apparent across

the fault zone. The lack of contrasting structures and metamorphic grade between the hanging wall and footwall is consistent with a small amount of dextral strike-slip displacement associated with the d'Abbadie fault zone.

4.5 Correlation of Eastern and Western Domain Structures

The Dycer Creek area was divided into the eastern and western domains to compare and contrast timing and style of deformation on both sides of the d'Abbadie fault zone. Detailed geologic mapping has revealed a similar style and relative timing of deformation events in these two domains.

D₁ deformation

Due to their scarcity, map-scale D_1 structures cannot be correlated across the d'Abbadie fault zone. There is better control on map-scale D_1 structures within the western domain, but this almost certainly reflects the amount of mapping conducted. Although there is a lack of correlative map-scale structures, minor structures and regional foliations can be correlated. An early cleavage, S_1 , that is axial planar to rare, F_1 , rootless, isoclinal, minor folds is identified in both domains. The nature of the D_1 event on both sides of the d'Abbadie is unclear, although it is interpreted to represent an early regional transposition event, that produced S_{T1} in the western domain. Therefore, D_1 is interpreted to have been more intense in the western domain.

D₂ deformation

Minor-scale and map-scale D_2 structures in both domains are correlative. Isoclinal, gently northwest and southeast plunging (Fig. 4-4 b i and Fig. 4-6 b iv), variably oriented F_2 minor folds are the dominant minor structures in both domains. A map-scale, northeast-

vergent, isoclinal fold system is identified on both sides of the d'Abbadie fault zone (Fig. 4-5 and 4-7). The fold system consists of variably oriented, tight to isoclinal, shallowly northwest-plunging F_2 folds with 250 to 750 metre wavelengths. This fold system, on both sides of the d'Abbadie, resulted in the rotation of regional foliations (S_{T1} in the western domain and S_0 and S_1 in the eastern domain) into parallelism with S_2 axial planar cleavage along the limbs of F_2 folds (Fig. 4-2 ii). A regional transposition foliation S_{T2} , representing the limbs of map-scale F_2 folds, resulted from this folding event.

The similarity of minor and major fold styles and polyphase foliation relationships in both domains is consistent with the entire Dycer Creek area having undergone a common D_2 deformation event, in which the regional foliation on both sides of the d'Abbadie developed coevally as a transposition foliation during D_2 isoclinal folding. The regional foliation in the eastern domain, based on regional arguments, is interpreted to have developed as flat-lying or gently dipping transposition foliation. It is reasonable to assume that the regional foliation of the western domain, that developed during the same deformation event, also developed with low dips.

D_3 deformation

F_3 minor folds have identical geometry in the two domains. These open to close, moderately northwest or southeast-plunging structures (Fig. 4-4 b ii and Fig. 4-6 b v) fold the regional S_{T2} transposition foliation, and have a variably developed axial planar or crenulation cleavage S_3 associated with them. The minor folds also have a consistent geometry across the fault zone; looking down-plunge, the folds are of Z-asymmetry with moderately northeast-dipping axial planes. F_3 folds have rotated S_{T2} foliations from

originally gently to steeply southwest-dipping orientations, albeit on different scales. In the western domain, these observations are limited to the outcrop scale (Plate 3-2). In the eastern domain, the rotation is observed on the regional scale (Plates 4-12, 4-13, 4-14; Fig. 4-7).

Both domains have undergone a common folding history that post-dates regional foliation development. Map-scale folds of this generation are responsible for the steeply dipping regional foliations observed in the Dycer Creek area. In the eastern domain the Two-Cub antiform is responsible for the steeply dipping regional foliations. The interpretation for the western domain is not as clear. Regional scale folding that post-dates S_{T2} development is responsible for the steeply dipping foliations, although it is not clear whether it is the Two-Cub antiform or another regional scale fold that is responsible.

4.6 Conclusions

Detailed geologic mapping of the eastern and western domains of the Dycer Creek area has revealed a similar structural geometry in both domains. This is consistent with the entire area having undergone a common deformation history. Regional foliations of the two domains, S_{T2} , are correlated and are interpreted to have developed synchronously during a single deformation event. This deformation event, D_2 , consisted of the development of a northeast vergent isoclinal fold system resulting in a regional transposition foliation S_{T2} . The transposition foliation, based on regional arguments, is interpreted to have developed as a horizontal or gently southwest-dipping foliation. Late large-scale folding, that post-dates regional foliation development, controls the geometry of these foliations and is ultimately responsible for the steeply dipping foliations observed

today. The eastern and western domains of the Dycer Creek area are separated by the north-south trending d'Abbadie fault zone that cross cuts all major regional structures. The common deformation history observed on both sides of the fault zone is not consistent with major displacement along the fault.

Figure 4-2 i and ii: A series of diagrams depicting the development of the S_{T1} transposition foliation, S_0 compositional layering parallel to S_1 cleavage, that developed from F_1 isoclinal folding. The end result is a transposed stratigraphy (S_{T1}) that contains intrafolial isoclinal and rootless F_1 folds.

Fig. 4-2 i: S_0 compositional layering originated as primary bedding or layering

Fig 4-2 ii: D_1 deformation, consisted of isoclinal folding, and resulted in thickened S_0 layering in the hinges and attenuated or discontinuous S_0 layering in the limbs of F_1 folds. S_1 axial planar cleavage also developed during D_1 folding.

In the limbs of F_1 folds - S_0 is parallel to S_1 , though S_0 and S_1 remain distinguishable from one another. S_0 and S_1 form a composite foliation (S_{T1}) on the limbs of F_1 folds.

In the hinges of F_1 folds - S_0 and S_1 are perpendicular to each other.

The D_1 transposition event resulted in the development of a composite foliation (S_{T1}) that consists of transposed S_0 compositional layering parallel to S_1 axial planar cleavage.

Fig 4-2 iii: D_2 deformation, involved reclined isoclinal folding that folded S_{T1} and developed the S_2 axial planar cleavage. In the limbs of F_2 folds, S_{T1} was rotated parallel to S_2 and resulted in the regional S_{T2} transposition foliation observed today.

In F_2 limbs - S_{T1} (S_0 and S_1) is parallel to S_2 . The three fabrics become difficult to differentiate from one another and are therefore referred to as the S_{T2} transposition or composite foliation.

In F_2 hinges - S_{T1} (S_0 and S_1) is perpendicular to S_2 axial planar cleavage.

Figure 4-3: Schematic block diagram depicts the relationship between oblique polyphase foliations (S_{T1} and S_2) and L_{T1-2} intersection lineations. Intersection lineations may be mistaken for stretching lineations if the relationship between different generations of foliations is not understood.

Quartz and feldspar, contained in S_0 , were deformed during D_1 transposition and occur as oblate spheroids. The flattening plane of these spheroids is parallel to S_1 cleavage; together they define the S_{T1} transposition foliation that is at a slight angle (exaggerated in this diagram) to S_2 cleavage.

The intersection of oblate quartz and feldspar grains with the S_2 cleavage surface results in a prolate ellipse that closely resembles a stretching lineation. If fabric relationships are not properly understood, these intersection lineations can be mistaken for true stretching lineations that, in this case, parallel F_2 fold hinges.

Figure 4-4a: Equal-area, lower-hemisphere stereograms of poles to western domain foliations.

i) S_{T1} Transposition Foliation ii) S_{T2} Transposition Foliation iii) S_3 Axial Planar Cleavage

iv) Shear plane Orientations in the Dycer Creek Shear Zone

Figure 4-4b: Equal-area, lower-hemisphere stereograms of western domain lineations.

i) F_2 Fold Hinges ii) F_3 Fold Hinges iii) L_{T1-2} Intersection Lineations iv) L_{T2-3} Intersection Lineations v) L_3 Crenulation Lineations

vi) Stretching Lineations in the Dycer Creek Shear Zone

Figure 4-5: Cross section of the Western domain, looking NW, from A to A' (See Fig. 4-1). The section displays a multiply refolded lithologic sequence dominated by regional F_2 isoclinal folding. Parallel limbs of the F_2 isoclinal folds define the regional transposition foliation. The marble marker unit, by which most of the structures were identified, is shaded. The intrusive Last Peak Granite (Lpg) and Mendocina orthogneiss (Mo) are shaded.

Figure 4-6a: Equal-area, lower-hemisphere stereograms of poles to eastern domain structural foliations.

i) S_{T2} Transposition Foliation ii) S_3 Axial Surface Cleavage iii) Flattening Planes in
the Little Lake Granite

Figure 4-6b: Equal area, lower-hemisphere stereograms of eastern domain lineations.

i) $L_{1.2}$ Intersection Lineations ii) F_3 Crenulation Lineations

iii) Stretching Lineations

iv) F_2 Fold Hinges v) F_3 Fold Hinges

Figure 4-7: Cross section of the eastern domain, looking NW, from B to B' (See Fig. 4-1). The section displays a similar structural geometry to that of the western domain. Regional foliations represent the highly attenuated limbs of map scale F_2 isoclinal folds. The orientation of these foliations are controlled by the Two-Cub antiform, a regional-scale F_3 fold structure. Variably deformed intrusive bodies, the Little Lake Granite (Llg) and the Dycer Creek Stock (Dcs) are shaded. The d'Abbadie fault zone (DFZ) is located at the southwestern most extent of the cross section.

Plate 4-5: Looking southward, a possible refolded F_1 hinge in the marble unit of the Last Peak assemblage.

Plate 4-6: Looking north at a typical F_2 fold within the marble unit. The folded foliation in the hinges is S_{11} and the attenuated foliation in the limbs is S_{12} .

Plate 4-7: Looking northwest, down-plunge at an open F_2 fold that folds S_{12} . Note moderately east-dipping S_3 cleavage cross cutting S_{12} .

Plate 4-12: Looking north at the shallowly west-dipping S_{T2} transposition foliation within the marble unit of the eastern domain.

Plate 4-13: Looking north at moderately west-dipping foliation in the eastern domain within the marble unit.

Plate 4-14: Looking north towards the steeply west-dipping regional foliation of the eastern domain.

Plate 4-15: Looking west towards a recumbent F_2 fold in the marble of the eastern domain. The axial plane is dipping away from the photo at approximately 20 degrees.

Plate 4-16: Looking westward at the pavement surface of a road in the field with a crenulation cleavage rich in graphite at the village.

Plate 4-17: Looking southward and up plunge at a moderately west-dipping S_1 crenulation cleavage. Open, sigmoidal F_2 folds present in S_1 microlithons fold the S_1 foliation.

Table 4a: Summary of structural style associated with the three generations of deformation observed within the western domain of the Dycer Creek Area

Generation	Fabrics	Folds	Lineations
D ₁	-defined by flattened quartz, feldspar and minor preferentially oriented muscovite	- rare, isoclinal folds of variable orientation - no map scale closures identified	- none observed
D ₂	- sub-vertical transposition foliation - defined by syn-kinematic metamorphic chlorite, biotite and muscovite - axial planar to F ₂ folds	- dominant mesoscopic and macroscopic folds - upright tight to isoclinal - shallowly plunging NW and SE - vergence reversals across macroscopic hinges are common	- S ₀₋₁ /S ₀₋₂ intersection lineation parallel to F ₂ fold axis - metamorphic actinolite mineral lineation parallel to F ₂ fold axis
D ₃	- crenulation cleavage defined by muscovite and biotite - moderately dipping to the NE - cross-cuts sub-vertical polyphase S ₁ /S ₂ fabric - axial planar to F ₃ folds	- common tight to open upright folds - axial planes dip moderately to the NE - shallowly plunging to the NW - form rare symmetrical box folds - consistent Z-asymmetry	- S ₀₋₂ /S ₃ intersection lineation parallel to F ₃ fold axis - crenulation lineation parallel to F ₃ fold axis

Table 4b: Summary of structural style associated with the three generations of deformation observed within the eastern domain of the Dycer Creek Area

Generation	Fabrics	Folds	Lineations
D ₁	- not well preserved - locally preserved parallel to F ₁ isoclinal folds	- rare isoclinal folds of variable orientation - no map scale closures identified	- none observed

D₂	<ul style="list-style-type: none"> - variably west-dipping transposition foliation - defined by syn-kinematic metamorphic muscovite and biotite - axial planar to F2 folds 	<ul style="list-style-type: none"> - dominant mesoscopic and macroscopic folds - upright to recumbent tight to isoclinal folds - shallowly plunging to the NW and SE - vergence reversals across macroscopic fold hinges are observed 	<ul style="list-style-type: none"> - S0-1/S0-2 intersection lineation parallel to F2 fold axis
D₃	<ul style="list-style-type: none"> - moderately north-east-dipping crenulation cleavage - defined by muscovite - cross-cuts variably west-dipping S0-1/S0-2 fabric - axial planar to F3 folds 	<ul style="list-style-type: none"> - rare upright open to close folds - axial planes dip moderately to the north-east - moderately plunging to the NE - confined to steeply-dipping S0-2 fabric - commonly form symmetrical box folds 	<ul style="list-style-type: none"> - S0-2/S3 intersection lineation parallel to F3 fold axis - L3 crenulation cleavage parallel to F3 fold axis is common within the graphitic phyllite

CHAPTER 5: METAMORPHISM OF THE DYCER CREEK AREA

5.1 Introduction

With the exception of de Keijzer and Williams (in press) few metamorphic petrology studies have been completed on the Teslin Zone and neighboring eastern Cassiar Platform. The majority of the studies have been focused on the pods of eclogite in the Teslin Zone (Creaser *et al.*, 1997; Erdmer *et al.*, 1996).

Qualitative analysis of metamorphic assemblages of the Teslin Zone are consistent with upper greenschist- to lower amphibolite-facies metamorphism (Stevens and Erdmer., 1996; Hansen, 1989; de Keijzer and Williams, in press). Stevens *et al.* (1996), working to the south of the Big Salmon River, documented the assemblages muscovite + chlorite $\pm$ albite $\pm$ biotite $\pm$ garnet in semi-pelitic schist and chlorite + actinolite + epidote + albite $\pm$ hornblende in mafic schists. It is generally accepted that the regional metamorphism was syn-kinematic with regional deformation, although each author interpretation of the deformation event differs.

Quantitative analyses of Teslin Zone metamorphic assemblages conducted by Hansen (1992b), yielded pressure-temperature estimates ranging from 575 - 675°C and 0.8 - 1.6 GPa in dip-slip domains to 425 - 550°C and 0.5 - 0.8 kbar in strike-slip domains and 1.0 to 1.3 Gpa and 675-800°C in Cassiar Platform rocks (Hansen, 1987a, 1992b). Hansen (1992) used these P-T estimates as evidence of the preservation of regional-scale subduction-zone metamorphic conditions.

5.2 Metamorphism of the Western Domain

A regional metamorphic event, characterized by greenschist-facies conditions, is recognized in the western domain. Minerals associated with peak metamorphism are referred to as the M_1 assemblage. For this study, descriptions of metamorphic assemblages and their temporal relationships to deformation events have been divided into three groups on the basis of lithology: i) the easternmost quartzofeldspathic psammites and semi-pelites; ii) the central quartzites, phyllites; and calcareous graphitic schists of the Last Peak assemblage; iii) the westernmost chlorite schists and psammites of the Mendocina assemblage.

5.2.1 Quartzofeldspathic psammites and semi-pelites of the western domain

Metamorphic assemblages observed

Psammites and semi-pelites are limited to the western portion of the domain (Fig. 3-1) adjacent to the Dycer Creek shear zone. The peak M_1 metamorphic assemblage consists of albite - biotite - muscovite - garnet that has been partially replaced by chlorite $\pm$ muscovite. These replacement textures are consistent with the peak greenschist-facies assemblage having undergone a chlorite overprint.

Albite is in contact with chlorite and biotite, but not with garnet. Where albite is in contact with chlorite and - or biotite, no replacement or reaction textures are observed and it is considered stable. Biotite is the most common phyllosilicate mineral observed, and occurs as anhedral to subhedral grains. Biotite displays straight grain boundaries with quartz and albite.

Overprint textures within the assemblage are common, with chlorite and biotite present as replacement products at the expense of garnet and muscovite. Sample YT95-

BS031a, taken from the quartz muscovite psammite, contains the only occurrence of garnets in the western domain. They occur as skeletal, anhedral to subhedral porphyroblasts that lack pressure shadows and contain no internal foliation. Biotite is observed as a replacement product of garnet along the grain boundaries and within internal fractures, while chlorite occurs as 'matted' grains around the garnet boundaries. Muscovite, commonly surrounded by biotite and chlorite, displays serrated grain boundary terminations. Chlorite consistently occurs as wispy grains with serrated boundaries associated with biotite; it is rarely observed separately as subhedral grains. Chlorite and biotite are consistently observed in contact with one another, sharing ragged or serrated grain boundaries. The relationship between chlorite and biotite remains unclear, although based on other retrograde reactions observed, chlorite is thought to be replacing biotite.

Timing of metamorphism

The M_1 assemblage, defined by preferentially aligned biotite, chlorite and muscovite, is concordant with and defines the S_{T2} transposition foliation. Chlorite, surrounding garnet porphyroblasts, is discordant to the foliation. These observations are consistent with upper greenschist-facies M_1 metamorphism being syn-tectonic with respect to D_2 deformation. Chlorite is both concordant and discordant to the transposition foliation suggesting that at least part of the retrogression post-dated D_2 deformation.

Contact metamorphism

Samples of psammite located near the Last Peak Granite contact show anhedral to

ehedral muscovite that is unaltered and overprints the coarser-grained M_1 muscovite grains. These grains are not concordant with the regional S_{T2} fabric observed in thin section. New muscovite growth is interpreted to represent localized metamorphism associated with intrusion of the Last Peak Granite (96 Ma; U-Pb monazite, this study).

Alteration

Iron staining, both in hand sample and thin section, is observed in the psammitic and semi-pelitic rocks and is spatially related to the Last Peak granite. In thin section, iron staining is most intense around muscovite and biotite crystals, but is observed along the grain boundaries of all mineral phases. Muscovite that has overprinted the stained micas of the M_1 assemblage is not stained. Ilmenite and rutile, limited to the chlorite and biotite rich layers, overprint the M_1 assemblage. There is no evidence eluding to a relationship between the iron staining and the growth of ilmenite and rutile.

5.2.2 Quartzite, phyllite and calcareous graphitic schist of the western domain

Metamorphic assemblages observed

The calcareous graphitic schist, graphitic phyllite and graphitic quartzite are located to the west of the quartzofeldspathic units and constitute the bulk of the western domain (Fig. 3-1). The peak M_1 assemblage of the quartz-rich metasedimentary rocks consists of plagioclase - quartz - biotite $\pm$ muscovite. Chlorite is present as a reaction product of biotite. The metamorphic assemblage and mineral relationships are consistent with peak greenschist-facies conditions, partially overprinted by chlorite.

Quartz and plagioclase are the dominant stable metamorphic minerals observed; they display straight grain boundaries and are in contact with all other mineral phases.

Biotite is less abundant and occurs as small anhedral grains in the cores of larger chlorite grains. Chlorite is interpreted to be a reaction product of biotite. Chlorite constitutes ~95% of the phyllosilicates, suggesting that biotite has almost been completely overprinted by chlorite. Sparse muscovite occurs as fine-grained interstitial material in the quartz-rich layers. Calcite is observed as fine- to medium-grained, anhedral, interstitial material that is altered to a 'skuzzy' brown or is intensely stained by iron oxide.

Timing of metamorphism

The M_1 assemblage chlorite - biotite $\pm$ muscovite is concordant with and defines the S_{T2} transposition foliation. Chlorite that has replaced biotite is also concordant with the foliation, suggesting that D_2 deformation most likely outlasted the observed retrograde reactions. These observations are consistent with a pre- to syn- D_2 greenschist-facies metamorphism.

Alteration

Ilmenite and rutile associated with the phyllosilicate layers overprint the regional metamorphic assemblage. Iron staining can be seen in thin section, localized around the ilmenite grains, and does not appear to be as pervasive as in the quartzofeldspathic rocks. Sample YT95-BS002 contains euhedral pyrite porphyroblasts associated with the phyllosilicate-rich layers that overprint the dominant metamorphic assemblage.

5.2.3 Mafic schists and psammities of the western domain

Metamorphic assemblages observed

The mafic schists are located along the western edge of the Dycer Creek area and are separated from the rest of the metasedimentary rocks by a possible post- D_1 , pre- D_2 fault (Fig. 3-1; Harvey *et al.*, 1996). The peak M_1 assemblage observed within the map area is albite - zoisite - chlorite $\pm$ calcite. The presence of albite, zoisite and chlorite and lack of replacement or reaction textures is consistent with greenschist-facies conditions. Albite - biotite - zoisite - calcite forms the M_1 metamorphic assemblage south of the map area (south of Dycer Creek; Harvey, 1996, unpublished mapping), where biotite is observed to replace chlorite. The presence of albite, zoisite and chlorite overgrown by biotite, is consistent with greenschist-facies metamorphic conditions.

Fine- to medium-grained, subhedral to euhedral zoisite constitutes up to 25% of the mafic schists in the map area. There are no replacement textures associated with zoisite and it is in contact with all other mineral assemblages. Two phases of chlorite are present within the unit: fine grained euhedral chlorite, and coarse grained poikiloblastic chlorite that contain zoisite inclusions. Neither the fine grained nor poikiloblastic chlorite has been replaced and they are in contact with all other mineral phases.

Rocks to the south of the Dycer Creek area consist of albite - chlorite - zoisite - calcite $\pm$ biotite. Albite is interpreted to be stable, as it is in contact with all mineral assemblages and there are no replacement or reaction textures associated with it. Zoisite within these rocks is less abundant, and occurs as fine-grained, anhedral to subhedral grains that are confined to the phyllosilicate layers. Medium-grained chlorite is the dominant phyllosilicate.

Biotite is common at grain boundaries of large chlorite grains and is interpreted to

be a reaction product of chlorite. There is no evidence of any mineral phase, such as plagioclase, that has replaced or partially replaced the zoisite.

Timing of metamorphism

Chlorite and zoisite are concordant with the dominant foliation which is interpreted to be the regional S_{T2} transposition foliation. Reaction products such as biotite are also concordant with the foliation. These observations are consistent with the M_1 metamorphic event being syn- D_2 in the mafic schists of the western domain.

Alteration

Iron oxide staining is not observed in these units. Veins, filled with euhedral calcite crystals that are stained by iron oxide, are common. Oxide development along the vein walls is documented. Euhedral pyrite porphyroblasts associated with the chlorite - zoisite rich layers overprint the M_1 assemblage.

5.2.4 Conclusions

M_1 metamorphism of the western domain is characterized by pre- to syn- D_2 greenschist-facies metamorphic assemblages. Psammitic and pelitic schists of the western domain preserve a prograde reaction of biotite replacing chlorite. Mafic schists to the west show possible prograde replacement of chlorite by biotite. No biotite isograd could be identified in the field. The interpreted metamorphic grade is consistent with observations made by previous workers in the Teslin Zone (de Keijzer and Williams, in press; Stevens and Erdmer, 1996; Hansen, 1989).

5.3 Metamorphism of the Eastern Domain

Metamorphic assemblage

A regional metamorphic event M_1 , consisting of greenschist-facies metamorphism, is documented in the domain. A second metamorphic event overprints the M_1 assemblage and is likely related to intrusion of Cretaceous plutons (Dycer Creek Stock and d'Abbadie pluton) into the eastern Cassiar Platform.

All units are characterized by the peak M_1 assemblage muscovite + chlorite $\pm$ albite, but local variations do occur. The assemblage muscovite - chlorite is consistently concordant with the S_{T2} transposition foliation. Not enough petrologic data has been recorded to confidently identify the metamorphic grade of the eastern domain, but preliminary estimates are consistent with greenschist-facies. This agrees with the observations of previous workers in the area (Tempelman-Kluit, 1979; Hansen, 1989); although, to the north, de Keijzer and Williams documented regionally extensive amphibolite-facies metasedimentary rocks consisting of the M_1 assemblage biotite - garnet - staurolite - kyanite.

A local outcrop of muscovite biotite pelitic schist (Sample CG97-704), located along the eastern limit of the map area is characterized by a biotite - muscovite - microcline assemblage that has been overprinted by An-rich plagioclase (andesine to labradorite), coarse grained muscovite and biotite. Fine-grained biotite, muscovite and microcline porphyroclasts are interpreted to represent the M_1 metamorphic assemblage. M_1 textural relationships are consistent with those observed elsewhere in the eastern domain. Coarse-grained, poikiloblastic biotite and muscovite overprint the M_1 assemblage. The majority of muscovite and poikiloblastic biotite are pseudomorphic after the M_1 biotite, but smaller grains do not show any preferred orientation. Plagioclase, with a

maximum anorthite content of ~60%, commonly occurs as coarse poikiloblastic grains that overprint the biotite. Plagioclase also occurs as pseudomorphs after biotite.

Cretaceous thermal overprint

Plagioclase (An₆₀), biotite and muscovite clearly overprint the earlier metamorphic assemblage. The random orientation of muscovite, biotite and plagioclase grains is consistent with the overprinting event post-dated M₁ metamorphism and D₂ transposition. Sample CG97-704 was taken from a station located ~300 metres from the Dycer Creek Stock (112 Ma; U-Pb zircon, this study). Thermal heating and circulation of related fluids within local country rocks, resulting from intrusion of the Dycer Creek Stock, is the most likely origin of this late event.

Late overprinting relationships are limited to one thin section, but are identical to those described by de Keijzer and Williams (1999). They documented a M₂ metamorphic event, characterized by oligoclase-andesine, biotite and muscovite, associated with intrusion of Cretaceous granitic plutons (d'Abbadie pluton). This event is distinguished by replacement textures at the expense of the M₁ mineral assemblage - most importantly Ca rich plagioclase that replaces biotite.

de Keijzer and Williams (1999) pointed out that this M₂ metamorphic event has important connotations with respect to thermobarometry studies conducted in the area (Hansen, 1992b). They argued that plagioclase and garnet growth may not have been contemporaneous and that 'unrealistically high pressure and temperature estimates are obtained using these minerals' (de Keijzer, 1999, page 128). Documentation of similar overprinting textures in the Dycer Creek area confirms that Cretaceous calcic

metasomatism was regionally extensive in the eastern Cassiar Platform and that caution must be used when using a thermobarometer involving plagioclase.

CHAPTER 6: PLUTONISM IN THE DYCKER CREEK AREA

Figures:

Figure 6-1: Fabric geometry of the Mendocina Orthogneiss

PLATES:

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Plate 6-2: asymmetrical extensional shear bands observed in the motion plane of the Mendocina Orthogneiss

Plate 6-3: K-feldspar sigma porphyroclast observed in the motion plane of the Mendocina Orthogneiss

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Plate 6-5: Little Lake Granite contact zoom

Plate 6-6: Little Lake Granite rock slab

Plate 6-7: S-C fabric observed in the motion plane of the Last Peak Granite

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Plate 6-10: Dynamic recrystallization along feldspar grain boundaries

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Plate 6-15: Pre to syn-D₂ garnets in skarn

Plate 6-16: Subgrain development along feldspar grains of Dycer Creek Stock

Plate 6-17: Annealed or polygonized quartz grains of the Dycer Creek Stock

TABLES:

Table 6-1: Mineralogical characteristics of the Granitoids

Table 6-2: Deformation and recovery of the Dycer Creek granitoids

6.1 Granitoid Rocks of the Western domain

Variably deformed granitoid rocks, the mylonitic Mendocina Orthogneiss (365 Ma; chapter 7) and the protomylonitic Last Peak Granite (98 ± 1 Ma; Harvey *et al.*, 1997a; chapter 7) are parallel, tabular, north- to northwest-trending sheets that together are ~750 metres thick and extend for at least 12 km along the eastern margin of the western domain. The contact between the two granitoid bodies is concordant with the regional S_{T2} foliation (Fig. 4-1) and contact relationships between the two rock types are not known (Harvey *et al.*, 1997a). The two granitoid sheets and a 50 metre strip of country rock to the west are located within the Dycer Creek shear zone, that is interpreted to be Early Cretaceous in age (Harvey *et al.*, 1997a). Within the shear zone is a north- to northwest-striking, near-vertical, mylonitic shear foliation concordant with S_{T2} and a gently northwest- or southeast-plunging stretching lineation. Table 6.1 provides a summary of the deformation and recovery mechanisms inferred in the major mineral phases of the granitoid rocks.

6.1.1 Mendocina Orthogneiss

Contact relationships and lithology

The Mendocina Orthogneiss is a two mica granite that is well exposed along the western margin of the Last Peak Granite (Fig. 4-1). The orthogneiss weathers from dark to light grey and displays tabular K-feldspar augen, up to 1 cm in length, in a dark grey, fine-grained matrix composed of quartz, K-feldspar, plagioclase, muscovite and biotite (Fig. 6-1). Contacts between the orthogneiss and the country rocks are concordant (Fig. 4-1), and there is apparently no chilled margin or contact metamorphism (Harvey *et al.*,

1997a). The contact is interpreted to be transposed (Harvey *et al.*, 1997a), but it is not possible to deduce if the original nature was intrusive or tectonic.

Structural fabrics

The dominant fabric in the orthogneiss is defined by discontinuous, north- to northwest-striking, near-vertical, C-planes (Fig. 6-1, 1) that are concordant with the regional S_{T2} foliation. This foliation forms part of a Type I S-C fabric (Twiss and Moores, 1992) and contains a northwest- to north-trending, shallowly plunging, stretching lineation that is defined by quartz ribbons, elongate feldspar porphyroclasts and elongate aggregates of biotite (Fig. 6-1, 4). C-planes occur as penetrative micro-scale shear planes consisting of discrete zones of very fine-grained, preferentially oriented biotite, quartz and insoluble residue (Plate 6-1). Discontinuous S-planes are defined by elongate or pulled-apart feldspar porphyroclasts, quartz ribbons, biotite fish and sigma K-feldspar porphyroclasts (Fig. 6-1, 2 a,b,c; Plate 6-1). S-planes are oriented at an angle of 45° or less counter-clockwise with respect to the C-planes and feed into the C-planes (Fig. 6-1). Asymmetric extensional shear bands cross-cut S-C fabrics and are oriented at an angle of $\sim 30^\circ$ clockwise with respect to the C-planes (Fig. 6-1, 3; Plate 6-2). A poorly developed set of micro-scale conjugate extensional shear bands, whose acute angle is bisected by the vertical shear planes, are observed in the YZ plane normal to the stretching lineation (Fig. 6-1, 5).

Mineralogical characteristics

K-feldspar occurs as matrix grains and tabular sigma porphyroclasts that average 1-1.5 cm in length (Plate 6-3). Core and mantle structures resulting from sub-grain

development, and dynamic recrystallization are common. Where grains have not been dynamically recrystallized, albite twinning within the core is common (Plate 6-3). Pressure shadows associated with porphyroclasts are composed of quartz, albite and biotite. Sericitized fractures within the porphyroclasts, normal to the long axis of the grains, are common, and a number of grains have been pulled apart. Intense sub-grain development and common pressure shadows are consistent with K-feldspar deforming in the ductile field. Serrate subgrain boundaries, Carlsbad twinning and undulose extinction preserved in the cores are consistent with little or no recovery within K-feldspar.

Quartz occurs as elongate aggregates or ribbons, up to 1 cm long, that define the stretching lineation contained in the S-plane (Fig. 6-1, 4; Plate 6-1). No fractures are observed within the quartz grains. Elongate sub-grains, with serrate boundaries, have a preferred orientation that is parallel to the S-planes of the rock. The development of such sub-grains is consistent with quartz deforming in the ductile field. Little or no annealing of quartz has occurred as sub-grains display undulose extinction and serrate boundaries.

Biotite occurs as poikilitic porphyroclasts and matrix grains (Plate 6-1). Porphyroclasts are kinked and display sweeping extinction with highly serrate grain boundaries. Matrix biotite is fine grained, averaging 0.5 mm in length, and displays kink banding consistent with a significant degree of recovery within biotite.

Kinematic indicators

Microstructures and their relationships are consistent with the orthogneiss having undergone general shear consisting of a component of strike-slip shear and a component of boundary normal shortening (Hanmer and Paschier, 1991). Within the motion plane,

S-C fabric relationships, asymmetric extensional shear bands, sigma porphyroclasts and biotite mica fish are consistent with dextral strike-slip motion (Fig. 6-1).

The plane normal to the stretching lineation preserves a more complicated history. Conjugate shear bands observed in the plane normal to the stretching lineation are consistent with a component of boundary normal shortening (Fig. 6-1, 5) while a weakly developed (or preserved) crenulated fabric suggests possible west-side-down dip-slip motion.

Conclusions

Transposed contacts are consistent with the emplacement of the Mendocina Orthogneiss, be it through igneous or tectonic processes, prior to or during regional transposition. The lack of a metamorphic aureole or chilled margin does not resolve the emplacement problem as it may be explained by tectonic thrusting or igneous intrusion into a middle crustal environment. Detailed kinematic analysis of fabrics within the orthogneiss reveals that it has clearly undergone general shear consisting of a major dextral strike-slip component and a minor, boundary-normal, shortening component. There may also be a component of west-side-down dip-slip motion recorded in the orthogneiss that is consistent with field observations to the north at Last Peak (de Keijzer *et al.*, 1988; in press). This weak asymmetry, observed in the YZ plane, may reflect a small component of dip-slip motion but may also reflect a pre-existing gently west-dipping fabric that was crenulated parallel to the vertical shear planes during general shear. An internal fabric that pre-dates Cretaceous strike-slip motion may have developed during regional D₂ transposition. The granitoid is therefore interpreted to contain fabrics

associated with both regional and local deformation.

6.1.2 Last Peak Granite

Field relationships and lithology

The Last Peak Granite is a resistant leucogranite that weathers light grey and consists of white, elongate, K-feldspar megacrysts set in a fine-grained, light-grey matrix of quartz, K-feldspar and plagioclase (Plate 6-7). The well exposed western contact between the granite and country rocks of the Last Peak assemblage is typically concordant with the regional transposition fabric S_{T2} , but undeformed dykes of granite intrude and cross-cut S_{T2} and S_3 (Plate 6-4, 6-5). A chilled margin at the western contact, along with a thin metamorphic aureole, are consistent with intrusion into rocks located in a cool upper-crustal environment (Harvey *et al.*, 1997a,b).

The Last Peak Granite can be divided into three domains on the basis of the style of structural fabrics and degree of mylonitization (Harvey *et al.*, 1997a). The granite in the western “contact zone” is fine grained, lacks structural fabrics and is characterized by an outcrop-scale igneous fabric (Harvey *et al.*, 1996). The main body of the granite is a protomylonite characterized by a well-developed, steeply dipping, shear foliation containing a shallowly north- to northwest-plunging stretching lineation. The eastern margin of the granite is marked by a ~100 m thick mylonitic C-S tectonite containing a shallowly northwest or southeast plunging stretching lineation.

Structural fabrics

The main body of the granite is characterized by a well developed foliation that is concordant with S_{T2} and is defined by comminuted, fine-grained biotite grains and quartz

ribbons. The eastern mylonitic zone is characterized by S-C tectonites and a northwest to north trending, shallowly plunging stretching lineation defined by quartz and feldspar rods. The plane parallel to the stretch and normal to the shear foliation displays near-vertical C-planes, defined by very fine-grained quartz, muscovite and biotite (Plate 6-6), concordant with S_{T2} . Discontinuous S-planes defined by elongate quartz aggregates, muscovite porphyroclasts and feldspar porphyroclasts are oriented at an angle of $< 45^\circ$ with respect to the C-planes (Plate 6-6). Asymmetric extensional shear bands form an angle of $\sim 30^\circ$ clockwise with respect to the C-planes (Plate 6-7). No structural fabrics are observed in the plane normal to the stretching lineation.

Mineralogical characteristics

The contact zone contains ductilely deformed K-feldspar porphyroclasts that display weakly developed core and mantle textures. Cores of the porphyroclasts contain poorly developed subgrains with serrate grain boundaries. There are no fractures observed in feldspar porphyroclasts. Quartz occurs as equant aggregates that preserve moderate to extensive preferentially oriented subgrains. Strongly undulose extinction is observed in quartz grains that do not contain subgrains.

K-feldspar megacrysts, up to 2 cm in length, contain well developed subgrains and are commonly pulled apart with quartz, biotite $\pm$ chlorite in fills. Pressure shadows associated with sigma and gamma K-feldspar porphyroclasts consist of fine-grained quartz and K-feldspar. Quartz occurs as well developed ribbons whose internal structure consists of very fine-grained subgrains with no preferred geometric orientation.

The mylonitic zone of the granite is characterized by highly sericitized K-feldspar

porphyroclasts set in a fine-grained matrix consisting of quartz, K-feldspar, and muscovite. Porphyroclasts commonly display core and mantle textures and subgrain development in the cores. Deformation twins are rare. Quartz typically occurs as elongate aggregates that partially define the flattening plane of the granite. Internal structure of the aggregates consists of subgrains that display polygonal and bilateral grain boundaries. Grain boundary migration within the aggregates is also commonly inferred. There are no fractures observed in the quartz grains. Syn-kinematic muscovite porphyroclasts contained within the flattening plane commonly occur as mica fish that are consistent with dextral shear (Plate 6-6). They typically exhibit strong undulose extinction and kink banding consistent with little or no recovery. Fine-grained muscovite and sericite observed in pressure shadows of feldspar porphyroclasts is typically undeformed.

Kinematic indicators

Kinematic indicators are consistent with dextral strike-slip shear, in the main body and the eastern margin of the granite. They are best developed in the mylonitic zone where well developed S-C fabrics, observed in the plane normal to the C-plane, are consistent with dextral strike-slip shear. Asymmetric shear bands, "mica fish" and sigma and gamma K-feldspar porphyroclasts, observed in the motion plane, are all consistent with dextral shear (Harvey *et al.*, 1996). There is no evidence of any boundary-normal shortening in the plane normal to the stretching lineation. Bulk shear associated with deformation of the Last Peak Granite is therefore interpreted to be of dextral strike-slip sense.

Conclusions

Cross-cutting relationships exhibited by the Last Peak Granite demonstrate an

intrusive relationship with the country rock; intrusion post-dated D_3 deformation.

Mylonitic fabric development within the granite must therefore post-date regional ductile deformation as well. Structural fabrics within the granite are interpreted to result from local dextral strike-slip motion. The west to east variation in strain observed in the granite is interpreted to represent intrusion of the magma just prior to or during displacement on the dextral strike-slip shear zone (Harvey *et al.*, 1997a).

6.1.3 Relationships between the Last Peak Granite and the Mendocina Orthogneiss

The Mendocina Orthogneiss, Last Peak Granite and local country rocks contain parallel mylonitic fabrics and stretching lineations that are unique to the Dycer Creek shear zone. It is therefore not unreasonable to assume these fabrics developed coevally in the shear zone following intrusion of the Last Peak Granite (the youngest intrusive body present). Geochronologic studies show that these granitoids are of vastly different ages (Mendocina Orthogneiss, 357 Ma and Last Peak Granite, 96 Ma), and the rocks share a common deformation history involving dextral strike-slip motion. Fabrics preserved in the Mendocina Orthogneiss are most complex and consistent with a component of boundary-normal shortening and possible west-side-down dip-slip motion. These fabrics are not observed in the Last Peak Granite and are therefore interpreted to have developed prior to intrusion of the Last Peak granite; possibly during regional D_2 deformation. Dextral strike-slip displacement within the Dycer Creek shear zone occurred some time after 98 Ma and the stretching lineations documented are therefore a local structure unrelated to regional deformation.

There is a spatial correlation between the granitoids and the Dycer Creek shear

zone (Harvey *et al.*, 1997a). Regional mapping has revealed that this dextral slip system is localized around the granitoids; to the north at Last Peak the dextral strike-slip system is not continuous along strike beyond the granitoids (de Keijzer *et al.*, 1988, in press). This raises the question of what came first: the shear zone or the intrusion of the Last Peak Granite. Igneous phases commonly intrude strike-slip systems where jogs create local areas of extension. If the Dycer Creek shear zone represents a jog in the d'Abbadie fault zone, its orientation is such that an area of local compression, not extension, would be created. This is not an ideal environment for the intrusion of magma. The author therefore proposes a model in which the magma intruded rocks of the western domain prior to development of the shear zone. Intrusion of the magma, whose shape would likely be constrained by the regional S_{T2} foliation, heated local country rocks and provided a vertical plane of weakness where local dextral shear manifested itself.

Field evidence is consistent with emplacement of the Mendocina Orthogneiss prior to regional foliation development, during D_2 folding, at which time an internal fabric developed as a result of transposition. In the Cretaceous, the Last Peak Granite intruded rocks of the western domain. The intrusion, through local heating, produced a plane of weakness within the country rocks where a local dextral strike-slip shear zone (Dycer Creek shear zone) manifested itself. The development of the shear zone resulted in a mylonitic fabric, concordant with, but unrelated to, the regional transposition fabric.

6.2 Granitoids of the Eastern Domain

Present within the Dycer Creek assemblage, are the weakly deformed Dycer Creek

Stock (112 Ma; chapter 7) and the moderately deformed Little Lake Granite (Fig. 4-1). To the north, the two granites are in contact while to the south the two granites are separated by a 50-100 m wide marble skarn. Variably deformed pegmatitic dykes, of similar mineralogy to the Dycer Creek Stock crosscut the mylonitic fabric of the Little Lake Granite. Top-to-the-east kinematic indicators are observed in both the granitoids. Table 6.1 provides detailed descriptions of the deformation-recovery mechanisms observed in the major mineral phases of the granitoids.

6.2.1 Little Lake Granite

Field relationships and lithology

The Little Lake Granite is a fine-grained two mica K-feldspar granite that is composed of K-feldspar porphyroclasts set in a light grey to white, fine-grained matrix of quartz, microcline, biotite and muscovite (Plate 6-8). The granite has a light grey fresh surface and weathers dark grey to brown. It is a 100-150 metre thick, northwest trending sheet (Fig. 4-1), with contacts parallel to S_{T2} . No chilled margin or contact aureole was observed. The western contact of the granite with adjacent marbles is well exposed and concordant with the regional transposition fabric S_{T2} (Plate 6-9). The eastern margin of the granite, which is in contact with both marble country rock and the Dycer Creek Stock, is not as well exposed. Contacts have been transposed parallel to the regional foliation and it is impossible to determine whether the contact is primary or tectonic in nature.

Structural fabrics

Structural fabrics in the granite consist of a southeast-striking, variably west-dipping foliation (Fig. 4-6 a iii) that contains a southwest-trending, down-dip stretching

lineation (Fig. 4-6 b iii). The foliation is concordant with the regional S_{T2} fabric. The transition in foliation orientation from moderately to steeply west-dipping is identical to that described of S_{T2} . The down-dip stretching lineation, defined by elongate quartz aggregates and elongate aggregates of muscovite and biotite, is contained in the flattening plane.

Domains of coarse-grained matrix, with the long axis of matrix grains averaging 0.75 mm in length, and fine-grained matrix, averaging 0.25 mm in length, also define the structural fabric in the granite (Plate 6-10). Finer grained domains are interpreted to represent zones of higher strain where grain size reduction has played a more important role in deformation. Weakly developed mesoscopic asymmetric extensional shear bands, observed in hand specimen, are present in the flow plane.

Mineralogical characteristics

The fresh surface of the K-feldspar granite consists of 2-4 mm diameter, anhedral K-feldspar porphyroblasts set in a fine-grained, light grey, matrix composed of quartz, K-feldspar, biotite, muscovite and chlorite. Compositional layering is defined by layers of phyllosilicates and alternating layers of fine-grained (1 mm) and coarse-grained (2-3 mm) quartz, K-feldspar and minor plagioclase. Accessory minerals consist of both rounded and prismatic zircon, apatite and pyrite.

Equant microcline porphyroclasts, averaging 2-3 mm in diameter, are highly fractured and sericitized. Grain size reduction of feldspar porphyroclasts is not observed in coarse grained domains but is recognised in the finer grained domains. In these domains, minor dynamic recrystallization and subgrain development along grain boundaries is

observed (Plate 6-11). Little or no annealing of feldspar porphyroclasts appears to have occurred as undulose extinction and strain twinning is commonly observed (Plate 6-11).

Quartz matrix grains occur as elongate aggregates that define the stretching lineation. Quartz subgrain geometry ranges from elongate to near granoblastic (Plate 6-10). Subgrain boundaries are typically bilateral and grain boundary migration is readily observed. These observations are consistent with a significant degree of recovery associated with the quartz grains. Fractures normal to the long axis of the grains are common, although no pulled apart grains or imbrication is observed.

In thin section, biotite is the dominant mafic mineral, present as remnant core material in chlorite. Biotite also occurs as porphyroclasts, 2 to 3 mm in diameter, with long axes parallel to the protomylonitic fabric. No kinking or folding of matrix or porphyroclastic biotite is observed. The majority of the porphyroclasts show regular extinction, though a small percentage do show undulatory extinction. Muscovite occurs solely as fine matrix grains that show no evidence of kinking or folding. Regular extinction is observed in most of the grains, which is consistent with a moderate degree of annealing or post-kinematic crystallization associated with muscovite. Myrkremites, associated with low temperature deformation, are observed in the matrix. Oxides and sulphides are concentrated along phyllosilicate layers.

Kinematic indicators

Mesoscopic kinematic indicators consist of weakly developed asymmetrical extensional shear bands observed in the motion plane and a moderately developed down-dip stretching lineation that are consistent with tops-to-the-northeast, dip-slip shear.

Discordant, southwest-dipping pegmatite dykes, present in the Little Lake Granite, are weakly boudinaged (Plate 6-12). This observation is consistent with a west-side-down shear sense within the granite. The scarcity of microstructural kinematic indicators most likely reflects primarily coaxial shear or a significant degree of post-kinematic recovery. Symmetrical pressure shadows around non to weakly deformed potassic feldspar and biotite porphyroclasts are consistent with this interpreted sense of bulk shear.

Conclusions

Transposed contacts are consistent with emplacement of the K-feldspar granite prior to regional transposition. The lack of contact metamorphism or a chilled margin is consistent with either tectonic emplacement or intrusion prior to regional cooling (~350-300°C by 118-115 and 110 Ma, muscovite and biotite; Hansen, 1989). It remains unclear whether or not the protomylonitic fabric of the granitoid body resulted from regional deformation or from the reactivation of regional fabrics (Cretaceous extension documented by de Keijzer and Williams; 1999, in press) in a local shear zone unrelated to regional deformation.

6.2.2 Dycer Creek Stock

The stock is a coarse-grained, biotite quartz monzonite and is exposed in the eastern domain (Fig. 4-1). It weathers dark grey and the fresh surface is light grey to white. The quartz monzonite is typically homogenous and does not contain phenocrysts. The variably exposed contact varies from highly discordant to concordant with S_{T2} . The western margin of the stock trends roughly parallel to the regional foliation while along an approximately 500 metre vertical exposure of the eastern margin, the vertical intrusive

contact sharply cross-cuts country rock containing moderately southwest-dipping S_{T2} (Plate 6-13).

Variably deformed pegmatite dykes, of similar composition to the Dycer Creek Stock, intrude the country rocks. The dykes vary from undeformed and discordant to deformed and concordant with respect to the regional fabric S_{T2} . Folded and boudinaged dykes are interpreted to have intruded the country rock prior to or during D_2 deformation (Plate 6-15), while undeformed discordant dykes are interpreted to have a post- D_2 intrusive relationship with the country rocks (Plate 6-14).

Locally, just west of S677 (Fig. 3-1), there is a distinct contact aureole consisting of a 10 metre wide calc-silicate skarn composed of coarse calcite, diopside, biotite and garnet. The skarn contains a well developed foliation that is concordant, on the outcrop scale, with S_{T2} . The foliation wraps around garnet porphyroblasts.

Structure

There is a weak, variably dipping, southeast-striking foliation defined by preferentially oriented biotite grains averaging 1 cm in diameter. The foliation is variably southwest dipping, from shallow in the east to steep in the west. This variation in foliation orientation is identical to that of S_{T2} and is interpreted to have been rotated by late F_3 folds. Southwest-plunging stretching lineations, defined by quartz rods and elongate aggregates of biotite, are limited to discrete, ~1m thick, gently west-dipping ductile shear zones. The stretching lineations lie in gently southwest-dipping S-planes that form a small clockwise angle with respect to the shear planes, and are consistent with a top-to-the-east shear sense.

Mineralogical characteristics

The stock consists of equigranular, coarse microcline, quartz, biotite with minor plagioclase and muscovite. Highly fractured and sericitized microcline grains show undulose extinction, tartan twinning and weakly developed core and mantle textures (Plate 6-17). K-feldspar grains are typically highly fractured and sericitized. Plagioclase grains display undulose extinction and deformation twins. No subgrains are observed, fractures are rare and there is little alteration associated with the grains. Quartz occurs as equant fine grained aggregates displaying 120° triple point junctions, bilateral grain boundaries and little or no undulose extinction that are consistent with a significant degree of recovery (Plate 6-16). Fractures within the quartz grains are rare. Biotite and chlorite are commonly observed together and biotite often forms the cores of chlorite grains.

Conclusions

Field observations associated with the Dycer Creek Stock provide evidence pertaining to the timing of intrusion into marbles of the Cassiar Platform. Cross-cutting contacts and pegmatite dykes along with contact metamorphism are consistent with post- D_2 intrusion into cold country rock. However, deformed pegmatite dykes and fabric relationships with contact aureoles are consistent with late- D_2 intrusive relationships. These observations are interpreted to be consistent with late syn- D_2 to post- D_2 intrusion of the quartz monzonite into the country rock. The weak foliation within the pluton most likely developed during late D_2 deformation or may have developed at a later time during reactivation of D_2 fabrics related to thermal softening at the time of pluton emplacement. Regardless, the foliation has been rotated by F_3 folding. We interpret this to be evidence

that the stock intruded the country rock prior to D_3 deformation.

Figure 6-1: Simplified block diagram of the structural fabric relationships in the Mendocina orthogneiss. The shaded plane represents the plane normal to the C-plane orientation of the rock.

1 Discontinuous micro-scale C-planes

2 Discontinuous S-planes defined by (a) mica fish, (b) quartz ribbons and (c) pulled-apart and sigma K-feldspar porphyroclasts

3 Cm-spaced, extensional shear bands

4 Intersection lineation observed on the C-plane surface

5 Conjugate shear bands in the plane almost normal to the stretching lineation

Plate 6-7: Polished rock slab of Last Peak Granite cut parallel to the stretching lineation and perpendicular to the C-plane. C-planes form a very small angle with S-planes and their orientation is consistent with dextral shear. Asymmetrical shear bands cross cut both C- and S-planes and their orientation are consistent with dextral shear.

Plate 6-8: Polished slab of Little Lake Granite cut parallel to stretching lineation and normal to shear foliation. The flattening plane is defined primarily by quartz, biotite and K-feldspar.

Plate 6-9: Looking south at the concordant contact between the Little Lake Granite and Cassiar Platform country rocks. The rock face is approximately 15 m in height.

Plate 6-10: Thin section of Little Lake Granite cut parallel to stretching lineation and normal to foliation. The foliation is primarily defined by flattened quartz, K-feldspar and muscovite grains.

Plate 6-11: K-feldspar grains along microcline grain boundaries in the Little Lake Granite. Grain development is interpreted to result from dynamic recrystallization due to ductile deformation.

Plate 6-12: Boudinaged felsic dykes in the Little Lake Granite. The extension direction of the discordant dyke and the dominant foliation is consistent with west side down shear.

Plate 6-16: Thin section of the Dycer Creek Stock showing characteristic straight to near polygonized quartz grain and subgrain boundaries.

Plate 6-17: Thin section of the Dycer Creek Stock showing very fine grained quartz and K-feldspar along Microcline grain boundaries. Quartz is interpreted to be neocrystic while K-feldspar is interpreted to result from dynamic recrystallization.

Table 6.1: Deformation and recovery mechanisms observed in major igneous mineral phases and shear sense of kinematic indicators in the plutonic rocks of the Dycer Creek area.

Mineral	Ductile Deformation Microstructures	Brittle Deformation Microstructures	Recovery and Annealing Textures	Kinematic Indicators
Mendocina Orthogneiss				
K-Feldspar Porphyroclasts	- core and mantle structure - subgrains in cores - undulose extinction and deformation twins	- fractures, imbricated grains and brittle pulled apart grains	- no recovery textures observed	- dextral sigma porphyroclasts - pulled apart and imbricated grains consistent with dextral shear
Quartz	- sutured elongate subgrains - quartz ribbons - undulose extinction	- no brittle deformation microstructures observed	- little to no recovery textures observed	- type I S-C fabric development consistent with dextral shear
Biotite	- undulose extinction	- kink bands	- no recovery textures observed	- dextral mica fish porphyroclasts
Last Peak Granite				
Potassic Feldspar	- core and mantle structures common - subgrain observed in cores - undulose extinction and deformation twins	- typically unfractured	- no recovery textures observed	- dextral sigma and gamma porphyroclasts
Quartz	- elongate subgrains	- none observed	- straight and polygonized subgrains boundaries	- quartz aggregates help define the dextral C-S fabric
Muscovite Porphyroclasts	- undulose extinction - rare subgrains	- kink bands - rare pulled apart grains	- none observed	- dextral mica fish porphyroclasts

Table 6.1 (cont.): Deformation and recovery mechanisms observed in major igneous mineral phases and shear sense of kinematic indicators in the plutonic rocks of the Dycer Creek area

Mineral	Ductile Deformation Mechanisms	Brittle Deformation Fabrics	Recovery / Annealing Textures	Kinematic Indicators
Little Lake Granite				
K-Feldspar	- very fine grained grains and subgrain along grain boundaries consistent with dynamic recrystallization - undulose extinction and deformation twins	- fractures are common	- none observed	- coaxial symmetrical pressure shadows
Quartz	- equant subgrains are common - undulose extinction	- none observed	- straight and polygonized subgrain boundaries	- defines flattening plane
Biotite / Muscovite	- undulose extinction	- kink bands	- none observed	- symmetrical pressure shadows - defines flattening plane
Dycer Creek Stock				
K-Feldspar	- minor fine grained K-feldspar and quartz along grain boundaries - undulose extinction and deformation twins	- fractures are common	- none observed	- none observed
Quartz	- extensive subgrains observed - undulose extinction	- none observed	- polygonized and bi-lateral subgrain boundaries are common	- none observed
Biotite / Muscovite	- none observed	- some kink bands	- none observed	-biotite megacrysts define foliation

CHAPTER 7: U-Pb ISOTOPIC AGE ANALYSES OF DYCKER CREEK AREA GRANITOIDS**Figures:****Figure 7-1: YT95-BS082 Concordia Diagram****Figure 7-2: YT95-BS107 Concordia Diagram****Figure 7-3: CG97-711 Concordia Diagram****Tables:****Table 7-1: Sample Descriptions****Table 7-2: Summary of analysis**

7.1 Analyses of geochronology samples from the Dycer Creek area

7.1.1 Analytical methods

Sample preparation, analytical techniques and data reduction were carried out at the Carleton University U-Pb geochronology laboratory using methods described by Krogh (1973, 1982), Roddick (1987), Roddick *et al.* (1987) and Parrish and Krogh (1987). Where available, zircon and monazite crystals of the best clarity, without visible cores, and with minimal opaque inclusions or fractures, were chosen for analysis.

Descriptions of zircon and monazite, outlined below, apply to the general population of crystals in each sample. A summary of sample descriptions and their field relationships are presented in Table 7-1 and U-Pb analytic data for the samples are presented in Table 7-2. Monazite data are reversely discordant, a common phenomenon in young monazite. This is interpreted to result from the incorporation of excess ^{230}Th during monazite growth. Monazite crystallization ages are therefore estimated from the 207/235 age, which does not reflect ^{206}Pb concentrations (Schärer, 1984; Parrish, 1990).

7.1.2 Mendocina Orthogneiss

The general population of zircons within YT95-BS082 (Mendocina Orthogneiss) occurs as 170 μm - 64 μm diameter, euhedral to prismatic, whole or fractured zircons of poor quality and clarity. Cores, preserved as cloudy inclusions within the crystals, are common and the majority of the zircons are altered to a dark rusty red. Opaque inclusions are present within most of crystals. Zircons of <64 μm diameter are of excellent quality and clarity. The <64 μm crystals contain no visible cores or fractures, opaque inclusions are rare and alteration is nonexistent to weak. We interpret the morphology of all the

zircons to reflect a magmatic origin.

Fractions A is a 6 grain fraction of equant, euhedral, $+149\ \mu\text{m}$ zircons and B is a 34 grain fraction of equant, subhedral to euhedral, $105 - 149\ \mu\text{m}$ zircons; both are highly discordant and are interpreted to have an inherited Pb component (Fig 7-1). Fraction E is a 14 grain fraction of $<64\ \mu\text{m}$ needles with a 3:1 length-to-width ratio. $207/206$, $206/238$ and $207/235$ ages derived from U-Pb analysis of fraction E are in agreement within error and the fraction is therefore interpreted to be concordant ($<0.3\%$ discordant; Table 1). The weighted average of the $^{207}\text{Pb}/^{235}\text{U}$ and $^{206}\text{Pb}/^{238}\text{U}$ ages yields a crystallization age of $357.8 \pm 1.2\ \text{Ma}$ and is interpreted to represent the crystallization age of YT95-BS082 (Mendocina Orthogneiss; Fig. 7-1).

7.1.3 Last Peak Granite

Both zircon and monazite were analyzed from YT95-BS107 (Last Peak Granite). Zircons, ranging from $160\ \mu\text{m}$ to $64\ \mu\text{m}$ in diameter, are of poor clarity and quality. Most euhedral to prismatic crystals contain cores and opaque inclusions and are altered to a dark rusty red. Zircon morphology is interpreted to reflect a magmatic origin. Equant, euhedral, monoclinic monazite crystals, with diameters of $>149\ \mu\text{m}$ were analyzed. Monazite is pale yellow to gold and of excellent clarity. Opaque inclusions within the monazite are common, although no cores were observed. Monazite morphology is consistent with an igneous origin.

Zircon fractions A, B, D, E, and F-1 are single grain to multigrain fractions ranging in size from $160\ \mu\text{m}$ to $74\ \mu\text{m}$ in diameter. All show various degrees of discordance and are interpreted to reflect components of inherited Pb (Fig. 7-2). The pattern of discordance is

consistent with multiple Proterozoic sources for the inherited Pb. Monazite fractions M1 - M3 are slightly reversely discordant (Fig. 7-2). The weighted average of the $^{207}\text{Pb}/^{235}\text{U}$ ages yield an age of 95.7 ± 0.6 Ma that is interpreted to represent the crystallization age of YT95-BS107 (Last Peak Granite).

7.1.4 Dycer Creek Stock

CG97-712 (Dycer Creek Stock) contains zircon and monazite, and both were analysed. Zircon occurs as 160μ to 64μ euhedral to prismatic crystals of poor clarity and quality. Crystals are predominantly highly altered and fractured. Large cores and opaque inclusions are common. Monazite is generally of good quality and occurs as pale yellow to gold equant euhedral crystals. Clarity of the crystals ranges from good to perfect. Cores are not observed although opaque inclusions are common. The morphology of zircon and monazite is interpreted to reflect a magmatic origin.

Multigrain zircon fraction G is discordant and is interpreted to contain a Proterozoic Pb component (Fig. 7-3). Monazite fractions M1 - M4 are all slightly reversely discordant with overlapping error ellipses (Fig. 7-3). A weighted average of monazite fractions M1 to M4 yields a $^{207}\text{Pb}/^{235}\text{U}$ age of 112.3 ± 1.3 Ma that is interpreted to represent the crystallization age of CG97-712 (Dycer Creek Stock).

7.2 Implications of U-Pb Age Analysis

Dating of igneous zircons from the Yukon-Tanana Terrane is problematic owing to Proterozoic inheritance. This is reflected in most of the zircon fractions analysed from the Dycer Creek Area. One exception is the multigrain fraction E from YT95-BS082 that consists of $<64\mu\text{m}$ prismatic needles. Inheritance of radiogenic Pb results from the

nucleation of new zircon crystals on remnant xenocrystic zircons of an older population.

The small grains selected for fraction E are interpreted to represent a population that

nucleated after all xenocrystic material had been incorporated in previous zircon crystals.

Such grains would have nucleated at new sites and are therefore free of inherited Pb. The

first stage of analysis has proved promising as analytical data is consistent with $<64\mu\text{m}$

grains containing no inherited Pb component. Small prismatic crystals, although difficult to

work with, may represent a solution to the inheritance problem.

A 357 ± 1.6 Ma U-Pb zircon age for the Mendocina Orthogneiss confirms the presence of highly deformed Late Mississippian to Early Devonian orthogneiss in the Teslin Zone. A Devonian magmatic event is widely documented throughout the Yukon-Tanana Terrane (Mortensen, 1992). The age of the Mendocina Orthogneiss is consistent with the interpretation that rocks of the Teslin Zone correlate with those of the Yukon-Tanana Terrane. A Devonian age does not place limits on protolith ages of the metasedimentary country rocks as it remains unclear whether the granitoid body was emplaced tectonically or as an intrusion.

Timing of intrusion of the Dycer Creek Stock (112.3 ± 1.3 Ma) is interpreted to pre-date large-scale F_3 folding and to be synchronous with or post-date the development of S_{T2} . We interpret intrusion of the Last Peak Granite (95.7 ± 0.6 Ma) into the western domain to post-date large-scale F_3 regional folding and to precede or be contemporaneous with development of the local dextral strike-slip shear zone. Age constraints on the timing of regional deformation can be drawn from these observations: i) regional S_{T2} transposition fabrics in the Dycer Creek area had developed by 112 ± 1.3 Ma; ii) large-

scale F₃ folding occurred between 112.3 ± 1.3 Ma and 95.7 ± 0.6 Ma; iii) development of the local dextral strike-slip shear zone, in the Teslin Zone, occurred after 95.7 ± 0.6 Ma

7.3 Conclusions

Timing of the D₂ transposition event remains poorly constrained, but regional foliations are thought to have developed by at least 112 Ma. Development of steep attitudes in the eastern Cassiar Platform occurred in the Cretaceous between 112 and 96 Ma, not in the Early Jurassic as previously proposed by Hansen (1989).

Future work includes further analysis of $<64\mu\text{m}$ zircons from YT95-BS082 to test the theory that this population does represent igneous crystals free of inheritance. To date, analysis of zircon fractions from CG96-711 (Little Lake Granite) have been problematic due to Pb inheritance. Analysis of a sub $74\mu\text{m}$ zircon population will attempt to constrain the age of the granitoid body.

Figure 7-1 a&b: Concordia diagrams of YT95-BS082 (Mendocina Orthogneiss) U-Pb zircon analyses (see text for interpretation). Figure 7-1b is a detailed concordia diagram for the concordant zircon fraction E.

Figure 7-2 a&b: Concordia diagrams of YT95-BS107 (Last Peak Granite) U-Pb monazite and zircon analyses (see text for interpretation). Figure 7-2b shows reversely discordant monazite fractions M1-M3

Figure 7-3 a&b: Concordia diagrams of CG97-712 (Dycer Creek Stock) U-Pb monazite and zircon analyses (see text for interpretation). Figure 7-3b shows reversely discordant monazite fractions M1-M4.

Table 7-1: Geochronology samples collected from the Dycer Creek area during the 1996 and 1997 field seasons.

Sample #	Location of Sample	Rock Type	Field Relationships	Constrains
YT95-BS082 Mendocina Orthogneiss	- southern limit of the exposed orthogneiss in the Dycer Creek shear zone	- mylonitic peraluminous orthogneiss	- concordant contacts with regional foliation - no chilled margin or contact aureole - mylonitic S-C fabric with shear planes parallel to regional foliation - mylonitic foliation interpreted to represent reactivated regional fabric	- no evidence of D_1 deformation - pre to syn-kinematic with respect to D_2 - places a lower age bracket on D_2 regional ductile deformation
YT95-BS107 Last Peak Granite	- north of the mapping limits in the Dycer Creek shear zone	- two mica potassic feldspar granite	- concordant to discordant contacts with regional S_{T2} and S_3 foliation - chilled margin and contact metamorphism is documented (Harvey <i>et al.</i> , 1996) - protomylonitic to mylonitic foliation that is concordant to regional S_{T2} foliation	- places an upper bracket on D_2 deformation - places an upper bracket on D_2 deformation - places a lower bracket on Dycer Creek shear zone
CG97-711 Little Lake Granite	- western margin of the granite	- protomylonitic peraluminous orthogneiss	- concordant contacts with regional S_{T2} foliation - internal foliation concordant to regional S_{T2} foliation - no chilled margin or contact metamorphism observed	- pre to syn-kinematic with respect to D_2 deformation - place a lower age bracket on D_2 regional deformation
CG97-712 Dycer Creek Stock	- approximately 700 metres northeast of CG97-711	- weakly foliated biotite-muscovite quartz monzonite	- concordant to discordant contacts with country rock - weak internal foliation parallel to regional foliation that is folded by F_3 - pre to syn-kinematic contact metamorphism associated with intrusion - undeformed-d discordant to deformed- concordant dykes of stock intrude the Little Lake Granite	- syn to post-kinematic with respect to D_2 - pre-kinematic with respect to D_3 - places an upper bracket on D_2 deformation and a lower bracket on D_3 deformation in the eastern domain

Table 7-2: U-Pb Analysis of Geochronology Samples from the Dycer Creek area

Fraction ^a	Wt. (ug) ^b	U (ppm) ^c	Pb ^e (ppm) ^c	²⁰⁶ Pb/ ²⁰⁴ Pb	Pb (pg) ^d	²⁰⁶ Pb %	Th/U	Corr. Coeff.	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁶ Pb/ ²³⁸ U Age (Ma) ^f	²⁰⁷ Pb/ ²³⁵ U Age (Ma) ^f	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma) ^g	% Disc.
YT95-B5082: Mendocino Orthogneiss															
A	91	281	33	5181	34	11.2	.4926	.9462	0.10913±.12	1.8577±.13	0.12346±.04	667.7±1.5	1066.2±1.7	2006±1.5	70.1
B	134	373	40	6404	51	8.4	.3419	.9636	0.10332±.32	1.4049±.32	0.09861±.09	633.9±3.9	891±3.9	1598.1±3.3	63.3
E	10	676	36	1484	17	4.1	.1444	.1346	0.03708±.39	0.4225±.26	0.03369±.44	357.8±0.7	357.8±1.6	357.9±19.9	0.0
YT95-B5107: Last Peak Granite															
A	94	366	14	4984	17	8.4	.3386	.8320	0.03769±.10	0.49971±.12	0.09613±.06	238.5±0.5	411.5±0.8	1150.6±2.3	86.1
B	145	701	14	375	386	5.6	.2042	.5034	0.02133±.16	0.18391±.75	0.06253±.69	136.1±0.4	171.4±2.4	692.4±29.3	81.2
D	52	293	5	884	21	5.7	.2020	.4916	0.01916±.09	0.14203±.37	0.03376±.33	122.4±0.6	134.9±1.1	361.0 ± 15.0	66.7
E	30	1313	24	2911	17	6.1	.2162	.6480	0.01900±.09	0.13945±.16	0.03323±.13	121.3±0.4	132.5±0.5	338.7±5.7	64.8
F-1	50	603	13	2019	21	6.1	.2212	.6388	0.02246±.47	0.18874±.43	0.06092±.39	143.2±1.4	175.5±1.4	636.4±16.6	78.4
M1	15	1458	36	1892	11	43.9	2.563	.5377	0.01546±.10	0.10086±.22	0.04730±	98.9±0.2	97.6±0.4	64.2±8.7	-54.7
M2	7	2876	385	1162	17	89.4	27.66	.4465	0.01559±.10	0.10085±.40	0.04691±	99.7±0.2	97.6±0.7	44.7±17.4	-124.
M3	17	809	109	1959	7	89.6	28.27	.5929	0.01540±.18	0.10038±.30	0.04733±	98.6±0.4	97.3±0.6	66.8±11.5	-48.0
CG97-712: Dycer Creek Stock															
G	11	844	27	942	20	6.8	.2579	.8815	0.03209±.28	0.31841±.32	0.07193±.15	203.6±1.1	280.7±1.6	984.6±6.2	80.6
M1	8	3006	363	3210	9	86.6	21.12	.8379	0.01784±.49	0.11747±.53	0.04774±.29	114.0±1.1	112.8±1.1	86.6±13.9	-31.9
M2	7	2741	377	1446	16	88.2	24.40	.6776	0.01791±.24	0.11717±.34	0.04743±.25	114.5±0.5	112.5±0.7	71.1±12.0	-61.5
M3	3	2706	476	301	33	90.8	32.21	.5579	0.01786±.31	0.11568±.11	0.04697±.98	114.1±0.7	111.1±2.4	47.8±47.0	-140.
M4	3	2736	513	1369	8	91.2	34.11	.4658	0.01807±.18	0.11838±.49	0.04749±.44	115.5±0.4	113.6±1.1	74.1±20.7	-56.3

^a s=single crystal, Letters (A-G)=zircon, M=monazite; ^b weighing error is 1 microgram; ^c radiogenic Pb; ^d = Radiogenic Pb, corrected for spike and Pb fractionation of 0.09% ± 0.03%/amu; ^e total common Pb in analysis corrected for spike and fractionation; ^f corrected for Pb and U laboratory blank where 208/204 : 207/204 : 206/204 are 19.02:15.65:38.23, errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma; ^g corrected for common Pb and laboratory blank, errors are 2 standard errors of the mean in Ma.

CHAPTER 8: DISCUSSION

FIGURES:

Figure 8-1: Proposed tectonic model of the Yukon-Tanana Terrane and Teslin Zone

Figure 8-2: Simplified block diagram of Teslin Zone and eastern Cassiar Platform geometry

8.1 Introduction

Detailed geologic and structural mapping within the Dycer Creek area (this paper), and at Last Peak (de Keijzer and Williams, 1998; de Keijzer *et al.*, in press), reveal polyphase fabric relationships and an overall structural geometry that is not consistent with previous models of evolution proposed for the Teslin Zone (Tempelman-Kluit, 1979; Hansen, 1989, 1992 a,b; Stevens and Erdmer, 1996, 1997). What is the significance of these new findings? Is it likely that the Dycer Creek area represents a suture zone or structurally divergent shear zone; or must an alternative hypothesis pertaining to the structural evolution of the Teslin Zone be proposed?

8.2 Comparison of Structural Data with Previous Models of Evolution

8.2.1 Does the Teslin Zone represent a preserved suture zone?

Previous authors have based the suture zone model on four key observations (Tempelman-Kluit, 1979; Hansen, 1989, 1992a): i) the presence of eclogite facies metamorphic rocks within the Nisutlin assemblage of the Teslin Zone; ii) the apparent presence of steeply west-dipping mylonitic fabrics containing two penetrative but variably oriented stretching lineations; iii) an apparent lack of stratigraphic continuity within the Teslin Zone; iv) the presence of tectonic klippen, of Yukon-Tanana affinity, preserved to the east of the Teslin Zone. None of these observations require a suture zone to be present in the Teslin Zone, and most can be explained with a model that is consistent with the structural geometry observed in the Dycer Creek and Last Peak areas.

i/ Significance of eclogite facies metamorphic rocks within the Teslin Zone

The presence of these high pressure rocks is interpreted as evidence that subduction related processes played a role in the tectonic evolution of the Yukon-Tanana Terrane (Tempelman-Kluit, 1979; Hansen, 1989). How important a role remains an interesting question that has been addressed by several authors (Erdmer *et al.*, 1996; Stevens and Erdmer, 1996; Hansen, 1992a). This question may be answered by determining the relationship between the origin of the eclogitic rocks and their host. Eclogite within the Teslin Zone may have 'primary' contact relationships; in this scenario, they would share contact relationships that developed within a subduction channel in Permian time. An alternative possibility is that the pods' contact relationships may be 'tectonic' in nature, and were incorporated into the assemblage at some time after Permian peak metamorphism occurred.

The Last Peak eclogite is a compositionally layered pod, ~10 m in length, set in a mylonitic fabric of the Nisutlin assemblage metasedimentary rocks. Eclogite from the Last Peak area yielded a metamorphic zircon age of 269 ± 1 Ma (Creaser *et al.*, 1997). Isotopic evidence, in the form of $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite cooling ages (236 ± 1 Ma; Erdmer *et al.*, 1996), provides important clues to the early history of the eclogite pods. Teslin Zone rocks (Hansen, 1989) and the rest of the Yukon-Tanana Terrane show much younger cooling ages (through 350°C) ranging from Mid Jurassic to Early Cretaceous time (Wilson *et al.*, 1985; Armstrong, 1988; Dusel-Bacon *et al.*, 1989). Creaser *et al.* (1997) pointed out that the muscovite from within the Last Peak eclogite was not isotopically reset during Mesozoic accretion (the temperature of the eclogite did not reach the diffusive temperature of argon (350°C) in the muscovite) though they did not offer an explanation as to why this did not occur.

Discrepancies in the ages of deformation and metamorphism between the eclogite and its surrounding rocks also exist. The mylonitic foliation observed in the eclogite is interpreted to have developed within the subduction zone channel, synchronous with Permian peak metamorphic conditions (Erdmer *et al.*, 1998); regional fabrics and metamorphism of the Anvil assemblage, also thought to have developed in the subduction zone channel, are interpreted to be mid-Jurassic in age (Hansen, 1989).

Erdmer *et al.* (1998) reported that the relationship of many high-pressure rocks to the host rocks in the Yukon-Tanana Terrane is commonly unclear. This is observed in the Teslin Zone, as field evidence of primary contact relationships have been obliterated by mid-Jurassic (Hansen, 1989) regional ductile deformation. Field data remains inconclusive as to whether the eclogite is 'primary' or 'tectonic' in nature. A 'tectonic' model better explains the contrasting cooling and deformation ages between the eclogite and its surrounding metasedimentary rocks.

A possible 'tectonic' model involves eclogite-facies metamorphism and associated deformation that occurred in rocks of unknown origin that peaked in the Permian (Fig. 8-1 i). Exhumation of these rocks, that must have involved a major tectonic event or continent-continent collision (possibly D_1 thought to have occurred in the late Permian; Mortensen, 1992), resulted in cooling of the eclogite below 350°C. The eclogite remained "refrigerated" below 350°C in an upper crustal environment until it was incorporated into rocks of the Teslin Zone. Incorporation of the eclogite most likely occurred during regional thrust faulting that postdated D_1 deformation (Fig. 8-1 ii; Mortensen, 1992). The fact that these are brittle features supports a cool upper crustal environment that would "refrigerate" the eclogite. The eclogite and surrounding Nisutlin rocks were deformed during D_2 deformation (Fig. 8-1 iii), at which time the entire Teslin Zone

underwent greenschist- to lower amphibolite-facies metamorphism (de Keijzer *et al.*, in press; Stevens *et al.*, 1996; Hansen, 1989; this study). Hansen (1989) obtained rapid cooling rates of up to 50°C / Ma in the Late Jurassic that post dated regional deformation. Therefore, the eclogite may not have had sufficient time to heat up past 350°C before the entire Teslin Zone cooled below 350°C - the closure temperature of muscovite.

A tectonic model describing the incorporation of eclogite facies rocks into the Teslin Zone, as proposed in Figure 8-1, would best explain the discrepancy in cooling ages and age of deformation. Most importantly, this model suggests that rocks of the Teslin Zone did not necessarily undergo subduction-related metamorphism and deformation and that subduction-related processes did not play a major role in the evolution of the Teslin Zone.

ii/ Presence of an apparent steeply dipping mylonitic foliation containing two variably oriented stretching lineations

Mylonitization of rocks of the Teslin Zone within a west-dipping subduction zone in Jurassic time is identified as the primary mechanism of fabric development in the Teslin Zone (Tempelman-Kluit, 1979; Hansen, 1989, 1992a). The steeply west-dipping mylonitic fabric is thought to represent the original orientation of the shear zone (subduction channel) when the regional fabrics were developed. Down-dip and shallowly plunging stretching lineations were reported throughout the Teslin Zone and were interpreted to represent zones of dip-slip and strike-slip shear (Hansen, 1989).

Rocks of the Teslin Zone are not characterized by mylonitic foliations, nor do they contain two variably oriented stretching lineations. The regional foliation has been clearly demonstrated to be a transposed stratigraphy resulting from isoclinal folding (de Keijzer *et al.*, in press; this study).

Shallowly and steeply plunging stretching lineations are observed within the Dycer Creek area, although their geographic distribution greatly differs from those reported by Hansen (1989). In total, L-S tectonites interpreted to have developed within small-scale shear zones, underlie a maximum of 20% of the mapping area. This is consistent with observations made by de Keijzer (pers. comm., 1998) to the north.

Observations made by Hansen (1989), concerning the structural fabric of the Teslin Zone, are invalid and subsequently any tectonic model based on these observations is invalid as well. This does not mean that a suture zone cannot exist in the Teslin Zone, it simply means that the style of deformation in the zone is not as Hansen interpreted it.

iii/ Lack of stratigraphic continuity in the Teslin Zone

Previous authors have suggested that the discontinuous nature and apparent random distribution of Teslin Zone lithologic map units is a result of chaotic deformation in the subduction channel. They suggest that the map patterns observed are representative of a tectonic melange in which different rock types are commonly tectonically bounded and no coherent geometry is preserved (Seyfert, 1987).

A coherent geometry and units that are mappable along strike for ~15 km have been documented within the Dycer Creek area (this thesis), as well as to the north (de Keijzer *et al.*, in press) and to the south (Stevens *et al.*, 1996). Contacts between the different units of the Nisutlin assemblage, especially the calcareous graphitic schist, graphitic phyllite and quartzite that make up the bulk of the assemblage, are typically gradational and are interpreted to represent primary gradational lithologic contacts. Primary lithologic contacts within Nisutlin assemblage rocks have been documented by Stevens *et al.* (1996) and de Keijzer *et al.* (in press). There is no evidence of

a “knocker” type melange in which lithologies are tectonically bounded.

The complex and seemingly irregular distribution of units in the Teslin Zone may be explained by the fact that the geometry of the Teslin Zone is controlled by F_2 isoclinal folds that fold a transposed stratigraphy rather than deformation within a trench melange.

iv/ Presence of tectonic klippen to the east of the d'Abbadie fault zone

Allochthonous rocks, similar to those observed in the Teslin Zone, are exposed as klippen lying on gently west-dipping thrust faults are documented as far as 400 km north-east of the Teslin Zone. Previous authors have suggested that the Teslin Zone represents the root of these klippen (Tempelman-Kluit, 1979; Hansen, 1989, 1992a).

The fact that these klippen exist cannot be disputed, though they do not represent strong evidence of subduction. In fact, the presence of tectonic klippen observed on North American cratonal strata is direct evidence of *obduction* of Anvil rocks onto the North American platform (Fig. 8-1). Obduction-related deformation is identified by de Keijzer *et al.* (1998, in press) as the primary mechanism of fabric development in the Teslin Zone.

Conclusions

Field data collected in the Dycer Creek area are not consistent with the suture zone model. Mechanisms of fabric development are primarily related to transposition and polyfolding rather than mylonitization associated with chaotic deformation. The presence of eclogite-facies rocks are better explained by a model involving tectonic incorporation of the eclogites into the Teslin Zone late in its history, rather than a model in which the entire Nisutlin assemblage was subducted and then rapidly exhumed.

8.2.2 A Structurally divergent Teslin Zone of Stevens and Erdmer, 1996

Stevens and Erdmer (1996) proposed a tectonic model in which the Teslin Zone represents a structurally divergent zone resulting from Late Triassic to Early Jurassic transpression and tectonic wedging. Structural and geochronologic data collected in the Dycer Creek area are not consistent with these interpretations. The following discussion involves the presentation of the key observations and interpretations that led to this particular model, and why they are not valid for the Dycer Creek area.

Apparent structural divergence in the southern Teslin Zone

Stevens and Erdmer (1996), working to the south of the Dycer Creek area, documented northeast and southwest directed shear on regional foliation planes, which they interpreted as evidence of structural divergence in the Teslin Zone. In contrast, structural divergence is not documented in the Dycer Creek area. Rather, F_2 macroscopic folds have a consistent northeast vergence throughout the Dycer Creek area and the Two-Cub Antiform is most likely part of a southwest verging structure. There is no single generation of structures that show opposing vergence in the Dycer Creek area, with the exception of F_2 parasitic folds on the opposite limbs of larger-scale F_2 folds. Minor folds similar in style to F_2 folds in the Dycer Creek area were documented by Stevens and Erdmer (1996) although they do not mention the presence of larger-scale folds of the same generation. It is possible that the apparent structural divergence observed in the southern Teslin Zone is a result of microstructures and fabric relationships on opposite limbs of a larger-scale fold system of the same generation that has not been recognized.

Regional foliation development

According to Stevens and Erdmer (1996, p. 1345), rocks of the southern Teslin Zone are

characterized by “their strong, ductile, S and L-S tectonite fabrics”. In contrast, the regional foliation documented in the Dycer Creek area is a result of transposition during D_2 deformation (Chapter 4; this thesis). Fabric relationships and microstructures cited by Stevens and Erdmer (1996) are not characteristic of rocks of the western domain. Lineated protomylonitic and mylonitic tectonites are documented in the Dycer Creek area, but are limited to local Cretaceous shear zones that overprint regional fabrics. It is possible that most of the foliations described by Stevens and Erdmer (1996) are associated with Cretaceous strike-slip shear (analogous to the Dycer Creek shear zone) and postdate regional ductile deformation.

Many of the structural domains, including all of transect AA' and most of BB' show variably developed foliations that are axial planar to tight isoclinal (rootless) folds (Table 1; Stevens and Erdmer, 1996). This of course is similar to relationships observed in the Dycer Creek area and is interpreted as evidence of transposition during isoclinal folding. Alternatively, Stevens and Erdmer (1996), suggested that the folds themselves were transposed.

Large-scale late folding in the southern Teslin Zone?

Stevens and Erdmer (1996; p. 1355) suggested that the variably steeply dipping regional fabrics are “a product of the deformation that produced the L-S tectonites”. These variably steep-dipping regional fabrics are observed in the Dycer Creek area, but are interpreted to have developed as a near-horizontal transposition foliation that was rotated into a steep attitude by late, large-scale folding.

Stevens and Erdmer documented local open mesoscopic- to macroscopic-scale folds that postdate development of regional L-S tectonite fabrics. As described, these folds and associated crenulation lineations and cleavage seem to be identical to those F_3 structures documented in the

Dycer Creek area (Stevens and Erdmer, 1996, Fig. 4 c,d). Structural domains (A2, B2, C4), in which these folds are common, show variably oriented regional foliations (Table 1; Stevens and Erdmer, 1996).

Stevens and Erdmer point out that “unfolding” of late folds does not bring mineral lineations into a common orientation. They use this as evidence that this late folding event, which postdates regional foliation development, is not responsible for the variation in dip of regional foliations. Rotation of mineral lineations (about a late fold axis) into a parallelism, would only occur if the mineral lineations shared a common orientation prior to late folding. By their own admittance (Stevens and Erdmer, p. 1355), “The difference in mineral lineation orientations from one domain to the next suggests that each records a specific kinematic regime ...”. These mineral lineations most likely did not develop with a common orientation. Therefore, late folding is still a possible and likely mechanism to explain the variable and steeply dipping regional fabrics of the Teslin Zone.

Conclusions

Data from the Dycer Creek area are not consistent with the interpretations made by Stevens and Erdmer (1996). Furthermore, the model presented by Stevens and Erdmer (1996) requires the eastern Cassiar platform to have undergone a completely different deformation history, a requirement that is not met in the Dycer Creek area. Therefore, a transpressional zone of structural divergence does not appear to be a plausible model for the evolution of the Dycer Creek area or the Teslin Zone itself. It appears that Stevens and Erdmer, as did Hansen, underestimated the role of folding in the evolution of the Teslin Zone. This again may reflect the style of the study that consisted of widely spaced transects or traverses instead of traditional

mapping of geologic units and structures. It also appears that many of the field observations made by Stevens and Erdmer (1996) may be explained by folding processes interpreted to have occurred in the Dycer Creek area.

8.2.3 The Grizzly Synform of de Keijzer *et al.* (in press)

F_1 and F_2 folding events documented in the Dycer Creek area can be correlated with similar events to the north at Last Peak (de Keijzer, pers. comm.). However, inconsistencies between the two areas occur when comparisons of the large-scale, post-transposition folding events are made. de Keijzer *et al.* (1998, in press) suggest that the F_3 Grizzly synform, located on the west side of the d'Abbadie fault zone (Fig. 8-2), is most likely part of a north-east vergent structure, whereas the F_3 Two-Cub antiform, located on the east side of the d'Abbadie fault zone (Fig. 8-2), is most likely part of a southwest vergent structure. At Last Peak, de Keijzer *et al.* suggested that the Grizzly synform is responsible for the steeply dipping regional foliation of the Teslin Zone, and in the Dycer Creek area, the Two-Cub antiform is responsible for the steeply dipping regional foliation of the Cassiar Platform (Chapter 4, this thesis). What is the relationship of these two fold structures? Are they a conjugate pair of folds that form a regional-scale box fold - or are they of different generations?

There is no evidence of a change in regional dip from the Last Peak area to the Dycer Creek area (unpublished mapping, Harvey *et al.*, 1996). The western domain of the Dycer Creek area must therefore be situated on the steep overturned short limb of the Grizzly synform (Fig. 2-3). F_3 minor folds documented in the western domain (Chapter 4, this study) cannot be parasitic folds of the F_3 Grizzly synform, as they are of the incorrect vergence. A correlation between F_3 minor folds documented in the Dycer Creek area and F_4T and F_4cT folds documented

in the Last Peak area (de Keijzer *et al.*, Table 1, in press) can be made. These minor folds of the western domain are correlated with the Two-Cub antiform located in the eastern domain.

F_3 folding documented in the Last Peak area is not observed in the Dycer Creek area. Development of F_3 parasitic folds of the Grizzly synform decreases away from the hinge area and ~3 km west of the hinge area there are significantly fewer F_3T structures (de Keijzer *et al.*, in press). It is conceivable that regional foliations of the western domain were rotated into a steep attitude by the Grizzly synform, but minor folds associated with the event did not develop as a result of their distal relationship with the hinge area. The hinge area of the Grizzly synform is not observed in the western domain of the Dycer Creek area because it is most likely truncated by the north - south d'Abbadie fault zone.

8.2.4 Conclusions

Field data recorded in the Dycer Creek area do not support the interpretations and models presented by previous workers (Tempelman-Kluit, 1979; Hansen, 1989, 1992a and Stevens and Erdmer, 1996). The data are consistent with a model of northeast directed thrusting, transposition and late folding presented by de Keijzer *et al.* (in press). Data presented by Hansen (1989) and Stevens and Erdmer (1996) is also compatible with the model presented by de Keijzer *et al.* (in press).

8.3 Tectonic Implications of the Dycer Creek area Structural Geometry

Tectonic significance of the Teslin Zone

It has been proposed in the literature that the Cassiar Platform and Yukon-Tanana Terrane represent distinct terranes with significantly different origins (Wheeler and McFeely, 1987).

Juxtaposition of these two terranes in the Teslin Zone would require a terrane boundary

separating the Yukon-Tanana Terrane from the eastern Cassiar Platform (Fig. 8-1 iv). This boundary, exposed to the north at Last Peak, is interpreted to have developed as a low-angle detachment associated with northeast-directed thrusting of Yukon-Tanana Terrane rocks over the Cassiar Platform (Fig. 8-2 iii; de Keijzer *et al.*, in press). The terrane boundary was then subsequently deformed and folded by several generations of regional deformation, including Cretaceous dextral strike-slip shear and extension (Gallagher *et al.*, 1999; de Keijzer *et al.*, in press).

Temporal relationships between the development of the near horizontal regional transposition foliation and northeast thrusting along the detachment remain unresolved (de Keijzer *et al.*, in press). Regional, post-transposition folding that rotated the regional foliation into a steep attitude, is possibly related to northeast-directed transport as well. One possible explanation for the formation of the Grizzly synform and the Two-Cub antiform is the rheologic contrast that occurred at the terrane interface at the time of folding. The steep belt coincides with a major facies change from competent carbonate platformal rocks in the east to incompetent graphitic slates and siliciclastics in the west. We suggest that the localization of the steep fabrics may be a result of this rheologic contrast, such that the stiff carbonates acted as a back stop during northeast directed transport, and resulted in large-scale northeast and southwest vergent folding.

Location of the suture zone

If the Teslin Zone does not represent a suture between allochthonous North American rocks and accreted terranes, then where does the suture occur? The Yukon-Tanana Terrane contains eclogite-facies rocks (Creaser *et al.*, 1997; Erdmer *et al.*, 1998) and is therefore related, either indirectly or directly, to subduction processes. If one is to use the presence of eclogites as

evidence of a suture zone and attempt to locate it, they must first restore the eclogites back to their original location. Klippen of Yukon-Tanana terrane rocks, such as the St. Cyr klippe, lie up to 100 km to the east of the present day Teslin Zone, which is interpreted to be their root (Erdmer *et al.*, 1998; Fig. 1). Restoration of these klippen places the Last Peak eclogite ~100 km to the southwest of its present day location. This is a minimum estimate as there has been significant shortening associated with regional D₂ and D₃ deformation as well. It is therefore reasonable to propose, as did Brown *et al.* (1998), that the suture zone lies a significant distance to the southwest of the Teslin Zone and has been overprinted or obliterated by younger deformation events.

The westernmost extent of the North American craton in the northern Cordillera must also extend a significant distance to the southwest of the Teslin Zone's present day location. This is in contrast with interpretations made by workers such as Tempelman-Kluit (1979) and Hansen (1989, 1992a) who suggest that the Teslin Zone marks the westernmost extent of the craton. Presence of North American basement underneath the Yukon-Tanana Terrane is supported by i) possible documentation of Cassiar Platform rocks to the east of the Teslin Zone (de Keijzer *et al.*, in press); ii) the presence of Proterozoic inheritance in many of the Yukon-Tanana Terrane's plutonic assemblages (Mortensen, 1992); iii) preliminary crustal seismic velocity profiles conducted by LITHOPROBE (Creaser and Spence, 1999).

Proterozoic Pb inheritance in zircons from many of the plutonic assemblages of the Yukon-Tanana Terrane has been documented by several workers (Mortensen, 1992; Chapter 7, this thesis). The source region for these xenocrystic zircons, interpreted to be between 1.66 and 3.05 Ga, can only be a crustal body (presumably the North American craton). This supports a

model in which the North American craton extends in the subsurface beneath the Yukon-Tanana Terrane.

Preliminary crustal seismic velocity analysis from LITHOPROBE SNORCLE Line 31 reveals that the Moho occurs as a significant velocity discontinuity (Creaser and Spence, 1999). The Moho shows topography, but no significant change in dip is observed along strike of the Cassiar - Yukon-Tanana boundary. In the preliminary model, there are no steeply dipping reflectors that propagate all the way to the Moho. Instead, a set of gently southwest-dipping reflectors, as predicted by Brown *et al.* (1998), is observed. There is therefore no evidence present in the seismic survey that supports a crustal-scale shear zone separating two contrasting tectonic plates at the Cassiar - Yukon-Tanana Terrane boundary.

Figure 8-1: A simple tectonic model depicting the possible evolution of the Yukon-Tanana terrane from the mid-Permian to present day.

i) Permian:

Roll back of the eastward subducting slab rifted a fragment of the YTT away from the North American craton; this resulted in the development of a back arc basin. Rocks of the Anvil assemblage formed the "Anvil" ocean that separated the arc from the North American craton. Cassiar Platform protoliths were deposited as passive continental shelf sediments. Sedimentary rocks of the Yukon-Tanana terrane are interpreted to have a bimodal source that consisted of distal North American and back-arc basin sediments. (Creaser, pers. comm., 1999). Eclogite-facies metamorphic rocks developed within the underplated material and was rapidly exhumed (Erdmer *et al.*, 1997).

ii) Early to Mid-Triassic:

Collision of a crustal body or other arc, outboard of the Permian arc, resulted in the obduction of Anvil and eclogite-facies rocks (Erdmer *et al.*, 1997) onto the North American shelf. This collision event is possibly correlative with regional D₁ deformation documented in YTT (Mortensen, 1992). Incorporation of the eclogite-facies rocks into the YTT possibly occurred during post-D₁ brittle thrust faulting that has been reported throughout the YTT (Mortensen, 1992). The eclogite remained below 350°C while situated in a cool upper crustal environment.

iii) Jurassic:

Continued obduction and thickening of the thrust sequence onto the Cassiar Platform resulted in a transition from brittle to ductile deformation that consisted of the transposition of CP and YTT rocks (D₂ deformation). Syn-D₂ upper greenschist- to lower amphibolite-facies metamorphism occurred at this time, although it was not long enough lived to isotopically reset the eclogite (Chapter 8, Section 8.1 i).

iv) Early Cretaceous (pre-Cretaceous dextral translation and extension):

Large-scale folding of the transposed lithologic sequence occurred at the Cassiar Platform - Yukon-Tanana terrane boundary. This folding event resulted in the development of a steeply dipping belt of rocks - that is now known as the Teslin Zone. Rocks of YTT and Anvil assemblage are exposed as klippen, up to 100 km east of the Teslin Zone, which lies structurally above the Cassiar Platform.

Magmatic Arc

Back-Arc Basin

W

i) Permian

ii) Early to
Mid Triassic

iii) Mid-Jurassic

iv) Early
Cretaceous

LEGEND

Cassiar Platform - Proximal North American sediments

Yukon-Tanana Terrane - Distal North American and back-arc basin sediments

Anvil Assemblage - 'MORB' back-arc mafic volcanics

Underplated material

Mantle

Eclogitic Rocks

TESL

Present day location
of the Teslin Zone

E

10 km

0 km amsl

10 km

20 km

30 km

Location of iii and iv

Incorporation of eclogite
into YTT rocks

'terrane'
boundary

'terrane'
boundary

TESLIN ZONE

Figures are not to scale

Figure 8-2: Simplified block diagram of the projected geometry of the Dycer Creek area (this study) and the Last Peak area (de Keijzer *et al.*, in press). The attitude of regional foliations and transposed stratigraphy are controlled by late large-scale folding. These events, the Grizzly synform in the Teslin Zone, and the Two-Cub antiform in the eastern Cassiar Platform may represent a conjugate set of folds.

CHAPTER 9: CONCLUSIONS

The origin and tectonic history of the Yukon-Tanana Terrane, the Northern Cordillera's largest tectonostratigraphic terrane, remains a controversial issue. Its location between slightly displaced North American strata and demonstrably allochthonous terranes has gained it much attention, as it is an important key to unraveling the early accretionary history of the Northern Cordillera. This study, in conjunction with de Keijzer and Williams to the north, has contributed to the overall understanding of this terrane's tectonic history by the addition of valuable structural and geochronologic data to the growing database. The major findings in the Dycer Creek area are as follows:

1) Three different generations of regional structures are correlated on both sides of the d'Abbadie fault zone (Section 4.3). This is taken as evidence that the entire Dycer Creek area has undergone a common deformation history. The lack of contrasting structures and metamorphic grade between the eastern and western domains is consistent with little dextral strike-slip displacement associated with the d'Abbadie fault zone. The d'Abbadie fault zone is therefore not considered to represent a major tectonic boundary.

2) The S_{T2} regional foliation of the Dycer Creek area is a transposed, previously folded (F_1), lithologic sequence. It is interpreted, based on regional arguments, to have developed as a flat lying or gently west-dipping transposition foliation during F_2 isoclinal folding (Chapter 4) and possibly northeast directed thrusting (de Keijzer *et al.*, in press). S_{T2} was then locally rotated into a steep attitude during regional F_3 (Grizzly synform) and F_4 (Two-Cub antiform) folding. The Teslin Zone represents a fold and thrust belt that developed during northeast directed transport of the Yukon-Tanana Terrane onto the Cassiar Platform.

3) Timing of D_2 regional deformation remains poorly constrained as little detailed geochronology has been conducted in the area. U-Pb geochronologic analysis completed in this study confirms the presence of Late Mississippian orthogneiss (Mendocina orthogneiss; 357.8 ± 1.6 Ma) and Cretaceous granite (Last Peak Granite; 95.7 ± 0.6 Ma) in the Teslin Zone. Studies also suggest that very small ($<64\mu$) prismatic zircons may offer a solution to the Proterozoic inheritance commonly encountered in the Yukon-Tanana Terrane. Studies have bracketed the formation of the Two-Cub antiform between 112 Ma and 96 Ma and placed a lower bracket of 96 Ma on the Dycer Creek shear zone.

4) The suture between the North American craton and allochthonous terranes likely exists a significant distance to the southwest of the Teslin Zone's present day location. The craton is therefore interpreted to extend underneath the Yukon-Tanana Terrane and neighboring allochthonous terranes to the west; an interpretation that is supported by preliminary LITHOPROBE seismic surveys.

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UMI

WESTERN DOMAIN

Last Peak Granite - protomylonitic to mylonitic

Mendocina Orthogneiss - biotite muscovite K

Last Peak Assemblage

Biotite - Chlorite Psammite

Calcareous graphitic schist gradational to phyllite

Graphitic phyllite gradational to quartz graphitic

Quartzite gradational to quartz graphite psammite

Quartz muscovite psammite to schist

Compositionally layered marble and minor calcite

Mendocina Assemblage

Chlorite psammite to chlorite calcite schist (unmapped)

Unmapped or undifferentiated rocks

13

12

6 8 1 1 0 0 0

to mylonitic muscovite bitoite leucogranite

muscovite K-feldspar granite

onal to phyllite (undivided)

artz graphite psammite (undivided)

phite psammite (undivided)

ist

d minor calc silicates

schist (undivided)

;

43

44

45

46

47

48

49

CG97-712
112.3±1.3 Ma

CG97-711

EASTERN

50

51

52

53

CG97-712
12.3±1.3 Ma

CG97-711

EAST RIVER

14

13

12

68
11000

52

53

54000

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UMI

68 18° 00'

17

16

A

15

Figure 4-1: Structural map of the Dyke region showing foliations and lineations of shear zone.

- Pole to foliation S_{T2} or S_M
- Pole to lineation L_{T1-2} or stre

A ——— A
Line of Cross Section

Contacts

Mapped Assumed Inferred

~~~~~ Ductile or Brittle Fault

Topographic features and contours are digitized from the Livingston Creek map sheet 105 E/8. Published by the U.S. Geological Survey.

1984 Magnetic Declination: $29^\circ 32'$ East of Grid North
Annual Change: $-6.8'$
Grid North: $1^\circ 6'$ East of True North

map of the Dycer Creek area compiled from 1995 and 1997 field seasons. Included
d lineations of the eastern and western domains as well as a plot of measurments from

S_{T2} or S_M
n L_{T1-2} or stretch
A'

s Section

ts

ed Inferred

e or Brittle Fault

ntours are digitized from the
105 E/8. Published in 1984.

29° 32' East of Grid North

e North

Stereonet 1: Plot of all S_{T2} foliations and L_{T1-2} intersection
lineations measured in the western domain.

Stereonet 2: Plot of all S_M mylonitic foliations and L_5 stret
lineations measured in the Dycer Creek shear zone.

Stereonet 3: Plot of all S_{T2} foliations and L_{1-2} intersection
lineations measured in the eastern domain.

and 1997 field seasons. Included are stereoplots of
as well as a plot of measurements from the Dycer Creek

S_{T2} foliations and L_{T1-2} intersection
in the western domain.

S_M mylonitic foliations and L_5 stretching
in the Dycer Creek shear zone.

S_{T2} foliations and L_{1-2} intersection
in the eastern domain.

N

6 8 18° 0 0

17

16

15

14

Stru

13

F_1, F_2, F_3 Fold Hinges
with S, Z or M asymmetry
indicated

12

S_1, S_2, S_3 Cleavage

Intersection Lineation

Stretching Lineation

Gneissosity

L3 Crenulation Lineation

6 8 11 0 0 0

5 41 0 0 0

42

Structural Legend

F₂ Antiformal Axial Surface Trace

F₂ Synformal Axial Surface Trace

F₃ Antiformal Axial Surface Trace

Contours: 100 meters

43

44

45

46

46

47

48