Geology of the Tincup Lake area, southwestern Yukon:
Implications for the tectonic evolution of the western
margin of the Yukon-Tanana terrane

by

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ABSTRACT

The Tincup Lake area (NTS 115G/11 14) comprises 12 map units of polyfolded upper greenschist to lower amphibolite facies metasedimentary and metavolcanic rocks correlated with assemblages of Yukon-Tanana terrane. Mapping at a scale of 1: 20,000 was carried out, resulting in the identification of four generations of Late Permian (?) to Late Cretaceous folding and two generations of Late Cretaceous, or younger, faulting in the Tincup Lake area.

Well exposed Cal marble marker units constrain the orientation and geometry of map units in this structurally complex area. In the southern part of the Tincup Lake area the map pattern is controlled by F_2 long limbs that are moderately to steeply S dipping. In the northern part of the Tincup Lake area, S-dipping or gently NW-dipping F_3 limbs control the map pattern.

South-dipping, brittle normal faults crosscut folds and are in turn cut by a network of near-vertical, dextral strike-slip faults striking 305° - 335° and with 500 - 2000m spacing. The steep faults are inferred to be related to the Denali fault and were coeval with regional strike-slip faults that cut the Late Cretaceous to Early Paleocene Ruby Range Batholith to the south.

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ORIGINAL CONTRIBUTIONS

Work on the Tincup Lake area, northern Canadian Cordillera, began in the summer of 2009 with an introduction to rocks of the Yukon-Tanana terrane while working as a member of a Yukon Geological Survey regional mapping party. Work consisted of regional-scale mapping of the Kluane Lake map sheet (NTS 115G). Familiarization with rocks in the Tincup Lake area was carried out during the last week of the 2009 field season.

During the summer of 2010, 12 weeks of geological mapping, at 1:10,000 scale, were carried out in the area bound by Tincup Lake to the east, Nuntaea Creek to the south, Toshingermann Lakes and the Kluane River to the west and 32 Creek to the north. Mapping was carried out by means of 9 fly camps in the study area. Mapping focused on identifying lithologic units (Chapter 2, Map 1), boundaries between units, and defining the structural style of the area (Chapters 3 and 4). During mapping, representative samples of rock types and structural measurements were collected. Map 2 shows the distribution of observations made during the 2009 and 2010 mapping campaign.

Mapping in the Tincup Lake area defined 13 lithologic units (Figure 2.1, Chapter 2; Map 1). Careful observation of contact relationships is the basis for defining a stratigraphic sequence for the area (Chapter 2). Interpretation of this stratigraphic sequence has provided new insight into the evolution of the Permian and older basin into which these sediments were deposited. Marker horizons within the stratigraphic sequence allowed for interpretation of structures in the Tincup Lake area. Analysis of ductile (Chapter 3) and brittle structures (Chapter 4) has provided new insight into the structural evolution of the western Yukon-Tanana terrane between Permian and Cretaceous time.

METHOD OF PRESENTATION

The body of the thesis is presented as three chapters (Chapters 2 – 4). Each chapter addresses a separate aspect of the geological evolution of the Yukon-Tanana terrane in the Tincup Lake area, southwest Yukon. They are prefaced by a general introduction (Chapter 1) and followed by a summary of conclusions (Chapter 5).

Chapter 1 - Introduction to the Yukon-Tanana terrane and geological setting of the Tincup Lake area. This chapter presents an overview of the current understanding of the Yukon-Tanana terrane with a focus on the regional geology of southwestern Yukon, northeast of the Denali fault. This chapter also outlines the goals of this thesis and the methods of analysis used to achieve these goals.

Chapter 2- Lithostratigraphic evolution of the Yukon-Tanana terrane, Tincup Lake area, southwest Yukon. Map units of the Tincup Lake area are described, protoliths are interpreted, and assemblages are defined based on relationships between units. Correlations are made between Tincup Lake area assemblages and assemblages of the Yukon-Tanana terrane based on lithologic and stratigraphic similarities.

Chapter 3 - Geometry and kinematics of polyphase deformation in the Tincup Lake area, Yukon: Implications for the evolution of the western margin of the Yukon-Tanana terrane during the Mesozoic. Four generations of folding, D_1 - D_4 , are documented in the Tincup Lake area. The structures and their geometry are described for each generation of folding are described. An argument constraining the timing of folding events is then presented based on regional-scale crosscutting relationships.

Chapter 4 - Kinematic and geometric analysis of an array of strike-slip faults in the Tincup Lake area, southwest Yukon. Two generations of brittle faulting, D_5 and D_6 , are documented in the Tincup Lake area. Descriptions of the faults and associated fault rocks are presented, followed by a kinematic analysis of slickenfibres and stepped surfaces. Results of the kinematic analysis are then applied in a geometric analysis of faults and shear fractures, for cases that lack kinematic indicators, to determine geometric compatibility with the estimated stress field.

Chapter 5 - Summary of conclusions: geological history of the Tincup Lake area

Maps and Appendices: A geologic map, station map, cross-section and digital database of the Tincup Lake area are included in the back pocket. Maps and digital data are projected using the UTM NAD 83 Zone 7N datum.

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Chapter 1 - Introduction to the Yukon-Tanana terrane and geological setting of the Tincup Lake area

1.1 Introduction

The pre-Late Devonian to Late Permian Yukon-Tanana terrane is a major component of the accreted pericratonic terranes of the Intermontane belt of the Canadian Cordillera. Parts of the Yukon-Tanana terrane have been identified over a distance of 1,500 km from west-central British Columbia, through central Yukon and into east-central Alaska (Figure 1.1). The largest part of the Yukon-Tanana terrane occurs in southern Yukon, between the Tintina fault to the northeast and the Denali fault to the southwest.

In mid- to late Paleozoic time, western Laurentia underwent extension and rifting of the continental margin (Colpron et al. 2006a). As the Laurentian margin rifted the Slide Mountain Ocean opened, forming the Slide Mountain terrane (Colpron et al. 2006a). One of the fragments rifted from the Laurentian margin would become the Snowcap assemblage, the basement assemblage of the Yukon-Tanana terrane (Colpron et al. 2006b). Overlying the Snowcap assemblage are the Finlayson, Klinkit and Klondike assemblages of the Yukon-Tanana terrane. These latter three assemblages record the formation of volcanic arcs, rifted back-arcs, and associated basins, during the opening and subsequent closure of the Slide Mountain Ocean.

This understanding of the Yukon-Tanana terrane is the result of recent studies focusing on the eastern and central parts of the Yukon-Tanana terrane (e.g. Mortensen 1992, Colpron et al. 2006b, Devine et al. 2006, Murphy et al. 2006). Recent work by Murphy et al. (2009), Murphy (2010), Murphy et al. (2011) and Israel et al. (2011) focused on and started to improve our understanding of the western margin (present coordinates) of the Yukon-Tanana terrane.

This study expands on the work of Murphy (2010) in the northern Kluane Lake map area (Figure 1.2). Regional-scale mapping by Murphy (2010) documented east-west-striking belts of

metasedimentary and metavolcanic rocks correlated with the Yukon-Tanana terrane and a structurally overlying assemblage of mafic and ultramafic rocks of the allochthonous Doghead assemblage. In the area between Tincup Lake and Toshingermann Lakes (139° 20' W, 61° 45' N) rocks of the Yukon-Tanana and Doghead assemblage are well exposed.

This study focuses on rocks of the Yukon-Tanana terrane in the Tincup Lake area (Map 1). The broad goals of this study are four-fold. The first goal is to define and characterize map units providing the basis for the second goal, which is to examine potential correlations with previously documented stratigraphic units of Yukon-Tanana terrane. The third goal is to characterize ductile deformation and to determine the relative timing of folding events with respect to the accretion of the outboard terranes and other phases of regional deformation documented elsewhere in the terrane. The final goal is to characterize the brittle deformation and attempt to relate map-scale faulting to regional-scale strike-slip faults.

To achieve these goals, detailed lithological and structural mapping of the Tincup Lake area was carried out at a scale of 1:10,000. Lithological mapping led to the characterization of 10 metasedimentary and 2 metaigneous map units of the Yukon-Tanana terrane and 1 undifferentiated unit of mafic and ultramafic rocks associated with the Doghead assemblage (Map, Chapter 2). Within the rocks correlated with the Yukon-Tanana terrane, 3 distinctive and extensive marble units provided key marker horizons. These marker horizons extend up to approximately 10 km along strike and can be traced through polyphase folds from the southern limit to the northern limit of the study area (Cross-section, Chapter 3). Detailed structural mapping, in conjunction with lithological mapping, led to the characterization of four phases of folding (D_1 - D_4) and two phases of brittle faulting (D_5 - D_6). D_1 - D_4 are interpreted to have occurred between Late Permian and Early Cretaceous time. The timing of D_5 is constrained to post-date

folding and pre-date D_6 , which offsets map-scale normal faults of D_5 (Map, Chapter 4). Mafic-intermediate dikes resembling the Donjek volcanics (Murphy, personal communication, 2011) are mutually crosscutting with respect to strike-slip faults of D_6 , suggesting D_6 is Late Cretaceous or younger.

1.2 Geological Setting

Of the accreted terranes in Yukon, the Yukon-Tanana terrane is the largest, covering an area of approximately 91,000 km² within Yukon (Figure 1.1, Colpron and Nelson 2011). The majority of the Yukon-Tanana terrane is situated between the Tintina fault to the northeast and the Denali fault to the southwest. An outlier of the Yukon-Tanana terrane lies northeast of the Tintina fault and southwest of the Inconnu thrust fault in southeastern Yukon (Murphy et al. 2006).

The Yukon-Tanana terrane has been the focus of geological studies since the late 1800's. A general lack of fossils and the pervasive resetting of the Ar^{40}/Ar^{39} geochronological system made correlation of units and assemblages within the Yukon-Tanana terrane difficult. This difficulty led to a variety of unit and assemblage nomenclatures being used in historical Yukon-Tanana terrane literature. A summary of Yukon-Tanana terrane nomenclature, past and present, is presented in Colpron et al. (2006a). It was not until the application of U-Pb geochronology in conjunction with detailed mapping and geochemical studies that seemingly unrelated assemblages were correlated. This new understanding of the Yukon-Tanana terrane led to the definition of four regional-scale tectonic assemblages within the Yukon-Tanana terrane. These assemblages are the Snowcap assemblage (basement to the Yukon-Tanana terrane, continental margin), Finlayson assemblage (rift to subduction-generated arc), Klinkit assemblage (subduction-generated arc), and Klondike assemblage (subduction-generated arc). This study

follows the nomenclature of Colpron et al. (2006a) in correlating assemblages identified in the Tincup Lake area with those of the Yukon-Tanana terrane.

The following section provides a description of the four assemblages that compose the Yukon-Tanana terrane. Following the description of Yukon-Tanana terrane assemblages is an overview of the geology in the northern Kluane Lake map area, southwest Yukon.

1.2.1 Eastern and central Yukon-Tanana terrane

As defined in Colpron et al. (2006a), the Yukon-Tanana terrane is composed of four assemblages of polymetamorphosed and polydeformed pre-Devonian to Late Permian metasedimentary and metavolcanic rocks. The Snowcap assemblage is a pre-Late Devonian, possibly as old as Neoproterozoic (Piercy and Colpron 2009), rifted part of the Laurentian margin that is the basement to the Yukon-Tanana terrane. Overlying the Snowcap assemblage are the Finlayson, Klinkit and Klondike assemblages. These overlying assemblages consist of arc-related rocks which record the evolution of the Yukon-Tanana terrane while it was outboard of the rifted Laurentian margin. A brief description of each Yukon-Tanana terrane assemblage is presented below.

1.2.1.1 Snowcap Assemblage

The Snowcap assemblage is heterogenous, consisting of pre-Late Devonian quartzite; psammitic, pelitic and calc-silicate schists; marble and local amphibolites and ultramafic rocks (Piercey and Colpron 2009, Colpron et al. 2006a). Units of the Snowcap assemblage are the oldest identified units in the Yukon-Tanana terrane in British Columbia and Yukon (Colpron et al. 2006a). Detrital zircon ages in the Snowcap assemblage have major peaks at c. 1870 Ma and c. 2720 Ma and secondary peaks at c. 2080 Ma and c. 2380 Ma. (Piercey and Colpron 2009). These ages are characteristic of a Precambrian cratonic source and have been interpreted to reflect sediment

input from the Laurentian craton (Colpron et al. 2006b, Piercey and Colpron 2009). These zircon ages, combined with Nd-Hf isotope analyses have led to the conclusion that the Snowcap assemblage represents a rifted fragment of the Late Proterozoic-Early Paleozoic Laurentian margin (Colpron et al. 2006b, Piercey and Colpron 2009). This rifted fragment became the basement on which the overlying Yukon-Tanana terrane arcs formed.

Snowcap assemblage rocks commonly occupy a low structural level in the Yukon-Tanana terrane. Snowcap assemblage rocks are typically polydeformed and metamorphosed to amphibolite facies (Colpron et al. 2006a). At least one phase of deformation in the Snowcap assemblage predates deposition of overlying Carboniferous strata (Colpron et al. 2006b). Late Devonian-Early Mississippian tonalite, granodiorite, and granite plutons are common in the Snowcap assemblage. These plutonic suites are inferred to be the subvolcanic equivalents of overlying arc-related rocks of the Finlayson and Klinkit assemblages (Colpron et al. 2006a).

1.2.1.2 Finlayson assemblage

The Finlayson assemblage unconformably overlies the Snowcap assemblage (Colpron et al. 2006a). The Finlayson assemblage consists of vertically and laterally variable stratigraphic successions, all containing magmatic products of arc and back-arc character (Colpron et al. 2006a, Piercey et al. 2006). In the Finlayson Lake district, from which the Finlayson assemblage takes its name, the Finlayson assemblage consists of variably deformed and metamorphosed lower greenschist to amphibolite facies Late Devonian-Early Mississippian mafic to felsic metavolcanic rocks, carbonaceous pelite, chert, minor quartzite, volcanoclastic rocks, and marble (Colpron et al. 2006a, Murphy et al. 2006). West of the Tintina fault, the Finlayson assemblage is characterized by greenstone and plutons of arc character and laterally equivalent black chert, argillite and quartzite (Colpron et al. 2006a, Roots et al. 2006, Mihalynuk et al. 2006). In the

Teslin, Quiet Lake and eastern Laberge map areas, dark siltstone, quartzite and phyllite are inferred to record synvolcanic sedimentation (Roots et al. 2006). Early Mississippian metavolcaniclastic rocks, mature orthoquartzite, intermediate metavolcanic rocks and bioclastic marble are characteristic of the Finlayson assemblage in the Glenlyon area (Colpron et al. 2006b).

Rocks of the Finlayson assemblage record the evolution of a Late Devonian-Early Mississippian continental arc to back-arc system (Colpron et al. 2006a). The part of the Finlayson assemblage northeast of the Tintina fault is interpreted as a back-arc facies (Murphy et al. 2006; Nelson et al. 2006). Southwest of the Tintina fault, Finlayson assemblage rocks consist primarily of arc magmatic rocks surrounded by basinal sedimentary deposits (Colpron et al. 2006a). These sedimentary deposits were derived from both arc and recycled cratonic sources (Colpron et al. 2006a). This is inferred to reflect deposition in an active volcanic arc environment built on basement of continental character (Colpron et al. 2006a). The most likely source for the recycled cratonic material was exposed rocks of the Snowcap assemblage (Colpron et al. 2006a).

1.2.1.3 Klinkit assemblage

The Klinkit assemblage unconformably overlies both the Snowcap and Finlayson assemblages (Colpron et al. 2006a). The Klinkit assemblage consists of intermediate to mafic volcaniclastic and volcanic rocks and limestone of middle Mississippian-Early Permian age (Colpron et al. 2006a). The Klinkit assemblage takes its name from the Klinkit Group which straddles the border between Yukon and British Columbia (Simard 2003). The Klinkit Group is dominated by volcaniclastic metasedimentary rocks but also contains Mississippian to Lower Pennsylvanian carbonates in the lowermost part (Simard et al. 2003). Roots et al. (2002) report an Early

Permian age (U-Pb zircon; 281 ± 2 Ma) on volcanoclastic rocks in the Teslin Lake area. Roots et al. (2002) infer that Klinkit volcanism began in the Carboniferous and continued into the Permian.

Intermediate and felsic volcanic rocks correlated with the Klinkit assemblage in the Glenlyon map area have the geochemical character of continental arc magmatism whereas basalts are alkalic (Colpron, 2001). In the Glenlyon area, these volcanic rocks are interpreted as the products of intra-arc rifting (Colpron, 2001). Within the Klinkit assemblage, basal conglomerates containing Early Mississippian detrital zircons occur. This is interpreted to reflect recycling of the Finlayson assemblage (Colpron et al. 2006a).

The volcanoclastic metasedimentary rocks of the Klinkit assemblage are primarily distal facies and carbonate units are composed of debris rather than carbonate reefs (Roots et al. 2006). Simard et al. (2003) identified upward-fining sequences in the Butsih and Mount McCleary formations resembling megaturbidite deposits (Simard et al. 2003). Roots et al. (2006) interpret these variations within the Klinkit assemblage to reflect the subaqueous distal facies of an extensive arc system that is on-lapping and overlying Carboniferous limestone.

1.2.1.4 Klondike assemblage

Unlike the Snowcap, Finlayson, and Klinkit assemblages, the Klondike assemblage is not widespread. The Klondike assemblage has been identified in the Stewart River and Dawson areas in the Klondike region of the Yukon (Colpron et al. 2006a). The Klondike assemblage consists of Middle to Late Permian calc-alkaline felsic and minor mafic metavolcanic rocks (Colpron et al. 2006a). The geochemical character of these rocks indicate emplacement in a continental arc setting (Piercey et al. 2006). Rocks of the Klondike assemblage are interpreted to

be the product of subduction of the Slide Mountain terrane beneath the Yukon-Tanana terrane (Mortensen 1992, Nelson et al. 2006).

1.2.1.5 Deformation in eastern and central Yukon-Tanana terrane

At least two generations of widespread deformation have been documented in eastern and central parts of the Yukon-Tanana terrane. Colpron et al. (2006b) report pre-Devonian deformation that affected only the Snowcap assemblage. Berman et al. (2007) attribute Early Triassic thickening of the Yukon-Tanana terrane in the Stewart River area to collision of the Yukon-Tanana terrane with the North American craton. Late Permian brittle and ductile deformation is widely reported in the eastern and central parts of the Yukon-Tanana terrane (e.g. Mortensen 1990, Murphy et al. 2006, Berman et al. 2007, Mackenzie et al. 2007).

1.2.2 Western Yukon-Tanana terrane of southwestern Yukon

Recent geological investigations in southwestern Yukon have begun to advance our understanding of the western margin of the Yukon-Tanana terrane to the level of understanding in the central and eastern parts of the Yukon-Tanana terrane. Previous mapping of southwestern Yukon by Muller and Christie (1966) subdivided the Yukon Complex (Yukon-Tanana terrane) based on lithologic characteristics. It was not until mapping by Murphy et al. (2009) and Murphy (2010) that the Yukon Complex assemblages of Muller and Christie (1966) were accurately correlated with the modern Yukon-Tanana terrane framework.

Mapping by Murphy (2010) also advanced the work of Mezger et al. (2000). Mezger et al. (2000) identified the Kluane Schist as a Cretaceous back-arc basin, marking the boundary between the pericratonic Yukon-Tanana terrane and outboard Alexander-Wrangellia terranes. Mezger et al. (2000) noted mafic and ultramafic bodies within the Kluane Schist in the Burwash Landing area. Murphy (2010) and Murphy et al. (2011) correlated the mafic and ultramafic bodies of Mezger

et al. (2000) with mafic and ultramafic bodies in the Nuntaea Creek and Lynx Creek map areas to the north. Together, the mafic and ultramafic bodies of the Burwash Landing and Lynx Creek map sheets define the Late Triassic and older Doghead assemblage (Murphy et al. 2011). This interpretation by Murphy (2010) and Murphy et al. (2011) is the first to suggest obduction of oceanic crust onto the western margin of the Yukon-Tanana terrane and subsequent deformation of the Yukon-Tanana terrane and Doghead assemblage following obduction.

West of the Donjek River Late Permian dikes occur both as foliated, foliation parallel bodies and as weakly foliated, folded and boudinaged bodies. This variation in deformation suggests Late Permian deformation in the western parts of the Yukon-Tanana terrane was localized. Little else is known about deformation in the western parts of the Yukon-Tanana terrane. The following summarizes the current understanding of the western margin of the Yukon-Tanana terrane in southwestern Yukon.

1.2.2.1 Northern Kluane Lake map area

The Kluane Lake map area (115G) is divided into two parts by the Cordilleran scale right-lateral Denali fault. The Denali fault separates the Wrangellia and Alexander terranes to the southwest from the Yukon-Tanana terrane and Cretaceous Kluane schist to the northeast (Israel et al. 2011). In the northern Kluane Lake map area rocks correlated with the Yukon-Tanana terrane occur between the east-west striking Early Cretaceous Nisling Range Granodiorite to the north and Paleocene Ruby Range Batholith to the south. Roof pendants of Yukon-Tanana terrane rocks occur within the Ruby Range Batholith (Murphy 2010, Israel et al. 2011). The Ruby Range Batholith obscures the contact between the Yukon-Tanana terrane and the underlying Kluane schist. However, regional mapping suggests the Yukon-Tanana terrane was thrust over the Kluane schist in the mid-Cretaceous (Israel et al. 2011).

Mafic and ultramafic rocks of the Late Triassic and older Doghead assemblage structurally overlie both the Yukon-Tanana terrane and the Kluane schist (Murphy et al. 2011). In the Duke River (115G06) and Burwash Landing (115G07) map-areas, the Doghead assemblage occurs as rare, local discontinuous bodies within the Kluane schist (Murphy 2010, Israel et al. 2011). Mezger (2010) interprets these bodies of mafic and ultramafic rock as parts of oceanic crust that were tectonically emplaced within the Kluane schist during Late Cretaceous subduction. Murphy (2010) and Murphy et al. (2011) inferred that in the northern parts of the Kluane Lake map area the Doghead assemblage was thrust over the Yukon-Tanana terrane (the Nuntaea Creek fault of this study) during Jurassic time (see Chapter 3). In the Nuntaea Creek (115G11) and Lynx Creek (115G12) map-areas, the Doghead assemblage occurs as a discontinuous body in the core of an east-west-trending, moderately-to-the-south-dipping regional-scale synform (Murphy et al. 2009, Murphy 2010). This synform folds rocks of both the Yukon-Tanana terrane and the Doghead assemblage.

Regional-scale right-lateral strike-slip faults crosscut all generations of folding and plutonism in the northern part of the Kluane Lake (115FG) and southern part of the Stevenson Ridge (115JK) map areas. These strike-slip faults extend from the Ruby Range Batholith in the south to, and beyond, the Yukon-Alaska border in the northwest. The age of these faults is constrained to Late Cretaceous or younger by crosscutting relationships with the Ruby Range Batholith in the Kluane Lake map area and the Donjek volcanics in the Stevenson Ridge map-area.

In the northern Kluane Lake map-area, the Yukon-Tanana terrane is composed of rocks correlated with the Snowcap, Finlayson and Klinkit assemblages. Rocks correlated with the Finlayson and Klinkit assemblages have been previously mapped as undifferentiated White River formation (Murphy et al 2009, Murphy 2010). Detailed mapping in the Tincup Lake area

suggests the rocks previously mapped as the White River formation are lithologically similar to the Klinkit and Finlayson assemblages. Rocks correlated with the Snowcap assemblage consist of homogenous quartz psammite, pelite, minor amphibolite and marble (Murphy 2010). Rocks correlated with the Finlayson assemblage consist of quartzite, marble and graphitic phyllite. Rocks correlated with the Klinkit assemblage consist of metavolcaniclastic psammite and schist, quartzofeldspathic psammite, Ms-Bt-Chl schist, minor Chl-Pl-Act amphibolite and marble.

Figure 1.1: Terranes of the Canadian-Alaskan Cordillera (after Colpron et al. 2007). Rocks correlated with the Yukon-Tanana terrane occur in northern British Columbia, southwest Yukon, and east-central Alaska. An outlier of the Yukon-Tanana terrane occurs northeast of the Tintina fault in southeastern Yukon. Abbreviations for major post-accretionary faults: BSF—Big Salmon fault; CSF—Chatham Strait fault; CSZ—Coast shear zone; FRF—Fraser River fault; KF—Kechika fault; NFF—Nixon Fork–Iditarod fault; PF—Pinchi fault; SRMT—southern Rocky Mountain trench; TkF—Takla-Finlay-Ingenika fault system; YK—Yalakom fault. Other abbreviations: AB—Alberta; AK—Alaska; B.C.—British Columbia; N.W.T.—Northwest Territories; YT—Yukon.

Figure 1.1

Figure 1.2: Terranes and assemblages of southwestern Stevenson Ridge (115JK) and Kluane Lake (115FG) map-areas, northeast of the Denali fault (Murphy et al. 2011). Study area is denoted by thatched area. Here, the Late Cretaceous Kluane schist occupies the lowest structural level. Overlying the Kluane schist is the Devonian to Permian Yukon-Tanana terrane. In the Kluane Lake map area, mafic and ultramafic rocks of the Doghead assemblage overlie both the Yukon-Tanana terrane and Kluane schist. In the Stevenson Ridge map-area, mafic and ultramafic rocks of the pre-Paleozoic Slide Mountain terrane structurally overlie the Yukon-Tanana terrane. Overlying both the Yukon-Tanana and Slide Mountain terranes are the Upper Cretaceous Donjek volcanics. Jurassic to Eocene intrusions, correlated with the Coast Plutonic Complex intrude rocks of the Yukon-Tanana terrane and overlap assemblages. Doghead assemblage: DR=Donjek River, DL=Dogpack Lake, DP=Doghead Point; Slide Mountain terrane: HP=Harzburgite Peak, WL=Wellesley Lake, D=Deadman Hill, EM=Eikland Mountain, SP=Sanpete Hill, KM=Koidern Mountain

Figure 1.2

1.3 References

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Chapter 2- Lithostratigraphic evolution of the Yukon-Tanana terrane, Tincup Lake area, southwest Yukon

2.1 Introduction

Previous regional-scale mapping in the polydeformed and polymetamorphosed Tincup Lake area, southwest Yukon, has resulted in identification of a heterogeneous package of psammites, schists and marbles correlated with the Yukon-Tanana terrane (Muller and Christie 1966, Murphy 2010). The structural complexity of this area and regional-scale of mapping precluded identification of an internal stratigraphy. Without an understanding of the stratigraphy of this heterogeneous package, previous mappers have not been able to propose a correlation between assemblages of the Tincup Lake area with those identified elsewhere in the Yukon-Tanana terrane. Therefore, detailed mapping of the Tincup Lake area was undertaken in an effort to identify map units and define assemblages within the heterogeneous package of psammites, schists and marbles. Twelve map units, including 3 marbles and an unconformable contact which provided marker horizons (Map 1). Thus a stratigraphic succession was defined in the Tincup Lake area, and is consistent with the results of 1:20,000 scale structural studies (Chapters 3 & 4).

The Tincup Lake area consists of a 12 km structural thickness of upper greenschist to lower amphibolite facies metasedimentary rocks with minor metaigneous rocks. Exposure in this area allows for marble marker horizons to be easily traced between ridge tops across the study area. Valleys are poorly exposed and heavily vegetated. Rocks are penetratively foliated and have been deformed by four phases of folding (D_1 - D_4 ; Chapter 3) and faulting (D_5 , D_6 ; Chapter 4, Map 1). Although folding and metamorphism have modified the thicknesses of map units and primary textures, primary sedimentary features such as compositional layering and graded bedding are still locally discernable.

Map units typically dip moderately to the south (Map 1). The orientation of map units varies from moderately dipping to the south, south of Conglomerate Plateau to gently dipping to the north in the immediate hanging wall of the Marble Top fault (Map 1). North of the Marble Top fault, map units dip gently to moderately to the south.

Map units occur in three lithostratigraphic domains defined on the basis of composition and relationships between map units. This chapter describes the map units and the relationships between them. The protoliths of these units are interpreted based on textural and mineralogical observations. Map units are grouped into assemblages reflecting composition and breaks in apparent stratigraphy. The relationships between assemblages and information from interpreted protoliths are used to infer the evolution of the basin of deposition. Assemblages of the Tincup Lake area are correlated with recognized assemblages of the Yukon-Tanana terrane (Colpron et al. 2006a).

2.1.1 Conventions used

Mineral abbreviations used in this chapter follow the recommendations of Whitney and Evans (2010). Mineralogy is listed following the guidelines of Schmid et al. (2007), where the least abundant mineral is listed first and the most abundant mineral is listed last.

2.2 Map Units

2.2.1 Overview

The Tincup Lake area is divided into 3 fault-bound lithostratigraphic domains. (Figure 2.1, Map 1). Domain 1, in the north, and Domain 2, in the central area, are separated by the brittle, moderately to the south-dipping D₅ Marble Top fault. The Marble Top fault is inferred to have a normal sense of motion, based on the change from upper greenschist facies metamorphic assemblage in the hanging wall to an amphibolite facies metamorphic assemblage in the

footwall. This change in metamorphic assemblage coincides with the juxtaposition of Snowcap assemblage rocks in the footwall and Klinkit assemblage rocks in the hanging wall. Domains 1 and 2 comprise predominantly metasedimentary rocks that have been correlated with the Yukon-Tanana terrane (Murphy, 2010). A pre-D3 reverse fault is inferred to separate Domain 2 and Domain 3 at the southern limit of the study area. Domain 3 is composed of an allochthonous mafic-ultramafic complex, the Doghead assemblage (Murphy et al. 2011). The mafic-ultramafic complex will not be addressed in this chapter. For more information on the Doghead assemblage, see Murphy et al. (2011).

2.2.2 Domain 1

Domain 1 is composed of units 1, 2 and 3 (Map 1). Domain 1 has a minimum structural thickness of 500 m. This domain is bound by the Marble Top fault to the south and continues to the north, beyond the extent of mapping (Murphy, 2010).

2.2.2.1 Unit 1 (*UDps*)

Unit 1 comprises tan- to grey-weathering, fine- to medium-grained, Gt-Ms-Bt-Qz-Fsp psammite interlayered with medium-grained Bt-Ms schist. It is the lowest exposed unit in Domain 1 in the map area based on its position in a map-scale synformal closure (cross-section). The lower contact of Unit 1 is not exposed within the study area. Unit 1 is a minimum of 100 m thick and has been traced approximately 10 km along strike.

Psammitic layers range from 1-30 cm thick. Schistose layers are 1-10 cm thick. Contacts between psammitic layers and schistose layers are typically sharp (Figure 2.2a) with rare gradational contacts (Figure 2.2b). Tan- to grey-weathering psammitic layers are composed of fine- to medium-grained muscovite, biotite, quartz and feldspar with euhedral to subhedral garnet porphyroblasts up to 1 cm in diameter and up to 10% quartz and/or feldspar pebbles (?) up to 1

cm in diameter (Figure 2.2c). Schistose layers are typically black- to grey-weathering. Schistose layers are composed primarily of fine- to medium-grained muscovite and biotite.

2.2.2.2 Unit 2 (vDm)

Unit 2 consists of lean, white- to grey- to tan-weathering calcite marble and calcareous psammite. Unit 2 can be traced ~10 km along strike, with a minimum thickness of ~100 m. The contact between unit 1 and unit 2 is gradational. The base of unit 2 is defined by the first appearance of marble interlayers within unit 1 quartzofeldspathic psammite.

The lower ~5 m of unit 2 is composed of interlayered clean, fine-grained calcite marble and fine-grained calcite marble with up to 20% medium-grained quartz, feldspar and mica. Clean calcite marble layers are 5-20 cm thick and are white- to grey-weathering. Layers containing quartz, feldspar and mica are typically less than 0.5 cm thick and are most common within 1 m of the contact between unit 1 and unit 2. Quartz-, feldspar- and mica- bearing layers typically weather a grey- to rusty-brown colour.

Beyond 5 m of the contact with unit 1, unit 2 is characterized by clean calcite marble in layers 5-20 cm thick. This clean marble is grey- to white weathering. This part of unit 2 contains black, stylolitic dissolution surfaces spaced between 5 and 10 cm.

2.2.2.3 Unit 3 (vDa)

Unit 3 is dark green- to black-weathering, fine- to medium-grained, massive to well foliated Chl-Pl-Act amphibolite. Unit 3 is in sharp contact with unit 1 and is interlayered with parts of unit 2 (see below). It is unclear whether this interlayering reflects volcanic flows and calcareous interflow sedimentation or intrusion parallel to compositional layering. Unit 3 has a minimum thickness of approximately 50 m and can be traced for 2 km along strike.

The amphibolite occurs in layers up to 5 m thick, separated by layers of chlorite schist, up to 10 cm thick. At the basal contact with unit 2, amphibolite and calcite marble are interlayered with sharp contacts (Figure 2.3) at a scale of 5 cm to 200 cm. The contacts of marble interlayers are commonly tan- to rusty brown-weathering where they are adjacent to amphibolite.

Pillow-like features have been observed at one outcrop. At the mesoscopic scale, these pillow-like features are teardrop-shaped areas of coherent outcrop surrounded by cm-thick areas of recessively weathering material (Figure 2.4a). The “top” of this layer is defined by fine-grained amphibolite with mm-scale, round to oblate, recessively weathering features resembling relict amygdules (Figure 2.4b). On fresh surfaces these recessively weathering features are filled with a soft, chalky white carbonate material. Near the “bottom” of this layer the simplectite-filled features decrease and plagioclase-actinolite amphibolite is medium-grained.

The eastern extent of unit 3 (587000 E, 6851500 N) is inferred from talus and limited outcrop exposure. Outcrops and talus clasts in this area are more schistose than western exposures, rarely containing coarse-grained amphibolite.

2.2.3 Domain 2

Domain 2 is divided into two subdomains, 2a and 2b (Map 1). Domain 2a consists of units 4-7, which are characterized by quartzite and marble. Domain 2b consists of units 8-12, which are characterized by mafic-intermediate metaigneous and quartzofeldspathic metasedimentary rocks. Domain 2b is interpreted to unconformably overlie domain 2a along the lower contact of unit 8. The lower amphibolite-upper greenschist facies metasedimentary rocks and minor mafic metaigneous rocks of Domain 2 have a minimum structural thickness of 12 km.

Domain 2a

2.2.3.1 Unit 4 (vDLMq)

Unit 4 is composed of peach- to grey-weathering, locally massive to layered quartzite. Quartzite is the lowest stratigraphic unit in Domain 2. The lower contact is not exposed in the study area. The maximum thickness of exposed quartzite is approximately 50 m. This unit can be traced approximately 1 km along strike through discontinuous outcrops.

Massive quartzite is most common in the lower parts of this unit. Unlike the layered component of this unit, massive quartzite does not contain layers of calcite marble. Heavy mineral layers have not been observed. The lowest appearance of calcite in the quartzite is as intergranular material in the massive quartzite. This calcite weathers recessively, exposing rounded quartz grains with positive relief. Calcite matrix becomes less common up section, whereas discrete lenses of recessively weathering calcite up to 2 cm thick become more common (Figure 2.5).

In the top 5 m of this unit the amount of calcite continues to increase until continuous 0.5 to 5 cm-thick layers and pebble- to cobble-sized bodies of grey- to white-weathering fine-grained calcite marble within quartzite compose more than 20% of the modal mineralogy of the rock (Figure 2.6).

Petrographic investigation of this unit indicates quartz grains have seriate-interlobate grain boundaries with a preferred long axis orientation, concordant with compositional layering. Where quartz grains are surrounded by calcite matrix, they are typically well rounded. Acicular tremolite crystals, up to 50 μm long, also show a long-axis preferred orientation concordant with compositional layering.

2.2.3.2 Unit 5 (uDLMqm)

Unit 5 is composed of interlayered quartzite and calcite marble. 1-10 cm-thick layers of grey- to white-weathering, fine- to medium-grained calcite marble with lenses of beige-weathering dolomitic marble characterize the lowest 5-10 m of unit 5. Marble layers are interlayered with 1-5 cm-thick layers of peach- to white- weathering quartzite. Locally, quartzite layers appear to grade upward into marble. Unit 5 ranges from 500-1000 m-thick and can be traced approximately 4 km along strike. Unit 5 is in gradational contact with unit 4. This transition occurs over a thickness of approximately 5 m.

Approximately 5 m above the gradational lower contact there is a transition from interlayered quartzite and calcite marble to calcite marble and rare dolomite clasts, up to 30 cm in diameter, in quartzite matrix, giving unit 5 a conglomeratic appearance (Figure 2.7a). For this reason, unit 5 is inferred to have a clastic protolith. The composition of the rock changes up section from matrix-supported pebble to boulder metaconglomerate to framework-supported pebble metaconglomerate (Figure 2.7b). Quartzite matrix also decreases up section approximating an inverse relationship with marble clast grain size. The uppermost 20-25 m of this unit contains discontinuous layers of quartzite less than 1 cm-thick (Figure 2.8a). In the same interval are layers less than 1 cm-thick of quartz sand and angular dark grey dolomite marble pebbles in a grey to white calcite matrix (Figure 2.8b). The upper limit of this unit is defined by the transition to massive marble.

2.2.3.3 Unit 6 (uDLMm)

Unit 6 is composed of grey- to white-weathering, fine- to coarse-grained calcite marble. The lower contact is gradational, occurring over a thickness of approximately 25m. The minimum structural thickness of unit 6 is 500m. Unit 6 is a marker unit that extends across the entire study area, both north-south and east-west.

The lower 50-100 m of this unit is composed of impure calcite marble. Above this interval, layered calcite marble with minor impurities is characteristic. Other characteristics of the lower 50-100 m of unit 6 include 1-5 cm-thick layers of grey- to beige-weathering dolomitic marble interlayered with 5-30 cm-thick layers of white-weathering calcite marble, angular clasts of dolomite up to 5 cm in diameter in calcite marble, and white- to yellow-weathering layers of locally pebbly calcareous quartzite. Above the lower 50-100 m, unit 6 is dominated by layered calcite marble (Figure 2.9). Layers of calcite marble are <1-15 cm- thick with rare <1-30 cm- thick boudins or lenses of tan- to beige- to grey- weathering, fine- to coarse-grained dolomitic marble. There are also rare 1-5 cm-thick layers of fine- to coarse-grained calcareous quartzite.

2.2.3.4 Unit 7 (UDLMph)

Unit 7 is composed of well foliated, black- to dark grey-weathering $\pm$ Py-Bt-Ms-Qz-Gr carbonaceous phyllite with rare calcite marble layers. Unit 7 is in sharp contact with Unit 6. Unit 7 is discontinuous across the study area. The maximum continuous strike length of unit 7 is approximately 3 km. The maximum thickness of this unit is 10 m.

Carbonaceous phyllite layers consist of $\pm$ pyrite, biotite, muscovite, graphite, and quartz. With the exception of calcite marble layers, unit 7 lacks any discernable compositional layering at the mesoscopic scale. Calcite marble layers, up to 15 cm thick, make up approximately 5% of unit 7 (Figure 2.10).

Domain 2b

2.2.3.5 Unit 8 (MMUPmv)

Unit 8 consists of black- to green- to tan-weathering, undifferentiated, cm-dm-scale, layered $\pm$ Mag $\pm$ Gt $\pm$ Act $\pm$ Ep $\pm$ Cal-Bt-Ms-Chl-Pl-Qz psammite, psammitic schist, schist and Ep-Bt-Ms-Chl-Act-Pl amphibolite. Unit 8 is in sharp contact with units 6 and 7 (e.g. 588000 E, 6843000 N, Map 1).

Unit 7 is discontinuous along the contact between units 6 and 8. This discordancy is unique in this stratigraphic sequence. Unit 8 has a structural thickness of at least 700 m. South of Division Valley, unit 8 can be traced 7 km along strike, outcropping along ridge tops. North of Division Valley, unit 8 can be traced for 10 km across the width of the study area.

Basal beds of unit 8 are generally schist with conglomeratic pockets. The schistose component in basal beds consists of chlorite schist. The conglomeratic part of unit 8 consists of poorly sorted, framework-supported conglomerate with elongate pebble- to cobble-sized clasts of hornblende metagabbro (Figure 2.11). Also within this basal schist are rootless, isoclinally folded discontinuous layers or lenses of rusty brown calcite marble.

Beyond approximately 10 m from the base of this unit, interlayered psammite and schist are characteristic. Interlayered psammite and schist, with rare quartzite, comprise the majority of this unit. Psammitic layers are tan- to dark grey- to green-weathering. Psammitic layers range from 5-20 cm thick to massive with no discernable layering at the mesoscopic scale. Psammitic layers are composed of fine- to medium-grained angular to subrounded, $\pm$ garnet, $\pm$ actinolite, $\pm$ biotite, muscovite, chlorite, quartz, and plagioclase in a chloritic matrix. Schistose layers are typically green-weathering but may also be dark grey- or rusty brown. Schistose layers have a well developed foliation and range from <1 cm to ~5 cm thick. Schistose layers are composed primarily of chlorite, with trace actinolite, garnet and pyrite.

The upper 50 m of this unit is calcareous. In the lowest part of the calcareous stratigraphy calcite is interstitial in schistose layers. In the upper 15-20 m of Unit 8 there are 1-2 m-thick layers of calcite marble interlayered with >10m-thick layers of garnet-muscovite-chlorite schist.

2.2.3.6 Unit 9 (MMuPm₁)

Unit 9 is composed of grey- to white-weathering calcite marble, rusty brown-weathering mica (clintonite ?)-calcite schist and rare medium-grained quartzite. Locally, this unit is massive, fine- to coarse-grained calcite marble that lacks mica and quartz. The contact between Unit 8 and Unit 9 is poorly exposed. The best exposure of this contact occurs at 587100 E, 6849100 N. Although the contact is generally obscured by talus, the transition from psammite and schist of unit 8 to calcite marble of unit 9 occurs over an interval of 1-2 m. The structural thickness of unit 9 varies from a minimum thickness of approximately 50 m (586950 E, 689810 N) to approximately 1500 m (583100 E, 6849250 N). Exposure of unit 9 is limited to the area north of Division Valley (Map 1). Outcrops of unit 9 can be traced along strike for approximately 7 km, between West Ridge and East Ridge (Map 1).

The composition of unit 9 varies from east to west. Trace medium-grained quartz grains in calcite matrix and rusty brown white-mica (clintonite?)-calcite schist is prevalent in eastern exposures of this unit. Conversely, exposures of this unit at the south end of West Ridge contains up to 5m of 5-20 cm-thick layers of peach- to white-weathering, fine-grained to massive quartzite. Layers of clintonite(?)-calcite schist up to 1 m thick make up 10-15% of this unit along West Ridge. A 3 m-thick exposure of ankeritic marble also occurs along West Ridge (583000 E, 6849100 N).

2.2.3.7 Unit 10 (MMuPps)

Unit 10 is composed of undifferentiated, interlayered Bt-Ms-Chl psammite, psammitic schist and $\pm\text{Gt}\pm\text{Chl-Bt-Ms-Fsp-Qz}$ schist. The lower contact is gradational. The lower 100 m is calcareous, containing up to 40% calcite as matrix. This unit has a minimum structural thickness of 1000 m and can be traced approximately 4 km along strike between West Ridge and Marble Top (Map 1).

Contacts between psammitic layers and schistose layers follow an alternating pattern of sharp-graded-sharp (Figure 2.12). Psammitic layers in unit 10 are 1-20cm thick and are made up of tan- to grey-weathering, fine- to medium-grained biotite-muscovite-chlorite psammite. Schistose layers in this unit are 1-15 cm thick and are made up of black- to brown- to green-weathering ±garnet±biotite-muscovite-chlorite schist.

2.2.3.8 Unit 11 (MMUPm₂)

The stratigraphically and structurally highest unit in this area is Unit 11, a 600 m-thick white- to grey-weathering calcite marble. It is fine- to coarse-grained and is made up of layers up to 30 cm thick. Unlike other marbles in the study area, Unit 11 does not contain siliciclastic material. The lower contact with unit 10 is not exposed.

Within unit 11 is a chlorite-plagioclase-actinolite amphibolite body. This amphibolite body is approximately concordant with compositional layering. It is unclear whether this body is a volcanic flow or a pre-D₁ (Chapter 3) intrusion. This amphibolite body thickens to the east, reaching a maximum thickness of approximately 2 m. To the west the amphibolite body narrows and pinches (Figure 2.13). The contact between the marble (unit 11) and the amphibolite is sharp and lacks evidence of a chilled margin. Within 1 cm of the marble-amphibolite contact, the marble weathers tan-beige, possibly reflecting a change in mineralogy.

2.2.3.9 Unit 12 (MMEPa)

Unit 12 is a 100-200 m thick body of massive to coarse-grained Gt-Pl-Act amphibolite and recessively weathering chlorite schist that lies between units 10 and 11. The upper and lower contacts of this unit are not exposed. It contains orange-weathering, discontinuous, cm-thick lenses of marble in the chlorite schist that are congruent with the dominant foliation.

2.3 Interpretation

2.3.1 Inferred Protoliths

The following section presents interpretations of protoliths for each metasedimentary map unit, and depositional settings of domains 1, 2a and 2b. Primary sedimentary structures are not well preserved in the Tincup Lake area. This limits the interpretation of protolith and depositional setting to generalities. This is followed by an interpretation of amphibolite bodies, units 3 and 12, and their relationship to metasedimentary units. Finally, an interpretation of the lower contact of unit 8, as an unconformity, is presented. The general interpretations presented assume, perhaps simplistically, that the units are conformable rather than in fault contact. Although the ages of units are not known, this assumption is reasonable given the lack of evidence for concordant faults.

2.3.1.3 Unit 1 – Gt-Ms-Bt-Qz-Fsp psammite and psammitic schist

Psammitic layers in this unit are interpreted to have an arkose arenite protolith. Psammitic schists are interpreted to have an arkose wacke protolith. Bt-Ms schists are interpreted to have a mudstone protolith. The observation of grading from pebbly psammite and psammite into psammitic schist and schists is consistent with an interpretation of deposition by turbidity flows. Turbidity flows occur in a variety of settings and data are insufficient to further constrain the depositional environment. The gradational contact with the overlying marble of unit 2 suggests a marine setting for unit 1.

2.3.1.4 Unit 2 – Cal marble

Calcareous marble of unit 2 is interpreted to have a limestone protolith. Clastic sedimentation also occurred in the limestone's depositional environment early on as reflected by rare thin siliciclastic beds in the lower part of the unit. Unit 2 is interpreted to have been deposited in a continental shelf environment.

2.3.1.5 Unit 3 – *Chl-Act* amphibolite

The interlayered amphibolite within marble (Figure 2.3), with sharp contacts, may be interpreted two ways. One interpretation is that the interlayered amphibolite and marble represents mafic sills that intruded along marble bedding planes. Alternatively, they may be mafic flows with interflow carbonate sedimentation. In this second scenario, contacts may have been modified by strain to create sharp contacts. The outcrop depicted in figure 2.4a could be interpreted as an amygdule-bearing pillow, suggesting the amphibolite is extrusive. Unit 3 is interpreted as a syn-depositional flow rather than a post-depositional intrusion.

Unit 3 is not continuous along strike. Mapping suggests it is also restricted to the southern limb of the syncline in the footwall of the Marble Top fault (584500 E, 6852500 N). This may reflect either original limits to deposition or attenuation of the unit during deformation.

2.3.1.6 Domain 1 setting

Domain 1 is characterized by limestone and interlayered sandstone, wacke and mudstone. This character suggests that the rocks in domain 1 were deposited in a marine setting. The turbiditic nature of Unit 1 suggests deposition in a basin with topographic relief sufficient for the development of turbidity currents. The graded contact between units 1 and 2 suggests a relative increase in calcareous input synchronous with a relative decrease in siliciclastic input until siliciclastic input was negligible. This may reflect a change in sediment source, a deepening of the basin, or progradation of shallow water carbonate environments into the basin. There is insufficient evidence to suggest one scenario over the other. The local occurrence of unit 3 along the contact of units 1 and 2 suggests local volcanism synchronous with this change in environment.

2.3.1.7 Unit 4 – Tr- Cal Quartzite

The protolith of Unit 4 is interpreted to be a mature, well sorted quartz arenite. A clastic sedimentary protolith is inferred rather than a chert or biogenic source because of the presence of rounded quartz grains preserved in calcareous and dolomitic lenses.

2.3.1.8 Unit 5 – Interlayered quartzite and Cal-marble

Interlayered quartzite and calcite marble in the lower 5-10 m of this unit are interpreted to reflect original compositional layering of quartz arenite and limestone. The irregular shape of marble lenses in a quartzite matrix above the interlayered quartzite and marble is interpreted to reflect limestone clasts in a quartz-arenite matrix. Alternatively, these calcite marble lenses could be dismembered or boudinaged layers of calcite marble (limestone). No neck folds are present, making the boudinage hypothesis less likely.

An interpretation of the dolomite and calcite marble lenses as clasts of limestone and dolostone is favored. This interpretation explains the variation of limestone clasts from pebble to boulder sized, the homogeneity of the quartz arenite matrix, and the variation from framework to matrix supported, poorly sorted, metaconglomeratic appearance of the unit.

2.3.1.9 Unit 6 – Qtz-Dol-Cal marble

Unit 6 is interpreted to have a limestone protolith. The presence of quartz sand and pebble layers along with angular dolomitic clasts suggests contribution from a clastic source. These characteristics suggest deposition in a shelf setting.

2.3.1.10 Unit 7 – $\pm$ Py-Bt-Ms-Gr-Qz Phyllite

Unit 7 is interpreted to have a carbonaceous quartz mudstone protolith on the basis of mineralogy. The graphite in this rock is inferred to be organic in origin. This suggests the conditions in the basin were either anoxic or anerobic when these sediments were deposited. Anoxic conditions may have been the result of water depth within the basin or the sediments

could have been buried quickly enough to keep the organic material from being broken down in an oxygenated environment.

2.3.1.11 Domain 2a setting

Based on the gradational contacts between the units in this domain, units 4 through 6 are interpreted to record continuous deposition. The dominance of quartz and carbonate sediments suggests these units were deposited in a shelf setting.

The sharp contact between unit 6 and 7 suggests a rapid change in the depositional environment. The carbonaceous nature of unit 7 is attributed to high organic content. The rapid change to a carbonaceous shale-dominated setting is interpreted to reflect a relative rise in sea level and deepening of the basin. There is insufficient data to determine whether this change is the result of a regional transgressive event or the result of local normal faulting in the Yukon-Tanana terrane.

2.3.1.12 Unit 8 – $\pm\text{Mag}\pm\text{Grt}\pm\text{Act}\pm\text{Ep}-\text{Cal}-\text{Bt}-\text{Ms}-\text{Chl}-\text{Pl}-\text{Qz}$ psammite, schist and $\text{Ep}-\text{Bt}-\text{Ms}-\text{Chl}-\text{Act}-\text{Pl}$ amphibolite

The lower contact of Unit 8 is interpreted as an unconformity (see section 2.3.3) as unit 8 overlies both unit 6 and unit 7. The lower 10 m of Unit 8, containing poorly sorted clasts of hornblende metagabbro in a chlorite schist matrix, is interpreted as a basal conglomerate, further supporting the interpretation of an unconformable contact. This basal conglomerate has a mafic source and is interpreted to reflect the onset of mafic volcanic activity in this part of the Yukon-Tanana terrane. The upward progression into interlayered psammite and schist may reflect deepening of a regional or local basin or progradation of a volcanic apron.

The predominance of chlorite and hornblende (actinolite) combined with an absence of alkali feldspars, indicates a mafic to intermediate volcanoclastic source for this unit. Angular grains in

psammitic layers are interpreted as crystal fragments. Alternating layers of volcanoclastic psammite and schist with rare quartzite layers suggests the volcanoclastic pile was periodically the locus of deposition from a different source. The gradational contact between unit 8 and the overlying marble suggest the depositional setting of unit 8 was a continental shelf/slope or volcanic apron.

2.3.1.13 Unit 9 – Middle Cal Marble

Unit 9 is interpreted to have a limestone or marl protolith. The presence of quartzite and calcite-muscovite schists in the unit along West Ridge suggests that siliciclastic input was significant at times. This siliciclastic input however did not include mafic epiclastic material such as that of Unit 8. Possible explanations for this change include, but are not limited to the following scenarios: volcanism ended by the time Unit 9 was being deposited, the transport direction of volcanoclastic sediments changed, or the sediment source for this part of the basin changed. With the available evidence, any of these scenarios is possible. Unit 8 is interpreted to have been deposited in shelf setting where either carbonate production was abundant or clastic carbonate material was readily available.

2.3.1.14 Unit 10 – *Bt-Ms-Chl* psammite and $\pm Gt \pm Chl$ -*Bt-Ms-Fsp-Qz* psammitic schist

The pattern of sharp-graded-sharp contacts between layers of psammite and schist supports the interpretation that psammite and schist layers are relict graded beds. The sharp contact between schists and psammite represents the basal contact between beds. Graded contacts between psammites and schists are inferred to reflect compositional changes within a single bed. Grading from psammite to schist is interpreted as an upward fining of grain size. These graded beds are correlated with the A facies of Bouma's turbiditic sequence (Bouma 1962).

2.3.1.15 Unit 11 – Upper Cal Marble

The protolith of this marble is interpreted to be limestone. This clean marble unit is unique in that it neither contains metasiliciclastic nor dolomitic material. This is inferred to indicate the cessation of siliciclastic sedimentation in this part of the Yukon-Tanana terrane. There is insufficient evidence to propose a specific depositional environment.

2.3.1.16 Unit 12 – Chl schist and Gt Pl Act amphibolite

There are two possible interpretations of Unit 12. The first interpretation is that Unit 12 represent mafic flows. These flows would have been extruded onto unit 10 and been overlain by unit 11. If this is the case, then calcareous lenses in the chlorite schist represent synvolcanic calcareous sedimentation. An alternative interpretation is that Unit 12 is a mafic intrusion that intruded along the contact between units 10 and 11. If this is the case, then calcareous lenses in the chlorite schist represent xenoliths incorporated into the intrusion.

There is no evidence to suggest submarine emplacement of this unit (i.e. pillows or quench features). For this reason, Unit 12 is interpreted as a basic intrusion.

2.3.1.17 Domain 2b setting

Domain 2b consists of four metasedimentary units. Unit 8 is composed of layered metavolcaniclastic sediments and is inferred to have been deposited near a volcanic apron or slope. Unit 9 is composed primarily of calcite marble with minor quartzite and calc-schist, suggesting deposition in a marine setting following the cessation of volcanism. Unit 10 is composed of graded quartzofeldspathic psammite and biotite-muscovite-chlorite schist suggesting deposition by turbidity flows. Unit 11 is composed of clean calcite marble lacking metasiliciclastic and dolomitic material.

The lack of sedimentary structures in marble units of Domain 2b limit the interpretation of basin evolution. It is unclear whether the turbidites in units 8 and 10 were deposited within the same sedimentary system, i.e. the same channel-fan complex. If units 8 and 10 were deposited within the same turbidite system, then unit 9 is likely a calcareous turbidite system. If unit 10 represents a renewed turbiditic system, then the system must have changed from a stable carbonate platform or shelf to a basin with new bathymetric character. This change in basin configuration could reflect faulting, progradation of a delta or a transgressive event.

Unit 11 is unique in the study area because it is the only marble that does not contain metasiliciclastic material. The lack of siliciclastic material suggests this unit may have been deposited in a relatively lower energy system than units 8-10. The underlying units are inferred to record a deepening basin. If this interpretation is correct, then unit 11 may represent a distal facies relative to the underlying marble units.

2.3.2 Amphibolites and their Significance

Unit 3 is interpreted as a volcanic rock. The occurrence of unit 3 only on the northern limb of the synform in the footwall of the Marble Top fault may reflect the lateral edge of the flow or attenuation of the unit during subsequent deformation. To the east (587000 E, 6851500 N) the extent of unit 3 is poorly constrained and may not occur to the extent depicted. The reason for the increased chlorite content, relative to western exposures, is unclear at this time.

It is unclear whether unit 12 is a volcanic horizon or mafic intrusion. If unit 12 is a volcanic horizon, then it would indicate renewed volcanism in this part of the basin between the deposition of units 10 and 11. If unit 12 is an intrusive body, then it would indicate post depositional magmatism. Since the chlorite schist of unit 12 is foliated, unit 12 must pre-date deformation.

2.3.3 Unconformity

The sharp contact of unit 8 over both units 6 and 7, as well as the local occurrence of a basal conglomerate in unit 8 suggests that the lower contact of unit 8 is an unconformity. Units below this contact are typical of continental margin environments – dominated by calcareous and siliceous sediments. Units above this contact represent mafic-intermediate volcanoclastic and siliciclastic sedimentation punctuated by calcareous sedimentation. The base of these siliciclastic units contains rocks interpreted as basal conglomerates. Graded beds in the siliciclastic units suggest a change from shelf environment to an environment capable of generating graded beds. Graded beds could indicate a slope, a steepened part of the shelf, or a volcanic apron.

2.4 Discussion

2.4.1 Basin Evolution

The faulted contact between domains 1 and 2 and the lack of age constraints precludes an integrated interpretation. For this reason, the evolution of Domain 1 and Domain 2 are treated separately.

Sedimentary features (e.g. cross beds, fossils) are not preserved in Domain 1, precluding a definitive assignment of a depositional setting. The presence of layered psammite, psammitic schist and marble in Domain 1 suggests a continental margin environment. Further interpretation of primary depositional setting requires further investigation.

Some primary sedimentary features (i.e. graded bedding) are preserved in rocks in Domain 2, allowing for a better interpretation of the depositional environment than for Domain 1. Domain 2a is composed of quartzite and marble. Their inferred protoliths of quartz arenite and limestone suggest deposition in a continental margin setting. It is unclear whether this margin was passive or active at the time of deposition.

The stratigraphically highest unit in Domain 2a is Unit 7, a graphitic phyllite. The sharp contact of this unit with the underlying marble is interpreted to reflect a rapid transgressive event. This event is inferred to have drowned the carbonate-producing environment. It is unclear whether this transgression reflects local faulting or a relative rise in sea level that affected the entire Yukon-Tanana terrane.

The angular nature of the contact between rocks of domains 2a and 2b indicates that prior to the deposition of the rock units of Domain 2b, the rocks of Domain 2a were eroded. The cause of this erosional event is unclear at this time. Throughout the Yukon-Tanana terrane a Devonian-Mississippian unconformity has been identified (Colpron et al. 2006b, Murphy et al. 2006). The unconformity separating domains 2a and 2b may correlate with this regional unconformity.

Unit 8 reflects the onset of volcanism in the area, and may correlate with the onset of regional volcanism in Yukon-Tanana terrane following the period of erosion. The abundance of mafic to intermediate volcanoclastic material suggests deposition of unit 8 in association with a volcanic arc. Deposition in an arc or back-arc setting could have provided the bathymetric relief needed for density flows to form. Rare quartzite layers in unit 8 are inferred to reflect terrigenous input into the basin during volcanism.

The voluminous limestone and minor quartzite and calc-schist areas of unit 9 reflect a mixing of calcareous and siliciclastic sediments. Siliciclastic sediments were dominated by quartz sand and mud. As these siliciclastic sediments are not interbedded with mafic-intermediate volcanoclastic sediments, volcanism associated with unit 8 volcanoclastic sediments is inferred to have ceased by the time unit 9 was being deposited.

Unit 10 reflects a return to siliciclastic-dominated sedimentation in the basin. Graded beds in unit 10 reflect deposition from turbidity flows. This is interpreted to reflect transgression in the basin, drowning the carbonate platform.

Unit 11 is interpreted to have a limestone protolith. This limestone does not contain siliciclastic material. This lack of siliciclastic material is interpreted to reflect the cessation of siliciclastic sedimentation in this part of the Yukon-Tanana terrane.

2.4.2 Correlation with Yukon-Tanana terrane assemblages

The following section presents a correlation of the assemblages found in the lithostratigraphic domains of the Tincup Lake area with Yukon-Tanana terrane assemblages of Colpron et al. (2006b; Table 3). Quartzofeldspathic psammite, psammitic schist, schists and marble dominate Domain 1. These rocks have been correlated with the Snowcap assemblage, the basement assemblage of Yukon-Tanana terrane (Murphy et al. 2009). There is no evidence to suggest a tectonic break separating Domain 1 from the rocks correlated with the Snowcap assemblage. For these reasons, Domain 1 is correlated with the Snowcap assemblage.

Rocks of Domain 2a and Domain 2b are similar to those of Finlayson and Klinkit assemblages reported by Colpron et al. (2006b) in the Glenlyon area, approximately 200 km northeast of Tincup Lake. In the Glenlyon area, Finlayson assemblage is composed of the lower Mississippian Drury formation, mid-lower Mississippian Pelmac quartzite and late-lower Mississippian Little Kalzas formation (Colpron et al. 2006b). The Drury formation is a succession of coarse-grained arkose psammite, quartzite, psammitic schist, local marble lenses and carbonaceous phyllite that unconformably overlie the Snowcap assemblage (Colpron et al. 2006b). The Pelmac quartzite unconformably overlies the Drury formation. The Pelmac quartzite is an extensive unit, consisting of massive to well-bedded, mature quartz arenite (Colpron et al. 2006b).

Gradationally overlying the Pelmac quartzite is the Little Kalzas formation (Colpron et al. 2006b). The Little Kalzas formation consists of arc-related volcanic and volcanoclastic rocks divided by a carbonate horizon (Colpron et al. 2006b). In the Glenlyon area, the Klinkit assemblage comprises the Little Salmon formation which unconformably overlies rocks of the Finlayson assemblage (Colpron et al., 2006b). The middle Mississippian-Pennsylvanian Little Salmon formation consists of a basal conglomerate, felsic to intermediate volcanoclastic rocks and minor massive mafic and felsic volcanic rocks, a prominent carbonate horizon in the lower part of the formation, and local quartzite (Colpron et al. 2006b; Simard et al 2003).

Rocks of Domain 2a are lithologically similar to, and therefore correlated with, the Finlayson assemblage and those of Domain 2b resemble, and are therefore correlated with, the Klinkit assemblage. Furthermore, as in Tincup Lake area, the contact between the Finlayson and Klinkit assemblages near Carmacks is inferred to be a profound unconformity, affirming the correlation between these two parts of Yukon-Tanana terrane.

2.5 Summary and Conclusions

The Tincup Lake area comprises 3 fault-bound lithostratigraphic domains (Map 1). Domains 1, 2a and 2b are composed of metasedimentary units (Figure 2.1). Domain 3 consists of the allochthonous Doghead assemblage (Figure 2.1), which was not addressed in this chapter. Within domains 1 and 2, 12 map units, with a total structural thickness of approximately 12 km, have been defined. Of these 12 map units, 4 consist of marble and 8 are metasiliclastic or metavolcanoclastic. Marble units are excellent marker horizons that can be traced across large parts of the study area (Map 1). Despite the structural complexity of the Tincup Lake area, the order and relationships between map units are maintained throughout the study area. This consistency allows for a lithostratigraphic sequence to be defined (Figure 2.1).

The rocks of Domain 1 (units 1-3) are correlated with the Snowcap assemblage of Yukon-Tanana terrane. Unit 1, Gt-Ms-Bt-Qz-Fsp psammite, psammitic schist, and schist (Figure 2.2), is gradationally overlain by Unit 2, Cal marble. This is interpreted to reflect deposition in a continental margin setting. Unit 3 is interpreted as a mafic flow. A moderately south-dipping normal fault (D₅) separates Domain 1 from Domain 2, precluding any interpretation of stratigraphic relationship between these domains.

Domain 2 consists of rocks correlated with the Finlayson (Domain 2a, units 4-7) and Klinkit (Domain 2b, units 8-12) assemblages of the Yukon-Tanana terrane (Figure 2.1). In the Tincup Lake area, the Finlayson assemblage consists of unit 4-quartzite, unit 5-interlayered and conglomeratic quartzite and Cal marble, unit 6-Qtz,-Dol-Cal marble, and unit 7-± Py-Bt-Ms-Gr-Qz phyllite. The predominance of quartzite (metamorphosed quartz arenite) and Cal marble (metamorphosed limestone) is interpreted to reflect deposition in a shallow marine setting. The stratigraphically highest unit in Domain 2a, ± Py-Bt-Ms-Gr-Qz carbonaceous phyllite, is interpreted as a metamorphosed carbonaceous mudstone. This progression from quartz arenite to limestone to carbonaceous mudstone is interpreted to represent a transgressive event. An unconformity is inferred to separate rocks of the Finlayson assemblage from rocks of the Klinkit assemblage. This is interpreted to reflect a regression event, allowing for erosion of the Finlayson assemblage.

The Klinkit assemblage consists of unit 8-±Mag±Grt±Act±Ep-Cal-Bt-Ms-Chl-Pl-Qz psammite, schist and Ep-Bt-Ms-Chl-Act-Pl amphibolite (Figure 2.11), unit 9-Cal marble, unit 10- Bt-Ms-Chl psammite and ±Gt±Chl-Bt-Ms-Fsp-Qz psammitic schist (Figure 2.12), unit 11-Cal marble (Figure 2.13), and unit 12 – Chl schist and Gt Pl Act amphibolite. Metavolcaniclastic mafic to intermediate rocks of unit 8 are interpreted to record the onset of volcanism in the Yukon-

Tanana terrane. These rocks are interpreted to have a slope or volcanic apron depositional environment. As volcanism and volcanoclastic sedimentation waned, carbonate sediments with a siliciclastic component (unit 9) became predominant.

Quartzofeldspathic turbidites (unit 10) overlie the dirty limestone of unit 9. This is interpreted to reflect continued deepening of the basin. The highest stratigraphic unit, unit 11, is a clean marble, interpreted to represent deposition of lime mud in a deep basin. Unit 12 is interpreted as a post-depositional, pre-tectonic intrusion.

The cause of the transgression-regression-transgression sequence, inferred from this study, is unclear at this time. The unconformable contact between the Klinkit and Finlayson assemblages provides a horizon for regional correlation, specifically with regards to the onset of volcanism in the western Yukon-Tanana terrane between the Middle Mississippian and Late Permian.

Figure 2.1: Stratigraphic units and relationships in the Tincup Lake area (Map 1). See text for discussion. Ages of domains are inferred based on correlation of domains with Yukon-Tanana terrane assemblages (Table 2.1). Chul. – Chulitna; Dog. – Doghead; Snow. – Snowcap.

Figure 2.1

Figure 2.2: (A) Westward view of a vertical rock face of unit 1. Quartzofeldspathic psammite and feldspar pebbly psammite are isoclinally folded by F_2 . Top direction is inferred to be to the right; tops are inferred on the basis of decreasing pebble size. (B) Southwest view of a vertical outcrop of beige-weathering psammite and grey-weathering Grt-Bt psammitic schist of unit 1. (C) Westward view of a vertical outcrop of graded light grey to white psammite and rusty-brown to green Grt-Bt schist of unit 1. Fining upward from psammite to schists suggests top direction is to top right.

Figure 2.2

Figure 2.3: Southwest view of a vertical rock face of interlayered unit 2, Cal marble and unit 3, amphibolite. Hammer handle is 30 cm long.

Figure 2.4: (A) Westward view of outcrop of unit 3 amphibolite. The edges of the outcrop are fine grained and the grain size increases towards the centre. Outcrops are surrounded by recessively weathering, chlorite-rich selvages. These outcrops are interpreted as pillows and interflow sediments. Hammer handle is 30 cm long. (B) Westward view of a vertical rock face of fine-grained unit 3 amphibolite. The pitted appearance is due to recessive weathering of grey-white carbonate which may be may be relict amygdules. Orange and white patches on the top and bottom of the outcrop are lichen.

Figure 2.3

Figure 2.4

Figure 2.5: Westward view of a vertical rock face of peach-weathering quartzite with black to grey, recessively weathering lenses of calcite (unit 4). Calcite lenses are inferred to reflect primary bedding.

Figure 2.6: Northwestward view of beige-weathering quartzite with finely layered, grey-weathering Cal marble (unit 4). The overlying light grey marble contains 5 cm-thick layers of quartzite. This is interpreted as a gradational contact between the quartzite (unit 4) and marble (unit 5). Hammer handle is 30 cm long.

Figure 2.5

Figure 2.6

Figure 2.7: A boulder of monomictic, framework- to matrix-supported metaconglomerate of unit 5 containing pebbles and cobbles of grey-weathering Cal marble in beige- to white-weathering quartzite matrix. (A) View is perpendicular to the long axis of elongate clasts. (B) View is perpendicular to the trace of the flattening plane determined from oblate marble rods. The interlayered marble and quartzite are interpreted as representing compositional layering, albeit tranposed.

Figure 2.7

Figure 2.8: (A) Westward view of a vertical outcrop of unit 5; light- grey-weathering Cal marble with: tan-weathering quartzite interlayers up to 1 cm-thick; quartzite clasts up to 5 cm long; and, (B) angular to subrounded clasts of medium- grey-weathering dolomite clasts up to 4 cm in diameter. (C) Quartzite layers are deflected around dolomitic clasts on the lower side of the clast. This wrapping of clasts is inferred to reflect primary compositional layering, albeit transposed. Compositional layering, S_0 , and the dominant foliation S_1 dips moderately to the south.

Figure 2.8

Figure 2.9: Eastward view of a north-dipping contact between grey- to white-weathering massive to medium-bedded Cal marble (to the right of hammer) and white- to grey- weathering finely layered (S_0) Cal marble and dolomitic marble (unit 6). Hammer handle is 30 cm long.

Figure 2.9

Figure 2.10: Northwest view of a vertical rock face of (G) black Gr phyllite (unit 7) with interlayers of grey-weathering Cal marble (M) up to 15 cm-thick -. S_0 dips to the southeast and is folded by F_3 .

Figure 2.10

Figure 2.11: Conglomeratic layers within 10 m of the lower contact of unit 8. (A) Westward view of a vertical rock face of bimictic, matrix-supported, metaconglomerate typical of the base of unit 8. Clasts are up to pebble sized, consisting of Cal marble, not pictured, and Hbl metagabbro. The matrix consists of Chl schist. (B) Metasiliciclastic metaconglomerate consisting of grey-weathering pebbles of psammitic schist (indicated by arrows) in a pelitic schist matrix.

Figure 2.11

Figure 2.12: Northwest view of a rock face of unit 10 containing 4 layers which grade from grey to green weathering Chl-Ms psammite to Ms-Chl schist. These gradational contacts are interpreted as right-way-up graded bedding. Layer A is composed of Ms-Chl schist. It is separated from the overlying Chl-Ms psammite by a recessively weathering contact (left side of image, obscured by lichen in middle of image). Layer B is quartzofeldspathic psammite. The lighter grey areas contain relatively less muscovite and chlorite than the darker grey-green parts of these layers. This gradual upward increase of muscovite and chlorite is interpreted to reflect graded bedding, with grain size decreasing towards the top of the figure. The base of layer C consists of quartzofeldspathic pebbly psammite and black to dark grey mm-scale micaceous clots. The latter are inferred to be rip-up clasts suspended in high concentration turbidity flows. The white band is a quartz vein concordant with compositional layering. The composition of layer D is the same as layer C, consisting of pebbly quartzofeldspathic psammite at the base, grading to Ms-Chl schist at the top.

Figure 2.12

Figure 2.13: Northward view of a vertical rock face of (M) grey- to white-weathering calcite marble of unit 11. Layering in the marble is defined by resistantly weathering, 5-15 cm-thick layers of medium- to coarse-grained marble separated by 1-10 cm-thick layers of recessively weathering fine-grained marble. The dominant foliation, S_2 , is concordant with recessively weathering layers and is oriented 139/47 [SW]. (D) A green-weathering Chl-Act-amphibolite dike, oriented 199/22 [W] crosscuts layering in the marble. Like the marble, the dominant foliation in the amphibolite is S_2 . This body is interpreted as a pre-tectonic dike or sill.

Figure 2.13

Tectonic Assemblage	Age	Characteristics	Tectonic Setting	This Study
Klondike	Middle to Late Permian	Felsic metavolcanic rocks, calc-alkaline volcanoclastic and volcanic rocks, coeval intrusions of the Sulphur Creek suite	Continental arc; Yukon-Tanana terrane	---
Klinkit	Middle Mississippian – Early Permian	Mafic to intermediate calc-alkaline volcanoclastic and volcanic rocks ; minor alkali basalt; limestone/marble basal conglomerate	Island arc; Yukon-Tanana terrane	Domain 2b
Finlayson	Late Devonian – Early Mississippian	Mafic to felsic metavolcanic rocks of arc and back-arc affinities; carbonaceous pelite, chert, minor quartzite ; volcanoclastic rocks, marble	Continental arc system (including arc and back-arc); Yukon-Tanana terrane	Domain 2a
Snowcap	Pre-Late Devonian	Polydeformed and metamorphosed quartzite, psammite, pelite and marble ; N-MORB to OIB amphibolite ; intruded by Devonian-Mississippian plutons	Metamorphosed continental margin; Basement to Yukon-Tanana terrane	Domain 1

Table 2.1: Correlation of Tincup Lake area assemblages with pericratonic assemblages of the Yukon-Tanana terrane (After Colpron et al. 2006a). Bolded rock types are present in the Tincup Lake area.

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3.1 Introduction

Chapter 3 - Geometry and kinematics of polyphase deformation in the Tincup Lake area, Yukon: Implications for the evolution of the western margin of the Yukon-Tanana terrane during the Mesozoic

INTRODUCTION

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3.1 Introduction

The structural evolution of the Yukon-Tanana terrane has been the focus of ongoing research (e.g. Mortensen and Jilson 1985, Mortensen 1990, Murphy 2004, Colpron et al. 2006a). The majority of these studies have focused on the deformation history of the eastern and central parts of the Yukon-Tanana terrane. In the eastern and central parts of the Yukon-Tanana terrane widespread pre-Mississippian deformation has been identified (Colpron et al. 2006b) as well as widespread Late Permian brittle and ductile deformation (e.g. Mortensen 1990, Murphy et al. 2006, and Berman et al. 2007, Table 3.1). In contrast, deformation of the western parts of the Yukon-Tanana terrane is poorly understood.

Prior to this study, the most recent mapping in the Yukon-Tanana terrane of southwestern Yukon, in the vicinity of the Tincup Lake area, was by Murphy et al. (2009), Murphy (2010), and Murphy et al. (2011). Murphy et al. (2009) undertook regional-scale mapping of Yukon-Tanana and neighbouring terranes north of the Denali fault, from the Yukon-Alaska border to Kluane Lake, identifying rocks of the Yukon-Tanana terrane and overthrust mafic to ultramafic bodies (Figure 3.1). In the northern Kluane Lake map area a mafic and ultramafic body, the Doghead assemblage, is in thrust contact with interlayered marbles and schists of the Yukon-Tanana terrane (Murphy, 2010). The fault separating the Doghead assemblage and Yukon-Tanana terrane rocks is tentatively named the Nuntaea Creek thrust fault. The Doghead assemblage and the metasedimentary assemblage are, together, folded by an east-west trending regional-scale synform (Murphy et al. 2009). Aside from the identification of a regional-scale synform affecting rocks of both the Yukon-Tanana terrane and the Doghead assemblage, no attempt has been made to characterize or constrain the timing of folding along the western margin of the Yukon-Tanana terrane.

This chapter presents the results of a 1:20,000 scale structural mapping campaign of the well exposed upper greenschist to lower amphibolite rocks in the footwall of the Nunataea Creek fault, Tincup Lake area, southwest Yukon. The geometry of folds (Map 1, cross-section) and the description of associated tectonic fabrics in the footwall are presented first (Section 3.2). An argument for the relative timing of deformation and possible regional correlations is then presented (Section 3.4.1). Finally, an interpretation of the post-Permian to pre-Late Cretaceous structural evolution of the Yukon-Tanana terrane's western margin is proposed. (3.4.2).

3.2 Observations - Folding and Fabrics

3.2.1 Overview

Four generations of folds, D_1 - D_4 , occur in the Tincup Lake area. Ordering of the generations of structures is determined by map- (Map 1 and Cross-section) and outcrop-scale interference patterns and crosscutting relationships of penetrative foliations of different generations. Ductile structures include folds with axial planar foliations, intersection lineations, crenulations, mineral lineations, and clast elongation lineations. Terms used to describe structures and fabrics follow the nomenclature of Twiss and Moores (2007). The following describes folds in the Tincup Lake area (Map 1).

3.2.2 Pre- D_1 deformation: Garnet inclusion trails

As documented by petrographic analysis, subhedral garnets in units 1, 8 and 10 contain quartz and feldspar inclusion trails. Inclusion trails are termed S_i . Unit 1 is correlated with the pre-Late Devonian Snowcap assemblage. The Snowcap assemblage is the only assemblage in the Yukon-Tanana terrane to record a Devonian-Mississippian deformation event (Colpron et al. 2006b). Therefore, S_i in unit 1 may be older than that of units 8 and 10, which are correlated with the

Middle Mississippian to Early Permian Klinkit assemblage. It was not possible to determine whether or not these inclusion trails were coeval from this study.

The angle between S_i and the dominant foliation (S_1 , see below) of the surrounding matrix is variable. In some garnets, S_i is nearly orthogonal to the dominant foliation throughout the garnet (Figure 3.3a). In other cases S_i is sigmoidal, varying from a high angle in the core of the garnet to a low angle with the dominant foliation at the edges (Figure 3.3b). No evidence of this earlier fabric, S_i , is encountered beyond the microscopic scale. For this reason the nomenclature of macroscopic structures and associated metamorphic events will begin at D_1 .

3.2.3 D_1

Macroscopic F_1 folds are rare in the Tincup Lake area. It is unclear if this is a reflection of a genuine paucity of macroscopic F_1 folds, poor exposure or a lack of distinctive markers that can be used to identify F_1 folds. F_1 folds are only encountered in rocks of Yukon-Tanana terrane in the Tincup Lake area. Map-scale F_1 folds have a minimum wavelength of 1.5 km and amplitude of approximately 1 km. Evidence of F_1 folds at the map-scale comes from apparent repetition of approximately 500 m of older units over younger units in the hanging wall of the Division Valley fault (cross-section). The units are in sharp contact; however, there is no reduction in grain size near these contacts nor are the rocks brecciated. Thus these contacts are likely primary contacts rather than fault contacts. In all cases where these stratigraphic repetitions occur, F_1 folds are north vergent.

At the outcrop scale, F_1 isoclinal folds are preserved as rootless, cm-scale fold hinges that fold S_0 compositional layering (Figure 3.4). These hinges are most commonly observed in siliciclastic

layers with minor interlayered marble, where the contact between siliclastic layers and marble layers or lenses can be easily traced. F₁ folds are rarely preserved in marble units.

The dominant foliation is axial-planar to F₁ folds and is concordant with compositional layering (S₀) outside of F₁ fold hinges. For this reason, the dominant foliation is termed S₁. S₁ is a rough to moderately rough, mm-scale disjunctive foliation. In siliclastic units, cleavage domains are defined by the preferred orientation of $\pm$ graphite $\pm$ biotite $\pm$ muscovite $\pm$ chlorite. In siliclastic units microlithons show a strong to complete preferred orientation of $\pm$ amphibole $\pm$ feldspar – quartz. In marble units, cleavage domains are defined by the preferred orientation of rusty-brown-weathering white mica and fine-grained calcite. In marble units, microlithons are defined by $\pm$ quartz and calcite.

The relationship between porphyroblasts and S₁ suggests either two generations of metamorphism; one post-S₁ and pre-S₂; and the other being syn- to post- S₁ development, or two phases of mineral growth in the same metamorphic event. Garnet is most common in micaceous layers, where S₁ wraps around subhedral, poikilitic garnets up to 1 cm in diameter and euhedral, inclusion-free garnets up to 0.5 cm in diameter. Poikilitic garnets are most common in layers with greater than 20% quartz and feldspar. Similarly, subhedral actinolite, up to 0.5 cm in length, is common in chlorite-rich micaceous layers. The long axis of actinolite is concordant with S₁. Unlike garnets, S₁ does not wrap around actinolite.

3.2.4 D₂

The S₁ foliation and quartz veins concordant with the S₁ foliation are folded by F₂ folds. F₂ folds are characteristically similar in shape, tight to isoclinal and overturned. F₂ fold hinges trend east-

west, plunging primarily to the west (Figure 3.5a). The orientation of F_2 axial planes varies from moderately southward to northward dipping (Figure 3.5b).

Map-scale F_2 folds are defined by rocks of the Yukon-Tanana terrane (Map 1, cross-section).

Map-scale F_2 folds have a wavelength of 200-500 m and amplitude of >1000 m. The wavelength of F_2 folds is the smallest near the Nuntaea Creek fault and increases towards the Division Valley fault (Map 1). North of Conglomerate Plateau, map-scale F_2 folds do not occur.

At the outcrop scale, F_2 folds are best preserved in siliciclastic units where S_1 can be traced around F_2 fold closures (Figure 3.6). F_2 folds are rare in marble units. F_2 folds in marble are defined by colour bands inferred to be bedding (Fig. 3.4a). Second-order F_2 folds typically have amplitudes ranging from 5-30 cm and wavelengths ranging from 1-10 cm. At the outcrop scale, psammitic layers show slight thickening in the hinge zone and thinning in the limbs. Schistose layers typically have thickened hinges and thinned limbs.

A penetrative axial planar foliation occurs in association with F_2 folds. This foliation is termed S_2 . S_2 is rough to moderately rough with a spacing of 1-15 cm. In marble units, S_2 is defined by the long axis of recrystallized calcite. In siliciclastic layers, S_2 is composed of preferentially oriented muscovite and chlorite. In F_2 fold hinges, S_2 wraps around subhedral garnets up to 5 mm in diameter. In F_2 fold hinges, S_2 is at a high angle to and crenulates S_1 (Figure 3.6b). On F_2 limbs, S_2 is concordant with S_1 , making it difficult to distinguish between the two foliations. S_2 is poorly developed in psammitic layers. In all cases, S_2 wraps around garnet, amphibole and magnetite porphyroblasts.

Petrographic investigation indicates garnet, amphibole, and magnetite porphyroblast growth predated the growth of chlorite associated with S_2 (Figure 3.7). In Unit 8, garnets are irregularly

shaped and are mantled by chlorite that is concordant with S_2 foliation. Amphiboles in Unit 8 are preserved either as coherent actinolite-hornblende intergrowths with a chlorite mantle or fractured actinolite with chlorite growth both in the fractures and as a mantle around the grain. Magnetite porphyroblasts are euhedral and are wrapped by the S_2 chlorite.

Mineral lineations, crenulation lineations, clast elongation lineations, and intersection lineations occur in association with F_2 . Mineral lineations (L_2) are defined by elongate calcite aggregates in units 6 and 9, quartz aggregates in unit 9, amphibole in unit 8, and quartz-feldspar aggregates in unit 10. Crenulation lineations (L_2) are best preserved in Unit 10 where S_1 is crenulated in the hinge zone of F_2 folds. Clast elongation lineations (L_2) are defined by elongate, angular clasts of calcite marble in unit 5, rounded dolomite clasts in unit 6, and rounded clasts of hornblende metagabbro rocks in unit 8. These clast elongation lineations typically plunge gently to the west and less commonly to the south (Figure 3.8). West-plunging clast elongation lineations are concordant with F_2 fold hinges, leading to their classification as L_2 . Occurrences of the south-plunging clast elongation lineation in Yukon-Tanana terrane rocks are restricted to within approximately 3 km of the fault contact with the overthrust Doghead assemblage (Figure 3.8B, C, red symbols; Figure 3.9).

3.2.5 D_3

Map-scale F_3 folds affect both Yukon-Tanana terrane rocks and rocks of the Doghead assemblage (Murphy et al., 2009). F_3 folds are gentle to open, similar folds. Map-scale F_3 folds have a wavelength of 5 km and amplitude of 1-2 km. At the map scale, F_3 long limbs dip gently to the north or south while overturned short limbs dip moderately to steeply to the south, defining a macroscopic northwest- to northeast-vergent fold train. Map-scale F_3 axial planes dip

moderately to the south. At all scales, F_3 fold hinges are approximately coaxial with F_2 fold hinges – plunging gently to the west and less commonly gently to the east (Figure 3.5c).

The majority of mesoscopic F_3 folds vary continuously between northwest- to northeast-vergent. Where mesoscopic north-vergent F_3 fold trains are encountered, their enveloping surfaces typically dip gently to the south to gently to the north. Southwest- to southeast-vergent F_3 folds occur in relatively narrow (<1 km-thick) bands. The enveloping surfaces of southwest- to southeast-vergent F_3 fold trains dip moderately to steeply to the south. The orientation of mesoscopic F_3 axial planes varies from gently dipping to the northwest (~33%) to gently to moderately dipping to the southwest (~40%) to gently dipping to the southeast (~23%) (Figure 3.5d). At the outcrop scale, F_3 folds have a wavelength of up to 5 m and amplitude of 5-50 cm.

The geometry of mesoscopic F_3 folds is influenced by rock type. In quartz- and feldspar-rich layers, F_3 fold hinges are typically rounded and short limbs are rarely overturned. Conversely, in folded schistose layers, F_3 fold hinges are typically angular and short limbs are regularly overturned (Figure 3.10).

The axial planar fabric of F_3 folds, S_3 , is not as well developed as S_2 , the axial planar foliation of F_2 folds. S_3 occurs both as a penetrative foliation and as a cleavage surface with fine-grained mica. This foliation is defined by the preferred orientation of muscovite and chlorite. The penetrative S_3 foliation occurs most commonly where schistose layers sit in cusped F_3 fold hinges, appearing as a crenulation of the S_1 and S_2 foliations. The penetrative S_3 foliation is not well developed in lobate F_3 fold hinges. In quartz- and feldspar-rich layers, an axial planar cleavage forms rather

than a penetrative foliation. In psammitic schist layers, fine-grained muscovite and chlorite are present on cleavage surfaces

Aside from fold hinges, the only lineations associated with D_3 are intersection lineations. Two examples of the intersection between S_1 and S_3 , L_{1-3} , were documented in the field. One measurement plunges gently to the southeast while the other plunges moderately to the northwest. These two measurements define a great circle that also contains L_{1-2} (Figure 3.11a). Three examples of the intersection between S_2 and S_3 , L_{2-3} , were encountered in the field. L_{2-3} consistently plunges gently to the west, with an average orientation of $15 \rightarrow 285$ (Figure 3.11a).

3.2.6 D_4

F_4 folds are the least commonly observed folds in the study area. These folds affect both Yukon-Tanana terrane rocks and rocks of the Doghead assemblage (Murphy et al. 2009). F_4 folds are gentle to open, upright, parallel folds. F_4 axial planes dip steeply to the east. The fold hinges of F_4 folds plunge gently to the south. Map-scale F_4 folds have a wavelength of approximately 5 km and amplitude of approximately 0.5 km. This estimate of wavelength comes from variations in the orientation of F_2 and F_3 fold axes which have been warped by F_4 .

At the outcrop scale, F_4 folds rarely occur. Mesoscopic F_4 folds have a wavelength at least 2 m and amplitude of approximately 0.5 m. These folds lack an axial planar foliation, cleavage or an associated lineation. The asymmetry of these folds suggests F_4 is east-vergent at the macroscopic scale.

3.2.7 Map and Fold Interference Patterns

Map-pattern and cross-section (Map 1) interpretations indicate one map-scale F_1 synform-antiform pair with a wavelength of at least 500 m and amplitude of approximately 1000 m. The axial planes of these F_1 folds dip steeply to the south and fold hinges plunge gently to the east.

These map-scale F_1 folds lie south of the Division Valley fault. Both synform-antiform pairs indicate that F_1 folds are part of a north-vergent fold train.

Also south of the Division Valley fault, 100m-scale F_2 folds with south-dipping axial planes, have a north-vergent asymmetry. F_2 folds have the smallest interlimb angle and shortest wavelength near the fault contact between Yukon-Tanana terrane and the Doghead assemblage. The interlimb angle and wavelength of F_2 folds increases, from less than 1 km to greater than 2 km, with distance from the Doghead assemblage. North of Conglomerate Plateau no map-scale F_2 folds were encountered. It is unclear whether this is because F_2 folds do not occur north of Conglomerate Plateau or as a result of poor exposure.

Across the entire map area, map-scale F_3 folds are north vergent. Similar to F_2 folds, F_3 folds are more common in the south of the study area, near the Doghead assemblage. Map-scale F_3 folds are most common between the Doghead thrust fault and Conglomerate Plateau. However, unlike F_2 folds, F_3 folds do not show a variation in wavelength or amplitude in this area. Between Conglomerate Plateau and the Marble Top fault, F_3 folds control the map pattern. In this area, mesoscopic F_3 folds are north vergent. The exception to this is within approximately 1 km of the Marble Top fault, where F_3 folds are more commonly south vergent. This change in vergence indicates a map-scale F_3 fold hinge occurs within 1 km of the Marble Top fault.

F_2 and F_3 are approximately coaxial at all scales, leading to the development of type 3 interference patterns (Ramsay and Huber 1987). South of the Division Valley fault, the map pattern is best interpreted as a type 3 fold interference pattern (Map 1). North of Division Valley the map pattern does not indicate an interference pattern between F_2 and F_3 folds. A major F_3 syncline lies between Conglomerate Plateau-Anticline Ridge and Bear Valley. Between Bear

Valley and the Marble Top fault, the dominant foliation dips gently to moderately to the north and mesoscopic F_3 folds are north vergent. This area is interpreted as a major long limb of F_2 and F_3 . Since this area is a long limb, there are few parasitic folds to create an interference pattern.

Evidence of F_4 folds occurs in both Yukon-Tanana terrane rocks and Doghead assemblage rocks (Section 3.2.6, Murphy, personal communication, 2011). This generation of folds, F_4 , does not show a spatial relationship to any geological features in the study area, i.e. it affects all units and does not appear to be related to faulting.

Type 2 and type 3 fold interference patterns are encountered at the outcrop-scale (Figure 3.12).

Type 2 interference patterns between F_1 and F_2 folds, occur on vertical, east- or west-facing outcrops north of the Division Valley fault. The map pattern does not indicate type 2 interference at the map-scale. The lack of map-scale type 2 interference patterns may be the result of erosion, poor exposure, poor observation coverage, lack of marker horizons, or a genuine lack of map-scale type 2 fold interference between F_1 and F_2 . If the mesoscopic type 2 fold interference is representative of the map pattern north of Division Valley, then it would suggest that F_1 fold axes and axial planes are at an angle to the F_2 and F_3 fold axes and axial planes.

3.3 Interpretation

Four phases of folding have been identified in the Tincup Lake area. F_1 , F_2 and F_3 are overturned north-vergent folds with east-west trending fold hinges. Upright F_4 folds are east vergent and plunge gently to the south. The following section addresses each fold generation in turn, interpreting macroscopic structures, and conditions of deformation, and discussing the significance of the structures.

3.3.1 D₁

The repetition of >250 m-thick panels of older over younger stratigraphy is interpreted in the south-facing slope of “Division Valley”. Here Unit 5 structurally overlies and is overlain by Unit 8, a geometry that is interpreted as an overturned F₁ synform-antiform pair. Map-scale F₁ synform-antiform pairs and the map pattern south of Division Valley suggest north-vergent F₁ are prevalent. North of the “Division Valley” fault and south of the “Marble Top” fault, evidence of map-scale F₁ folding is lacking. Possible explanations for the lack of F₁ folding include: 1) this area lies on the long limb of a major F₁ fold, 2) overprinting by D₂ and D₃ was sufficient to obliterate evidence of large-scale F₁ folds, or 3) a lack of marker horizons makes it difficult to identify the F₁ folds.

3.3.2 D₂

Both the relatively high frequency of map-scale F₂ folds near the Nuntaea Creek fault and the variation in wavelength and amplitude of map-scale F₂ folds with respect to distance from the Nuntaea Creek fault (Section 3.2.4) are inferred to reflect a genetic relationship between F₂ folds and thrust emplacement of the Doghead assemblage. Folding in the footwall could be accomplished if the Nuntaea Creek fault were a ductile shear zone, syntectonic with the formation of F₂ folds. Alternatively, the Doghead assemblage could have been overthrust along a discrete fault-which locked. After the fault locked, displacement would have then been transferred into footwall strain, resulting in folding. This study does not provide enough evidence to rule out either of these hypotheses.

The presence of an unconformity between rocks correlated with the Finlayson and Klinkit assemblages (Chapter 2, Section 2.3.3) may have also played a role in the development of prevalent fold trains in the area between the Nuntaea Creek fault and the Division Valley fault.

Irregularities along an unconformable surface may serve as a locus for fold development (cf. Hudleston and Treagus, 2010). North of Conglomerate Plateau, the unconformity is interpreted to project into the subsurface. The unconformity, along with the overlying units, approximates a homocline that dips gently to moderately to the north.

On the basis of mineral assemblages that grew during folding, and are consistent with the fold style, Yukon-Tanana terrane rocks are interpreted to have been at greenschist conditions during thrust faulting of the Doghead assemblage over Yukon-Tanana terrane rocks. Ductile flow of Yukon-Tanana terrane rocks during emplacement of the Doghead assemblage is inferred to have resulted in the L_2 clast elongation lineation. In order to accommodate shortening, local coaxial flow is inferred to have occurred. This coaxial flow led to the elongation of marble clasts parallel to F_2 fold axes in unit 5 and metamafic clasts in unit 8. The crenulation of S_1 in F_2 fold hinges is interpreted to reflect the reorientation of biotite and muscovite into the F_2 axial plane. The prevalence of chlorite growth in the S_2 foliation suggests D_2 occurred at greenschist facies conditions. Chlorite selvages on garnet, actinolite and biotite, along with hornblende being partially replaced by actinolite, support this interpretation. Variations in the orientation of S_2 are attributed to folding of S_2 during the F_3 event.

The occurrence of south-plunging lineations (L_2) within 3 km of the Nuntaea Creek fault implies a relationship between this south-plunging lineation and emplacement of the Doghead assemblage. If we can assume the Doghead assemblage emplacement approximates orthogonal emplacement rather than transpressive emplacement, then these lineations can be used to infer north or south transport direction (Tikoff and Greene 1996). A southward transport direction can be ruled out since there are no intraoceanic arc terranes north of the Tincup Lake area. For this reason, the transport direction of the Doghead assemblage is interpreted to be to the north.

Thus, emplacement of the Doghead assemblage and development of F_2 folds are consistent with the observation of map-scale F_2 folds forming a north-vergent fold train.

3.3.3 D_3

F_3 folds, like F_2 folds, are most common between the Nuntaea Creek fault and the Division Valley fault. The prevalence of map-scale F_3 folds in this area is interpreted to be the result of F_3 folds nucleating on F_2 fold hinge zones. The frequency of higher-order F_3 folds in this area may also be influenced by its proximity to the regional-scale F_3 hinge zone that contains the Doghead assemblage (e.g. 586500E, 6847250 N, Map 1).

Like F_2 folds, map-scale F_3 folds define a north-vergent long limb fold train in this area. The area between the Division Valley fault and Division Valley contains predominantly south-vergent mesoscopic F_3 folds. This area is interpreted as a map-scale F_3 short limb. Up to approximately 1 km south of the Marble Top fault, south-vergent mesoscopic F_3 folds are also common. This area is also interpreted as a map-scale F_3 short limb. The scarcity of F_3 folds north of Conglomerate Plateau and the area of south-vergent mesoscopic F_3 folding near the Marble Top fault are interpreted as a map-scale F_3 long limb. This pattern of map-scale F_3 long limbs and short limbs indicates a macroscopic, north-vergent F_3 fold train in the Tincup Lake area.

Although map-scale F_3 axial planes typically dip moderately to the southwest, mesoscopic observations indicate F_3 axial planes vary from dipping gently to moderately to the northwest, to gently to moderately to the southwest to dip gently to moderately to the southeast. The reason for this variation in F_3 is unclear at this time. Possible explanations include an undocumented phase of folding that is post- F_3 but pre- F_4 , disharmonic folding, or the orientation of S_3 may have

also been influenced by the orientation of compositional layering as a result of earlier folding events.

The re-orientation of chlorite in cusped F_3 hinge zones and the growth of fine-grained chlorite and muscovite on cleavage planes suggest F_3 folds formed under greenschist conditions. The reason for the difference in development of the S_3 axial planar fabric; a crenulated foliation in schistose layers and fine-grained muscovite and chlorite on the axial planar cleavage in psammitic layers, is unclear at this time.

3.3.4 D_4

F_4 folds are the only folds with a fold axis that is not east-west trending and verge to the east. The different orientation of F_4 from the three earlier documented folding events suggests a change in the direction of shortening from a north-south directed maximum shortening to east-west directed maximum shortening. This phase of deformation is the cause of east-west variations in F_2 and F_3 fold hinges. F_4 folds did not develop an axial planar cleavage and no metamorphic minerals can be correlated with this phase of deformation.

3.3.5 Relationship between metamorphism and deformation

S_1 is interpreted to record garnet growth that is syntectonic with formation of S_1 . However, since S_1 wraps around both subhedral poikilitic garnets and inclusion free, euhedral garnets, S_1 is inferred to have continued to develop after metamorphic conditions no longer favored garnet growth. Alternatively, the wrapping of garnets by S_1 could reflect a tightening of S_1 during D_2 . There are insufficient data from this study to rule out either of these hypotheses.

Mafic schists contain amphibole. While the majority of these amphiboles are now actinolite, the intergrowth of actinolite and hornblende (Section 3.2.4) suggests actinolite was originally hornblende. The long axis of actinolite crystals is typically oriented either concordant with or at

a low angle to S_1 . It is unclear whether this is the result of rotation of actinolite crystals into the foliation plane or growth of actinolite in nutrient-rich layers while the rocks were being deformed. The lack of pressure shadows favours syn- to post-tectonic growth rather than rotation during F_1 . In either case, actinolite must be pre- D_2 since it remains concordant with S_1 around F_2 fold hinges.

The relationship of garnet with S_1 , the presence of amphibole porphyroblasts, and the development of a biotite-muscovite-chlorite foliation suggest D_1 occurred at upper greenschist to lower amphibolite conditions.

3.4 Discussion/Implications

The following discussion attempts to constrain the timing of deformation events in the Tincup Lake area based on regional-scale crosscutting relationships. Three generations of north-vergent folds have been identified in the Middle Permian and older rocks of the Tincup Lake area. A fourth generation of folding, an open warp of earlier structures about a gently-to-the-south-plunging fold axis, is also identified. Once the timing of these folding events has been constrained, possible correlations between events in the Tincup Lake area and events documented elsewhere in the Yukon-Tanana terrane are proposed.

3.4.1 Constraints on timing of folding and metamorphism

The timing of folding and metamorphism in the Tincup Lake area is broadly constrained by the inferred depositional age of metasedimentary rocks (Chapter 2) and by cross-cutting brittle faults (Chapter 4). The youngest metasedimentary rocks in the area are correlated with the Middle Mississippian to Early Permian Klinkit assemblage (Chapter 2). Since these rocks have undergone all documented phases of folding, folding must be younger than Early Permian. Steep

strike-slip faults are interpreted to be Early Cretaceous or younger (Chapter 4). This constrains folding to be older than late Early Cretaceous (Israel and Westberg, 2011). The following discussion applies the result of reconnaissance mapping by Murphy (2010) and Israel and Westberg (2011) to further constrain the timing of deformation events in the Tincup Lake area.

3.4.1.1 Possibility of pre- D_1 deformation in Snowcap Assemblage

This study has focused on the map area between the Nuntaea Creek fault and the Marble Top fault. North of the Marble Top fault are rocks correlated with the Snowcap assemblage of the Yukon-Tanana terrane. Inclusion trails in garnet, S_i , indicate that these rocks record a pre-Mississippian deformation event (e.g. Colpron and Reinecke, 2000; Colpron et al., 2006b; Berman et al. 2007). From this study, there are insufficient data to evaluate this possibility.

The possibility of a pre-Mississippian deformation event in rocks correlated with the Snowcap assemblage has implications for the proposed significance of S_i . The inclusion trails in unit 3 garnets may be the same age as those in units 8 and 10. Alternatively, inclusion trails in unit 3 garnets may reflect a deformation event that affected only the Snowcap assemblage. If S_i in unit 3 is older than that in units 8 and 10, then it could be as old as Devonian (Colpron et al. 2006a). If the correlation of Units 8 through 12 with the Klinkit assemblage is accurate, then S_i in units 8 and 10 must be younger than Early Permian.

3.4.1.2 D_1

F_1 affected all metasedimentary rocks in the Tincup Lake area. The youngest rocks in the study area are correlated with the Klinkit assemblage (Chapter 2, Section 2.4.2), constraining D_1 to Late Permian or younger. Late Permian deformation, both brittle and ductile, has been widely documented in the Yukon-Tanana terrane (e.g. Mortensen 1990, Mortensen 1992, Murphy et al. 2006, Berman et al 2007, Mackenzie et al. 2007). Late Permian deformation has been attributed

to west-dipping subduction of the Slide Mountain terrane under the Yukon-Tanana terrane (Berman et al., 2007). D_1 may be related to this Yukon-Tanana terrane-wide deformation event. However, without further evidence to support this hypothesis D_1 can only be constrained to post-Late Permian and pre- D_2 .

3.4.1.3 D_2

F_2 affected all metasedimentary rocks in the Tincup Lake area. It is unclear whether F_2 structures occur in the Doghead assemblage but the change in orientation and shape of F_2 folds toward the thrust fault suggests that these structures are contemporaneous with the overthrusting of the Doghead assemblage. Limited observations of the Doghead assemblage showed it to contain bodies of medium-grained, unfoliated gabbro. No evidence of F_2 structures or associated fabrics was observed in the Doghead assemblage. Therefore F_2 must be pre- to syntectonic with emplacement of the Doghead assemblage. If this is the case, then F_2 must be post-Late Triassic.

Post-Late Triassic deformation is widely documented in the Yukon-Tanana terrane. North of Tincup Lake, Berman et al. (2007) attribute deformation and metamorphism in the Stewart River area in the Early Jurassic (ca. 195-187 Ma) to thickening and imbrication of the Yukon-Tanana terrane as the result of collision with the North American craton. Also to the north of Tincup Lake, Mackenzie et al. (2007) report north-directed thrust faulting and isoclinal folding within thrust sheets in the Klondike area during the Jurassic. Israel and Westberg (2011) correlate southwest-vergent isoclinal folding in the Aishihik area, to the east of the study area, with SW directed thrusting documented by Johnston and Erdmer (1995). At this time it is unclear how the Tincup Lake area fits in to the regional Jurassic deformation story.

3.4.1.4 D_3

F_3 folds affect both Yukon-Tanana terrane and Doghead assemblage rocks in the Tincup Lake area (Murphy et al. 2009). There are two constraints on the timing of D_3 . At the southern limit of the study area the Doghead assemblage is inferred to sit in a F_3 synform (Murphy et al. 2009). Since gabbro in the Doghead assemblage is Late Triassic in age (Murphy et al. 2011), D_3 must be younger than Late Triassic. Southwest of Tincup Lake, the F_3 synform that contains the Doghead assemblage is crosscut by the undeformed 99-105 Ma Nisling Range granodiorite (Murphy 2010). Therefore, D_3 is constrained to post-Late Triassic and pre-105 Ma.

3.4.1.5 D_4

The timing of F_4 is poorly constrained in the Tincup Lake area. F_4 folds cause a plunge reversal of F_2 and F_3 fold hinges. S_2 , the F_2 axial planar foliation, is also warped about south-trending, gently plunging F_4 fold hinges. This is similar to observations made by Israel and Westberg (2011) in the Aishihik Lake area where the dominant foliation is gently warped about a north- to northeast-trending axis. Israel et al. (2011) infer this warp of the dominant foliation to be Late Cretaceous and likely coincides with ca. 84 Ma thrusting of Yukon-Tanana terrane over the Kluane Schist. F_4 is cautiously correlated with this Late Cretaceous deformation in the Kluane Lake area of Israel et al. (2011).

3.4.2 Regional variations in post-Permian deformation history

In the area surrounding Tincup Lake, two phases of ductile deformation have been identified and constrained by radioisotope dating. Between the Donjek River and Tincup Lake are rocks that record post-Late Permian deformation. Between the Donjek River and the Kluane River, south of the Doghead assemblage, quartz feldspar metaporphyrries have been dated at ca. 258

Ma and 254 Ma (Murphy, personal communication, 2011). The metaporphyrries are foliation parallel and are strongly foliated.

To the west of the Donjek River, post-Middle Permian deformation does not appear to be as intense as that recorded to the east. West of the Donjek River, ca. 260 Ma aphyric felsic dikes are discordant with the dominant foliation. These dikes are gently folded, weakly foliated and boudinaged (Murphy, personal communication, 2011). These observations suggest that this part of the Yukon-Tanana terrane has undergone multiple phases of ductile deformation. However, the younger phases of deformation affect only certain parts of the Yukon-Tanana terrane.

3.5 Conclusions

Four phases of Late Permian to Late Cretaceous folding have been characterized in the Tincup Lake area.

- F_1 and F_2 affected only Yukon-Tanana terrane rocks, which lie structurally below the allochthonous Doghead assemblage, units 1-12 (Map 1). F_3 and F_4 affected both the rocks of the Yukon-Tanana terrane and Doghead assemblage (Map). At the map scale F_1 , F_2 and F_3 are all north vergent. F_4 warps all earlier fabrics about a gently to the south plunging fold axis.
- F_1 is tentatively correlated with Late Permian deformation that has been documented across the majority of the Yukon-Tanana terrane. The dominant foliation, S_1 is axial planar to F_1 folds. S_1 is continuous with some garnet inclusion trails, suggesting garnet growth was pre- to syntectonic with the development of S_1 under lower amphibolite to upper greenschist conditions. The development of S_1 outlasted garnet growth.

- F_2 has a spatial relationship to the Doghead assemblage, with amplitude and wavelength of F_2 folds decreasing in Yukon-Tanana terrane rocks with distance from the Doghead assemblage. This suggests a genetic relationship between overthrusting of the Doghead assemblage over the Yukon-Tanana terrane in the Jurassic and the development of F_2 folds. S_2 wraps garnet and amphibole. D_2 is inferred to have occurred under greenschist facies conditions. D_2 is younger than Late Triassic, the age of the Doghead assemblage.
- F_3 folds are coaxial with F_2 folds. F_3 is constrained to older than post-Late Triassic, since the Late Triassic Doghead assemblage cores a mesoscopic F_3 synform in the Nuntaea Creek map area. F_3 folds are constrained to older than Late Cretaceous, since this mesoscopic F_3 synform which is cored by the Doghead assemblage is crosscut by regional-scale dextral-strike slip faults (Murphy 2010). F_3 is inferred to have occurred at lower greenschist facies conditions.
- F_4 is inferred to be Late Cretaceous, reflecting the correlation of F_4 with similar structures observed by Israel and Westberg (2011) in the Aishihik area. No metamorphism or tectonic fabric can be correlated with F_4 . This phase of deformation is inferred to reflect thrusting of the Yukon-Tanana terrane over the Kluane Schist in the Late Cretaceous.
- Late Paleocene or younger brittle northwest-southeast trending strike-slip faults cut all generations of folds (see Chapter 4). This indicates the Tincup Lake area was above the brittle-ductile transition zone by this time.

Figure 3.1: (A) Yukon map: shaded area near the Denali fault delineates Map B, “Terranes and assemblages of southwest Stevenson Ridge and northwest Kluane Lake map areas, southwest Yukon” (Escayola et al. in press). Bedrock in this area is dominated by rocks of the Yukon-Tanana terrane and rocks of the Selwyn basin. Rocks of the Yukon-Tanana terrane are separated from the Selwyn Basin by the ophiolitic Harzburgite Peak-Eikland Mountain complex. Along the southern side of Yukon-Tanana terrane rocks, the mafic to ultramafic Doghead assemblage structurally overlies rocks of the Yukon-Tanana terrane. The study area, marked by hatching ($139^{\circ} 20'W$, $61^{\circ} 45' N$), comprises Yukon-Tanana terrane rocks located north of the Doghead assemblage ultramafic rocks. Doghead assemblage: DR=Donjek River, DL=Dogpack Lake, DP=Doghead Point; Slide Mountain terrane: HP=Harzburgite Peak, WL=Wellesley Lake, D=Deadman Hill, EM=Eikland Mountain, SP=Sanpete Hill, KM=Koidern Mountain

Figure 3.1

Figure 3.2 Lithological units and their relationships in the Tincup Lake area. The polydeformed, polymetamorphosed Tincup Lake area is composed of twelve map units, correlated with the Yukon-Tanana terrane, and one map unit correlated with the Chulitna terrane (Map 1, Chapter 2). Marker horizons within rocks correlated with the Yukon-Tanana terrane (e.g. unit 9) can be traced across the study area, aiding in the delineation of map-scale structures. See chapter 2 for unit descriptions. See Map 1 for list of mineral abbreviations.

Figure 3.2

Figure 3.3: Photomicrographs (plane-polarized light) of garnet-actinolite-biotite-muscovite psammitic schist, showing relationship of garnet (Gt) and actinolite (Act) porphyroblasts to the dominant foliation (S_1). (A) Quartz and feldspar inclusion trails in the garnet (S_i) are at a high angle to S_1 suggesting S_i represents a foliation that predates garnet growth and is therefore older than S_1 . (B) Some garnets have internal inclusion trails that are continuous with, and morphologically similar to, the external foliation. The continuity and morphological similarity of the inclusion trail with the surrounding foliation suggests garnet started growing after the foliation started to develop. The garnet continued to grow during the evolution of the external foliation, overgrew the warped S_i , then stopped growing while the foliation continued to evolve, wrapping the garnet. These examples suggest there may be two phases of garnet growth recorded in the Tincup Lake area, one pre- D_1 and the other syn- D_1 . Actinolite is generally oriented with long axes in the S_1 plane. Here actinolite lies within the S_1 foliation or may have overgrown S_1 . Actinolite is interpreted as late syn- to post-tectonic with respect to development of S_1 . Samples (A) 10SMS118B, (B) 10SMS093.

Figure 3.3

Figure 3.4: (A) Westward view of a vertical outcrop face of marble with cm-scale F_2 tight to isoclinal folds folded by coaxial gently westward-plunging, north-verging F_3 folds. The layers comprises compositional layering (S_0) of white and grey layers of recrystallized marble and concordant resistantly weathering quartz veins or psammitic layers. S_0 and S_1 are approximately concordant on the limbs of isoclinal F_2 folds. F_3 folded $S_{0=1}$ and S_2 into open to close folds that are typically north-vergent and plunge gently to the west. The F_3 axial planar foliation, S_3 , is typically poorly developed. In this photo S_3 is represented by a dissolution cleavage. (B) Looking northwest at a vertical outcrop of calcite-chlorite schist with an isoclinally folded layer of tan-weathering calcite marble. The significance of the relationship between marble and F_1 folds is unclear at this time. Stations (A) uncertain, (B) 10SMS313.

Figure 3.4

Figure 3.5: Lower hemisphere, equal area stereograms show fold hinge and axial plane orientations of coaxial F_2 and F_3 folds occur throughout the study area. (A) F_2 fold hinges (average = $15 \rightarrow 256$), (B) poles to F_2 axial planes, (C) F_3 fold hinges (average = $14 \rightarrow 261$), and (D) poles to F_3 axial planes. F_2 and F_3 folds are coaxial. The orientations of F_2 and F_3 axial planes are variable, with a bias for poles to lie in the northeastern quadrant giving a northwest general strike. The dip of F_2 axial planes, S_2 , varies from moderately to the southwest to gently to the northeast. The dip of F_3 axial planes, S_3 , is variable. See text for discussion.

Figure 3.5

Figure 3.6: (A) Northwest view of a fold interference pattern, F_2 folded by a north-verging open S_3 fold. (B) Westward view of interlayered quartzofeldspathic psammite and albite-garnet-biotite-chlorite schist in hinge of F_2 isoclinal fold. $S_{0=1}$ is folded by F_2 . An S_2 foliation, axial planar to F_2 , is formed by crenulated phyllosilicates. Chlorite in S_2 wraps garnet in the schist, suggesting greenschist facies conditions prevailed during F_2 . Stations (A) 10SS039A, (B) 109SS034.

Figure 3.6

Figure 3.7: Composite photomicrographs of (A, B) garnet in chlorite-muscovite psammitic schist, unit 8, and (c) magnetite in chlorite schist, unit 8. Garnets are pre- to syntenctonic with respect to S_1 . (A) In some cases inclusion trails and garnet breakdown products (S_i) are at a high angle to S_1 , (B) and in other cases S_i is at a low angle or continuous with S_1 . (C) Magnetite porphyroblast in chlorite psammitic schist. The S_1 foliation is weakly deflected around, and overgrown by, the magnetite porphyroblast. This indicates magnetite is syn- to post-tectonic with respect to S_1 .

Abbreviations: Gt – garnet, Chl – chlorite, Ms – muscovite, Mag – magnetite, Qz – quartz.

Samples: (A) 10SMS215C, (B) 10SMS183B, (C) 10SMS021.

Figure 3.7

Figure 3.8: Representative photographs of S_2 crenulated foliations, L_2 mineral lineations, and clast elongation lineations. (A) Looking north at a crenulated foliation in an F_3 hinge zone. S_2 (235/15 [NW]) is crenulated. Crenulation lineations typically plunge gently to the west. Station 10SMS097. (B) Stereogram of L_2 mineral and clast elongation lineations. Unlike crenulation lineations which have a constant plunge, the orientation of L_2 lineation varies from gently to the west to moderately to the south. South-plunging mineral lineations and clast elongations occur only within approximately 3 km of the Nuntaea Creek fault (denoted by red measurements). (C) Stereogram of clast elongation lineations and clast flattening planes. Clast elongation lineations are interpreted as L_2 , showing the same variation in orientation to the L_2 mineral lineation. Red symbols within 3 km of Nuntaea Creek fault. (D) Elongate calcite marble clasts in a quartzite matrix are the most common clast elongation lineation, L_2 . Elongate metagabbro clasts in chlorite schist and dolomite clasts in calcite marble were also encountered. Station: (A) 10SMS100, (D) Boulder on south central ridge.

Figure 3.8

► Clast Elongation (L_2) ● Clast Flattening Plane (S_2)

* red denotes measurements within 5 km of Doghead fault

Figure 3.9: (A) North-northwest view of the SW dipping layers (S_0) of interlayered tan-weathering quartzite and grey-weathering calcite marble of unit 5 approximately 200 m north of the contact of unit 5 with the Doghead assemblage. (B) Northwest view of south-plunging lineation on a dip-slope surface approximating S_0 . Note the L_2 elongation lineation defined by the long axis of tan-weathering quartzite clasts (Q) in the grey-weathering marble matrix (M). The long axis of resistantly weathering quartzite clasts shows a preferred orientation, plunging gently to moderately to the south. Stations: (A) 10SMS366, (B) 10SMS366.

Figure 3.9

Figure 3.10: Looking west at a south verging F_3 fold in a vertical outcrop of interlayered (P) quartzofeldspathic psammite and (S) biotite-muscovite-chlorite schist of unit 10. Psammite layers maintain their thickness through the fold hinge while schist layers are thickened in the hinge zones and thinned in the limbs. Station: 10SMS101.

Figure 3.11: (A) Looking west along L_{312} in subcrop of psammitic schist from unit 10. Pencil is pointing down dip of S_2 and plexiglass board is oriented along S_3 . (B) Lower hemisphere, equal area stereogram of intersection lineations between S_2 and S_3 . L_{312} plunges gently to the northwest. This is consistent with field observations of F_2 and F_3 being coaxial and F_3 fold hinges plunging gently to the west. The intersections of S_2 - S_1 and S_3 - S_1 lie on a great circle with a pole that plunges gently to the southwest. This suggests these intersection lineations have been folded, most likely by D_4 . Station: (A) 10SS045.

Figure 3.10

Figure 3.11

Figure 3.12: Fold interference patterns in the Tincup Lake area. (A) Composite photo, looking west at macroscopic type 3 ($F_2 \times F_3$) fold interference. Isoclinally folded grey and beige calcite marble $\pm$ quartzite of unit 5 and 6 and black weathering graphitic phyllite of unit 7 folded by F_2 have been refolded by a coaxial macroscopic north vergent F_3 fold. Field assistant is circled at bottom right of image, outcrop is approximately 150 m high. (B) Looking southwest at a vertical outcrop of dark-green-weathering chlorite schist of unit 9 with brown-weathering marble forming a type 2 ($F_1 \times F_2$) interference pattern. (C) Looking northeast at an outcrop of white and dark grey weathering marble showing type 1 fold interference pattern between F_1 and F_2 . (D) Looking east at type 2 interference pattern, between F_1 and F_2 , in a subcrop of interlayered dark grey-weathering calcite marble and light grey-weathering calcite marble. Stations: (A) 10SMS244, (B) 10SMS041D, (C) 09SS049, (D) 10SMS372.

Figure 3.12

A

Table3. 1 Studies in western Yukon used to constrain deformation events in the Tincup Lake area.

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Chapter 4 - Kinematic and geometric analysis of an array of strike-slip faults in the Tincup Lake area, southwest Yukon

4.1 Introduction

The North American Cordillera formed, in part, as a result of Early Triassic to Paleocene interactions between the western margin of Laurentia, outboard plates and entrained terranes (Beranek et al. 2010, Colpron et al. 2007, Evenchick et al. 2005, Monger et al. 2005, Monger and Price 2002). During accretion of allochthonous terranes and crustal thickening, rocks of the accreted terranes and the western Laurentian margin underwent shortening, thrust faulting, folding and metamorphism (Monger 1997). Strike-slip faulting accompanied and succeeded this deformation (Gabrielse 1985, Price and Carmichael 1986, Gabrielse et al. 2006). The timing and nature of accretion of peri-Laurentian and allochthonous terranes is still a contentious topic. This debate is complicated by an incomplete understanding of the timing and magnitude of strike-slip faults that transect the accreted terranes and may obscure original boundaries between them (Pope and Sears 1997, Reid et al. 2002, Zelt et al. 2006).

Cordilleran scale strike-slip faults occur from southern British Columbia to western Alaska (e.g. Northern Rocky Mountain -Tintina trench, Figure 4.1). Three major strike-slip fault systems transect the Yukon: the Tintina, Teslin, and Denali faults. Studies of the Tintina, Teslin and Denali faults evaluating offset of geological markers or structures, and geological relationships that bracket timing indicate Cenozoic dextral displacements ranging 130 km to > 400 km (Gabrielse et al. 2006, Zelt et al 2006, Irving et al 1996). The Tincup study area lies approximately 20 km northeast of the Denali fault, an active, continental-scale strike-slip fault extending from western Alaska, through southeastern Yukon, and may connect along strike with the Chatham Strait fault, and possibly the Coast Shear Zone (Gabrielse et al., 2006; Hickman et al, 1978).

Regional-scale strike-slip faults trace through the Kluane River valley to the west and the Tincup Lake valley to the east of the Tincup Lake study area (Murphy 2010). These strike-slip faults are continuations of faults mapped by Murphy (2010) in Stevenson Ridge area, approximately 30 km to the northwest of Tincup Lake, and have strike lengths of 30-50 km. In the Stevenson Ridge area these strike-slip or oblique-slip faults are generally parallel to the Denali fault but change orientation north of the Tincup Lake area, northern Kluane Lake map sheet, to an angle that converges at a low angle with the Denali fault south of the Tincup Lake area (Figure 1.2, Murphy 2010, Murphy et al. 2011).

Rocks of the Tincup Lake study area are cut by strike-slip faults. These map-scale strike-slip faults have strikes similar to regional-scale strike-slip faults (Figure 1.2, 4.2). The relationship between strike-slip faults in the study area, the regional-scale strike-slip or oblique-slip faults and the Denali fault is uncertain.

The purposes of this study are to: describe faults, associated fault rocks and fracture zones in the Tincup Lake study area; document the temporal sequence of faulting; and, determine, if possible, the kinematics and dynamics of faulting by combining fault geometry with kinematic information about the orientation of the fault motion from faults and shear fractures. The estimate of principle stress orientations is used to evaluate whether or not the faults in the Tincup Lake area may be kinematically and dynamically compatible with the regional-scale strike-slip faults, which trace through valleys to the east and west of the Tincup Lake area, and the Denali fault. In the absence of data constraining the absolute age of faulting in the area, such an analysis may give insight into whether or not faults in the study area may belong to a broader Denali fault 'system' or are unrelated.

4.2 Local Geology

The 10 x 12 km Tincup Lake study area is located in southwest Yukon, approximately 40 km northwest of Kluane Lake and 20 km northeast of the Shawkak trench, a geomorphic feature hosting, at this latitude, the Denali fault (Figure 1.2; Map 1, back pocket). Rocks in the Tincup Lake area comprise primarily a poly-folded and faulted assemblage of metasiliciclastic psammite, psammitic schist and schist with interlayered calcite marble (see Chapter 2, Figure 2.1; Map 1) which have been assigned to the Yukon-Tanana terrane (Murphy et al. 2009, Murphy, 2010).

Four phases of ductile deformation have been documented in rocks of the Yukon-Tanana terrane in the Tincup Lake area (see Chapter 3). The youngest phase of folding, D_1 , is preserved as rootless mesoscopic isoclinal folds. D_1 is syntectonic with the formation of the dominant foliation, S_1 , which is predominantly east-west striking and is concordant with compositional layering.

Near the southern limit of the study area, rocks of Yukon-Tanana terrane are structurally overlain by an assemblage of mafic and ultramafic rocks along the Doghead fault (Map 1); these rocks are part of the newly defined Doghead assemblage (Murphy et al. 2011). North-vergent, asymmetric, tight folding of the Yukon-Tanana terrane, in the immediate footwall of the Doghead fault, F_2 , is inferred to be a consequence of shortening of the Yukon-Tanana terrane during obduction of the Doghead assemblage (see Chapter 3). The Yukon-Tanana terrane and the Doghead fault were subsequently folded by co-axial, north-vergent, open to close folds, F_3 .

Two generations of map-scale brittle faults are documented in the study area based on orientation, fault type and cross-cutting relationships. The oldest generation of faults, termed D_5 , consists of east-striking, moderately southward-dipping normal faults. Normal faults are spaced

approximately 4 km apart. The traces of these map-scale normal faults extend up to 10 km along strike and crosscut $F_1 - F_4$ closures, axial planar foliations and fold hinges. Thus, normal faults are interpreted to post-date the penetrative deformation history described in Chapter 2, and must be late Early Cretaceous or younger.

The D_5 normal faults are crosscut by D_6 , steep northwest-southeast striking map-scale strike-slip faults. These strike-slip faults are steeply dipping and strike northwest-southeast (Figure 4.2). A detailed description of these strike-slip faults is presented in section 4.2.1. Strike-slip faults of D_6 are crosscut by east striking, shallowly dipping mesoscopic shear fractures, D_7 .

Planar, unmetamorphosed, fine- to medium-grained diorite and quartz diorite amphibole $\pm$ plagioclase porphyry dikes intruded rocks of the Yukon-Tanana terrane. These dikes are suspected to be syntectonic with D_6 strike-slip faults (Section 4.4.3.4). The unfoliated dikes, ranging from 0.5 – 4 m thick, are similar in composition and texture to Late Cretaceous dikes associated with the Carmacks volcanics (Murphy, personal communication, 2012), although there are no geochemistry or geochronology constraints at present to test this correlation.

4.2.1 D_6 Strike-slip faults

Strike-slip faults extend for up to 10 km along strike (Figure 4.2, Table 4.1). Where the faults cut calcareous units they typically weather recessively and fault walls are well exposed. Fault traces in metasiliclastic units are typically poorly exposed and the presence of faults is inferred from linear topographic depressions with the same strike as mapped faults in calcareous units. Two domains of steep northeast-southwest striking D_6 dextral strike-slip faults, Domain 1 and Domain 2, are defined on the basis of the strike orientation of map-scale faults (Figure 4.2). The domains are separated by fault SS 10, a northwest-southeast striking, steeply dipping strike-slip fault that extends approximately 8 km along strike. In the northern Domain 1, the average

orientation of strike-slip faults is 330/86 [NE] (Figure 4.3a). The average orientation of strike-slip faults in Domain 2 is 128/89 [SW] (Figure 4.3b).

Strike-slip fault spacing ranges from 250 – 2000 m. The tightest fault spacing is in the northeastern part of Domain 1 (Figure 4.2), where faults are spaced 250-500 m. Faults in the southwestern part of Domain 1 are less common and are spaced at 800 – 1000 m. The average fault spacing in Domain 1 is 500 m. In Domain 2 the average fault spacing is 1000 m.

The absolute ages of strike-slip faults in the Tincup Lake area are poorly constrained. Strike-slip faults and diorite dikes may be mutually crosscutting. One dike is cut at a low angle by a strike-slip fault. Another dike was observed discontinuously along strike in the area of station 10SMS176 (586064E, 6852378N) crossing topographic lineaments oriented parallel to strike-slip faults. If this dike is continuous and the lineament does reflect a strike-slip fault, then the dikes can be inferred to be mutually crosscutting with respect to strike-slip faults.

Faults in the Tincup Lake area lack piercing points that could be used to give a confident estimate of offset. In Domain 1 the contact between chlorite schist and calcite marble shows an apparent reverse-sense dip-slip component of displacement of approximately 120 m along fault SS6.3 (Figure 4.2, 4.3). In Domain 2, the thrust fault between the Doghead assemblage and rocks correlated with the Yukon-Tanana terrane has an apparent dextral offset of approximately 600 m along fault SS8 (Map, Figure 4.3). These two observations provide an estimate for the magnitude of displacement along strike-slip faults in the Tincup Lake area.

4.2.2 Fault Rocks and Damage zones

Descriptive terminology and classification of fault rocks are based on the textural classification of Sibson (1977, Table 4.2). Damage zones, composed of relatively densely fractured rock, commonly flank map-scale strike-slip faults. In marble, damage zones extend approximately 25

m on both sides of the fault trace. In metasiliciclastic units, damage zones are commonly less than 10 m-thick on both sides of the fault trace. Within the damage zones, shear fractures are oriented nearly parallel to the fault surface. Shear fractures are spaced at 15-30 cm in marble outcrops and 20-30 cm in metasiliciclastic outcrops. The spacing of these shear fractures increases with distance from the fault, and are uncommon beyond the damage zone.

In both Domain 1 and Domain 2, fault zones are typically 1-2m wide; however, one 15 m wide fault zone (SS10, Figure 4.4) was observed in Domain 2. Fault rocks in both domains consist of angular crush breccia to fine crush breccia (Figure 4.5). Where the faults cut units with more than approximately 30% calcite content, breccia clasts of wall-rock material ranging from 0.5 cm to 10 cm in diameter are set in fine-grained calcite cement. In metasiliciclastic rocks, unconsolidated boulders of wall-rock material fill lineaments along the fault trace and fault rocks are not exposed. It is unclear whether these boulders represent weathered crush breccias, incohesive fault breccia or boulders weathered from the walls of the fault zone.

Veins, shear-fractures and syn- and post-kinematic mineralogy of fault rocks were examined. Calcite coatings rarely occur on the fault surfaces or surfaces of fault breccia fragments. Quartz veins and selvages on metasiliciclastic angular crush breccia fragments in fault zones were not observed. Relative to the number of faults and shear fractures, slickenfibres are rare. Where slickenfibres are present, their mineralogy reflects that of the wall rock (i.e. there are calcite slickenfibres in marble units, chlorite slickenfibres in chlorite schist units).

4.2.3 Shear Fractures

Shear fractures commonly occur in the rocks adjacent to map-scale steep strike slip faults. There are two sets of fractures. The primary shear-fracture set, is approximately parallel to the adjacent fault. Within approximately 10 m of faults, the primary fracture set is tightly spaced

(15-30cm), and is more widely spaced (1-5 m) with increasing distance from the fault. The secondary fracture set is also steeply dipping but strikes at angles of 40°-80° with respect to the primary fracture set. The primary and secondary fracture sets were not observed crosscutting cataclastic rocks.

A tertiary fracture set, D₇, crosscuts fault rocks and both the primary and secondary fracture sets. The tertiary fracture set is typically E-W striking with dips that vary from shallowly to the south, to flat lying to shallowly to the north.

4.2.4 Stepped surfaces and slickenfibres

The mineralogy of slickenfibres reflects that of the wall rock (i.e. calcite slickenfibres occur in marble, and chlorite slickenfibres in chlorite schist). Stepped surfaces and slickenfibres are typically cm-scale structures. Stepped surfaces indicate both dip-slip and strike-slip movement along fault and shear fracture surfaces (Figure 4.6).

Stepped surfaces are used to interpret shear sense of a fault or shear fracture. Slickenfibres are interpreted to have grown during movement of the fault. The interpretation of shear sense is achieved by considering the orientation of slickenfibres and the smooth direction of the stepped surface. The smooth direction of the stepped surface is determined by moving your hand in the direction of the slickenfibre lineation. The smooth direction, the direction the eroded wall of the fault moved, feels smooth to the touch. Using this criteria, northwest-southeast striking, steeply dipping faults and shear fractures with a shallowly plunging lineation and stepped surfaces are interpreted to have a dextral sense of displacement. The interpretation of shallowly plunging slickenfibres and the stepped surfaces on southwest-northeast striking, strike-slip shear fractures yields a sinistral sense of motion.

Measurements were collected from 16 faults and shear fractures with stepped surfaces and slickenfibres (Figure 4. 7a). The faults and shear fractures are similar to the orientation of map-scale faults in the Tincup Lake map area (Figure 4.3). Faults and shear fractures correlated with D_5 normal faults are east-west striking and dip gently-moderately to the south (Figure 4.7b). Stepped surfaces associated with D_5 faults and shear fractures typically give a down dip sense of motion. Faults and shear fractures associated with strike-slip faults strike northwest-southeast to northeast-southwest and are steeply dipping (Figure 4.7c), and slickenfibres on stepped surfaces show one of three preferred orientations. The most common slickenfibre lineation plunges gently to the northwest on steep northwest-southeast striking faults and fractures, and stepped surfaces are consistent with a dextral sense of motion. Northeast-southwest striking faults and fractures also occur. These northeast-southwest striking faults have a slickenfibre lineation that plunges gently to the southwest and stepped surfaces consistent with a sinistral sense of motion. In addition to evidence for strike slip faults, there are faults, oriented similarly to the strike slip faults, yet they have a slickenfibre lineation that plunges steeply to the east and stepped surfaces that are consistent with a normal sense of motion.

4.3 Kinematic and Geometric Analyses

Faults and shear fractures are treated differently depending on whether or not there is field evidence for the transport direction and sense of displacement, in the form of lineations and stepped surfaces. Thus, the analysis of faults and shear fractures in this study comprises two parts. The first part is a kinematic analysis of faults and fractures, in the cases where the sense of displacement could be determined. The purpose is to estimate the orientation of σ_1 during D_5 and D_6 faulting, respectively (Section 4.3). Part 2 is a geometric analysis of faults and fractures that have similar orientations to the aforementioned structures; however, information about

sense of displacement is lacking. The structures are tested for geometric compatibility with the estimated σ_1 orientations from part 1 (Section 4.4). For the purposes of this study the maximum principal stress is denoted by σ_1 . The minimum principal stress is denoted by σ_3 .

A number of assumptions are made in order to apply results of the geometric analysis of faults and shear fractures to the cases where there are constraints on the direction of fault motion.

Firstly, faults and shear fractures without kinematic data are correlated with faults and shear fractures of the same orientation that do have robust kinematic indicators. Secondly assumption is Coulomb Failure Criterion can be used to describe the failure of these rocks. Thirdly, this analysis assumes the faults have not undergone any significant reactivation and the lineations and stepped surfaces can be used to interpret the direction and sense of motion of the fault at the time of formation. The method of each analysis is described below.

4.3.1 Kinematic and Dynamic Analysis Method - Estimating the orientation of σ_1

For each fault or shear fracture with known direction of displacement, the orientation of σ_1 , the maximum principal stress, was estimated using the method of Angelier (1979). According to the Mohr-Coulomb failure criterion, there are two possible fault or shear-fracture orientations that can form for a given orientation of principle stresses (Figure 4.8a), albeit in homogeneous material. The conjugate faults or shear fractures are oriented symmetrically about σ_1 and σ_3 (Figure 4.8b). Since direction of displacement on each fault is controlled by σ_1 and σ_3 , the orientation of slickenfibres must lie on the plane defined by σ_1 and σ_3 (Figure 4.8c). The principle stresses are orthogonal, requiring σ_2 to be oriented within the fault plane (Figure 4.8c). The orientations of σ_1 and σ_3 are estimated by assuming the angle of internal friction. This study follows Angelier (1979, 1984) in assuming an angle of internal friction of 30° . In this study, the

σ_1 estimates precisely for a more pr

orientation of one of the shear planes and the direction of displacement are known. By plotting these data on a stereonet, a geometric estimate of σ_1 can be made.

First, the orientation of each fault or fracture and the orientation of its corresponding slickenfibres lineation, representing the displacement direction, are plotted on the lower hemisphere of an equal-area stereonet (Figure 4.9a). Next an auxiliary plane is plotted, containing both the pole to the fault plane and the slickenfibres lineation. Sigma 1 is then located within the auxiliary plane by finding the point that lies at an angle of 30° from the lineation, which plots at the intersection between the fault and auxiliary planes. It should be noted that there are two possible orientations of σ_1 , $+30^\circ$ and -30° from the lineation, or intersection between the fault and auxiliary planes, and that they have opposite sense of displacement. For a given fault, only one orientation of σ_1 is compatible with the observed sense of displacement. For D_5 and D_6 , the individual kinematically compatible estimated orientations σ_1 from faults of known sense of displacement are plotted on respective stereonets for each generation of faulting (Figure 4.9b, c). The compilation for each fault system is evaluated independently to determine whether there is a systematic pattern. If a systematic pattern is present, an average value for the estimated orientation of σ_1 is taken to represent a general estimate for the orientation of σ_1 for the generation of faulting and fault system under consideration.

4.3.2 Kinematic and Dynamic Analysis Results of D_5 and D_6 structures

D_5 , east-west striking normal faults and shear fractures, dip predominantly to the south. Estimates of σ_1 for D_5 faults and shear fractures vary from moderately plunging to the northeast to steeply plunging to the south. Using the method outlined above (Section 4.3), the orientation of σ_1 was estimated to be $77 \rightarrow 304$ for D_5 (Figure 4.9b). The small dataset ($N = 4$) and scatter of σ_1 estimates precludes a more precise estimate of σ_1 for this generation of deformation. The

scatter of estimates for σ_1 may reflect the rotation of fault blocks during subsequent deformation, the influence of local anisotropies on the orientation of the faults, or different generations of faults forming in different stress regimes.

D₆ strike-slip faults and shear fractures have strikes that range between northwest-southeast and northeast-southwest. Kinematic data for D₆ is limited to Domain 1 (Figure 4.7c). Estimates of the orientation of σ_1 for D₆ can be broken in to two groups, reflecting whether the sense of movement is strike-slip or dip-slip (Figure 4.9c). For D₆ faults with a strike-slip sense of motion; individual estimates of σ_1 either plunge gently to the north or gently to the south, reflecting variation through the horizontal. For D₆ strike-slip faults the average of σ_1 is estimated to be 06→186. This estimate for the orientation of σ_1 is consistent for faults and fractures with either sinistral or dextral sense of movement. The angle between the average plane for surfaces with dextral movement and the lone surface with a sinistral sense of movement is 62°. This orientation of σ_1 , 06→186, is the estimated σ_1 that will be used in the geometric analysis of northwest-southeast and northeast-southwest striking strike slip faults.

For D₆ faults with a dip-slip sense of motion, individual estimates of σ_1 are typically steeply plunging to the southwest. For D₆ faults with a dip-slip sense of motion the average of σ_1 is estimated to be 70→238. See section 4.3.2 for interpretation of different orientations of σ_1 for D₆ strike-slip and dip-slip fault.

4.4 Geometric Analysis

4.4.1 Geometric Analysis Method

In the cases for map-scale faults and mesoscopic shear-fracture sets where there are no kinematic data available, the orientation of structures is used to test for geometric compatibility with the kinematic analysis (Section 4.3) of similarly oriented structures with known

displacement sense. Compatibility of map-scale faults and mesoscopic shear fractures were tested by plotting the faults and shear fractures on a lower hemisphere equal area stereonet along with the estimated orientation of σ_1 , $06 \rightarrow 186$, obtained from the kinematic analysis of D_6 (Figure 4.10). If the estimated σ_1 orientation lies within 20° - 40° of a northwest-southeast or northeast-southwest striking fault or fracture then the structure is considered to be kinematically compatible with the estimated stress field. The range of angles ($30^\circ \pm 10^\circ$) between the fault or fracture and σ_1 is intended to allow for block rotation and local anisotropies in the rock which may have influenced fault or shear-fracture orientation. In this analysis, a northwest-southeast striking strike-slip fault is compatible with a dextral sense of motion and southwest-northeast striking strike-slip faults are consistent with a sinistral sense of motion.

4.4.2 Classification of shear fractures with uncertain sense of displacement

Each shear fracture is considered independently and categorized as a steep dextral or sinistral strike-slip fracture, a normal fracture or as a fracture with uncertain sense of displacement. Shear fractures categorized as dextral have a strike of northwest or southeast and the angle between the plane and the estimated orientation of $D_6 \sigma_1$ is 20 - 40° . Shear fractures categorized as sinistral have a strike of northeast or southwest and the angle between the plane and the estimated orientation of σ_1 is 20 - 40° . Shear fractures categorized as normal strike north or south, dip steeply, and the angle between the shear fracture and the estimated orientation of σ_1 must be $<10^\circ$. Shear fractures categorized as uncertain do not meet the aforementioned criteria.

4.4.3 Geometric Analysis Results

4.4.3.1 Map-Scale Normal Faults

Map-scale normal faults are correlated with D_5 structures of the Kinematic and Dynamic Analysis (Map, Section 4.3). These map-scale normal faults are predominantly moderately south dipping.

Like the outcrop-scale normal fractures, approximately 25% of these faults dip gently to moderately to the north. These north-dipping faults have relatively short surface traces compared to map-scale strike-slip faults.

4.4.3.2 Map-Scale D_6 Strike-Slip Faults

Domain 1 contains 25 faults and fault segments (Figure 4.2). Fault segments are defined as a part of the fault that differs by greater than 10° in orientation with respect to the main fault or an adjacent part of the fault. The orientation of fault segments varies in dip from moderately southwest, to vertical or moderately northeast. However, the fault segments are interpreted to belong to a single fault population, with an average orientation of $330/86$ [NE] (Figure 4.3a). The angle between this average fault plane and the estimated orientation of $D_6 \sigma_1$ is 32° . In the kinematic model (Section 4.3), this orientation is compatible with an interpretation of predominantly dextral sense movement on the faults.

Domain 2 strike-slip faults consist of 3 faults that are steeply southwest dipping or steeply northeast dipping. These faults are interpreted as a single population, with an average orientation of $128/89$ [SW] (Figure 4.3b). The angle between the $D_6 \sigma_1$ and the average fault for Domain 2 is approximately 54° . This suggests strike-slip faults in Domain 2 may not have formed under the same stress regime as those in Domain 1.

4.4.3.3 Shear Fracture Sets

Results of the geometric analysis of shear fractures are presented in Table 4.3. In Domain 1, 91 shear fractures were measured at 33 outcrops. Of these, 55 shear fractures are in an orientation that is kinematically compatible with the dynamic analysis and estimates of the orientation of $D_6 \sigma_1$. Similarly, in Domain 2, of the 79 shear fracture surfaces from 26 outcrops, 50 are in an

orientation that is kinematically compatible with the kinematic and dynamic estimate of the orientation σ_1 for D_6 .

4.4.3.4 Dikes

The average orientation of the 18 dikes encountered in the Tincup Lake area is 146/73 [SW] (Figure 4.11a). The estimated D_6 σ_1 orientation (Section 4.3.2) is oriented at an angle of approximately 35° with respect to the dyke orientation (Figure 4.12b). This is close to the angle between σ_1 and the average orientation of D_6 dextral strike-slip faults and shear fractures. This suggests dikes may have propagated along fractures compatible with D_6 faults or shear fracture orientations, rather than intruding into purely extensional features.

4.5 Interpretation and Discussion

4.5.1 Conditions of Faulting

In the Tincup Lake area, fault rocks associated with moderately southward-dipping D_5 normal faults are most commonly crush breccias. Fault rocks in D_6 strike-slip faults are angular to fine crush breccias. This suggests that these faults, with coherent fault rocks, formed in the brittle regime between about 4km and 13 km depth (Sibson, 1977).

Crush breccias associated with normal faults are commonly stained an ankeritic orange colour, suggesting fluids were present in these faults during or following faulting. Conversely strike-slip faults lack quartz veins and selvages on breccia fragments. This suggests strike slip faults were dry during and following faulting.

4.5.2 Fracture Sets

Primary and secondary fractures are spatially related to map-scale D_6 strike-slip faults. Primary and secondary fractures do not cross strike-slip faults. For these reasons, the primary and

secondary fracture sets are interpreted as syntectonic with respect to the formation of D_6 strike-slip faults. The tertiary fracture set, D_7 , occurs throughout the area without any localization or near any generation of faults and they crosscut tectonic breccias associated with strike-slip faults. For these reasons, the tertiary fracture sets are interpreted to be younger than D_6 strike-slip faults and are therefore the youngest generation of deformation, D_7 . No map-scale features can be correlated with D_7 on the basis of crosscutting relationships.

4.5.3 Discussion of kinematic analysis results

Outcrop-scale shear fractures have the same crosscutting relationships as map-scale features; that is south-dipping normal fractures are offset by steep strike-slip fractures (Figure 4.2). East-west striking faults and fractures with down-dip sense of motion are correlated with the older D_5 normal faults on the basis of orientation and sense of movement (Figure 4.3b). Northwest-southeast to northeast-southwest striking faults and shear fractures are correlated with the D_6 strike-slip faults that offset the D_5 south-dipping normal faults (Figure 4.3a). D_6 faults and shear fractures with sense of displacement information can be broken down into three, kinematically compatible groups. Northwest-southeast striking dextral strike slip faults and northeast-southwest striking sinistral strike-slip faults are compatible with the gently to the south-plunging estimated orientation of σ_1 ($06 \rightarrow 186$, Figure 4.9c). Sigma 1 lies within 1 degree of the plane bisecting the sinistral and dextral strike-slip faults and shear fractures, suggesting these faults and shear fractures formed as a conjugate pair. North-south striking normal faults are also consistent with the estimated orientation of σ_1 determined for D_6 dextral and sinistral strike-slip faults. The steep normal faults are interpreted as accommodation structures in a regional wrench system. For this reason, the steep orientation of D_6 σ_1 associated with the steep normal shear fractures reflects local structural variations and not a change in the overall stress regime.

The presence of the normal faults does not necessarily indicate that σ_3 was tensile at the time of strike-slip fault formation. Figure 4.12 is a graphical representation of this interpretation.

4.5.4 Discussion of geometric analysis results

4.5.4.1 Map-Scale D_5 Normal Faults

The geometric analysis of map-scale normal faults (N=12) shows a predominance of south-dipping faults with approximately 25% of the faults dipping gently to moderately to the north. According to Mohr-Coulomb theory of faulting, the orientation of these faults and their field classification as normal faults suggests a near vertical orientation of σ_1 . This is consistent with results of the kinematic analysis of normal faults which estimates σ_1 to be oriented $77 \rightarrow 304$.

4.5.4.2 Map-Scale Strike-Slip Faults

The orientations of strike-slip faults in Domain 1 (330/87 [NE]; Section 4.2.4.2, Figure 4.2, 4.3, 4.13a) more closely reflect the dominant orientation of dextral faults and shear fractures used in the kinematic analysis (153/79 [SW]) than do the orientation of map-scale faults in Domain 2 (128/89 [SW]). The paucity of kinematic data from Domain 2, N=3, makes it difficult to determine whether differences in map-scale strike-slip fault orientation is a bias in the data or if it does truly reflect a change in the orientation of the stress field resulting in the difference in average fault orientation between Domain 1 and Domain 2. No further correlations can be made on the basis of geometric analyses of map-scale strike-slip faults and kinematic analyses.

4.5.4.3 D_6 Shear-Fracture Sets

Unlike D_6 map-scale strike-slip faults, the orientation of primary and secondary shear fractures does not show any appreciable difference in orientation between Domain 1 and Domain 2 (Figure 4.13b, c). The orientation of primary and secondary shear fractures shows a stronger

geometric correlation with map-scale faults in Domain 1 relative to those in Domain 2. This suggests the shear fractures in Domain 1 and Domain 2 formed with the same orientation of σ_1 . Alternatively, this could indicate the measured D6 faults in Domain 2 are not representative of the shear fractures and could reflect reactivation of the faults.

4.5.4 Uncertain shear fractures

The group of shear fractures that were classified as “uncertain” (Figure 4.13) is dominated by east-west striking fractures with a range of dips. While these fractures are likely unrelated to map scale D₆ strike-slip faults, there is a possible correlation with the early generation of map-scale normal faulting, D₅. All the east-west striking shear fractures lie between 15°-75° of the estimated orientation of σ_1 for D₅. These orientations are within the range of shear fractures that would form due to anisotropies (i.e. cleavage or foliation plane) within the rock (Donath 1961). The orientation of these east-west striking shear fractures is consistent with the formation of shear fractures influenced by the orientation of planes of weakness within the rock volume exposed to a near-vertical σ_1 . While the orientation of these fractures is kinematically compatible with the estimate of σ_1 for D₅, a definitive correlation cannot be made with the information at hand.

4.5.5 Compatibility of dikes with σ_1

The average orientation of diorite and quartz-diorite dikes in the Tincup Lake area is not geometrically compatible with intrusion along extensional accommodation structures that would be expected to form under the interpreted stress regimes for D₅ or D₆. The orientation of the dikes, approximately 35°, with respect to the σ_1 estimated for D₆ suggests the dikes likely intruded approximately parallel to pre-existing planes of weakness rather than under an extensional regime (e.g. σ_3 orthogonal to dikes). The apparent mutually crosscutting relationship

of faults and dikes suggests syntectonic intrusion is likely the case. However, there is insufficient evidence to make a definitive conclusion.

4.5.6 Correlation with regional faults

4.5.6.1 Normal Faults

From this study, there are insufficient data to correlate D₅ normal faults in the Tincup Lake area with Upper Cretaceous normal faults in the Stevenson Ridge map area (Murphy 2010). The only constraint that can be placed on the age of south-dipping normal faults in the Tincup Lake area is that they are younger than the last phase of folding, interpreted to be pre- 103 Ma (Murphy 2010), and they are older than D₆ strike-slip faults.

4.5.6.2 Strike-Slip Faults

Map-scale D₆ strike-slip faults in the Tincup Lake area have not been dated but are suspected to be younger than late Early Cretaceous in age based on crosscutting relationships with map scale D₅ normal faults. The relative timing of map-scale strike-slip faults with respect to regional-scale strike-slip faults mapped by Murphy (2010) is uncertain. The strike-slip faults of Domain 1 and regional-scale strike-slip faults both strike north-northwest—south-southeast. The significance of the similarity, $\pm 5^\circ$, in strike is uncertain (Figure 4.14). Similar strike orientation could indicate faults in the study area are coeval with regional-scale strike-slip faults, representing a damage zone of en echelon faults between regional scale strike-slip faults. Alternatively, Domain 1 faults may have had a different orientation initially and been rotated into the current orientation.

Domain 2 faults strike northwest-southeast, approximately parallel with the southern contact of the Ruby Range Batholith with the Kluane schist (Figure 4.14). This strike is also approximately

parallel with a regional-scale fault northeast of the study area, shown as a dashed line on Figure 4.14. From this study it is not possible to determine the significance of these similarities.

Further investigation of regional strike-slip faults in southwestern Yukon is required in order to determine the relationship between map-scale and regional strike slip faults shown on Figure 4.14. This will prove challenging since many of the regional scale strike-slip faults occupy poorly exposed topographic lows and river valleys. A potentially fruitful strategy would be to map and date quartz-diorite dikes that are mutually crosscutting with respect to the faults. The dikes are potentially part of the Upper Cretaceous Donjek volcanic rocks (Murphy, personal communication, 2012).

4.6 Conclusion

Two generations of late Early Cretaceous or younger map-scale brittle faults, D_5 and D_6 , have been identified in the Tincup Lake area. Both generations of faulting are younger than emplacement of the Doghead assemblage over rocks correlated with the Yukon-Tanana terrane. Therefore, faulting is younger than the post-Late Triassic collision of the Chulitna (?) terrane with the Yukon-Tanana terrane.

Both D_5 and D_6 faults can be traced for up to 10 km along strike across the study area (Figure 4.2). The older generation of faults, D_5 , is characterized by east-west-striking, gently-moderately-to-the-south dipping normal faults with a spacing of approximately 4 km. The magnitude of displacement on D_5 faults is uncertain. The damage zones associated with these normal faults is typically on the order of meters. These normal faults contain a cohesive crush breccia, suggesting the faults formed between 4 km and 13 km depth. Ankeritic staining on fault rocks suggests fluids were present in these faults during or following faulting.

The younger generation of map-scale faults, D_6 , is dominated by northwest-southeast striking, steeply dipping strike-slip faults (Figure 4.7a). The spacing of D_6 faults ranges from 250 to 1000 m. Damage zones consisting of shear fractures flank both sides of the fault for up to 10 m. Offset along D_6 faults is poorly constrained. However, from the offset of map unit contacts along two faults (Section 4.2.1) vertical offset is expected to be in the order of tens of meters while horizontal offset is expected to be in the order of hundreds of meters. Like the normal faults, these strike-slip faults are interpreted to have formed at 4-13 km depth. Strike-slip faults lack evidence to suggest fluids were present.

Porphyritic (amphibole-plagioclase) quartz diorite and diorite dikes crosscut folds and foliations and evidence suggests these dikes may be mutually crosscutting with respect to D_6 faults. The average orientation of the 18 dikes encountered is similar to the orientation of dextral strike-slip faults of D_6 and is oriented at an angle of approximately 35° to the estimated σ_1 for D_6 . These dikes are interpreted to have propagated along planes of weakness congruent with D_6 faults.

Stepped surfaces and slickenfibres occur on both D_5 and D_6 faults (Figure 4.6). Analysis of this kinematic data was used to estimate the orientation of σ_1 for both D_5 and D_6 faults following the methods of Angelier (1979). For D_5 , $N=4$, σ_1 is estimated to be oriented $77 \rightarrow 304$ (Figure 4.9b). For D_6 , σ_1 is estimated to be oriented $06 \rightarrow 186$ for σ_1 (Figure 4.9c).

Faults and shear fractures which lack kinematic indicators were tested to see if they were geometrically compatible with these estimated orientations of σ_1 for D_5 and D_6 . Map-scale strike-slip faults are kinematically compatible with the estimated orientation of σ_1 (Figure 4.13a). Of the 170 shear fractures measured, 105 were in an orientation that is kinematically compatible with the estimated orientation of σ_1 for D_6 (Figure 4.13b, c). The remaining 65 shear fractures of

uncertain relation may be related to D₇, a poorly understood generation of fracturing which crosscuts D₆ breccias.

Strike-slip faulting in the Tincup Lake area is likely related to regional-scale faulting of the Yukon-Tanana terrane, however, there are insufficient data to test this hypothesis. The lack of fault ages and uncertainty of the relationship to regional-scale strike-slip faults precludes the development of a unique model relating map-scale faults to regional-scale faults and to the Denali fault.

Figure 4.1: Simplified map of Alaska, Yukon and British Columbia showing the main orogen-parallel strike-slip faults in the northern Cordillera. The red dot in southwestern Yukon, east of the Denali fault, denotes the study area. After Gabrielse et al. (2006) and Redfield and Fitzgerald (1993).

Figure 4.1

Figure 4.2: Strike-slip faults and representative shear fractures of the Tincup Lake area. The area is divided into two domains, based on fault orientation. North of fault SS10 is domain 1 where strike-slip faults characteristically strike 150° - 330° . South of fault SS10 is domain 2 where strike-slip faults characteristically strike 130° - 310° .

Figure 4.2

Faults and fractures of the Tincup Lake area

Reverse fault.....
ornamentation on hanging wall

Normal fault.....
ornamentation on hanging wall

Strike-slip fault.....
with sense of motion

Fractures.....
primary, secondary

Fault name..... SS1 U1

strike-slip, uncertain motion

Extent of mapping.....

Figure 4.3: Stereogram of map-scale strike-slip faults from the study area plotted as great circles. (A) Map-scale, steep, strike-slip faults in Domain 1 are considered as a single population despite the variation in strike from west-northwest – east-southeast to north-northwest – south-southeast. The average orientation of these faults is 330/86 [NE]. (B) Strike-slip faults in Domain 2 show a similar variation in dip but typically have a strike 20° closer to E-W than Domain 1. The average orientation of Domain 2 strike-slip faults is 128/89 [SW].

Figure 4.3

Figure 4.4: Southwest view of a 1 km long ridge (588750 E ,6849250 N) of chlorite schist (Unit 8) overlain by rusty brown calcite marble (Unit 9) which have been dissected by northwest-southeast striking, steep strike-slip faults. Stepped surfaces and shallowly plunging slickenfibres on fault SS6.4 are interpreted to give a dextral sense of motion. Offset of the contact along faults indicate an apparent component of reverse movement as well.

Figure 4.4

Figure 4.5: Photographs of fault outcrops. (A) Northwest view along the strike of fault SS10 (588300 E, 6843500 N). This fault separates a package of chlorite schist (S) to the northeast from marble (M) to the southwest (Map). The trace of this dextral strike-slip fault extends 2 km into the middle distance of the photograph as delineated by the dotted line. (B) Northwest view of a 1 m-thick, steeply dipping strike-slip fault in calcite marble, characterized by 0.1-0.5 m angular clasts of marble crush breccia in a calcite fault breccia. (C) Crush breccia with quartzite fragments in a calcite matrix, typical of fault zones in unit 5 (Chapter 2, Map 1).

Figure 4.5

Figure 4.6: Steep or shallow slickenfibres on faults and fractures. (A) Eastward view of a vertical shear fracture in calcite marble where stepped surfaces and shallowly plunging calcite slickenfibres yield a sinistral sense of motion. (B) Westward view of a vertical shear fracture in calcite marble. Stepped surfaces and steeply plunging calcite slickenfibres give a normal sense of motion.

Figure 4.6

Figure 4.7: Stereograms of data used for kinematic analysis of faults and fractures. Each colour represents the orientation of fault or shear fracture, plotted as a great circle, and the interpreted transport direction, plotted as a polygon. These data are used to estimate the orientation of σ_1 using the methods of Angelier (1979; see section 4.2.1 for explanation). (A) Stereogram of all data from faults or fractures with known kinematic sense; each pair is plotted in a different colour. (B) Stereogram of data from east-west striking normal shear fractures. (C) Stereogram of data from northwest- or southwest-striking faults or shear fractures: 7 are dextral or sinistral strike-slip faults or fractures with shallow lineations, and 3 are normal faults or fractures with steep lineation.

Figure 4.7

Figure 4.8: General example of relationships between Mohr-Coulomb failure criteria and principle stresses. (A) Mohr-Coulomb failure criteria predicts two possible orientations of faults forming in a stress regime. The faults could form at the angle, 2θ , when the Mohr's circle is at a critical state of stress (i.e. the radius of the circle, $\sigma_1 - \sigma_3$, is tangential to the failure envelope). In the "real world", θ is the angle between n , the normal to a fault or fracture plane under consideration, and σ_1 . The angle between n and σ_3 is $90 - \theta$. (B) As illustrated in a two-dimensional sketch, the maximum principal stress, σ_1 , bisects the acute angle of conjugate faults or shear fractures. The obtuse angle is bisected by σ_3 . (C) Block diagram of a fault surface with lineation indicating transport direction and stepped surface indicating sense of motion. The auxiliary plane, which contains both σ_1 and σ_3 , contains the displacement vector. The intermediate principal stress, σ_2 , is parallel to the fault plane and perpendicular to the displacement vector.

Figure 4.8

Figure 4.9: (A) Stereogram showing a general example of the plot used to estimate the orientation of σ_1 for faults or shear fractures in the case where the orientation and shear sense are known. A great circle, representing the fault or shear fracture, is plotted on a lower hemisphere, equal area stereonet. The auxiliary plane contains both the pole to the fault or shear fracture and the lineation taken to represent the transport direction. The maximum compressive stress, σ_1 , is plotted within the auxiliary plane, at an angle of 30° from the intersection between the auxiliary plane and the fault or shear fracture under consideration. There are two possible orientations for σ_1 , σ_{1a} and σ_{1b} . In this example σ_{1a} would result in a sinistral sense of motion and σ_{1b} would result in dextral sense of motion. The orientation of σ_1 that is consistent with the observed sense of motion is selected. (B) Colour coded interpretation of σ_1 orientation of pairs of faults or fractures and lineations presented in (Figure 4.7b). (C) Estimates of σ_1 colour coded to data presented in Figure 4.7c. For northwest and southwest striking faults and shear fractures the estimated orientation of σ_1 is estimated to plunge gently to the north or south. For north striking shear fractures, the estimated orientation of σ_1 (circled) is interpreted to plunge moderately to the southwest, steeply to the west and moderately to the northwest.

Figure 4.9

A: Fault Plane 345/75 [NE]

● pole to fault plane

B: Auxiliary Plane 190/17 [W]

L: Slickenfibres Lineation 07→347

s_{1a} : 14→317

s_{1b} : 02→196

Figure 4.10: Stereogram showing general case for the geometric analysis of shear fractures and faults in the cases where there are no direct constraints on the orientation or shear sense of the displacement vector. Great circles and poles to planes represent theoretical shear fractures of different orientations. Red represents a shear fracture compatible with σ_1 and with a dextral sense of motion. Green represents a shear fracture compatible with σ_1 , with a sinistral sense of motion. Blue represents a shear fracture compatible with σ_1 and with a normal sense of motion. The black point represents the orientation of σ_1 (06→186), estimated by the kinematic analysis (Section 4.2.2, Figure 4.6). Light grey areas are small circles at $30^\circ \pm 10^\circ$ constructed about a cone axis centered on σ_1 . If northwest-southeast or southwest northeast, steeply dipping shear fractures are tangential to a small circle within the shaded area, they are taken to be geometrically compatible with σ_1 , that is the fault could form with the estimated orientation of σ_1 . Dark grey areas are small circles at $60^\circ \pm 10^\circ$, where σ_1 is the cone axis. If the pole to a northwest-southeast or northeast southwest, steeply dipping fault or shear fracture lies in this shaded area, the fault or shear fracture may be interpreted as geometrically compatible with σ_1 . North-south striking, steeply dipping faults or shear fractures are considered to be geometrically compatible with σ_1 if the plane is within 5° of σ_1 or if the pole to the plane is shallowly plunging and lies within 5° of the plane that is orthogonal to σ_1 . These north-south striking normal faults are interpreted as analogous to longitudinal fractures and do not necessarily reflect extension.

Figure 4.10

Plane	Pole	
—	●	NW-SE striking fault or shear fracture geometrically compatible with estimated σ_1 , dextral sense of movement expected
—	●	NE-SW striking fault or shear fracture geometrically compatible with estimated σ_1 , sinistral sense of movement expected
—	●	N-S striking fault or shear fracture geometrically compatible with estimated σ_1 , normal sense of movement expected
	●	estimated orientation of σ_1 from kinematic analysis of D6 (06→186)
		Zone of possible shear fracture planes geometrically compatible with estimated σ_1
		Zone of possible poles to shear fracture planes geometrically compatible with estimated σ_1

Figure 4.11: Analysis of dikes. (A) Stereogram showing poles to 16 quartz diorite and diorite dikes plotted on a lower hemisphere, equal area stereogram. The average orientation of these dikes, shown as a great circle, is 149/73 [SW]. (B) Stereogram showing the average dike, plotted as a great circle, and the estimated orientation of σ_1 (06→186). The shaded area on the stereogram indicates $30^\circ \pm 5^\circ$ from the average dike. This shaded area contains σ_1 , suggesting a geometric relationship between the dikes and σ_1 the same as that between D_6 faults and fracture planes and σ_1 (see Section 4.2.4.2). This suggests the dikes intruded along fault surfaces rather than in an extensional regime.

Figure 4.11

Figure 4.12: Block diagram showing the three-dimensional relationships of faults that formed in a stress regime with σ_1 oriented $06 \rightarrow 186$. Northwest-southeast striking dextral faults are the dominant fault set whereas southwest-northeast striking sinistral faults are a secondary conjugate set. North-south striking normal faults are interpreted as analogous to longitudinal fractures and do not necessarily indicate extension.

Figure 4.12

Figure 4.13: (A) Map scale faults, (B) Domain 1 shear fractures, and (C) Domain 2 shear fractures for which there are no direct data on displacement vector or shear sense are plotted on stereograms and rose diagrams. Faults and shear fractures are colour coded to reflect geometric compatibility with estimated orientation of σ_1 . Red – compatible with σ_1 estimate for D_6 , sense of motion most likely dextral. Blue – compatible with σ_1 estimate for D_6 , sense of motion most likely normal. Green – compatible with σ_1 estimate for D_6 , sinistral sense of motion most likely sinistral. Grey – fault or shear fracture orientation incompatible with σ_1 estimate for D_6 , possibly related to D_5 or D_7 .

Figure 4.13

Figure 4.14: Strike-slip faulting in the Tincup Lake area and relationship to regional-scale faulting. Faults in the study area (grey area) are congruent with those at the regional scale. It is unclear whether this is the result of a genetic relationship between map-scale and regional scale-faults or the rotation of map-scale faults into an orientation similar to regional scale faults.

Figure 4.14

Fault	Observed/Inferred	Orientation	Strike Length (m)	Width Estimate (m)
SS1	Observed	150/77	5500	5
SS2	Observed	155/50	7500	10
SS3.1	Inferred	323/83	6500	2
SS3.2	Observed	134/56	4500	5
SS4.1	Inferred	323/76	2500	5
SS4.2	Inferred	156/66	3500	5
SS4.3A	Inferred	170/64	1000	2
SS4.3B	Inferred	164/70	3500	5
SS4.3C	Inferred	329/63	1000	2
SS5.1	Observed	344/71	6000	3
SS5.2	Inferred	334/59	1500	2
SS5.3	Observed	304/89	3000	3
SS5.4	Inferred	122/48	2000	3
SS6.2N	Observed	309/86	3500	5
SS6.2S	Inferred	336/83	2500	3
SS6.3	Observed	337/81	2500	3
SS6.4	Observed	327/74	5000	5
SS6.5	Observed	140/62	2500	2
SS6.6	Observed	146/78	1000	3
SS7	Observed	324/62	5000	4
SS8	Observed	310/87	9500	6
SS9	Inferred	132/64	9500	5
SS10	Observed	302/71	10000	15
SS11	Observed	333/84	2400	5
U1	Inferred	357/58	850	5
U2	Inferred	342/60	2500	5
U3	Inferred	338/66	1500	5
U4	Inferred	314/67	1000	5
U5	Observed	152/76	2000	5

Table 4. 1: Summary of information about map scale faults. Orientation recorded using right hand rule. Fault widths are estimated width of recessively weathering fault breccia at surface.

		Random Fabric	Foliated
Incohesive	Fault Breccia (visible fragments >30% of rock mass)	—	
	Fault Gouge (visible fragments < 30% of rock mass)	—	
Cohesive	Glass - devitrified glass	Pseudotachylyte	—
	Tectonic reduction in grain size dominates grain growth by recrystallization and neomineralization	Crush Breccia (fragments > 0.5cm) Fine Crush Breccia (fragments < 0.5 cm) Crush Microbreccia (fragments < 0.1cm)	0-10
	Cataclastic series	Pseudocataclasite	10-50
		Cataclasite	50-90
		Ultracataclasite	90-100
Grain growth pronounced		—	
			Blastomylonite
		Sibson (1977)	

Table 4. 2: Fault rock classification scheme of Sibson (1977).

Domain	Dextral			Sinistral			Average angle between dextral and sinistral	Estimated σ_1 orientation	Normal			Uncertain
	Nt	Average Orientation	Angle	Count	Average Orientation	Angle to σ_1			Count	Average Orientation	Angle	
1	29	335/81[NE]	31	19	209/83[NW]	23	57	17→182	7	185/82[W]	0	36
2	15	340/72[NE]	26	11	211/77[NW]	26	59	33→184	7	002/77[E]	6	17

Table 4. 3: Summary of geometric analysis of shear fracture surfaces compatible with D_6 . N: number of shear fractures included in average. Angle: angle between average shear fracture and σ_1 measured along orthogonal plane containing pole to average shear fracture and σ_1 . Uncertain: number of shear fractures with an uncertain relationship to D_5 and D_6

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Chapter 5 - Summary of conclusions: geological history of the Tincup Lake area

Yukon-Tanana terrane rocks in the Tincup Lake area, southwestern Yukon, record the evolution of a sedimentary basin from a shallow marine setting, to an erosional surface, to a deepening basin experiencing renewed sedimentary deposition in an arc setting. Between the Late Permian and Late Cretaceous four generations of folding are recorded in the Tincup Lake area, with at least one generation of folding, D_2 , interpreted as coeval with obduction of the Doghead assemblage, an allochthonous oceanic arc terrane, onto the Yukon-Tanana terrane. Late Cretaceous or younger brittle faults cross-cut all generations of folds.

The following is a synopsis of the conclusions reached in this study regarding the evolution of the Yukon-Tanana terrane in the Tincup Lake area, southwest Yukon.

5.1 Stratigraphy

- The Tincup Lake area comprises three fault-bound lithostratigraphic domains (Map 1). The northernmost of these domains is domain 1 which is bound to the south by the Marble Top fault (Map 1) and extends to the north beyond the extent of mapping. Domains 2a and 2b occupy the largest part of the study area, bound to the north by the Marble Top fault and to the south by the steeply south-dipping Nuntaea Creek fault. Domain 3 is the structurally highest domain, consisting of the Doghead assemblage. Domain 3 is bound to the north by the steeply to the south-dipping Nuntaea Creek fault and is open to the south beyond the extent of mapping.
- Domain 1 consists of quartzofeldspathic psammite, psammitic schist, schist and calcite marble (Figure 2.1; Section 2.2.2). Metasedimentary rocks of domain 1 are interpreted to have been deposited in a continental margin setting. Domain 1 is correlated with the Snowcap assemblage of the Yukon-Tanana terrane.

- Domain 2a consists of quartzite, calcite marble with rare dolomitic marble and carbonaceous pelite (Figure 2.1; Section 2.2.3). The metasedimentary rocks of domain 2a are interpreted to have been deposited in a shallow marine setting that underwent a transgressive event. Domain 2a is correlated with the Finlayson assemblage of the Yukon-Tanana terrane.
- Domain 2b consists of metavolcaniclastic psammite, psammitic schist, schist and amphibolite, calcite marble, and interlayered quartzofeldspathic psammite, psammitic schist and schist (Figure 2.1; Section 2.2.3). The metavolcaniclastic and metasedimentary rocks of domain 2b are interpreted to have been deposited in a deepening slope or volcanic apron environment. Domain 2b is correlated with the Klinkit assemblage of the Yukon-Tanana terrane.
- Rocks of Domain 2b unconformably overlie those of Domain 2a. The Doghead assemblage consists of mafic and ultramafic rocks. The Doghead assemblage has been tentatively correlated with the allochthonous Chulitna terrane (Murphy et al. 2011).

5.2 Ductile deformation

- Four phases of folding, D_1 - D_4 , have been identified in the Tincup Lake area. D_1 and D_2 affected only Yukon-Tanana terrane rocks (Section 3.2). D_3 and D_4 affected rocks of both the Yukon-Tanana terrane and the Doghead assemblage.
- D_1 is tentatively correlated with Yukon-Tanana-terrane-wide, Late Permian deformation (Section 3.4.1.2). This correlation is speculative and requires further work to test the hypothesis.
- D_2 folds show a spatial relationship with respect to the Doghead assemblage (Section 3.4.2). The amplitude of D_2 folds decreases and the wavelength of D_2 increases with

distance from the Doghead assemblage. D_2 is interpreted to reflect shortening of the Yukon-Tanana terrane syntectonic with overthrusting of the Doghead assemblage onto the Yukon-Tanana terrane. Since D_2 is interpreted to be syntectonic with overthrusting of the Doghead assemblage, D_2 is constrained to younger than Late Triassic. D_2 is inferred to have occurred under greenschist facies conditions (Section 3.3.2).

- A regional-scale F_3 synform, cored by the Doghead assemblage in the Nunataea Creek area, constrains D_3 to younger than Late Triassic. This regional-scale F_3 synform is offset by late Early Cretaceous or younger regional-scale dextral strike-slip faults (Section 4.4.6.2, Murphy et al. 2009), further constraining D_3 to be older than late-Early Cretaceous.
- D_4 is inferred to be Late Cretaceous, reflecting the tentative correlation of Tincup Lake area F_4 folds with similar structures observed by Israel and Westberg (2011) in the Aishihik area.

5.3 Brittle faulting

- Two phases of late Early Cretaceous or younger brittle faults, D_5 and D_6 , are recorded in the Tincup Lake area (Chapter 4). Both generations of brittle faults are younger than folding in the Tincup Lake area (Map 1). South-dipping normal faults are crosscut by northwest-southeast-striking, steeply dipping dextral strike-slip faults (Figure 4.2). D_6 faults offset the Nunataea Creek fault and therefore are younger than obduction of the Doghead assemblage onto the Yukon-Tanana terrane.
- The older generation of faults, D_5 , is characterized by east-west-striking, gently to moderately dipping normal faults that can be traced for up to 10 km along strike (Section 4.1.1). D_5 faults have a spacing of approximately 4 km. The amount of offset on

D₅ faults is uncertain. A cautious estimate of 5 kilometers of throw along the Marble Top fault is given based on missing stratigraphy between the footwall and hanging wall. D₅ faults contain a cohesive crush breccia, suggesting the faults formed between 4 km and 13 km depth. Ankeritic staining on fault rocks suggests fluids were present in these faults during or following faulting.

- The younger generation of map-scale faults, D₆, is characterized by northwest-southeast striking, steeply dipping strike-slip faults which can be traced for up to 10 km along strike (Section 4.1.1.1). The spacing of map-scale D₆ faults ranges from approximately 250 m in the northeast to approximately 2 km in the southwest. Offset along D₆ faults is a combination of dip-slip and strike-slip motion. Throw of D₆ faults is estimated to be on the order of 10's of meters for an individual map-scale fault while heave is estimated to be on the order of hundreds of meters.
- The estimated orientation of σ_1 for D₅ faults is 77→304. The estimated orientation of σ_1 for D₆ faults is 06→186 (Figure 4.9c)
- Fractures crosscutting D₆ fault material and shear fractures, D₇, are poorly understood. These fractures are likely decompression fractures.

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