Late Wisconsinan McConnell glaciation of the Big Salmon Range, Yukon

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ABSTRACT

The late Wisconsinan McConnell glaciation of the Big Salmon Range in the Pelly Mountains consisted of a four-phase ice-flow history. Phase 1 ice-flow consisted of local alpine glaciers advancing to the mountain front. During phase 2, or glacial maximum, the Cassiar lobe of the Cordilleran ice sheet advanced to the northwest and overtopped the range. Retreat of the Cassiar lobe during phase 3 of the glaciation resulted in ponding of meltwater in eastern drainage basins. The meltwater spilled over into western basins and caused significant erosion of surficial sediments. Phase 4 of the glaciation is marked by a limited late-glacial readvance of local alpine glaciers. This glacial history has several important implications for mineral and placer exploration in the area.

RÉSUMÉ

La glaciation de McConnell au Wisconsinien tardif dans la chaîne Big Salmon des monts Pelly a consisté en quatre phases d'écoulement glaciaire. Pendant la phase 1 des glaciers alpins locaux se sont avancés jusqu'à la marge de la chaîne de montagnes. Pendant la phase 2, ou maximum glaciaire, le lobe de Cassiar de l'inlandsis de la Cordillère s'est avancé vers le nord-ouest pour déborder la chaîne de montagnes. Le recul du lobe de Cassiar pendant la phase 3 de la glaciation a entraîné l'accumulation d'eau de fonte dans les bassins versants orientaux qui ont débordé dans les bassins occidentaux engendrant une importante érosion des sédiments de surface. La phase 4 de la glaciation a été une nouvelle avancée glaciaire tardive restreinte des glaciers alpins locaux.

Des études des minéraux lourds ont été effectuées afin d'évaluer les ruisseaux à placers éloignés dans la chaîne Big Salmon. Les faits saillants relevés dans les données sont des concentrations anormales en arsénopyrite dans un tributaire sans nom du lac Quiet et en ilménite dans le ruisseau Iron, qui présente une géochimie favorable pour les indicateurs de diamants.

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INTRODUCTION

During the last glaciation, southern Yukon was covered both by ice-cap complexes and the northern component of the Cordilleran ice sheet (Jackson and Mackay, 1991). The Cordilleran ice sheet consisted of an assemblage of semi-autonomous ice lobes that originated from distinct upland source areas, such as the Selwyn, Cassiar and eastern Coast mountains (Hughes *et al.*, 1969; Jackson and Mackay, 1991). Not all uplands in southern Yukon acted as accumulation zones for ice lobes, perhaps due to the nature of precipitation patterns (Ward and Jackson, 1992; Bond and Kennedy, 2005). Previous glacial history studies in the Pelly Mountains have shown that after a period of local alpine ice advance at the onset of the last glaciation, the upland was invaded by the Cassiar lobe of the Cordilleran ice sheet (Fig. 1; Kennedy and Bond, 2004; Bond and Kennedy, 2005). The result was an effective reversal of ice-flow in mountain valleys and a complete change in the style of glaciation for that area.

In 2005, a glacial history study was completed for the Big Salmon Range, which forms the southwest extension of the Pelly Mountains (Fig. 1). While a multi-phase glacial history was anticipated for the range, it was uncertain how the glaciation may have impacted valley bottom sediments, which are of particular interest due to the placer potential of the area. This paper characterizes the glacial history for the Big Salmon Range and addresses implications for mineral and placer exploration.

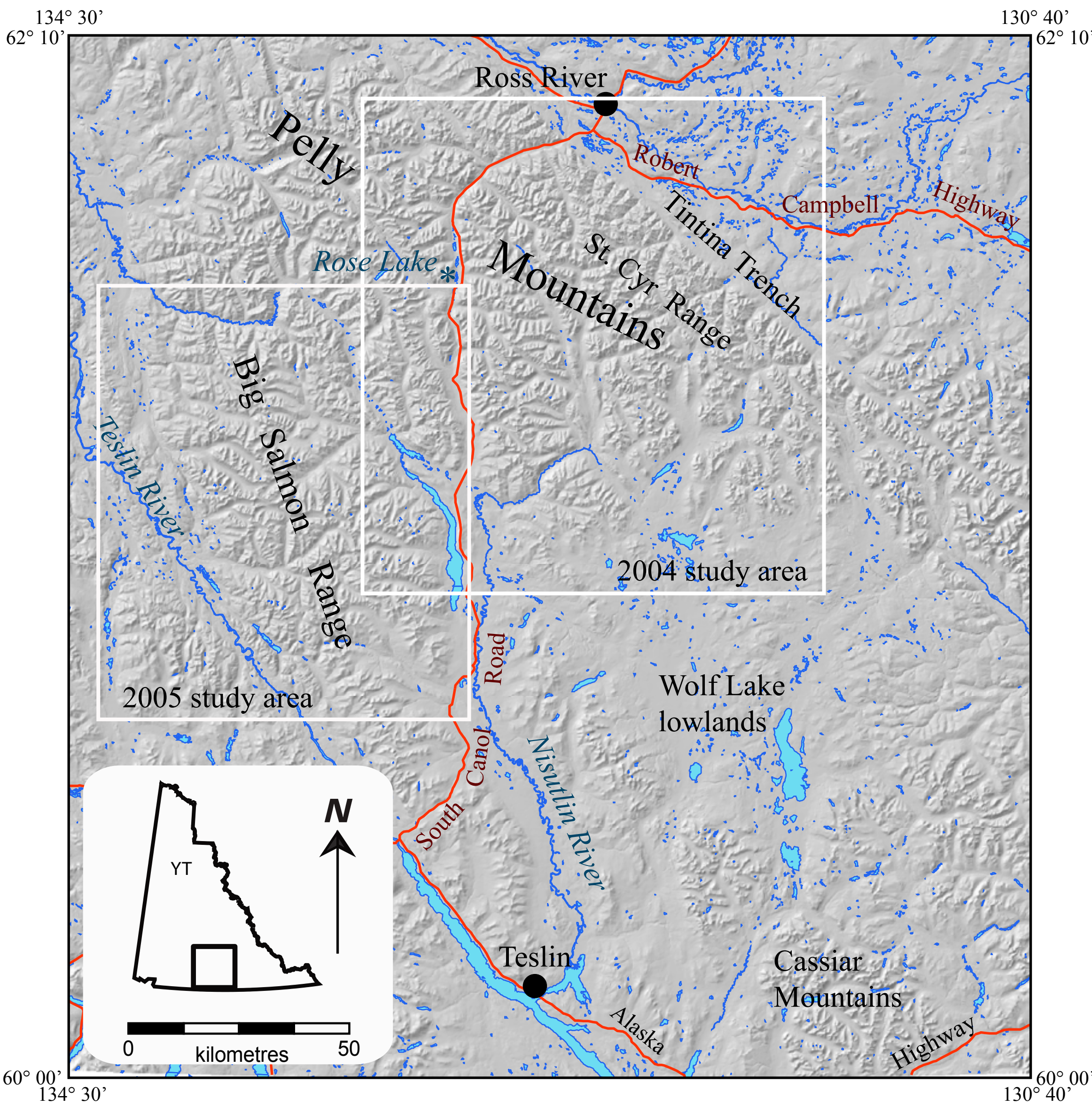


Figure 1. *Regional location map of the 2004 Pelly Mountain study area and the 2005 Big Salmon Range study area.*

PHYSIOGRAPHY, DRAINAGE AND BEDROCK GEOLOGY

The Big Salmon Range is bounded to the west by the Teslin River and to the east by the Wolf Lake lowlands and main ranges of the Pelly Mountains (Figs. 1 and 2). The main drainage divide within the range trends to the north-northwest and contains summits that reach up to 2174 m (7134 ft). Mountains and ridges near the fringe of the range typically reach elevations of around 1800 m (6000 ft). Streams draining to the east in the Big Salmon Range flow into the Nisutlin River, Quiet Lake, or the Big Salmon River; streams draining to the west flow into the South Big Salmon River, Boswell River, or directly into the Teslin River (Figs. 1 and 2).

The geology of the Big Salmon Range can be divided into two general groups: Yukon-Tanana Terrane and mid-Cretaceous intrusive rocks. Yukon-Tanana Terrane in this area consists of metasedimentary rocks belonging to the Snow Cap and Finlayson assemblages (Colpron, 2006a). The Snow Cap assemblage is largely composed of upper Devonian and older siliciclastic rocks, whereas the Finlayson assemblage in the Big Salmon Range consists of upper Devonian to lower Mississippian carbonaceous rocks (Colpron, 2006a). These assemblages have been intruded by the mid-Cretaceous Quiet Lake Batholith, which forms the core of the Big Salmon Range (Gordey and Makepeace, 2003).

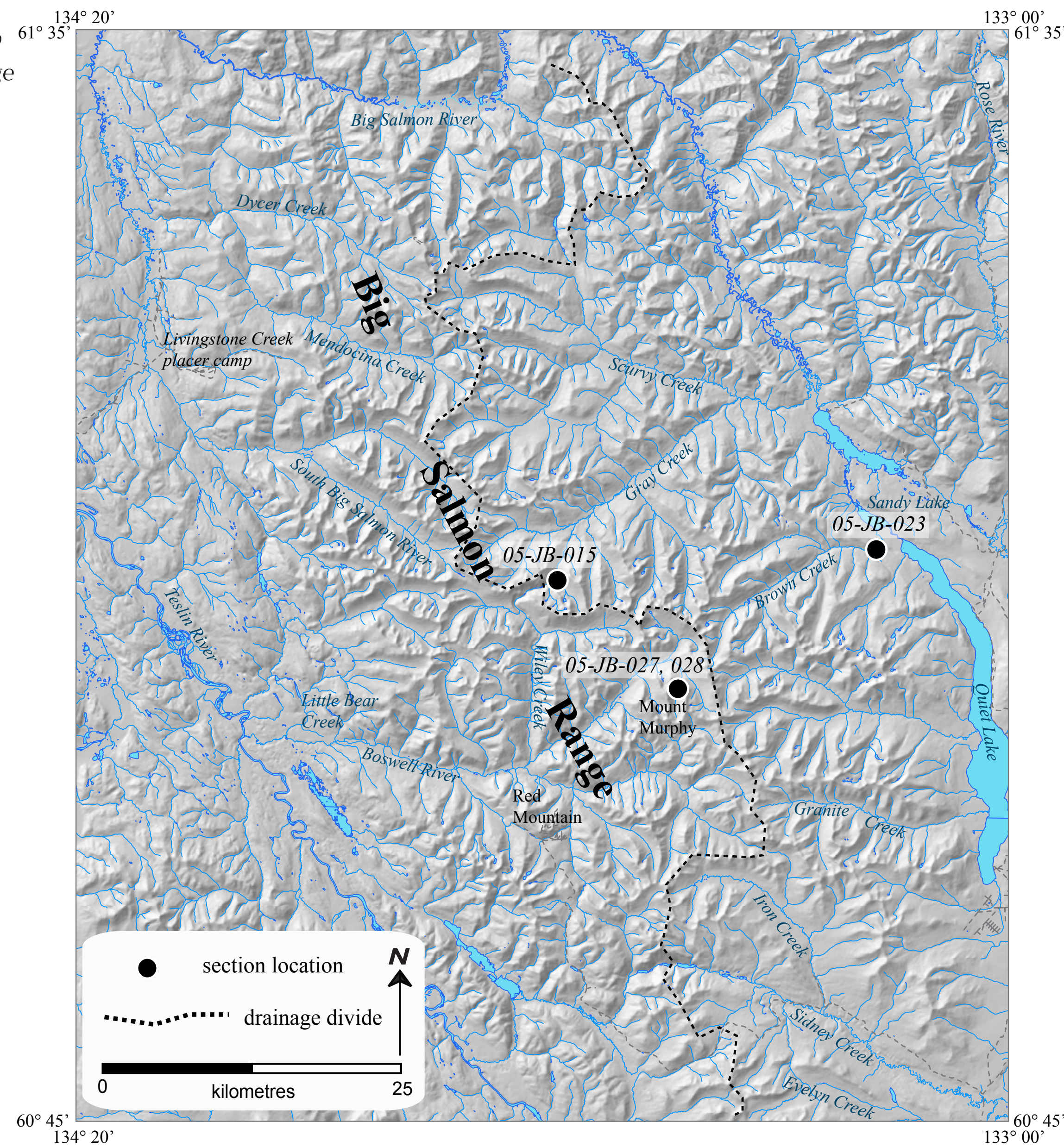


Figure 2. *Location map of the Big Salmon Range study area.*

MINERALIZATION

The Livingstone Creek placer district has produced an estimated 50,000 ounces of gold (1.6 million grams) since its discovery in 1898 (Fig. 2; Lebarge, 2006). The source of the Livingstone placer gold is uncertain, however it is thought to be attributed to skarn mineralization associated with an Early Mississippian metatonalite body that intrudes metasedimentary rocks of the Yukon-Tanana Terrane Snowcap assemblage (Colpron, 2006b). Other known placer gold-bearing streams in the Big Salmon Range include: Dycer, Little Bear, Brown, Iron and Evelyn creeks (Fig. 2; Bond and Church, 2006).

Significant hardrock mineralization in the Big Salmon Range includes the Red Mountain porphyry molybdenum deposit. Molybdenite is found in quartz stockwork that cuts a Late Cretaceous quartz-monzonite porphyry stock (Mortensen, 1992). The inferred resource consists of 187 million tonnes, grading 0.167% MoS_2 (Deklerk and Traynor, 2005).

REGIONAL QUATERNARY GEOLOGY

Yukon has been glaciated numerous times since the late Pliocene (Froese *et al.*, 2000; Duk-Rodkin and Barendregt, 1997; Duk-Rodkin *et al.*, 2001; Jackson *et al.*, 1991). Centres of ice accumulation have included: the St. Elias, interior Coast and Cassiar mountains in southern Yukon; the Selwyn Mountains in eastern Yukon; and the Ogilvie and Wernecke mountains in central Yukon. Ice from each of these accumulation zones flowed radially outward into lowland areas. Thickening ice masses on the landscape developed into separate ice lobes that eventually merged to form a continuous carapace of ice across southern and eastern Yukon (Fig. 3; Jackson *et al.*, 1991). This ice mass formed the northern component of the Cordilleran ice sheet (Fulton, 1991). The limited ice extent in central Yukon was likely a function of aridity (Ward and Jackson, 1992).

The chronology of the last (McConnell) glaciation in Yukon is pieced together from a variety of locations throughout south and central Yukon. Where pre-McConnell organic material has been identified, pollen assemblages suggest that a cooler climate was associated with the onset of glaciation approximately 29 600 years BP (Matthews *et al.*, 1990). In the Tintina Trench, at the foot of the Pelly Mountains, bone-bearing gravel beneath McConnell till was dated at 26 350 ± 280 BP, suggesting ice had not yet extended out of the Pelly Mountains by this date (Jackson and Harington, 1991). Similarly, twig fragments found in a silt unit beneath McConnell till near Watson Lake imply that McConnell glaciers had not yet reached southeast Yukon by 23 900 ± 1140 BP (GSC-2811; Klassen, 1987). The retreat of ice from glacial maximum had begun by 13 660 ± 180 BP as indicated by a radiocarbon age obtained near the terminus of the St. Elias Mountains piedmont lobe complex (GSC-1110; Rampton, 1971). Radiocarbon-dated macrofossils in Marcella Lake (Kettlehole pond) suggest ice-free conditions in southern Yukon by 10 700 BP (Anderson *et al.*, 2002).

Evidence of prior glaciations and interglaciations in the Big Salmon Range is limited. Previous sedimentological investigations by Levson (1992) identified buried pre-Wisconsinan gravel in the Livingstone Creek area. No dates have been published from the Big Salmon Range that constrain the Quaternary chronology specific to this area. A companion open-file map to this report was published for the Big Salmon Range and describes the ice-flow and placer activity history (Bond and Church, 2006).

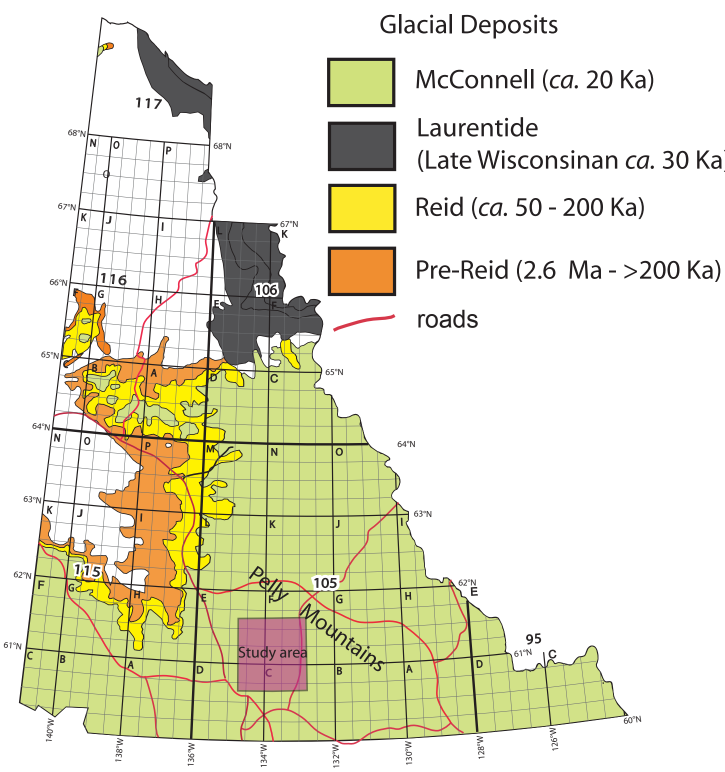


Figure 3. *Glacial limits map of the Yukon (after Duk-Rodkin, 1999).*

METHODS

GEOMORPHOLOGY

The reconstruction of ice-flow for the study area was achieved by examining surficial maps by Jackson (1993), Klassen and Morison (1987), Morison and Klassen (1997), and from new air photo interpretation (Bond and Church, 2006). Morphological attributes of specific glacial landforms contain information that can be used to interpret ice-flow and relative ice thicknesses. This study identified meltwater channels and moraines in order to reconstruct ice-flow during deglaciation, whereas high-elevation aligned landforms provided evidence used to reconstruct ice-flow at glacial maximum.

Meltwater channels

Meltwater channels represent the most common landform that was identified in order to resconstruct the glacial history in the Big Salmon Range. Their relative adundance compared to other glacial landforms is attributed to their erosional nature, particularly when they cut into bedrock as opposed to sediments, in which case they are more resistant to weathering. Meltwater channels that form adjacent to a glacier are called ice-marginal channels, and these commonly occur in clusters where structural and physical characteristics of the bedrock promote enhanced erosion. The dip of an ice-marginal channel reflects the former dip of the glacier and therefore indicates a general ice-flow direction (Fig. 4). A second type of meltwater channel is called a proglacial channel. These channels are commonly found originating at drainage divides where glacier meltwater was impounded on one side of a mountain pass and forced to drain into a lesser glaciated opposing basin (Fig. 5). These channels may erode their base for many kilometres away from the glacier source, depending on valley gradients and ice dams. The presence of these channels helps reconstruct relative ice distributions between neighbouring valleys.

Figure 4. *Lateral meltwater channels cut into bedrock on the flanks of Gray Creek valley. The channels dip in the up-valley direction and reflect the former profile of a receding glacier. View is to the southeast.*

Figure 5. *Proglacial meltwater channels in a tributary to Brown Creek. These channels flow away from a former ice margin and cut across a drainage divide into Gray Creek. View is to the north.*

Moraines

All of the moraines that were mapped represent recessional features and are therefore remnants from deglaciation. Lateral and end moraines reflect the profile of the former glacier margin, and from that, relative ice distribution and ice-flow can be determined (Fig. 6). Where moraines are present on valley sides, caution must be used to ensure that post-depositional mass movements have not occurred to alter the original dip of the landforms.

Aligned landforms

Aligned landforms are the most direct indicators of ice-flow, but they are the least abundant in the study area. In the Big Salmon Range, the only types of aligned landforms found are bedrock ridges that are oriented in the direction of ice-flow (Fig. 7). The combination of glacial erosion and bedrock attributes produce elongate ridges aligned in the direction of ice-flow. Where the ice-flow is parallel with bedding or foliation in bedrock, the aligned ridges are commonly more abundant and better accentuated. Aligned landforms are typically preserved at higher elevations on flat to gently sloping terrain. Their position near mountain summits makes them useful for understanding ice-flow that occurred at, or near, glacial maximum.

STRATIGRAPHY

The stratigraphy of Quaternary sediments was compiled to advance our understanding of the McConnell ice-flow history in the Big Salmon Range. Emphasis was placed on exposures that could potentially help understand ice accumulations prior to glacial maximum; landforms formed during glacial maximum are generally eroded or buried during later phases of glaciation.

Figure 6. *An end moraine in an unnamed alpine valley in the Big Salmon Range. This moraine was deposited by a late deglacial readvance of a local cirque glacier.*

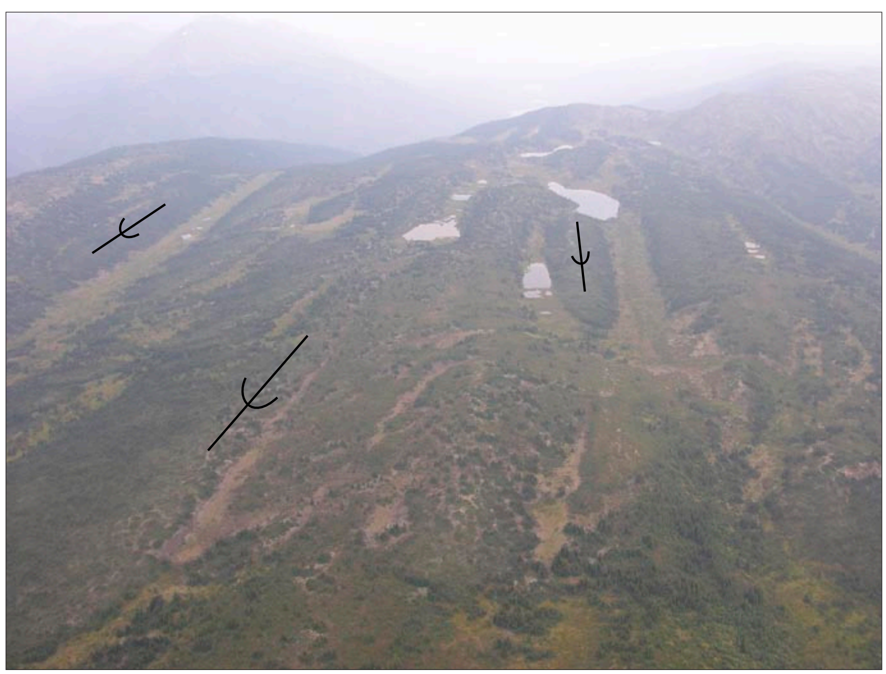


Figure 7. *Aligned till and bedrock landforms west of Livingstone Creek in the Semenof Hills. These landforms are evidence of paleo ice-flow in a northerly direction.*

Exposures of glacial deposits were examined in order to collect the following data: texture, thickness, colour, cohesion, structure and till fabric. Till fabric analysis measures the long-axis orientation of clasts in subglacial till in order to determine the ice-flow direction at the time of deposition. A minimum of 50 clasts are measured during till fabric analysis. The mean orientation of the clasts was also determined for each population, providing an indication of dominant ice-flow direction.

RESULTS

GEOMORPHOLOGY

The oldest landforms from the McConnell glaciation are associated with unconstrained ice-flow across the Big Salmon Range (Bond and Church, 2006). Aligned bedrock landforms were mapped on ridges at elevations up to 1676 m (5500 ft). These landforms record a northwesterly ice-flow direction across the range, and in many areas, they crosscut the topographic trend of the local valleys and ridges. Additional evidence is provided by glacial erratics on Mount Murphy, near the Boswell River. Boulders as tall as 1 m are found at elevations up to 2000 m (6561 ft). The direction of transport of this erratic train is uncertain, however, its elevation indicates ice thicknesses likely exceeded the mountain peaks in this area. Similar landforms were mapped further east near Rose Lake at 1524 m (5200 ft) and they record a northwesterly ice-flow across the Pelly Mountains (Fig. 1; Bond and Kennedy, 2005).

Abundant moraines and meltwater channels below 1676 m (5500 ft) of elevation record topographically controlled easterly to westerly ice-flow in the Big Salmon Range (Bond and Church, 2006). Easterly draining valleys experienced up-valley ice-flow, whereas westerly draining valleys experienced down-valley ice-flow. Evidence of this is provided by up-valley trending end moraines and meltwater channels preserved in Scurvy, Gray, Granite and Iron creek drainages (Fig. 2). On the fringes of the range, the ice flowed parallel to major bounding valleys such as the Teslin and Big Salmon river valleys. As a result, tributaries located on the periphery of these major bounding valleys were commonly glaciated by ice-flow moving perpendicular to the topography.

Up-valley flowing glaciers on the eastern side of the Big Salmon Range forced meltwater to breach mountain passes and descend into the western drainages. In doing so, the meltwater cut deep channels that extended several kilometres (Fig. 8).

Figure 8. *An incised meltwater channel near the headwaters of Mendocina Creek. The depth of incision into the valley bottom is approximately 30 m. The channel extends for 35 km down to the South Big Salmon River. View is to the east.*

QUATERNARY STRATIGRAPHY

05-JB-023: Brown Creek

Brown Creek drains the eastern side of the Big Salmon Range and is a tributary to Sandy Lake (Fig. 2). The exposure is located at the intersection where Brown Creek flows out of the Big Salmon Range and enters the Quiet Lake valley. The section is illustrated in Figure 9 and consists of boulder-dominated gravel/diamicton (Unit 1) at the base, overlain by a compact grey till (Unit 2), and capped by a sorted sandy gravel (Unit 3).

Unit 1 is a coarse, poorly sorted gravel containing a matrix of muddy sand. The cobbles and boulders are mainly granitic in composition and have a crude imbrication suggestive of a paleo-flow that is concurrent with the modern Brown Creek. Unit 1 is interpreted as an ice-proximal, advance outwash gravel originating from a local alpine glacier in Brown Creek.

Unit 2 is a cohesive, matrix-supported, grey diamict with abundant jointing and an abrupt lower contact. The mean orientation of the clast fabric within the unit is 309°. Unit 2 is interpreted as a basal lodgement till derived from a Cordilleran/regional glacier flowing to the northwest in the Quiet Lake valley.

Unit 3 consists of >13 m of bedded sand and gravel. This unit was deposited by fluvial processes, however, no paleo-current information was obtained from this unit. Its stratigraphic position above the basal lodgement till

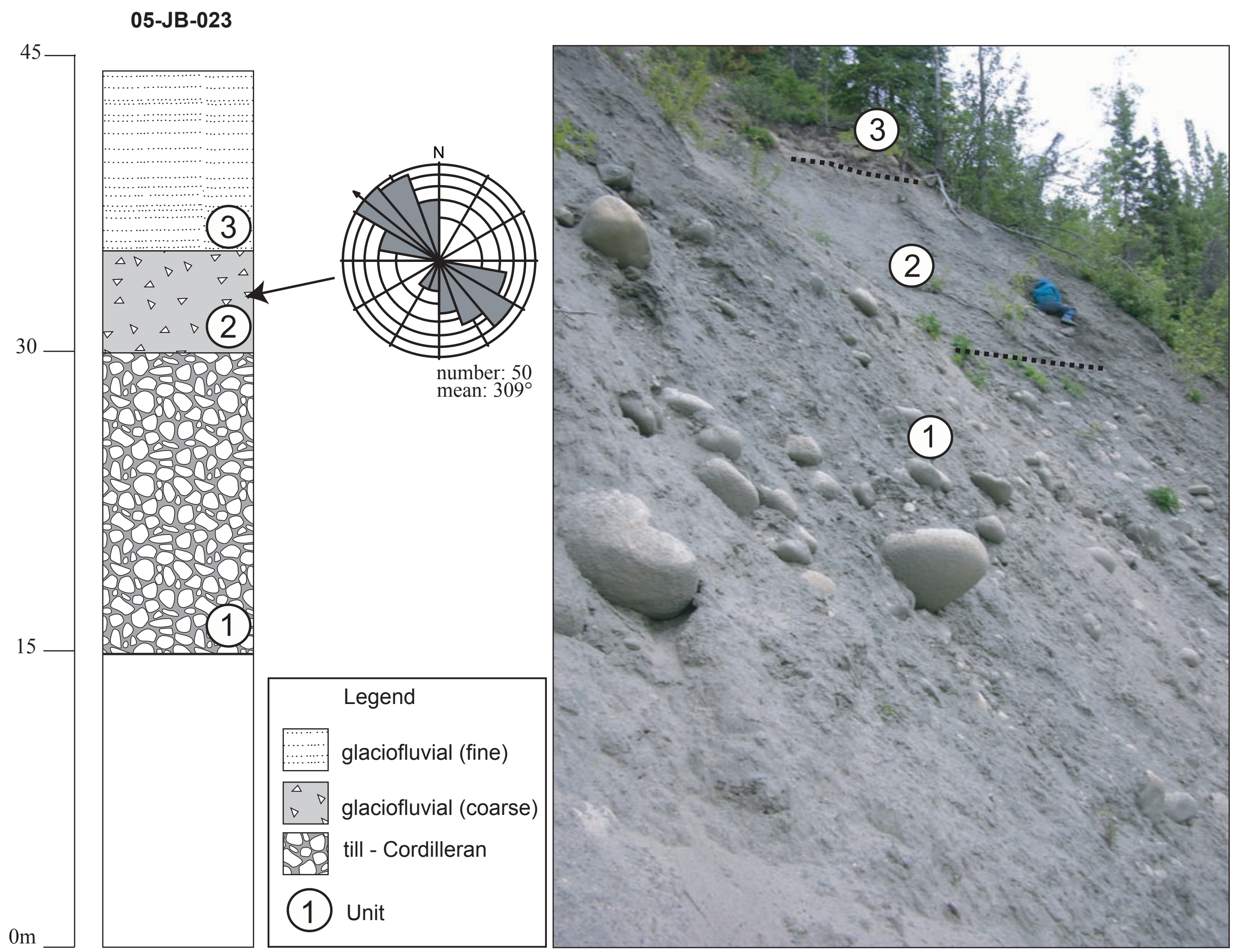


Figure 9. *Stratigraphic section 05-JB-023 consists of a sequence of glacial deposits from at least three phases of the last glaciation. The exposure is located near the mouth of Brown Creek.*

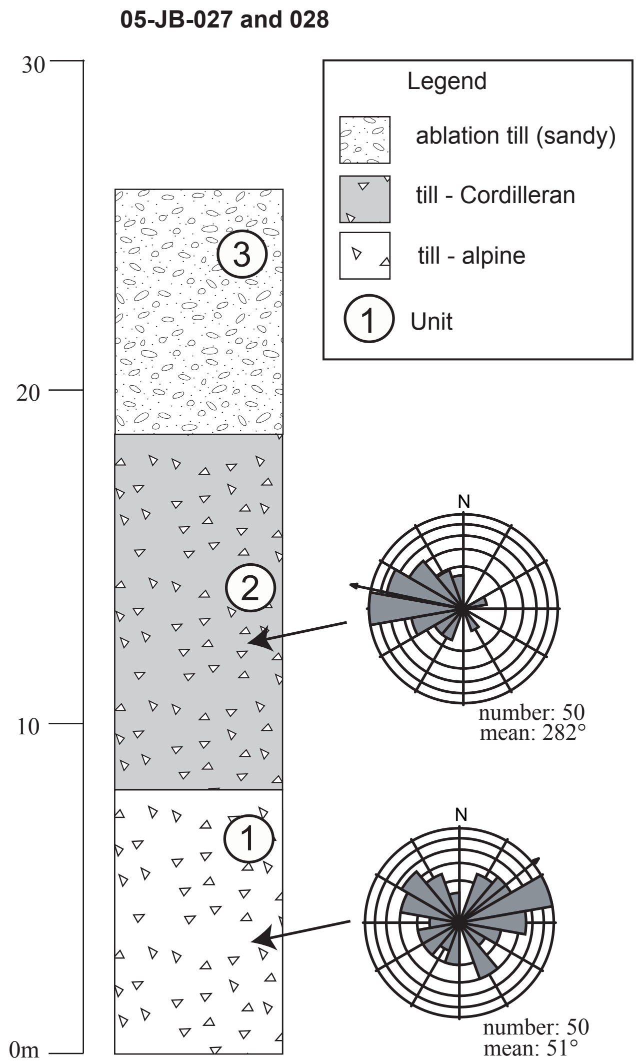
suggests it is an ice marginal, post-glacial outwash associated with the retreating Cordilleran ice sheet.

05-JB-027 and 028: Mount Murphy

A stream-cut exposure was investigated on an unnamed tributary in the headwaters of the Boswell River. The stream originates from a northeast-facing cirque on Mount Murphy (Fig. 2). The stratigraphy from the exposure is illustrated in Figure 10. The unit descriptions and correlations were acquired from a number of exposures within the valley, so the stratigraphy represents a composite reconstruction.

Unit 1 is an olive-brown diamict consisting of equal parts fine and coarse clastic fractions. Clasts are angular and the dominant lithology is intrusive. Bedded sand and washed granule deposits are present within the matrix component and jointing was also observed in less-sorted components of the deposit. The clast fabric analysis yielded two apparent ice-flow directions: the dominant fabric suggests an ice-flow toward the northeast and a secondary fabric records a northwesterly ice-flow. Unit 1 is interpreted as a basal melt-out till that was deposited from a local alpine glacier originating on Mount Murphy.

Unit 2 is a moderately cohesive and jointed olive-grey diamict. The diamict is matrix supported and contains



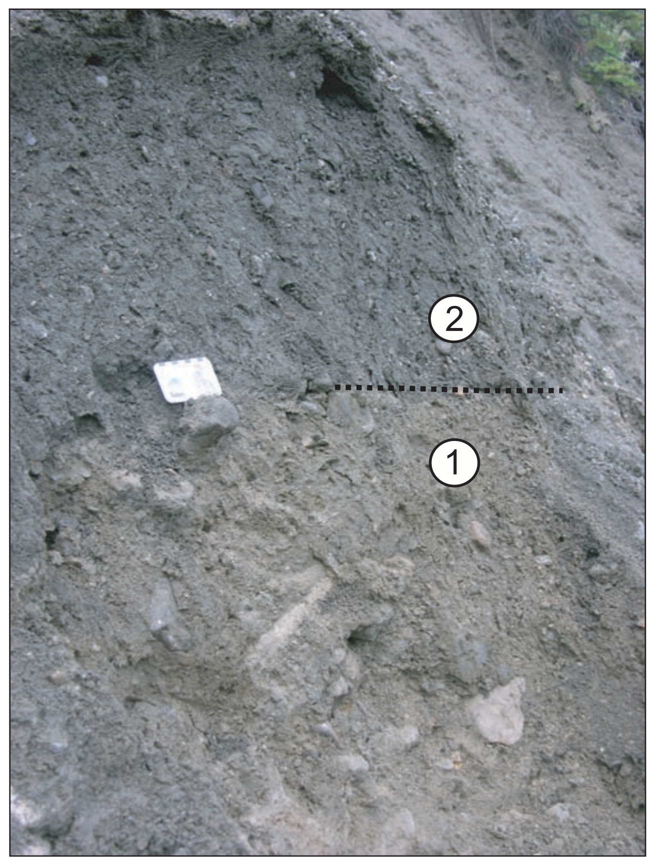


Figure 10. *A composite stratigraphic section from 05-JB-027 and 028 includes multiple tills exposed in a stream-cut near Mount Murphy. Photo is of till deposits exposed at station 05-JB-027.*

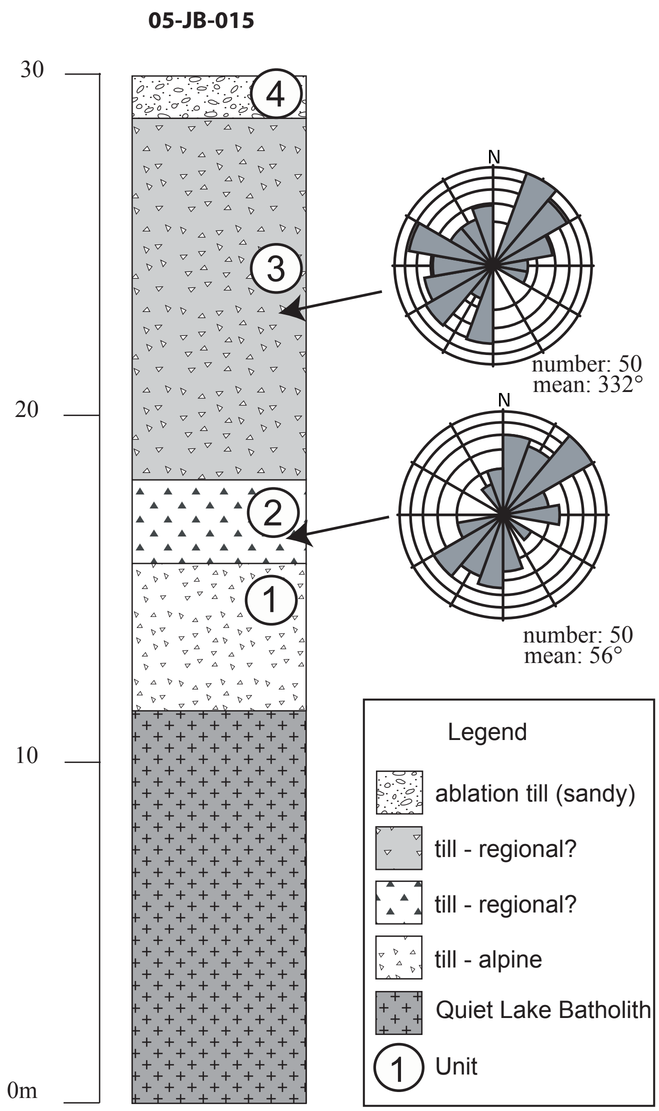
smaller clast sizes than observed in unit 1. The clast fabric within this unit reflects a consistent paleo ice-flow direction to the west (mean 282°). Unit 2 is interpreted as a basal lodgement till originating from a large Cordilleran valley glacier or ice sheet flowing to the west.

Unit 3 is a loosely consolidated, sandy-brown diamict that contains numerous boulders. The deposit is poorly exposed and likely correlates with an erratic train located above the section on the edge of the valley. Unit 3 is interpreted as an ablation till associated with retreat of the same Cordilleran ice sheet/glacier that deposited unit 2.

05-JB-015: near divide between Gray and Wiley creeks

A stream-cut exposure was examined on an unnamed tributary at the headwaters of Gray Creek (Fig. 2). The stream originates within a cirque located approximately 3 km to the southeast. The deposit consists of a succession of glacial diamicts of local and regional origins (Fig. 11).

Unit 1 is a coarse, clast-supported diamict consisting of angular granitic boulders. This unit is in contact with bedrock of the Quiet Lake Batholith. Unit 1 is interpreted as local alpine till. No fabric analysis was completed on this unit.



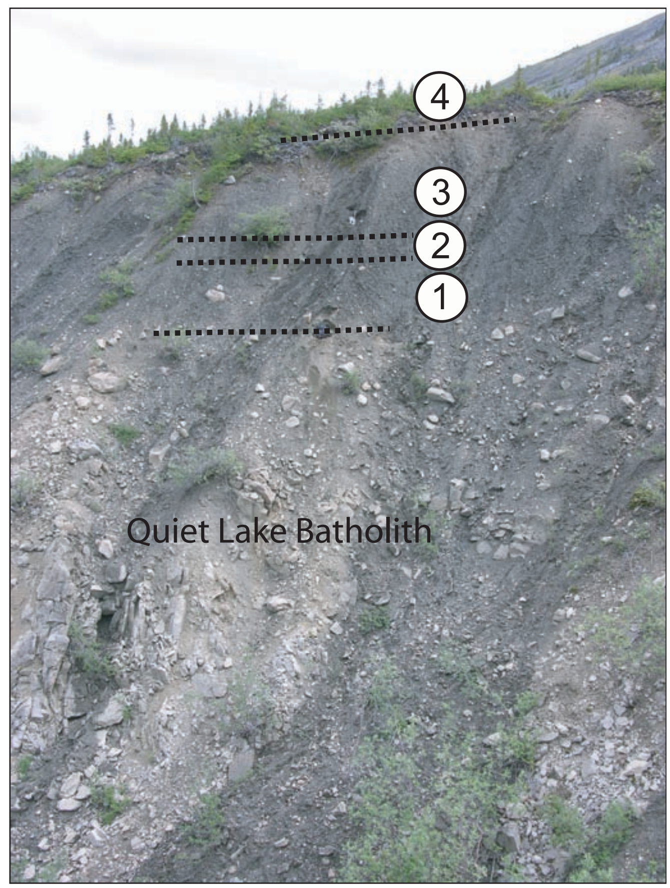


Figure 11. *Stratigraphic section 05-JB-015 consists of a sequence of glacial diamicts. The exposure is located near the divide between Gray and Wiley creeks.*

Unit 2 is a grey-brown cohesive, oxidized till. Equal amounts of clasts and matrix are present and minor jointing was observed. Polished and striated clasts were also present. The prominent mean orientation of clast fabric within this unit is 56°, or northeast. This unit is interpreted as a basal lodgement till, possibly originating from the Cordilleran ice sheet.

Unit 3 is a dark-grey, matrix-supported cohesive diamict. Jointing is present within the deposit and few boulders were visible. The clast fabric for this unit has a strong northeast, as well as southwest orientation. Westward, or up-valley-trending moraines noted in the local area likely correlate with this diamict unit and therefore the fabric likely suggests a southwesterly ice-flow.

Unit 4 consists of a veneer of sandy gravel, interpreted as an ablation deposit associated with unit 3.

DISCUSSION

The late Wisconsinan McConnell glacial history of the Big Salmon Range can be separated into four phases. McConnell ice originated from both local and regional sources.

McConnell advance: Phase 1

Evidence for the regional extent of this phase is limited to sporadic stratigraphic exposures. A more comprehensive reconstruction would require glacier modeling and additional Quaternary stratigraphic research.

Local alpine glaciers within the Big Salmon Range advanced from cirque accumulation zones with onset of the last glaciation and global cooling (Fig. 12a; Bond and Church, 2006). Stratigraphic evidence from Mount Murphy and upper Gray Creek indicate that local glaciers advanced at least 3 km from their source areas. However, stratigraphy from Brown Creek suggests that local alpine glaciers likely extended up to 16 km down-valley. A proglacial outwash deposit at the mouth of Brown Creek (Fig. 9, Unit 1) implies that a local alpine glacier had advanced near to the mountain front. No alpine till was observed at this site, so it is unlikely that the glacier extended into the Quiet Lake valley. This contrasts with the Pelly Mountains further east, where alpine glaciers advanced beyond the mountain front at the onset of the last glaciation (Bond and Kennedy, 2005).

McConnell glacial maximum: Phase 2

At glacial maximum, the Cassiar lobe of the Cordilleran ice sheet overtopped the Big Salmon Range (Bond and Church, 2006). This is supported by the presence of high-elevation, glacially aligned bedrock that records a west-northwest ice-flow direction throughout much of the range (Fig. 12b). This ice-flow trajectory shifts to a north-northwest direction upon converging with ice in the Teslin River valley on the west side of the range (Fig. 12b). A similar glacial history was interpreted for the Pelly Mountains to the east where the Cassiar lobe of the Cordilleran ice sheet advanced from the south and invaded the upland (Bond and Kennedy, 2005). The thickness of the ice sheet over the Big Salmon Range was sufficient enough to permit ice to flow over the range without being influenced and directed by the underlying topography. Erratics observed on Mount Murphy indicate a minimum ice sheet thickness of 760 m in the central part of the range, however, the actual thickness was likely much greater.

McConnell early deglaciation: Phase 3

The Cassiar lobe began to thin over the Big Salmon Range due to a warmer climate following glacial maximum. This reduction in ice thickness resulted in ice movement shifting more towards the west due to the greater influence of the underlying, local topography (Fig. 12c). As the Cassiar lobe retreated to the southeast, its ice-front eventually reached the Teslin River valley. Westward-flowing ice out of the Big Salmon Range would have resembled a system of converging valley glaciers. These glaciers were being fed by thicker accumulations of the Cassiar lobe on the eastern side of the range, where ice-flow was partially restricted by the topographic divide.

Retreat of the Cassiar lobe valley glaciers to the east of the drainage divide signified another change in the deglacial history. The up-valley ice-flow on the eastern side of the range resulted in the formation of proglacial lakes against the ice front. The accumulating meltwater was forced to drain over the divide into the western valleys. Glaciolacustrine and glaciodeltaic deposits at the headwaters of Gray Creek provide good evidence for this part of the glacial history, as do many channels that cut across the passes within the Big Salmon Range (Bond and Church, 2006). Erosion by the meltwater streams was significant enough in some western drainages (e.g., Mendocina Creek) to completely erode morainal deposits from the valley bottoms. In addition, the meltwater erosion created canyons in the valley floors; these are still present today and continue to control the modern floodplain morphology (Fig. 13).

With the development of proglacial lakes in the eastern drainages, the retreat of the Cassiar lobe may have been more rapid due to the process of calving at the ice front. Glaciolacustrine deposits were not observed in any of the eastern drainages with the exception of a small exposure at the headwaters of Gray Creek. Glaciolacustrine deposits would be expected in these eastern drainages had a consistent rate of ice retreat been maintained. Instead, morainal and glaciofluvial sediments appear to dominate the lower valleys (Jackson, 1993; Morison and Klassen, 1997) of the eastern drainages. The lack of glaciolacustrine sediment can only be explained by two processes: either rapid ice-retreat, or a change from

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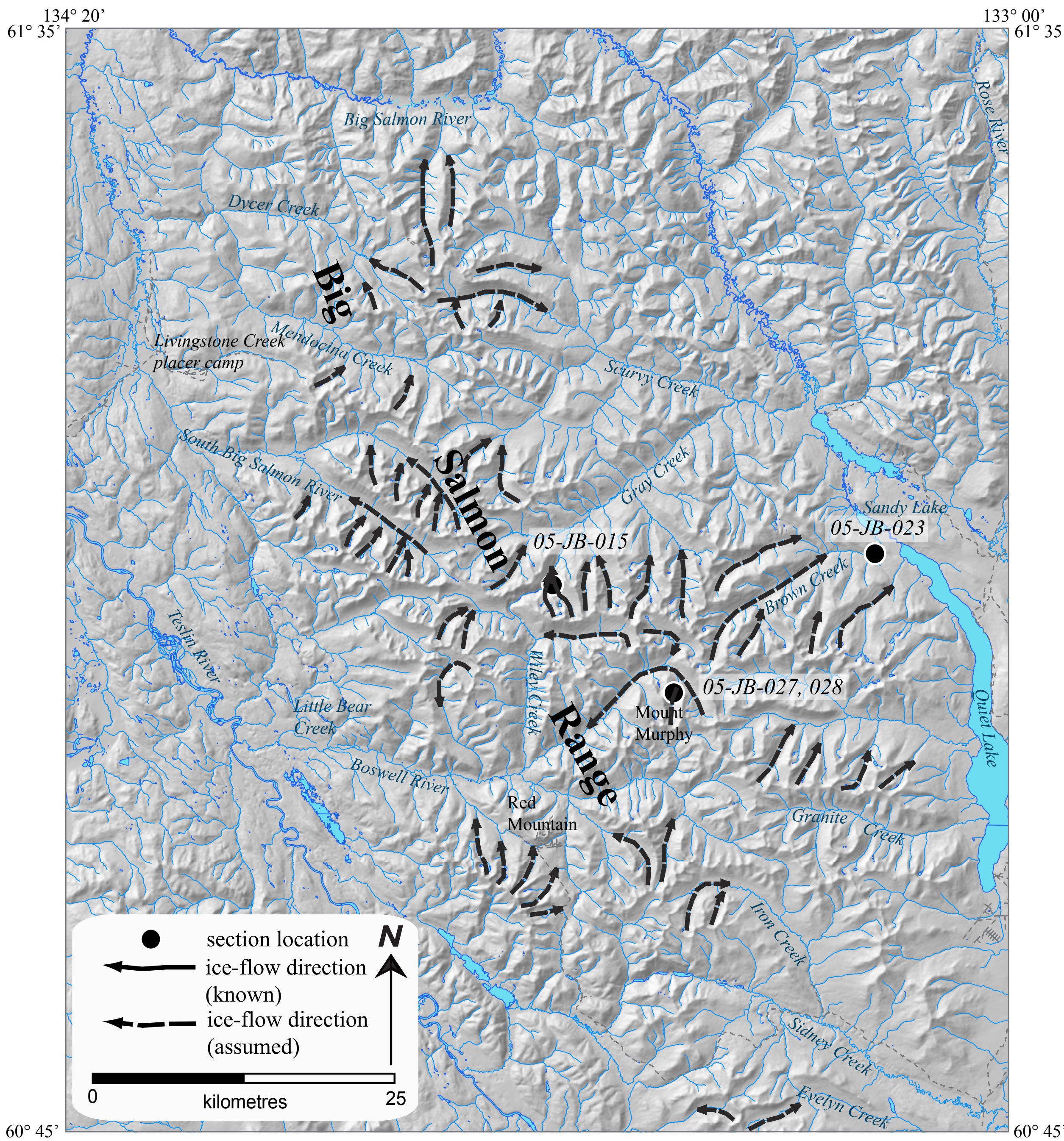

Figure 12(a). *Phases of the McConnell glaciation in the Big Salmon Range (after Bond and Church, 2006). Phase 1: glacial onset characterized by local ice-flow.*

ponded to unponded conditions at the ice margin. An increased rate of retreat could be explained by the process of glacial calving into the proglacial lakes. A change from ponded to unponded conditions could only occur if subglacial drainages were established that allowed water to flow down-valley against the ice-flow. At present, the mechanism of ice retreat in the eastern valleys remains uncertain.

b

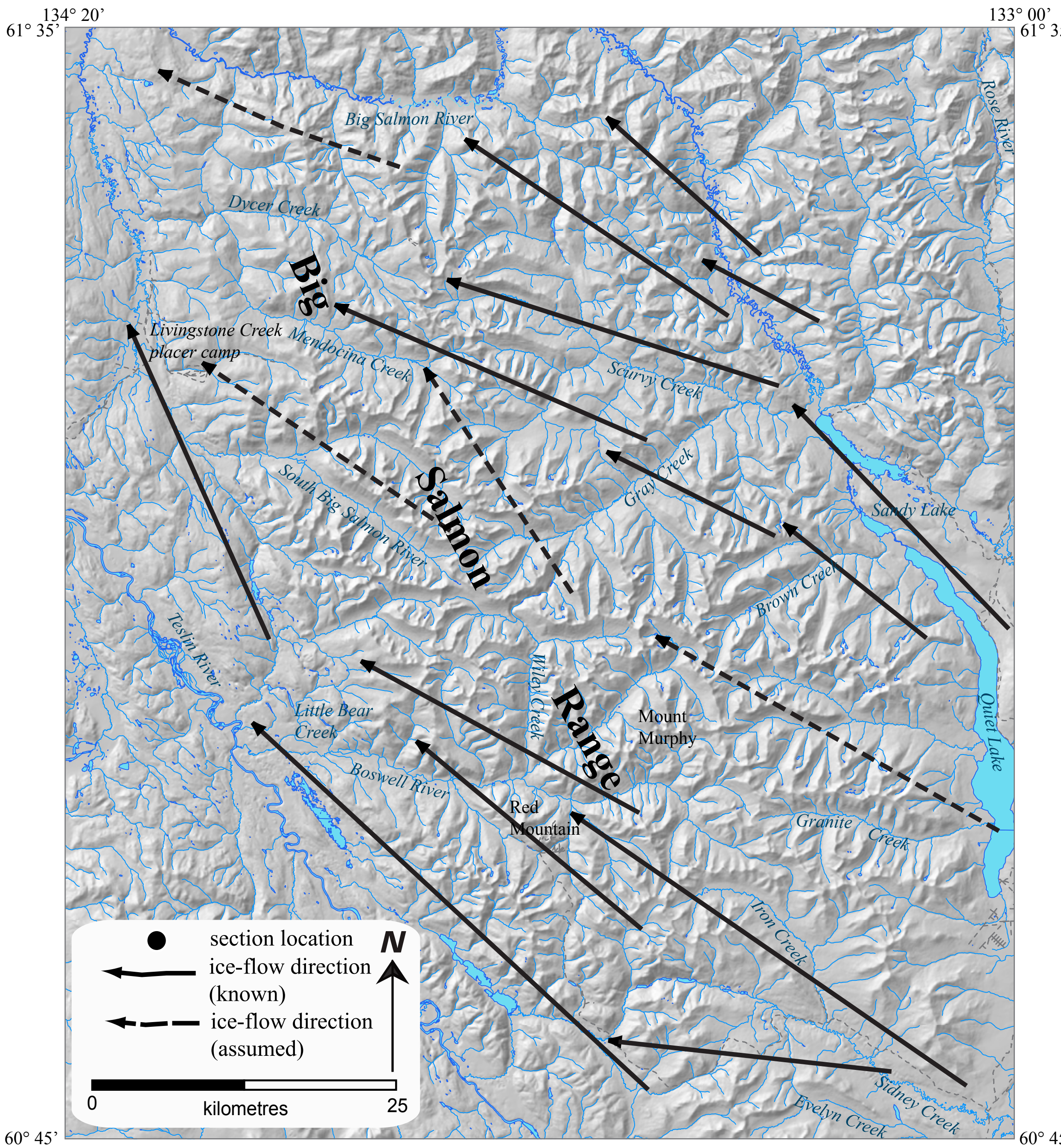

Figure 12(b). *Phases of the McConnell glaciation in the Big Salmon Range (after Bond and Church, 2006). Phase 2: glacial maximum ice-flow reconstruction; ice-flow is to the west-northwest and originates from the Cassiar lobe of the Cordilleran ice sheet.*

McConnell late deglaciation (local readvance): Phase 4

As the Cassiar lobe continued to retreat, a reactivation of local alpine glaciers occurred (Fig. 6; Bond and Church, 2006). End moraines preserved from this phase provide evidence that ice generally advanced less than 3 km from their source areas. The timing of this readvance during the deglacial period has not been determined. A similar readvance was noted in the Pelly Mountains to the east, however, it was much more extensive than in the Big Salmon Range (Bond and Kennedy, 2005). One reason for this could be attributed to the presence of larger cirque complexes in the main part of the Pelly Mountains. Larger upland basins would have had the ability to preserve

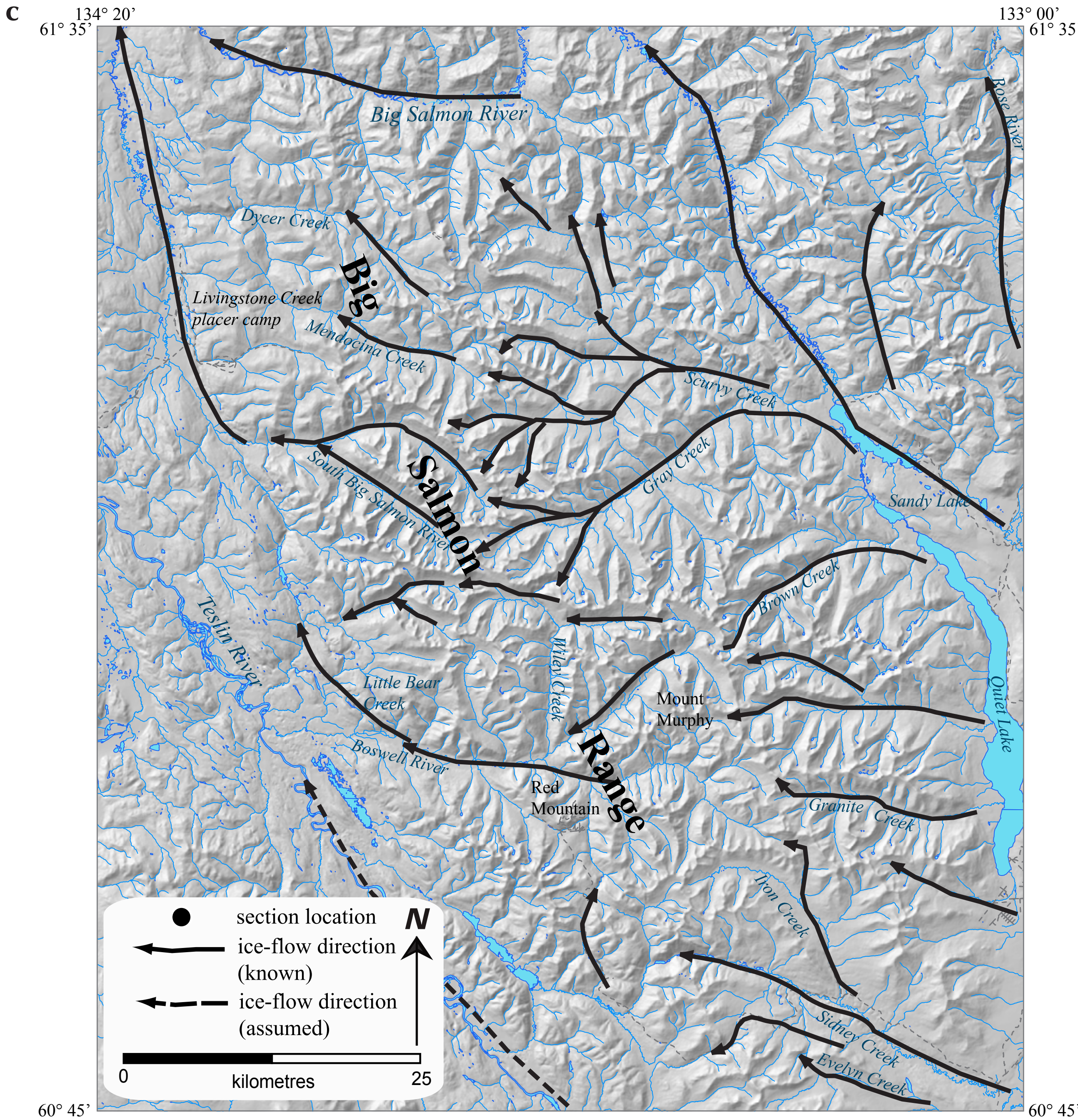

Figure 12(c). *Phases of the McConnell glaciation in the Big Salmon Range (after Bond and Church, 2006). Phase 3: early deglaciation characterized by the retreat of the Cassiar lobe.*

Figure 13. *A canyon located near the middle reaches of Mendocina Creek. The entire length of Mendocina Creek has been altered by the effects of glacial meltwater erosion. The canyon walls are approximately 10 m high.*

remnant ice bodies from the Cassiar lobe. These ice caps would then have the capacity to respond quickly to periods of cooler and wetter climatic conditions if they were to arise during deglaciation. Remnant ice bodies likely existed in the Big Salmon Range as well; however, they were probably less extensive and more isolated. Furthermore, differences between remnant ice bodies in the Big Salmon Range and the Pelly Mountains may be related to regional precipitation variations. This phase of late deglaciation is not illustrated in the Figure 12 compilation because of its limited nature.

IMPLICATIONS FOR MINERAL AND PLACER EXPLORATION

Understanding the glacial history of a region is critical for determining transport directions and provenance of erratic trains and soils derived from till. In addition, it provides invaluable details regarding the erosional and depositional history that has impacted a specific area. The multi-phase ice-flow history proposed for the Big Salmon Range highlights several implications that are important for exploration projects which utilize float, soil and stream geochemistry sampling methods.

MINERAL EXPLORATION

Ice-flow directions for the individual phases of the glacial history have been discussed in the previous section; however, the relative significance of these phases, in terms of drift transport, has not been discussed. As a general rule, the most recent ice-flow phase is the most significant transport direction to consider when tracing the origin of till and erratics. For example, upper drainages that were glaciated by the phase 4 alpine readvance are likely blanketed by surficial materials that are largely derived from the source cirque and surrounding valley walls. A component of foreign drift from a previous transport phase will also be present within the deposit, but will likely account for a smaller percentage of the total drift composition. Despite this general rule of thumb, it is of critical importance that the entire glacial history be taken into account when assessing the origin of soil and erratics. This was demonstrated at the mouth of Brown Creek, where phase 1 alpine glaciofluvial deposits were preserved under phase 2 and 3 till deposits. If anomalous float boulders were found in the modern floodplain of Brown Creek, all three phases would need to be considered in order to determine an origin for the clasts.

The transportation of drift by the Cassiar lobe at glacial maximum is difficult to quantify in the study area. Most of the ice-flow indicators for this phase are found on ridge and plateau summits that were susceptible to basal erosion by the ice sheet. The transport directions on these summit surfaces should parallel the ice-flow directions outlined for glacial maximum. The significance of this phase on sediment transport in mountain valleys is likely less important, due to the active deglacial processes that occurred in the lower areas.

The most widespread surficial materials in the Big Salmon Range are attributed to drift deposits from phase 3 of the glaciation. For most of the valleys, this was the last

Figure 14. *A boulder-dominated placer gold-bearing alluvium deposit observed near the mouth of Brown Creek. This deposit likely represents an alluvial lag deposited during post-glacial incision of Brown Creek. The pogo stick is divided into 25-cm intervals.*

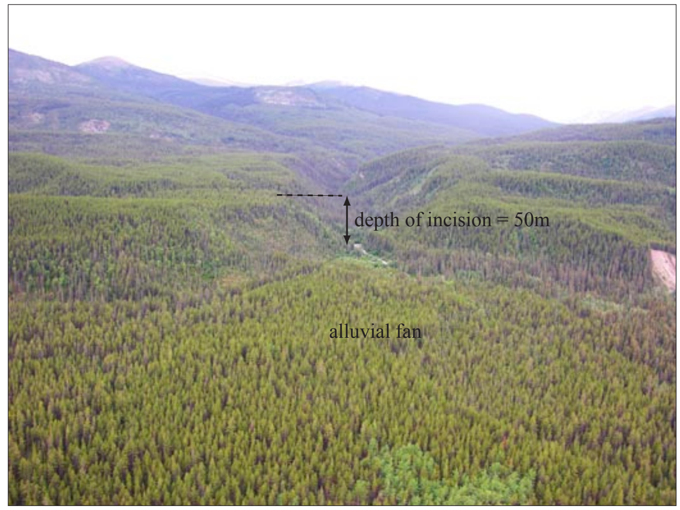

Figure 15. *A view to the west of the mouth of Brown Creek where it exits the Big Salmon Range. In the foreground is an alluvial fan that was deposited as Brown Creek incised into the underlying glacial deposits and bedrock during the early Holocene. The incision occurred in response to base-level change of the Quiet Lake valley.*

phase of the glaciation to erode and deposit material. The processes of erosion and deposition would have been enhanced because the ice was thinner and more easily directed by the underlying topography. The glacier dynamics may have also varied on either side of the Big Salmon Range. On the east side of the range, the ice was flowing up-valley and therefore may have flowed with less energy compared to the western valleys that contained down-valley flowing glaciers. The more vigorous ice-flow in the western valleys likely increased erosion of the bedrock. However, the increased flow would also have reduced the thickness of the glaciers, and therefore the area exposed to erosion. The slower ice-flow on the eastern side of the range may have increased glacial deposition in this area, resulting in greater drift thicknesses.

PLACER EXPLORATION

The Big Salmon Range contains a number of placer gold-bearing streams (Bond and Church, 2006). The Livingstone Creek placer camp contained the most significant deposits, which produced $1 million in gold in the first forty years of activity (Bostock and Lees, 1938). Previous geological research by Levson (1992) in the Livingstone camp area concluded that small tributaries oriented transverse to the former direction of ice-flow were dominated by glacial deposition rather than erosion. This permitted the preservation of pre-glacial placer deposits. Similar settings for placer gold were also noted in glaciated areas near Mayo (LeBarge *et al.*, 2002). These cases highlight the significance of this setting as a potential exploration target for placer gold deposits.

While stream orientation with respect to paleo ice-flow played a significant role in the Livingstone camp, other known placer streams in the Big Salmon Range were affected by ice-flow parallel to valleys. These include Dycer, Iron, Evelyn and Brown creeks (Bond and Church, 2006). The stratigraphy presented earlier for Brown Creek provides evidence that the drainage contained a local alpine glacier during phase 1 of glaciation. Despite the active glacial processes in this drainage, a potentially economic placer deposit has been identified near the mouth of the creek. The placer gold-bearing unit at Brown Creek is not a pre-glacial deposit, but rather a boulder-dominated alluvium that was deposited in response to significant down-cutting following the glaciation (Fig. 14). This local erosion, or base-level change, occurred in response to fluvial/glacial erosion that occurred in the Quiet Lake valley (Fig. 15). Similarly, a drop in base-level in Sidney Creek caused Iron Creek to cut a canyon at its mouth through thick glacial drift in the valley bottom. This undoubtedly contributed to the value of the placer deposit by reconcentrating placer gold onto a new bedrock surface. Future placer discoveries on the east side of the Big Salmon Range may occur where similar post-glacial erosion has taken place.

Streams draining the west side of the Big Salmon Range were more likely to have been affected by meltwater erosion rather than glacial deposition. As was described for phase 3 of the glaciation, meltwater that ponded in the eastern basins drained to the west over low passes. The streambed erosion associated with these meltwater channels was significant and may have eroded or redistributed any previously existing placer deposits (Fig. 13). Areas affected by meltwater erosion are easily identified by the abundant bedrock canyons in the valley bottoms. Basins that escaped the meltwater erosion should have a component of the glacial stratigraphy preserved in the valley bottom. A stratigraphy similar to that described by Levson (1992) for the Livingstone placer camp would be of particular significance for potential placer exploration.

CONCLUSIONS

A four-phase glacial history has been described for the Big Salmon Range. Local ice accumulations early in the glaciation were insufficient to redirect the Cassiar lobe that invaded the range from the southeast. By glacial maximum, the Cassiar lobe had overtopped the upland. During deglaciation, the retreat of the Cassiar lobe differed between the western and eastern drainages. The down-valley ice-flow, as well as high meltwater flow on the west side of the Big Salmon Range, resulted in the dominance of erosional processes. In contrast, the up-valley ice-flow on the eastern side of the Big Salmon Range caused depositional processes to dominate. The surficial geology throughout most of the Big Salmon Range is remnant from this phase of the glacial history.

While conducting mineral and placer exploration in the Big Salmon Range, one needs to consider all phases of the last glaciation to properly assess the genesis of the surficial materials. For example, it is important to be aware that ice-flow directions shifted during the glaciation, causing upland flow directions to possibly contrast with lowland flow directions in the same general area. Likewise, the preservation of a placer deposit is dependant on the degree of erosion that has occurred in the drainage. This is commonly a function of a valley's orientation relative to the paleo-ice flow. In addition, the placer economics of an area are significantly enhanced by post-glacial fluvial incision.

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