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FIELD GUIDE TO QUATERNARY RESEARCH
IN CENTRAL AND WESTERN YUKON TERRITORY

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Field Guide to Quaternary Research in
Central and Western Yukon Territory



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Introduction

The last 5-6 years has seen a resurgence in studies of the rich Plio-Pleistocene record preserved in central and western Yukon. The exceptional exposure of Plio-Pleistocene deposits by Yukon placer miners in the Klondike and Mayo areas, the recognition of new exposures, and the application of new dating methods (palaeomagnetism, tephrochronology - including fission track and Ar-Ar dating) along with soil-stratigraphic and sedimentologic data, and the diligent efforts of palaeoecologists, have led to a more accurate and precise understanding of this region's Ice Age history. These studies, by their nature, are interdisciplinary and this field trip will highlight these connections in the developing story of the evolution of environments in this region over three million years.

The trip will start in Whitehorse and proceed along a transect from the most recent glacial deposits of southern Yukon through progressively older terrain, into the unglaciated region of west-central Yukon. This five day trip will visit some of the classic locales within the Yukon's Quaternary stratigraphic framework, but will focus on new sites that have in some cases fundamentally changed, in others refined our understanding of late Cenozoic environments of the region. Day 1 stops illustrate the soil chronosequence on the McConnell (late Wisconsin), Reid (mid-Pleistocene) and pre-Reid (multiple late Pliocene-early Pleistocene) drift surfaces, and the Quaternary history of the Stewart River valley. Day 2 will focus on the preglacial-to-glacial transition (ca. 2.6 Ma) of the Klondike goldfields, and the record of distal tephra beds and their relation to Plio-Pleistocene environments. Day 3 will visit active mining exposures to look at megafaunal remains, early permafrost, and the relation between past climates and the Klondike placer deposits. On the fourth day, some participants will take a helicopter supported trip into the Tintina Trench and southern Ogilvie Mountains west of the Dempster Highway to look at the record of multiple glaciations preserved there. Other participants will use road access to look at the moraine sequences preserved along the Dempster Highway. The final day is spent traveling back to Whitehorse from Dawson City.

The guidebook is organized in two parts. At the beginning are a series of background summaries on climate, permafrost, regional Quaternary history, palaeoecology, palaeofauna, soils, tephrochronology and placer gold deposits. These are intended to provide a more complete background and context for the individual stops. This is followed by a day-by-day list of stops with associated background information.

Acknowledgements

Much of our understanding of Yukon's Quaternary legacy is the result of the exceptional exposures and support that have been provided by placer miners. In many cases these individuals have gone out of their way, taking time from their short mining season to clear exposures, collect fossils, or re-excavate trenches that may be of interest. Although only a very partial list can be presented here, we would like to acknowledge the contributions of John Alton, Jim Archibald, Tony Beets, Bob Cattermole, Jim and Tara Christie, Torfinn Djukasteinn, Chris Erickson, Paul and Guy Favron, Lance Gibson, Walter Hinnek, Bernie and Ron Johnson, Jerry Klein, Grant Klein, Marty Knutson, David Millar, Mike McDougall, Norm Ross, Stuart Schmidt, Frank Short, Wayne Tattlow, and Brendan White. Scott Smith (Agriculture Canada) is thanked for providing the background summaries on Climate, Permafrost, and Soils and Vegetation from the forthcoming Ecoregions of the Yukon volume.

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Yukon Climate

H. Wahl

The Yukon's climate is basically **subarctic** continental, relatively dry with major temperature variability on both a daily and a seasonal basis. Major **orographic** barriers oriented in a southeast to northwest direction through the Yukon strongly affect precipitation and temperature patterns (Wahl et al. 1987). These broad physiographic barriers, from south to north, are the St. Elias-Coast Mountains, the Pelly-Cassiar Mountains, the Mackenzie-Selwyn-Ogilvie Mountains and the Richardson Mountains (Figure 1). Annual precipitation on coastal Alaska varies between 2,000 to 3,500 mm, whereas within the Yukon, low

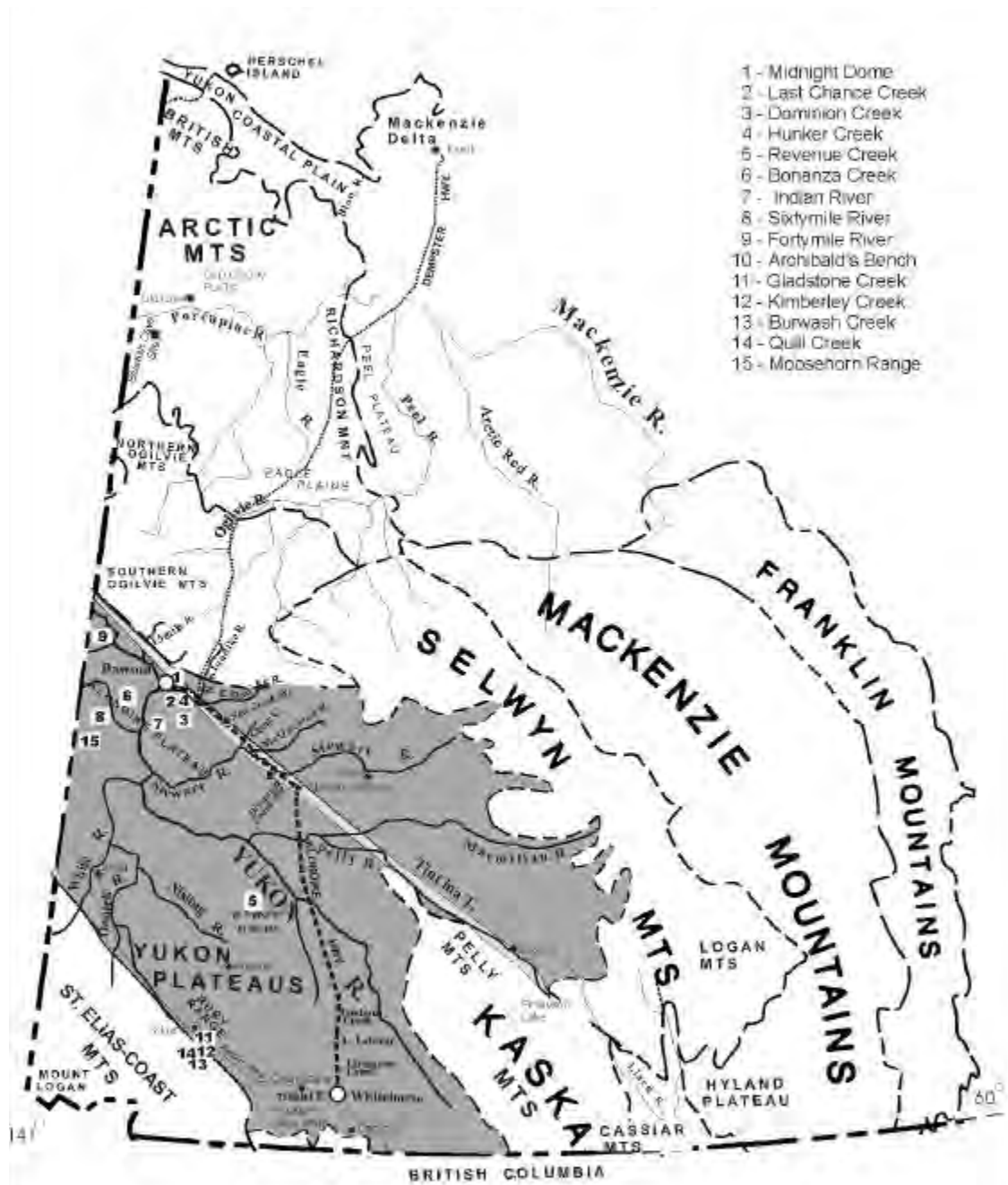


Figure 1. Physiography of Yukon and location of sites mentioned in text.

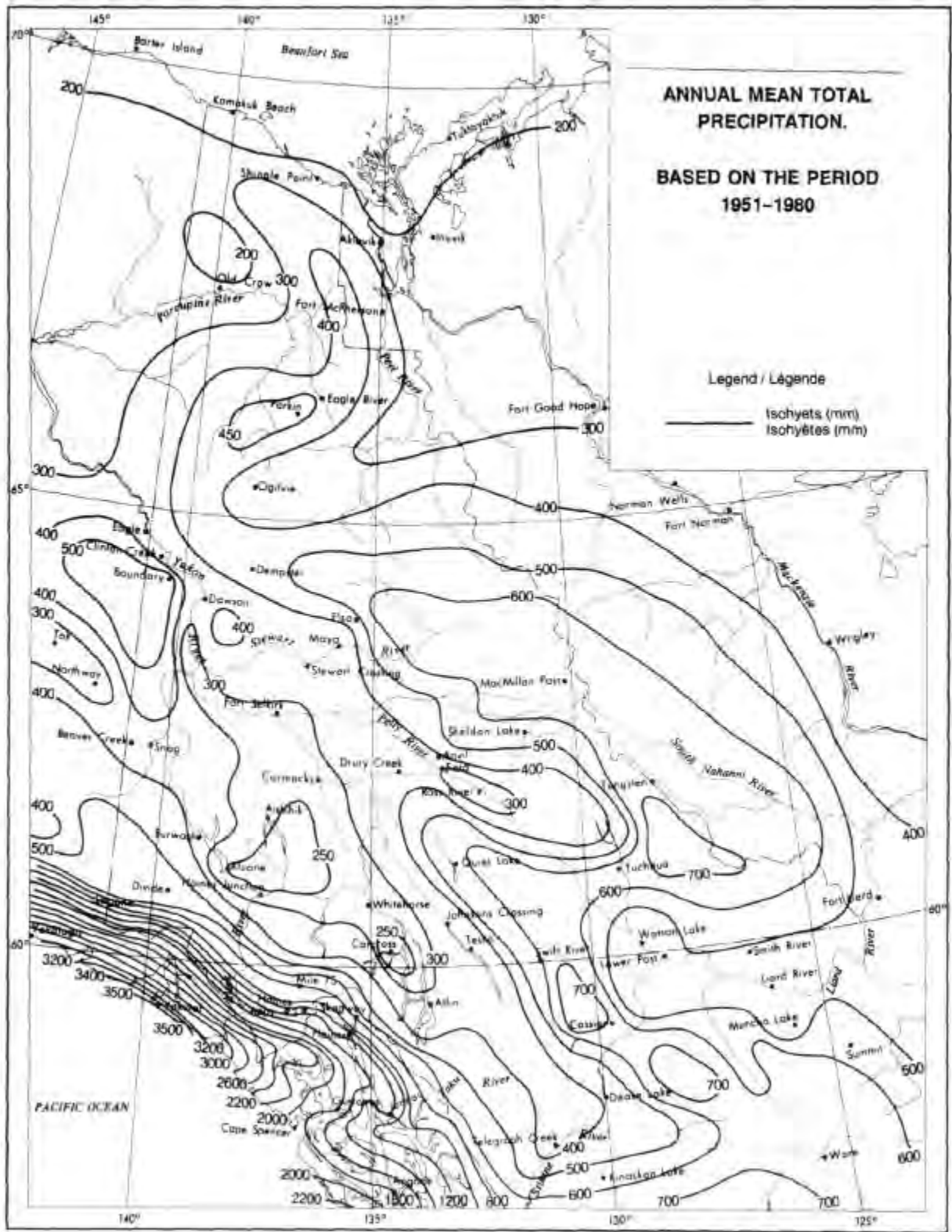


Figure 2. Annual mean total precipitation, Yukon Territory (from Wahl et al. 1987).

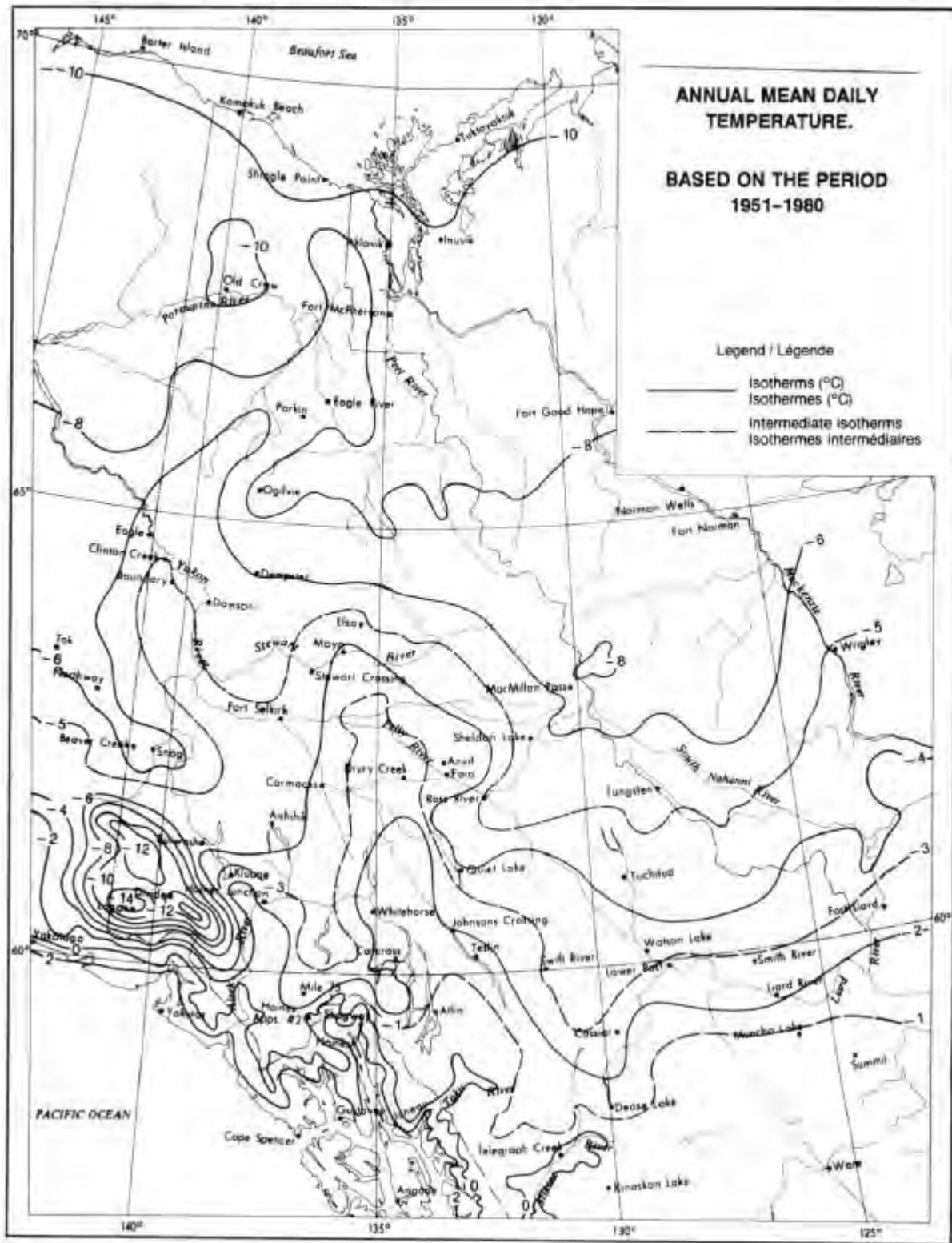


Figure 3. Annual mean daily temperature, Yukon Territory (from Wahl et al. 1987).

elevation valley floors have only 250 to 300 mm (Figure 2). Over the higher barriers within the Yukon, amounts are nearer 400 to 600 mm.

Temperature regimes are, however, much more complex due to both latitude and elevation (Figure 3). On an annual mean basis the latitudinal effect is evident, showing a range from near -2°C over the southern Yukon to below -10°C along the Arctic Coast.

Seasonal temperature variations in the Yukon are the most extreme in Canada, ranging from a minimum of -62.8°C at Snag to a maximum of 36.1°C at Mayo. Daily temperature variations of 20 to 30°C are not uncommon. Although summers are relatively cool, mean daily temperatures are generally above zero from May through September. July is the only month when temperatures below freezing do not occur at all in most of the Territory. As both the Arctic and the north Pacific Oceans are subject to frequent storms, the southwestern Yukon and the Arctic Coast are subject to more wind and cloud than the rest of the Yukon.

Climatic Controls

Latitudinal Effects and Solar Radiation

The Yukon lies between the latitudes of 60° and 70°N . At these latitudes, the hours of possible sunshine in the southern Yukon range from 19 hours per day on June 21 to less than 6 hours per day on December 21. At Herschel Island there is continuous sunshine from May 20 to July 23 and the sun is continually below the horizon from December 1 to January 3. The angle of the sun above the horizon is lower over the Yukon than in southern Canada, therefore the solar energy available to the Yukon averages only 60% of that in extreme southern Canada. Furthermore, the Earth itself is a radiating body and loses heat steadily through its own radiation in a process known as long-wave radiation cooling.

When the sun is well above the horizon, the solar radiation being absorbed exceeds the longwave cooling; the earth warms and the air temperature rises. During the period November to February when the sun is low above the horizon, more energy is lost than gained and the temperature will fall even during a clear, sunny day and, of course, more rapidly after the sun sets. When microclimates are being evaluated, it should be recognized that slopes facing to the east, south or west are more nearly perpendicular to the sun's rays and therefore absorb more of the sun's heat.

The Distribution of Land and Water Masses

Landmasses react quickly to radiation heating and cooling but only at a relatively shallow depth. Large waterbodies, however, with their high heat capacity, appear to react more slowly since through mixing, the heat is distributed through a greater depth and thus is available for a longer time period.

The Pacific Ocean, as the Gulf of Alaska in close proximity to the Yukon, has a great control on the territory's climate. Always a source of moist air, its relatively constant temperature is a potential heat source during the winter, and has a moderating influence on summer heat. Its effectiveness is, of course, dependent on weather patterns and wind direction. The predominant airflow over the Yukon is from the south and west.

The Arctic Ocean is a cold body of water and is predominantly covered with ice even in the summer. Its effect as a climatic modifier is therefore limited primarily to the immediate Arctic Coast.

Orographic Barriers and their Effect on Precipitation

The primary prerequisites for precipitation are a moisture source and a lifting mechanism. Air that is lifted cools and the moisture in the air mass turns into cloud and precipitation. Lifting can be caused by storms, by **convection**, or by collision with orographic barriers. Conversely, air that is forced to descend on the lee of an orographic barrier becomes warm and dry and produces what is known as a **rainshadow** effect.

The main source of moisture for the Yukon is the Pacific Ocean; the predominant air flow is from the south and west. In a typical storm system, southerly winds force air masses to rise over the massive St. Elias-Coast Mountains. Consequently, most of their moisture is precipitated on southern and western slopes of the barrier. The air then descends and dries resulting in a rainshadow over the Kluane, Aishihik and Southern Lakes areas. Air masses are then forced to rise over the Pelly and Cassiar Mountains, repeating a cycle of increasing cloud and precipitation, although to a lesser degree. A rainshadow is evident in the Finlayson Lake-Ross River portion of the Pelly valley. The Mackenzie, Selwyn, and Ogilvie Mountains act as a further orographic barrier. The British-Richardson Mountains, although not as formidable, also have enhanced precipitation compared to the surrounding lower **terrain**.

Less frequent are synoptic patterns that result in northerly or easterly wind trends. In these cases, the northern and eastern slopes receive the heavier precipitation.

In any case, orographic barriers result in enhanced precipitation patterns as is evident in Figure 2. Generally, precipitation increases with elevation with a maximum near the 2,000 m level. A review of Yukon data comparing precipitation amounts with elevation allows a crude approximation of an increase of eight percent precipitation for every 100 metre increase in elevation, up to a maximum at the 1,500 to 2,000 m level and then a slow decrease with increased elevation. Stations being compared should lie on the same side of the slope where possible.

Temperature and Elevation

Normally air temperature decreases with an increase in elevation at the rate of 6°C per 1,000 metres. This change, known as the **lapse rate**, occurs in the southern Yukon from April through October and in the northern Yukon from May through September. As days shorten during the winter months, surface heat loss increases due to the decrease in longwave radiation. Cold air develops over all surfaces, although on mountain slopes this cold air, being relatively heavy, slides into the valley bottoms. The result is a reversal of the normal lapse rate, known as an **inversion**. Air temperature, instead of being cooler with increased elevation, remains **isothermal** through a vertical portion of the atmosphere or, in some cases, the temperature actually rises with increased elevation.

In addition, inversions may be caused at lower elevations by very cold air masses from the Arctic. This Arctic inversion is generally in place over the Yukon from late October to early March and is at its extreme in January. For example, temperatures in the valley floors may range from -20° to -30°C, but increase at a rate of 3° to 5°C per 1,000 metres to temperatures near -10° to -15°C at the 1,500 m level. Air remains isothermal until it begins cooling again above 2,500 m at a more normal lapse rate of 5°C per 1,000 metres. These inversions can be temporarily destroyed by strong winds mixing the warm air from above into the colder valley floors. This occurs most frequently over the southwestern Yukon.

Air temperature changes are greater in the vertical profile than in the horizontal. A change of 6°C per 1,000 m in the vertical dimension is common. Yet in the horizontal dimension, changes of 5° to 10°C are normal over distances of 500 to 1,000 km.

Permafrost

Chris Burn

Permafrost is ground that remains at or below 0°C for two or more years. Permafrost occurs throughout the Yukon, but its thickness and the proportion of ground it underlies increase northwards (Figure 4). All terrain except rivers and lakes is underlain by perennially frozen ground in the northern Yukon, but the scattered permafrost of the southern Yukon is found under less than 25% of the ground surface. Permafrost terrain comprises a seasonally thawed **active layer**, underlain by perennially frozen ground. The active layer is the layer of ground above the permafrost that thaws in the summer and freezes in the winter.

Throughout the Yukon, except in the mountains south of Carcross where snow is deep, **ground ice** and permafrost present hazards to municipal and highway construction. Ground ice is most often found beneath **organic** soil, and is impressively preserved in the Klondike "**mucks**" (Fraser and Burn 1997; Kotler and Burn 2000). Except in the north, permafrost problems have commonly been managed by attempted obliteration of ground ice. For projects of relatively short duration such as mining, this approach has often been adequate. In the long run, however, more imaginative arrangements have proven necessary, for instance at Tatchun Creek, where ground ice melting has caused repeated failure of the Klondike Highway roadbed.

Thickness

Four ground temperature profiles from sites marked on Figure 4 indicate the variation in permafrost conditions across the Yukon. The profile from Blow River represents deep permafrost at 238 m, characteristic of glaciated environments on the Yukon Coastal Plain. Permafrost may be well over 300 metres thick in more westerly, unglaciated portions of this ecoregion, but it thins rapidly to the south, for it was absent beneath glacial ice and lakes; it is only 63 metres thick at Old Crow. The high geothermal flux in Cordilleran terranes of the central and southern Yukon helps to raise the permafrost base to 89 m at the North Cath drill site in Eagle Plains, and to 135 m in the mountains of the Yukon Plateau-North Ecoregion. In areas underlain by coarse glacial deposits, convective heat from groundwater circulation may also raise the base of permafrost locally. Thicknesses between 20 and 60 m have been reported from valley bottom sites in the Klondike Plateau near Dawson,



Figure 4. Permafrost distribution in Yukon, and locations of sites mentioned in text.

and between 25 and 40 m near Mayo. Drilling in the Takhini Valley, in the southern Yukon, has revealed 16 m of frozen sediments, while municipal excavations near Teslin encountered only two metres. In the Yukon, annual mean temperatures at the top of permafrost decline northward with a steep drop across the treeline, varying from -0.8°C in Takhini Valley to -2.8°C at Eagle Plains.

Very thin permafrost may degrade or be established in years or decades, while the time scale for thicknesses of over 15 m is on the order of centuries. Permafrost in the Yukon Coastal Plain has formed over millennia. Thus the permafrost zones are temporal, as well as spatial, units.

Distribution

Annual mean near surface ground temperatures below 0°C lead to permafrost growth. At macroscale, these are a function of air temperature, modified by the insulation of snow. In the Yukon, physiographic factors are responsible for the presence of permafrost, particularly the blocking of maritime air by the Boundary and Icefield Ranges, and topographic enhancement of winter inversions within the dissected Yukon plateaus (Burn 1994).

Snow cover affects local permafrost variation, as it acts as insulation against extreme temperatures. For example, a heavy snowfall in early fall may delay or reduce frost penetration. Permafrost in uplands of the central and southern Yukon is a result of short, cool summers, for in winter the ground is protected by a thick snow cover. In valleys, summer is commonly hot, but the winter may be extremely cold. Within the boreal forest, the snowpack is usually uniform with little drifting, because of interception of snow by the canopy and reduced wind speeds. Above or north of treeline, local conditions of snow cover are more variable and directly impact permafrost distribution, in some areas even leading to the absence of permafrost. For example, in the Mackenzie Delta, thick accumulations of snow in willow thickets around lakes and rivers may lead to eradication of permafrost.

Within the discontinuous permafrost zone, the specific location of frozen ground depends mainly on the thickness of the organic horizon which insulates the ground from higher summer temperatures, and on the moisture content of the active layer. A high rate of evapotranspiration dissipates solar energy that would otherwise warm up the soil and melt permafrost.

The combinations of factors at various scales that lead to permafrost imply that its response to climate change is complex. Changes in surface conditions, such as those wrought by forest fire, often alter the ground thermal regime more rapidly than fluctuations in climate, and are currently causing permafrost degradation in the Takhini Valley. However, climate change over decades, particularly if it alters snow accumulation, may also warm permafrost.

Ground Ice

The practical significance of permafrost largely relates to the growth and decay of ground ice. There is usually an ice-rich zone at the base of the active layer, which forms by ice segregation during downward migration of water into permafrost at the end of summer. Water may also be injected into near-surface permafrost in autumn. The growth of ice wedges, by snow melt infiltrating winter thermal contraction cracks, also contributes to high ice contents in the uppermost 10 metres of the ground. Ice wedge polygons are well developed in lowlands of the northern Yukon, but individual wedges have been reported further south.

Accumulation of ground ice leads to heaving of the ground surface. Thick, laterally extensive bodies of massive, near-surface ice, probably formed by ice segregation during permafrost growth, are found in the Yukon Coastal Plain and in the Klondike District. Glaciolacustrine sediments in the central and southern Yukon often contain beds of segregated ice, which may comprise over 80% ice by volume in the upper 10 metres of the ground. Over 400 open-system pingos have been identified in the central Yukon, mostly in unglaciated valleys, where coarse materials do not impede groundwater movement downslope. Numerous palsas, which are peat mounds with a core of segregated ice, have been identified in wetlands. Buried glacier ice is abundant near the termini of glaciers throughout the southern Yukon, and, at elevation, is a relict from the Neoglacial. Rock glaciers are also widespread in the alpine zone.

Thermokarst

Ground subsidence occurs during thawing of ice-rich terrain, with water pooling in enclosed depressions to form thermokarst lakes. Thermokarst lakes are currently widespread and actively developing in glaciolacustrine sediments deposited during the McConnell glaciation (Burn and Friele 1989; Burn and Smith 1990). Retrogressive thaw slumps may be observed in river banks where active erosion has ex-

posed ice-rich soil, and at some other sites including road cuts.

Drainage

Permafrost derives its ecological significance from cold ground temperatures within the active layer, and from the influence of a relatively impermeable frost table on drainage. Moisture and frost-tolerant species, such as black spruce and mosses, are often associated with permafrost, while deciduous forests usually grow in drier, permafrost-free soil. The relation between vegetation and permafrost is illustrated by the sequence of vegetation succession that commonly follows forest fire. Ground warming and thickening of the active layer in years shortly after fire improve drainage and allow the establishment of species suited to dry soils, especially pine. However, as a surface organic horizon redevelops, the active layer thins, segregated ice persists at the base of the active layer, and drainage is impeded, leading to replacement of pine by moisture-tolerant species such as spruce.

Yukon Vegetation

Adapted From Oswald and Senyk (1977)

The recent publication *Flora of the Yukon*, by William Cody (1996), lists vascular plant species for the Yukon. Although relative to other parts of Canada its plant diversity is low, Yukon still has a varied and in many ways unique flora.

Trees cover most of the plateaus and valleys in the southern Yukon, and form closed to open canopies, depending on site conditions. The southeastern Yukon has the greatest proportion of closed-canopy forests and the greatest number of tree species, which include white spruce, black spruce, larch, alpine fir, lodgepole pine, aspen, balsam poplar and paper birch. To the west and north, tree stands become more open and discontinuous. Larch, alpine fir, and lodgepole pine are generally absent, but larch occurs north of the Selwyn Mountains, and alpine fir occurs to a limited extent in the northwest Klondike Plateau.

White spruce and black spruce would normally be the climatic climax species on moderate to well-drained sites in the south-central and eastern Yukon, but as a result of fire, current stands contain lodgepole pine, and to a much lesser extent, aspen. Black spruce predominates in poorly drained areas. In the southwestern Yukon, black spruce or mixed black spruce and white spruce form the climatic climax due to the incidence of permafrost developing under mature for-

ests. Currently, stands consist of black and white spruce, aspen, balsam poplar and paper birch, in pure stands or mixed in various proportions. Succession after fire usually starts with willow, aspen and balsam poplar. Occasionally, black spruce may be an initial invader.

Alpine fir is the primary alpine timberline species throughout the south-central and eastern areas, but white spruce replaces it westward and northward. Several tree species occur near the Arctic treeline, but white and black spruce are the most prevalent. White spruce, aspen and balsam poplar extend in protected situations almost to the Arctic Ocean.

Shrub communities are restricted to recent alluvial sites, disturbed areas, and wetlands and sites near treeline in the southern Yukon, except where they occur under a forest canopy. Their incidence increases northward, especially on higher plateaus and protected slopes of mountains. Willow, shrub birch, soapberry and alder are the most prevalent species on better-drained sites; ericaceous shrubs (shrubs of the family Ericaceae, usually with thick, leathery leaves), frequently with willows and shrubby cinquefoil, occur on poorly drained sites.

Tussock fields, of either sedge or cottongrass, occur in the southern Yukon, but increase in ground coverage northward to form the predominant vegetation of the Arctic tundra. They occur on imperfectly drained sites where seasonal frost lasts for a significant portion of the year. The tussocks may be hummocked, especially where permafrost is present. High water tables supported by permafrost may permit tussock development on gentle slopes. Trees, particularly black spruce and larch, and shrubs, mainly ericads and willows, may occur in tussock fields with forbs, lichens and mosses usually present.

Grasslands are restricted to steep, dry, south-facing slopes along the Yukon and Pelly Rivers on moraine, colluvium and glaciofluvial material. Shrubs, such as sagewort and rose, and several forbs occur in the grasslands. These areas are very dry during the summer and, because of their position on steep slopes, are susceptible to erosion.

The Yukon, compared with most other areas of northern Canada, does not possess extensive wetlands (Zoltai et al. 1988). Wetlands are critical landscape components as hydrologic storage and filtering areas and are important wildlife habitat. The nature of wetlands varies within the territory. Within the Boreal Cordilleran Ecozone in the southwestern Yukon wet-

lands are relatively small and scattered with the exception of a few large marshes associated with active deltas. Wetlands there tend to be largely without peat formation and are often characterized by marl and fen development as is typical in the Whitehorse area. In the central and southeastern Yukon, wetlands are usually a complex of fen and plateau bog where the bog portion of the wetland is underlain by permafrost. As permafrost is continuous in the Taiga Cordilleran region of northern Yukon, plateau bogs and collapse scar fens are the most common wetland forms. In all cases shallow open water is often a major component of the wetland complexes.

Alpine tundra consists of several communities, ranging from sedge meadows and tussock fields to **pioneer** colonies of lichens on rocks. The wetter areas, common on gently sloping to depressional terrain with an accumulation of organic matter, possess vegetation similar to that described for tussock fields. The **mesic** alpine vegetation is characterized by a combination of prostrate shrubs, mainly ericads and willows, grass, sedge, forbs, lichens, and sphagnum and other mosses. **Mineral soils** at the surface are usually stony, and permafrost is either deep or absent. The soils are well-drained and tend to dry out during summer if the snow-free period is sufficiently long. Rock fields may have only **crustose** or **fruticose** lichens growing on the rock, and members of the mesic alpine vegetation community growing in **interstices** between rocks.

Yukon Glacial History

A. Duk-Rodkin and D.G. Froese

The glacial history of the Yukon Territory is old and complex. The Yukon Territory has been affected by Cordilleran and montane glaciers as well as by continental ice, the Laurentide Ice Sheet in late Pleistocene (Late Wisconsinan) time. Southwestern Yukon and southeastern Alaska is the only region in northwestern North America with an extensive and old glacial record that dates to the Miocene >5 Ma (Denton and Armstrong 1969; Lagoe et al. 1993). The history of glaciations in this particular region is related to the history of uplift of these coastal mountains which record initial uplift ca. 14 Ma (O'Sullivan and Currie 1996) and cold conditions around 5 Ma (Lagoe et al. 1993). These events were followed by regional erosion and renewed uplift associated with cold conditions and the late Cenozoic ice age (Raymo 1992; White et al. 1997). High rates of late Cenozoic uplift in southern Yukon and Alaska (St. Elias, Wrangell and Alaska ranges), which host the highest peaks on the continent, have served as a progressive barrier to

Pacific moisture, increasing continentality over this interval. The Cordilleran ice reflects the coalescence of ice from the continental divide and from local and multiple upland areas (St. Elias, Cassiar, Pelly and Selwyn mountains, principally) to form an ice sheet that covered up to 70% of Yukon (Duk-Rodkin 1999; Figure 5). The earliest glaciation occurred in west-central Yukon prior to 2.6 Ma (Froese et al. 2000; Duk-Rodkin and Barendregt 1997; Duk-Rodkin et al. 2001). This earliest glaciation was the most extensive and formed a continuous carapace of ice, leaving a small area of Dawson Range free of ice in west-central Yukon.

The chronology of Cordilleran glaciations for central Yukon was first established by Bostock (1966), who proposed a record of four successively less extensive glaciations from oldest to youngest named the Nansen, Klaza, Reid and McConnell glaciations. Nansen and Klaza glaciations were events that occurred during the early Pleistocene, while Reid is of middle Pleistocene age and the McConnell glaciation is late Pleistocene (Late Wisconsinan). Subsequent work was unsuccessful in recognizing the Nansen and Klaza events outside of the reference area, and we now know that the northern Cordillera was affected by more than four glacial events in the Mackenzie Mountains (Duk-Rodkin et al. 1996) and at least six events in the Tintina Trench (Duk-Rodkin and Barendregt 1997; Duk-Rodkin et al. 2001). The youngest two events, Reid and McConnell, have been maintained since they are useful and can be associated with mappable criteria for their recognition (surface expression and soil development principally), while the Nansen and Klaza events are grouped as pre-Reid. With the refinement of dating methods we can now say with confidence that the Reid is associated with marine isotope stage 8 (ca. 250 ka) based on the overlying Sheep Creek tephra and stage 7 interglacial material (Berger et al. 1996; Westgate et al. in press), while the McConnell is associated with marine isotope stage 2 (< 28 ka) (Matthews et al. 1990).

A correlative chronology to that of Bostock was established for the Southern Ogilvie Mountains by Vernon and Hughes (1966), whose Latest, Intermediate and Oldest glaciations are likely equivalent to the McConnell (Late Pleistocene), Reid (Middle Pleistocene) and pre-Reid (pre-Middle Pleistocene), respectively. The last two southwest Yukon glacial events were identified by Rampton (1969) as Macaulay Glaciation (Late Pleistocene) and Mirror Creek Glaciation (Middle Pleistocene).

The extent of glaciations in Yukon follows a gen-

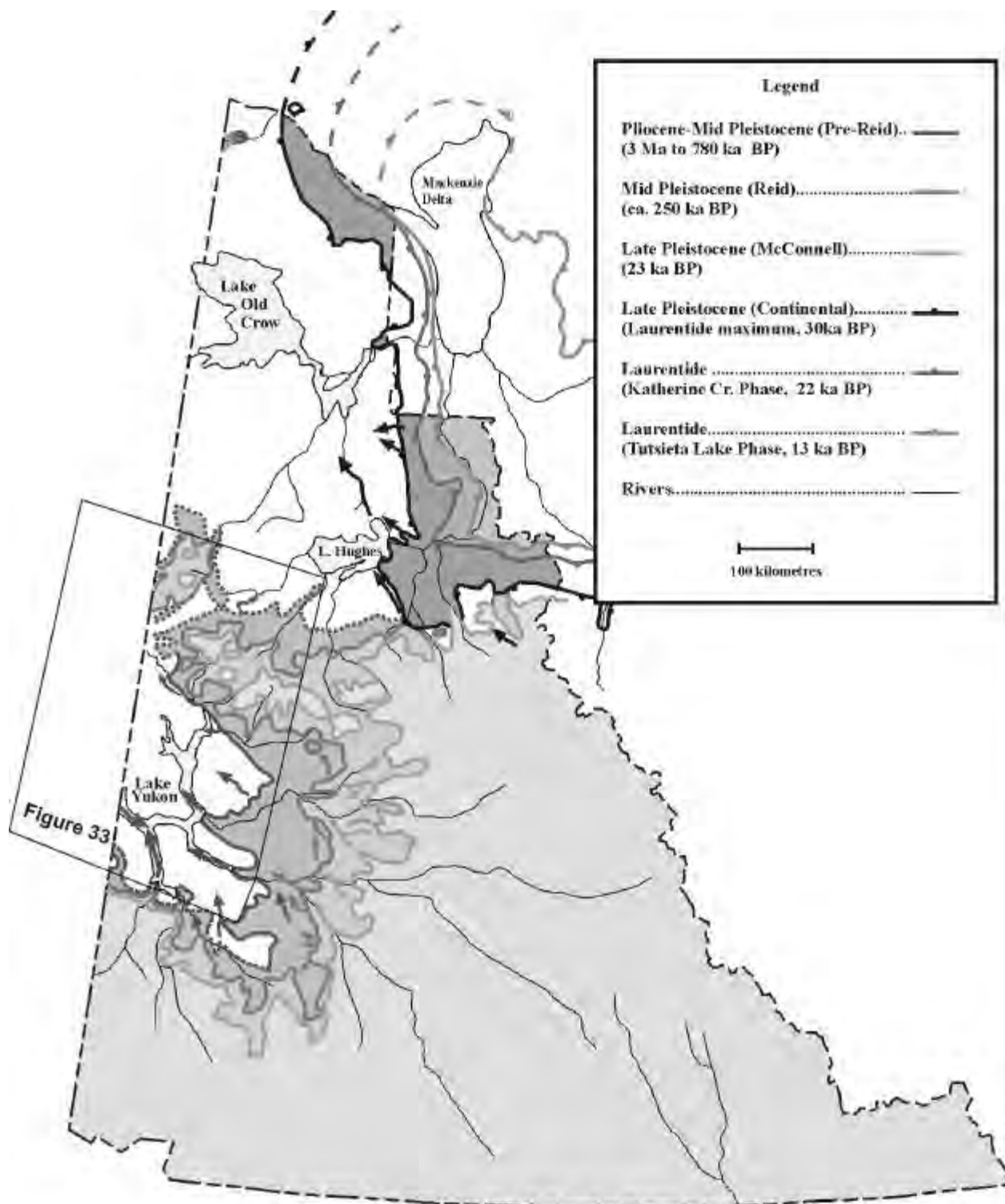


Figure 5. Extent of Plio-Pleistocene glaciations in Yukon.

eral pattern in which early Pleistocene glaciations were more extensive than Middle and Late Pleistocene glaciations. As mentioned above, this pattern is related to uplift of the southwestern coastal mountains in Yukon-Alaska and the corresponding lack of moisture in the interior plateaus. The pre-Middle Pleistocene glaciations, here grouped as one (Figure 5), formed a continuous carapace of ice from St. Elias Mountains to the continental divide and Southern Ogilvie Mountains. There were two locations where these glaciations appear as individual ice caps, the Northern Ogilvie Mountains and the British Mountains at the border with Alaska.

The extent of Middle Pleistocene glacial ice, namely Reid Glaciation, along the southern boundary comes close (about 2-3 kilometres) to the older glacial limits, while in west-central Yukon it reaches within 70 km, and within 12 km in Southern Ogilvie Mountains. Reid glaciers formed several coalescent ice caps in the Northern Ogilvie Mountains and formed a single glacier in Malcolm Valley of the British Mountains. Reid and pre-Reid glacial limits are marked by moraines and meltwater channels along the northern slopes of Wernecke Mountains. These glacial features were truncated by the Late Pleistocene (Wisconsinan) Laurentide ice sheet (Figure 5) and were subsequently cut across by Cordilleran valley glaciers during the latest Wisconsinan.

The extent of the Late Pleistocene glaciation (Wisconsinan) was further reduced forming a continuous carapace of ice between St. Elias Mountains and the continental divide. In the Southern and Northern Ogilvie Mountains several individual ice caps formed during this time. A similar development occurred in Malcolm Valley in the British Mountains.

The northern and northeastern Yukon Coastal Plain includes both glaciated and unglaciated terrain. The boundary of these two terrains is marked by the Late Wisconsinan and all-time limit of Laurentide Ice Sheet (ca. 30 ka BP; Hughes et al. 1981; Lemmen et al. 1994; Duk-Rodkin and Hughes 1995). The ice-sheet covered about 10% of Yukon. The limit follows the northeastern front of the Richardson, Barn and British mountains, descending westward towards Herschel Island. The ice sheet advanced across extensive late Cenozoic pediment surfaces which extend from the foothills to the coast, descending from approximately 850 m ASL at the east end of the coastal plain to almost sea level at the border with Alaska (Rampton 1982). Besides the Laurentide glacial maximum, there are two other well defined ice marginal positions in this part of Yukon: 1) Katherine

Creek Phase (ca. 22 ka BP; Duk-Rodkin and Hughes 1995; Lemmen et al. 1994), traceable from southern Mackenzie Mountain to the Coastal Plain; and 2) Tut-sieta Lake Phase (ca. 13 ka; Hughes 1987; Duk-Rodkin and Hughes 1995) which follows Peel River and the western edge of the Mackenzie Delta. The three former ice margins are considered correlative to Rampton's Buckland, Sabine phase and late Wisconsinan limit of glaciation respectively (Rampton 1982).

Major and minor drainage diversions occurred in northwest Canada because of glacial activity. The earliest is the diversion of the Yukon River, prior to 2.6 Ma, to the northwest into Alaska (Duk-Rodkin 1997; Duk-Rodkin et al. 2001; Froese et al. 2001). The largest drainage diversion, however, occurred during the Late Pleistocene (Wisconsinan) in the Mackenzie region, caused by the advance of the Laurentide Ice Sheet into the Mackenzie Valley. The ice sheet integrated the drainage of the eastern flanks of the northern Cordillera into one drainage system, the Mackenzie River (Duk-Rodkin and Hughes 1994; Lemmen et al. 1994). Western tributaries of the Mackenzie River from northern Yukon (Porcupine and Peel rivers) were diverted into the Old Crow basin causing formation of proglacial lakes Old Crow and Hughes. Overflow of Old Crow Lake into central Alaska followed by catastrophic drainage established the Porcupine River as a tributary to the Yukon River system (Thorson and Dixon 1983; Thorson 1989). Major sediment input is recorded in a subsiding basin (Yukon Flats) associated with the diversion of the Yukon River in the middle Pliocene (Froese et al. 2001) and the Porcupine River in the last continental glaciation. Overall, in preglacial time the drainage in northwest Canada exited in four directions: the Arctic Ocean (Peel and Porcupine rivers); Atlantic Ocean (palaeo-Mackenzie River); Pacific Ocean (Bering Strait: Kwikhpak River and Gulf of Alaska: palaeo-Yukon River). At present, the drainage of this area exits in only two directions: Pacific Ocean - Bering Strait (Yukon River) and Arctic Ocean (Mackenzie River). Overall, drainage has been changed by glaciation in over 95% of northwestern Canada.

Field Trip Route

The route of the field trip provides an opportunity to transect recent glacial features into progressively older mid-Pleistocene, early Pleistocene and unglaciated environments of western Yukon. The route is briefly described below:

Whitehorse to Lake Laberge

Whitehorse is located in the Yukon-Southern Lakes area dominated by glacial till, glaciofluvial gravel and glaciolacustrine sediments deposited during the last glaciation (McConnell). Ice flowed into the area from the Cassiar Mountains and the eastern Coast and St. Elias Mountains (Jackson et al. 1991). Trunk glaciers followed the major valleys and flowed northwestward across this region to terminate in central Yukon. The streamlined topography of this region was shaped by this flow. Glacial ice covered the lowland areas after 26 ka BP and was near its present distribution by 10 ka BP. Blockage of drainage and possible isostatic depression dammed extensive lakes in this area during deglaciation, such as Champagne, Whitehorse and Carcross. Lowland areas are underlain by extensive glaciolacustrine sediments as a result. During postglacial time, streams incised into the thick drift in the region, cutting steep sided canyons and leaving flights of terraces.

Lake Laberge to Tintina Trench

This area crosses from McConnell (to the south) to Reid terrain (to the north). McConnell glaciation features are visible along the road from Whitehorse until it crosses the McConnell limit five kilometres from Five Finger rapids (Stop 1), which is traceable along the right side of the road. McConnell glacial ice reached its maximum extent ca. 24 ka BP (Jackson and Harington 1991). Very sharp-edged features can be seen along the former ice-frontal position as well as nunataks further east. During Reid glaciation Cordilleran ice reached its maximum extent within the western part of this area, but features related to this glaciation are visible along the road all the way to the Tintina Trench. Well-defined glacial limits for this ice position can be traced by somewhat subdued moraines and meltwater channels.

Soils associated with the Reid surface (Diversion Creek) and McConnell (Stewart Neosol) developed during nonglacial periods (Smith et al. 1986). The Diversion Creek soil is a composite of up to three interglaciations (marine isotope stages 7, 5 and 1), while the Stewart Neosol (modern soil) is the product of the present interglaciation (Holocene- marine isotope stage 1). Commonly the Diversion Creek soils are buried under aeolian sands and loess that accumulated during the McConnell period.

Tintina Trench (Stewart Crossing) to Dawson City

The Klondike Highway follows the north side of

the Stewart River which drains through the middle of the Tintina Trench. Reid-age features are visible from Stewart Crossing to the confluence with McQuesten River, where the road crosses the Reid limit. Clear features of Reid age are found along the Stewart and McQuesten rivers where the type locality for the Diversion Creek palaeosol is found (Bond 1997; Smith et al. 1986), and to the south in a cut-bank of the Stewart River where the Sheep Creek tephra (190±20 ka; Berger et al. 1996) overlies Reid-age glacial material (Westgate et al. in press). The road turns to the northwest to follow the middle of the trench about 5 km northwest of the confluence of the Stewart and McQuesten rivers, and from here we enter an area dominated by pre-Reid glacial drift until close to Dawson City. This area between McQuesten and the Dempster Highway corner is called the Flat Creek beds (McConnell 1905; Bostock 1966) and consists of glacial outwash, till, loess and colluvial deposits. The Wounded Moose palaeosol (up to 2 m thick luvisol) is associated with these surfaces, which is characteristic of pre-Reid terrain. On its eastern extent, the Flat Creek beds likely contain deposits from more than one Cordilleran ice sheet advance from the southeast, while on their western extent Froese et al. (2000) argued that the outwash is part of the Klondike gravel, representing the first glaciation downstream in the Klondike valley (>2.6 Ma). Along the margins of the Flat Creek beds, where stream and rivers transect them, the deposits are extensively slumped.

Soils

C.A.S. Smith

Soils form at the earth's surface as the result of interactions among climate, geologic **parent material**, time, relief and living organisms. Soils in the Yukon have formed under a cold, semi-arid to moist subarctic climate on a range of geologic materials. The result is that most Yukon soils are only mildly chemically weathered, and many contain near-surface permafrost. Since much, but not all, of the Territory has been glaciated in the past, some soils have formed directly over local bedrock, whereas others have formed in glacial debris of mixed **lithology**. In mountainous terrain, soils form on a range of slope debris (colluvium) and are subject to ongoing mass wasting and erosion.

Within the Canadian System of Soil Classification, the most common soil orders in the Yukon are the Brunisols (mildly weathered forest soils), the Regosols (unweathered alluvial and slope deposits), and the Cryosols (soils underlain by near-surface permafrost). Each of these "soil orders" is

Soil Order	Occurrence	Description
Brunisol	Very common in Boreal Cordilleran Ecozone	Mildly weathered mineral soil, commonly forms under forest cover and grasslands in southwest and central Yukon. The most common subgroup of Brunisol in the Yukon is the Eutric Brunisol, which has a pH in the surface soil of >5.5. Dystric Brunisols are less common, acidic forest soils with pH<5.5
Cryosol	Very common in all northern ecozones	Permafrost-affected soils, may be associated with wetlands, tundra or taiga forest conditions. Turbic Cryosols are mineral soils strongly affected by frost churning, which generates various forms of patterned ground. Static Cryosols lack this frost churning process. Organic Cryosols are the soils of peatlands underlain by permafrost
Regosol	Scattered throughout all ecozones	Regosols are soils that have not been weathered and are associated with active landforms such as floodplains, colluvial slopes, and dunes, thaw slumps and debris flows. The soils do not exhibit weathering or horizon formation typical of other soils.
Luvisol	Restricted to regions in southeastern Yukon	Luvisols are the soils associated with fine-textured soils under boreal and temperate forests throughout Canada. In the Yukon, they only develop at lower elevations on clay-rich glacial deposits under relatively mild and wet conditions such as are found the Liard Basin, Hyland Highland and Muskawa Plateau regions.
Organic	Scattered wetland soils of Boreal Cordilleran Ecozone	In soil taxonomic terms, Organic refers to soils that are formed of decomposed vegetation (peat) rather than sand, silt and clay. Organics are associated with fen wetlands that are not underlain by permafrost.
Podzol	Rare	Podzols are associated with temperate, high rainfall forested areas. In the Yukon, they are occasionally found in Selwyn Mountains and Yukon Stikine Highlands regions.

Table 1. Simplified descriptions of major soil orders and subgroups in the Yukon. For more detailed definitions of these soils, see Canadian System of Soil Classification.

associated with a specific environment created by the soil-forming factors listed above. The distribution of major soil groups is shown in Figure 6 and a brief description of each soil order is given in Table 1. For more detailed treatments of the material presented here the reader is referred to the following: Tarnocai et al. (1993), White et al. (1992), Rostad et al. (1977), Smith et al. (1986) and Davies et al. (1983).

Those areas of the southern Yukon that lie in the

rainshadow of the St. Elias Mountains, such as the Ruby Ranges, Yukon Southern Lakes and Yukon Plateau-Central, are dominated by soils formed under a semi-arid climate on **calcareous** glacial parent materials. These soils tend to be alkaline and belong to the Eutric Brunisol great group of soils. These soils support mixed forests of aspen, pine and spruce, while grassland ecosystems appear on south facing slopes. Milder temperatures, higher precipitation and finer-textured parent materials in the valleys and basins of the Liard Basin and Hyland Highland areas of the southeast Yukon result in soils containing subsurface **clay** accumulations. These belong to the Grey Luvisol great group of soils. Where parent materials are coarse textured, Eutric Brunisols are formed.

In the main ranges of the Selwyn Mountains along the Northwest Territories border, and at higher elevations of the Pelly Mountains along the British Columbia border, high precipitation causes strong **leaching** of the soil. This results in the formation of Dystric Brunisols, i.e. forested soils with acidic soil horizons. These soils sup-

port extensive **conifer** forests composed of **alpine** fir, spruce and pine. Occasionally, in very coarse textured parent materials without any calcareous mineralogy, Humo Ferric Podzols may form. These have been reported in the Mackenzie Mountain and Selwyn Mountain regions under subalpine forest conditions. They may also occur sporadically in alpine environments in the Yukon-Stikine Highlands immediately adjacent to the Yukon/BC border.

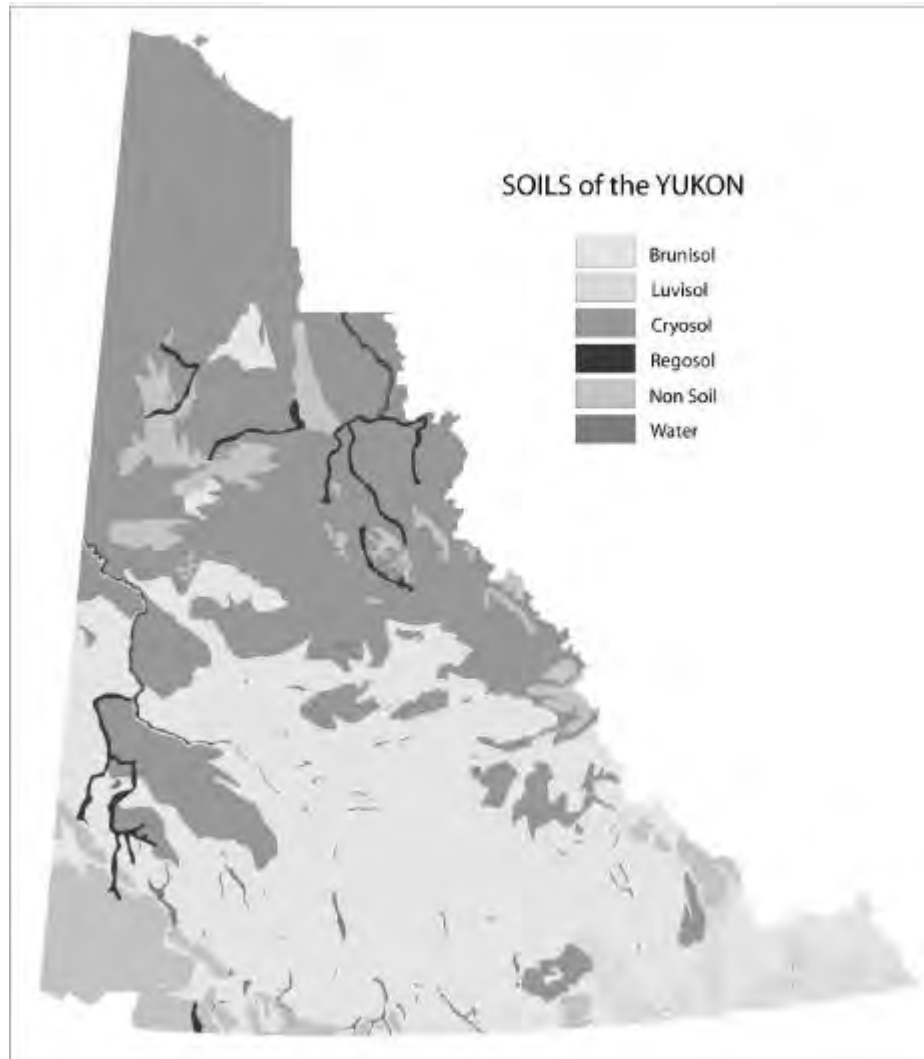


Figure 6: Distribution of major soil groups in Yukon.

Permafrost is widespread and discontinuous throughout the Central Yukon-North and Klondike Plateau regions. While many well-drained **upland soils** are free of permafrost and classify as mildly weathered Eutric Brunisols, poorly drained areas and north facing slopes are usually underlain by near-surface (i.e. active layer < 2m depth) permafrost. These often show evidence of **frost churning** in the upper **soil horizons**. These soils are classified as Turbic Cryosols. Open forests of paper birch and black spruce are associated with Cryosols in these areas. The Ogilvie, Selwyn and British-Richardson Mountains are characterized by rugged landscapes of **colluvium** and bedrock outcrops dissected by major valley systems. Here the upper slopes are composed of rubble, **scree** and bedrock. In unglaciated valleys, lower

pediment slopes are composed of fine-**textured** silt and clay and almost always underlain by permafrost. These soils are Orthic or Gleysolic Turbic Cryosols depending on the degree of saturation.

The alluvial parent materials deposited on river **floodplains** are often **braided** and may be scoured by the formation of auefis in northern ecoregions. There is generally little soil formation or permafrost on these active floodplains throughout the territory. The soils are usually classified as Orthic Regosols or Cummulic Regosols if there is evidence of multiple deposits an buried vegetative debris.

Many valleys have higher-level glaciofluvial terraces. Terrace surfaces are not subject to the same

erosional and depositional forces as those on active floodplains and soil development usually produces Brunisolic soils. In the Yukon Plateau-Central and Klondike Plateau areas, there are glaciofluvial terraces that formed during early pre-Reid glaciations, some dating back to over 2.5 million years. Some of these glaciofluvial surfaces have been subjected to weathering throughout most of the Pleistocene and the resultant soils are termed palaeosols. These palaeosols reflect both warm, almost temperate, weathering during interglacial periods and intense cold and periglacial weathering processes during subsequent and intervening glacial periods. One of these palaeosols (designated the Wounded Moose palaeosol after the Wounded Moose Dome where it was best preserved), associated with the oldest drift surfaces, exhibits deep soil horizon development and sand wedge formation, the latter associated with polar desert conditions. The soils do not fit the Canadian soil classification system for modern soils very well but have been termed palaeo-Luvisols because of the clayey nature of the subsurface horizons.

The areas that cover the large plateaus and plains of the northern Yukon are underlain by continuous permafrost and dominated by Cryosols. The unglaciated Old Crow Basin, Old Crow Flats and Eagle Plains regions have extensive permafrost with open stands of black spruce, birch and occasional larch. Turbic Cryosols and lacustrine parent materials dominate the pediment. Peat deposits underlain by permafrost, called Organic Cryosols, are common in the Old Crow Flats. The Peel Plateau area has been glaciated but the soil formation (Cryosolic) and vegetation cover are similar to adjacent unglaciated ecoregions.

Arctic tundra landscapes exhibiting extensive patterned ground characterize the Yukon Coastal Plain area. The soils all belong to the Cryosolic Order. Where surfaces show evidence of frost churning, soils are classified as Turbic Cryosols. On recently disturbed surfaces, such as alluvium, thaw slumps, or dunes where no churning is evident, soils are classified as Static Cryosols. Soils of lowland polygons may be underlain by perennially frozen peat and are classified as Organic Cryosols. In the western portion of this area, the unglaciated plain includes large fluvial fan formations composed of sandy soils that have active layers too thick to allow these soils to be classified as Cryosols. In these cases the soils are classified as Regosols.

The Mount Logan and St. Elias Icefields contain vast areas without soil formation or vegetation cover. Here high-elevation icefields and rock summits domi-

nate the landscape.

Quaternary Palaeoecology of the Yukon

C. Schweger

Events long ago and far away have determined much of the Quaternary history of the Yukon. Eight million years ago the northward traveling Indian plate impacted the southern edge of Asia starting a chain of events that were to have great consequence. The collision thrust up the highest landmass on earth, the Himalaya-Tibetan plateau, which influenced atmospheric circulation in the Northern Hemisphere (Zhisheng et al. 2001). On a grander scale, irregularities in the earth's orbit around the sun and the way the earth wobbles as it rotates on its axis result in periods when the earth receives more or less insolation (solar radiation) than usual (Berger et al. 1984). Emergence of the Isthmus of Panama and uplift of mountains such as Yukon's St. Elias Range further changed atmospheric and oceanic circulation (Haug and Tiedemann 1998). These changes interacted with the cycles of insolation resulting in periods of cold climate and the onset of Northern Hemisphere glaciations, the earliest of which dates to about 2.65 million years ago. Shortly after this date we see evidence for glaciation in the Yukon (Froese et al. 2000; Duk-Rodkin et al. 2001). From this point on, the earth's climate has cycled from glacial conditions when massive ice caps formed over northern Europe and North America and glaciers expanded down mountain slopes, to interglacial times very much like now with warm climate and restricted ice cover. The past million years has seen periods of glaciation about every 100 thousand years; the last, the Wisconsinan, is named after glacial deposits first described in the state of Wisconsin. The late Wisconsinan glaciation, 25 to 10 thousand years ago, saw ice cover virtually all of Canada from Atlantic to Pacific, and extend nearly as far south as St. Louis, Missouri. With so much ice on land the sea level dropped by 160 metres exposing large areas of continental shelf.

In Yukon, late Wisconsinan ice spread from centres of accumulation, the Mackenzie Mountains on the eastern border, the Cordillera (i.e., St. Elias, Kluane, Pelly and Logan mountains) in the south and the Ogilvie Mountains further north (Duk-Rodkin 1999). But glacial ice did not cover all of Yukon. Northern and west central Yukon were ice free, as was much of interior and northern Alaska, and with lower sea level, the exposed shelves of the Bering and Chukchi seas.

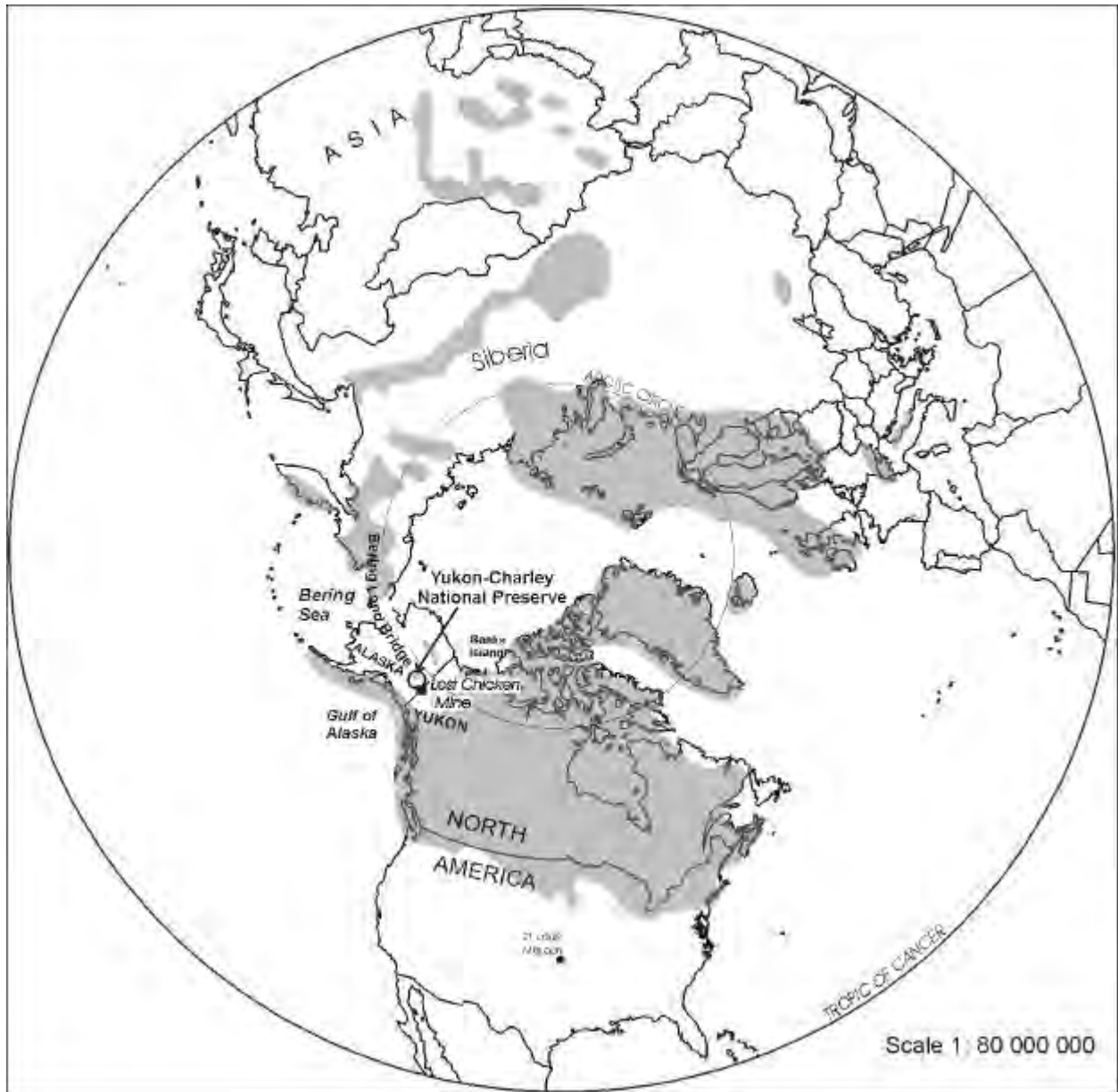


Figure 7. Circumpolar projection and location of Beringia and the distribution of Late Wisconsinan ice sheets (from various sources).

These exposed shelves formed the Bering Land Bridge, which connected western Alaska to ice free regions of eastern Siberia. This region, from eastern Siberia to central Yukon is known as Beringia (Figure 7). During glacial times, if you reflect on it, Yukon was actually part of far eastern Asia, cut off from the rest of North America by glacial ice (Figures 7 and 8). On the other hand, during interglacial periods water from the melted ice filled the oceans, sea level rose, the Bering Land Bridge flooded and became seabed again, and Yukon and Alaska became part of North

America. This change of geography no doubt took place multiple times during the Quaternary.

For clarification, the Quaternary is a period of geological time subdivided into the Pleistocene and Holocene epochs. The Pleistocene began 1.7 million years ago, although some researchers would place the beginning at the first evidence of glaciation, 2.65 million years ago. The Holocene, or postglacial, began about 10,000 years ago. We live in the Holocene; it is our interglacial.

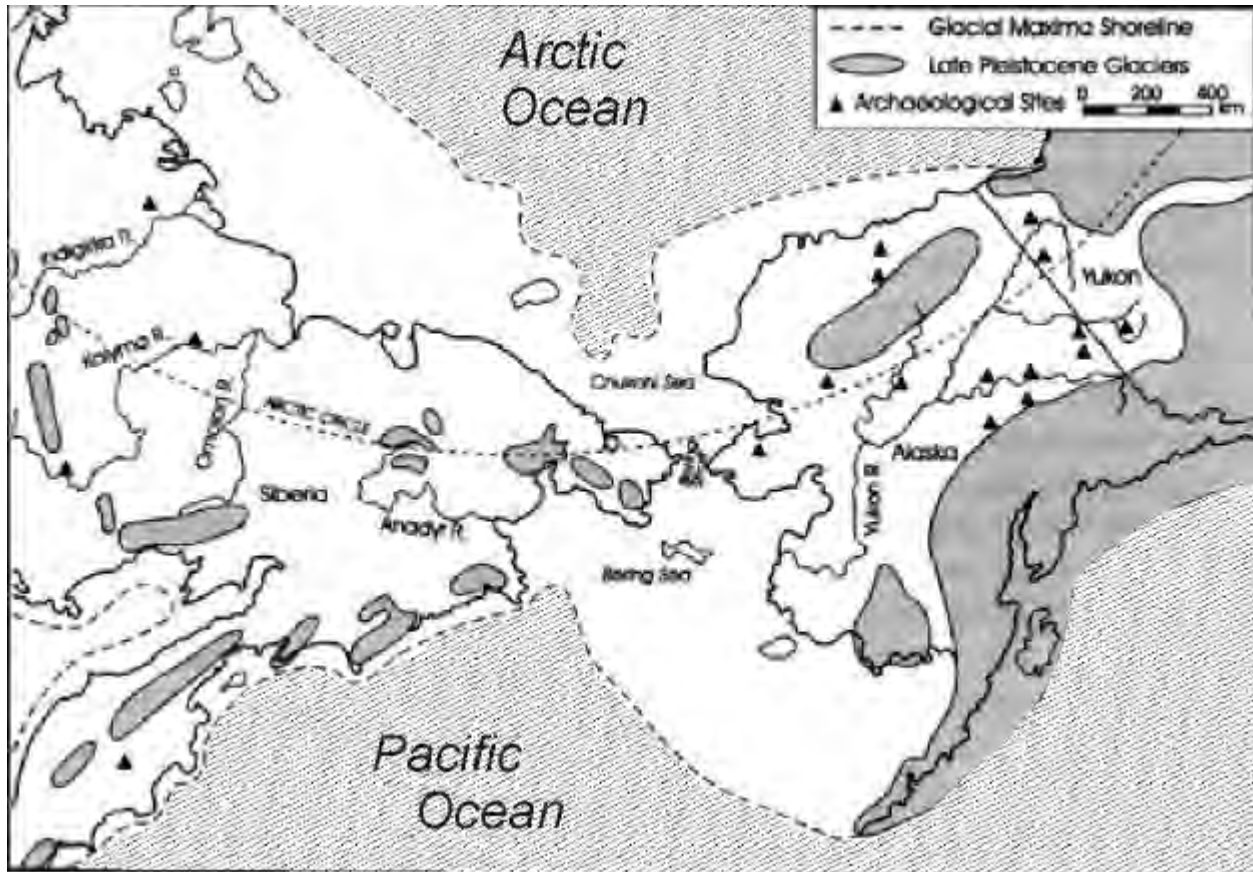


Figure 8. Beringia during Late Wisconsinan time (modified from Morlan 1997).

The Yukon portion of Beringia, sometimes called the Yukon refugium, is a special place, Canada's window on the Pleistocene ice ages (Figure 8). Here one can see the most deeply weathered soils of Canada complete with polygenetic characteristics and sand wedges, there are pediments, deposits of miners' "muck" and loess, and perhaps most importantly abundant fossils of the megafauna that draw attention to the extraordinary changes that have occurred over the past. For over a century, tons of bones have been exposed through gold mining operations, particularly in the Dawson City area. Recent controlled excavations at Bluefish Cave, northern Yukon, have revealed a similar fauna of woolly mammoth, horse, bison, caribou, musk-ox, mountain sheep and saiga antelope dated 25 to 10 thousand years ago (Cinq-Mars 1990). The megafauna, mostly grazers, are at the centre of any discussion about the Pleistocene environment of the Yukon. There were also spectacular predators such as the short-faced bear and lion. We know this fauna was present during the last glaciation, which raises the question of how such animals, a mammoth,

for example, could have survived glacial conditions for even at the present time subarctic Yukon with its boreal forest and tundra could not support the megafauna. There just aren't enough nutrients produced for them to succeed. The solution to this "production paradox" (Hopkins et al. 1982) requires that the ice age vegetation be reconstructed through the recovery and identification of a variety of fossil material.

Plant and insect remains have been recovered from screened Pleistocene sediments and researchers have cored lakes in order to recover fossil pollen trapped in the lake sediments. Antifreeze Pond, south west Yukon (Rampton 1971), displays fossil pollen assemblages dominated by grass (Poaceae), sedge (Cyperaceae), sage (*Artemisia*) and other herbs but not trees. Many such studies support the conclusion that during glacials Yukon was treeless and vegetated by what has been called steppe-tundra, arctic-steppe or herbaceous tundra. Steppe refers to cold grassland vegetation with a variety of herbs and *Artemisia*. The latter is a key indicator of modern steppes, such as

those found on the northern Plains or central Asia, which are excellent at cycling nutrients and support large grazing animals, including bison, horse and saiga. Since the same Pleistocene megafauna, dominated by woolly mammoth, existed westward across Beringia, northern Asia and into Europe, and is associated with similar *Artemisia* and grass rich pollen records, R.D. Guthrie (1990) has hypothesized the existence of a mammoth-steppe biome, without modern analogue, that dominated much of the Northern Hemisphere during glacial periods. Yukon was not only geographically part of Asia, the mammoth-steppe links Yukon to Asia ecologically as well. This hypothesis is strengthened by the presence of fossils of saiga antelope, an animal presently found only on the arid steppes of central Asia where winter snow depth does not exceed 10 cm (Banikov et al. 1967; Harington and Cinq-Mars 1995).

The mammoth-steppe hypothesis has been challenged by researchers who emphasized other characteristics of the fossil pollen record. It has been demonstrated that pollen production, known as influx (#grains/cm²/yr) and a measure of the amount of vegetation cover that existed in the past, was very low and that the pollen flora included tundra species (Cwynar and Ritchie 1980). A discontinuous, herbaceous tundra, likely unable to support megafauna, has been hypothesized. With the hope of discovering a modern analogue, fossil pollen assemblages have been statistically compared with large pollen data sets derived from different types of vegetation. A match suggests modern vegetation that is analogous to the fossil vegetation. With fossil *Artemisia* and grass pollen eliminated from the comparisons, a match with pollen assemblages derived from modern herbaceous tundra of northern Alaska and Banks Island can be made (Anderson et al. 1989). What to do about the *Artemisia* and grass so common in the fossil pollen assemblages remains a problem. The obvious solution is to collect and identify macrobotanical fossils which compared to pollen have finer taxonomic resolution and limited dispersal (Birks 2000). Peat from the Bering Land Bridge, encountered in marine cores (Elias et al. 1996), and plant remains excavated from a buried soil surface in western Alaska (Goetcheus and Birks 2001) have yielded glimpses of full glacial vegetation but without *Artemisia*. However, ongoing research in northern Yukon has recovered great quantities of *Artemisia* flowers and grass seeds (G. Zazula, University of Alberta) dated to the late Wisconsinan, keeping the controversial issue of non-analogous steppe vegetation in full glacial Yukon unresolved.

Since there were no weather stations in Beringia

what we know about the climate comes from the study of proxy, or substitute indicators. Insect fossils, beetles in particular, serve as fine proxy indicators of Beringian palaeotemperatures. It seems that summer temperatures were about 6.4°C cooler than modern for the last glacial period and 3.5°C warmer than modern for the last interglacial, 125 thousand years ago (Elias 2001). Precipitation is the other major component of the climate and by all indications the climate of Beringia in general and Yukon in particular was hyper-arid. Remember the saiga antelope? Its modern ecology points to very low winter snow depth for the areas it is known to have occupied during the full glacial. Comparisons between pollen records of northwestern Alaska and those from the Yukon suggests a west to east decline in precipitation values (Barnosky et al. 1987) while lake coring indicates that many lakes in the Yukon refugium were dry during the last glacial. Loess, or windblown dust derived from glacier meltwater streams, was another unique component in the environment. Its influence was most strongly felt in the type of soil that formed during glacial periods. The predominantly mineral soil had higher pH values, little organic matter, deeper permafrost depths and was likely more productive for grazing animals (Laxton et al. 1996) than at present. For comparison, interglacial soils are more acidic and have a thick layer of accumulated organic matter that serves as insulation allowing for colder soil temperatures and high permafrost tables. While organic production may be high it does not cycle within the ecosystem but is sequestered as accumulating organic material often frozen in permafrost (Schweger 1997). If you were a large grazer it appears that glacial periods were more suitable than interglacials.

The recession of glaciers and spread of birch shrub tundra between 14 and 10 thousand years ago were in response to increased warmth and precipitation, and define the late glacial. These changes are implicated in the disappearance of the mammoth steppe and the extinction of the megafauna (Guthrie 1984). Warmer climates of the early Holocene saw numerous plants arrive in the Yukon, their migrations from refugia documented in ¹⁴C dated pollen records. In southern Yukon, alder and birch increased simultaneously before the appearance of spruce. Elsewhere in the Yukon, the appearance of spruce (probably *Picea glauca*) at about 8 thousand years ago preceded that of alder, 7 to 6 thousand years ago. The mid Holocene saw the expansion of black spruce (*Picea mariana*) and green alder (*Alnus crispa*) over southern and central Yukon in response to cooler, more mesic climate, and the development of peatlands and permafrost (Cwynar and Spear 1995). Lodgepole pine

(*Pinus contorta*), an important Yukon forest species, first appeared in southern Yukon about one thousand years ago and migrated to its present limit south of Dawson City perhaps only a few hundred years ago (Schweger et al. 1987). One wonders if pine is still migrating and given enough time if it would spread further, especially under the influence of fire.

Field and laboratory research has documented eight different interglacials before the Holocene. These have been dated by palaeomagnetic correlation (based on changes to the earth's magnetic poles) and tephrochronology (dating of volcanic ash) as early as 2.3 million years and as recent as 125 thousand years ago (Schweger et al. n.d.). We know from pollen and macrofossil studies that boreal forests were well developed during these warm periods. Warmer than present interglacials saw dense spruce-fir (*Abies lasiocarpa*) forests spread as far north as Old Crow in northern Yukon. Cattails and insects now found in central British Columbia were present in northern Yukon during the last interglacial (Matthews et al. 1990). The Holocene stands out as the most unique of the Yukon interglacials probably because it is the only interglacial without the ecological influence of the megafauna and with the ecological influence of people who may have arrived in the Yukon as early as 25 thousand years ago but were certainly present during the Holocene. There is a very practical application to the study of interglacials as they provide a picture of what the future of the Yukon might be given the impact of predicted global warming.

Over the past 3 million years climatic changes have again and again transformed Yukon from dense forest to open forest and shrub tundra to herbaceous tundra or mammoth steppe and then back to interglacial forest. Now very much part of North America, at other times Yukon was geographically and ecologically part of Asia. As the easternmost province of the mammoth-steppe it supported a spectacular megafauna dominated by the woolly mammoth. When you travel through Yukon try to imagine what it would have been like 20 thousand years ago, when the short-faced bear would have you nervously looking over your shoulder as you walked across a dry and dusty tundra-steppe, or what it was like to come across a giant beaver feeding in a lake on a hot humid day surrounded by a dense forest of spruce and fir. In Yukon nothing is the same for very long.

Vertebrate Palaeontology of the Dawson City Area

J. Storer

The latest Pleistocene (late Wisconsinan) vertebrate fauna of the Klondike Goldfields is well known, at least in regard to the larger mammals. Nearly all these mammals have been discovered during placer mining activities, as miners have exposed Plio-Pleistocene gold-bearing gravels at various stratigraphic levels by stripping away a chiefly loessal overburden. The fine-grained deposits that are most widely distributed and produce most of the fossils date from the last (McConnell) glacial maximum about 27,000 to about 14,000 radiocarbon years BP (Kotler and Burn 2000), with a younger upper layer representing the end of the Pleistocene and a sporadically exposed lower layer perhaps dating to the Reid-McConnell interglacial period. Older fossiliferous deposits have been reported at Lost Chicken Mine in Alaska (Late Pliocene; Matthews and Ovenden 1990); Fifteenmile Creek northwest of Dawson City (Paleocene; Skwara and Kurtz 1988); and Revenue Creek, south of the Klondike (pre-McConnell; Jackson et al. 1996).

The content of the later beds, which span the advance and maximum extent of the last (McConnell, or late Wisconsinan) glaciers, is consistent across the Klondike. Steppe bison (*Bison priscus*, formerly called *Bison crassicornis*) is the most commonly preserved fossil mammal in the Klondike, as it is in the Old Crow section of Northern Yukon and the Fairbanks area of Alaska. A larger bison with flattened horn cores, *Bison alaskensis*, may also be common in these deposits, but most of today's workers assign all northern Ice Age bison to *B. priscus*.

Mammoth is not quite as common as bison, but it too is ubiquitous. The prevalence of grass-eating woolly mammoth over American mastodon in these deposits is generally taken to confirm the wide extent of treeless "Mammoth Steppe" vegetation, dominated by grasses, sedges, sage, and in some areas chenopods. The overall ratio of identified woolly mammoth to mastodon fossils is at least 100:1 in the Yukon. American mastodon, known in other parts of the continent to have inhabited moister habitat, eating woody and soft vegetation, was presumably excluded from the steppe-tundra. A recent radiocarbon date of $18,460 \pm 350$ on American mastodon from Gold Run Creek (YTG specimen; IsoTrace Lab number TO-7745), however, points out not only that mammoth

and mastodon were contemporaries throughout the Late Pleistocene in this area, but also that islands of mesic, presumably moist habitat persisted in the Klondike through the last glacial maximum, even during the time of greatest dominance of the "Mammoth Steppe."

The third most common Late Pleistocene mammal of this area, the small horse *Equus lambei*, has a less even distribution. Its bones and teeth are far less numerous in some placers than in others, although it is seldom missing altogether. A frozen carcass of this small horse was discovered in 1993, at 15 pup, Last Chance Creek. This fossil gave a first look at the animal's coat colour and mane length (not as similar to Przewalski's horse as expected), and yielded DNA samples clarifying the species' closer relationship to horses than to asses or zebras (Harington and Eggleston-Stott 1996).

Other large herbivores of the Klondike are caribou (*Rangifer tarandus*), helmeted musk-ox (*Bootherium bombifrons*) and tundra musk-ox (*Ovibos moschatus*), and mountain sheep (*Ovis* cf. *O. dalli*). Mountain sheep is one of the few large mammals known from the Klondike that are not found at Old Crow (Harington 1978), indicating some differences in habitat between the two areas.

Moose (*Alces alces*) and wapiti (*Cervus elaphus*) are present at many localities in the Dawson City area, but dated specimens are all latest Pleistocene or later. Saiga antelope (*Saiga tatarica*) is known from nearby Alaska (Porter 1988) and we may now have a specimen from Dominion Creek, but the identification is not yet firm. The western camel (*Camelops hesternus*) has been documented from Sixtymile River (Harington 1978, 1997), but is otherwise represented by only a YTG specimen from Hunker Creek dated at 19,770±640 BP (IsoTrace Lab number TO-7740).

Large carnivores are also well represented in the Last Glacial Maximum of the Klondike. The most commonly discovered in my experience is the lion *Panthera leo atrox*, and some excellent specimens have been discovered (Harington and Clulow 1973). Also known from several localities is the largest Ice Age carnivore, the giant short-faced bear (*Arctodus simus*). Wolf (*Canis lupus*) has a scattered record, and the scimitar cat (*Homotherium serum*) is a very rare fossil in the Klondike.

Smaller mammals are far less well known, chiefly because we have yet to use screen-washing and other special techniques that might produce them in quan-

tity. Burrows and occasional frozen carcasses of the Arctic ground squirrel (*Spermophilus parryi*) are widespread, and collared lemming (*Dicrostonyx torquatus*), badger (*Taxidea taxus*), hare (*Lepus* sp.), and in one instance even black-footed ferret (*Mustela nigripes*) have been found.

A special feature of these permafrost deposits is the frequent preservation of soft tissues including fragments of muscle on bison and horse bones, and preservation of bison horn sheaths. More spectacular and rare is preservation of partial or even whole carcasses, and although we cannot match the Fairbanks bison specimen called "Blue Babe" (Guthrie 1990), or the frozen mammoths of Siberia and Alaska, we do have horse (*Equus lambei*), Arctic ground squirrel (*Spermophilus parryi*), and black-footed ferret (*Mustela nigripes*) from the Klondike.

Older fossils from this area are less well known now. Harington (1989) has reported American lion, steppe bison, and woolly mammoth from presumed interglacial beds at Revenue Creek, northwest of Carmacks (Jackson et al. 1996). A single skull of *Equus* cf. *E. verae* came from Late Pliocene deposits at Lost Chicken Mine (J.V. Matthews, pers. comm 1999). The Paleocene deposits at Fifteenmile Creek produced a fossil bird (Skwara and Kurtz 1988). An emerging fauna from the "cover sands" at Midnight Dome, beneath the Mosquito Gulch Tephra and thus more than 1.5 million years old, includes an ancient vole (*Allophaiomys* sp.) and an ancestral lemming (*Predicrostonyx* sp.).

Prospects for extending this fossil record are excellent, as you will see on this field trip. Several beds described and dated by Froese, Westgate and others (Froese et al. 2000) will produce Plio-Pleistocene assemblages, and similar discoveries by Lionel Jackson and associates have already produced some voles. Current research in lithostratigraphy and tephrochronology will produce opportunities to make significant extensions to the Klondike fossil record in the next few years.

Late Cenozoic Tephrochronology of The Klondike Area, Yukon

J. Westgate and S. Preece

Tephra beds possess many of the attributes of index fossils, and like them, facilitate accurate correlation of strata over long distances, which, in some cases, are on a continental scale. They can be used to

establish relative stratigraphic sequences, as with fossils, but also temporally calibrated sequences, because they can be dated by several radioisotopic methods. Characterization and dating are best done by grain-specific methods with high spatial resolution because tephra beds - especially distal occurrences - are typically contaminated, rendering bulk analyses suspect. The essential toolkit of a tephrochronologist, therefore, includes an electron microprobe for major-element analyses, a laser ablation inductively-coupled plasma mass spectrometer for trace-element analyses, and $^{40}\text{Ar}/^{39}\text{Ar}$ (feldspar) or fission-track (glass, zircon) systems for dating, preferably augmented by palaeomagnetic data.

The Klondike region lies within the fall-out zone of large-magnitude volcanic eruptions from the Aleutian arc - Alaska Peninsula (AAP) and Wrangell volcanic field (WVF) so that its late Cenozoic sediments contain numerous silicic, distal tephra beds (Figure 9). Glass morphology, mineral content, and geochemistry of each tephra bed clearly reveal its provenance. For example, tephra beds from AAP (type I) have few crystals, mainly bubble-wall glass shards, abundant pyroxene, and rare-earth-element (REE) profiles with a well developed Eu anomaly. In contrast, tephra beds from WVF (type II) have many crystals, the glass is mainly in the form of highly inflated pumice, hornblende is abundant, and REE profiles are steep with a weakly developed Eu anomaly (Figures 10 and 11).

Twenty distinctive tephra beds have been identified in the Klondike area, although many samples remain to be studied. Their location and identity are shown in Figure 12. Each is readily identified by its physical and chemical properties (Figures 13 and 14). These tephra beds include, in order of increasing age (mostly by glass-fission-track dating method, Ma): White River Ash (1-2 ka), Dawson (c. 20 ka), VT (0.13±0.03), Old Crow (0.14±0.01), Dominion Creek (0.17±0.02), Sheep Creek (0.19±0.02), Midnight Dome (1.09±0.18), Flat Creek (1.23±0.18), Mosquito Gulch (1.45±0.14), Paradise Hill (1.54±0.13), Quartz Creek (2.97±0.24), and Dago Hill (3.18±0.41). Underlined beds come from the WVF. Important conclusions are that the gold-bearing Upper White Channel gravel is of late Pliocene age, as is the first extensive Cordilleran glaciation, and that prominent glaciers existed in the Ogilvie Mountains at 1.5 Ma.

Surprisingly, only four of these tephra beds occur in the detailed tephrostratigraphical sequence at Fairbanks, central Alaska: VT, Old Crow, Sheep Creek, and Mosquito Gulch tephra. However, better integration of the Fairbanks and Klondike tephrostratigraphi-



Figure 9. Location of the Klondike district in relation to the Aleutian - Alaska Peninsula arc, the Wrangell volcanic field, and the distribution of the widespread Old Crow and Sheep Creek tephra beds.

cal records will likely come from investigations in progress on the several sites with thick loess and multiple tephra beds that occur along the Yukon River in and near the Yukon-Charley National Preserve, Alaska (Figure 7).

Placer Deposits in Yukon

W.P. LeBarge and G. Lowey

Yukon placer deposits have been actively exploited for over one hundred years. Mining methods have evolved from hand-mining in the 19th century to modern mechanized methods in the 21st century. The Yukon placer gold mining industry continues to be a major contributor to the Yukon's economy as it has been since the Klondike Gold Rush of 1898. Today, most placer mining operations are family-owned and operated, and direct seasonal employment has varied between 500 and 700 in the last 15 years, not including significant employment spinoffs in the hospitality and support industries.

Since the first recorded gold mining in 1886, over 16 million crude ounces of placer gold have been produced. In 2000, 76,507 crude ounces worth \$25.3 million CDN were mined. Yukon placer deposits range from Pliocene to Holocene in age and occur in both glaciated and unglaciated areas. Aside from these major geographic subdivisions, most placer deposits are fluvial in origin and occur within a number of different stratigraphic and geomorphic settings. Approximately 85% of past and current production of placer gold has been derived from the unglaciated ar-

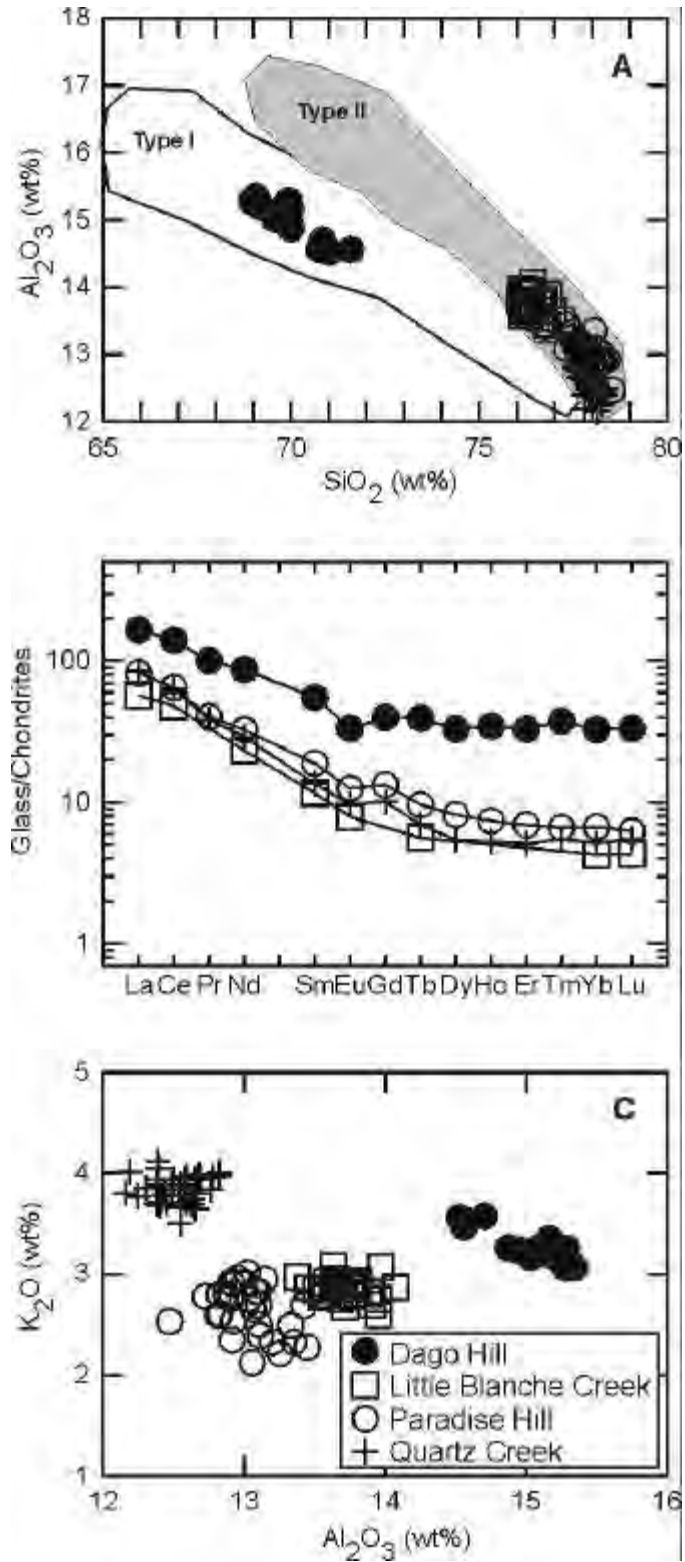


Figure 10. Major- and rare-earth-element composition of glass shards in Dago Hill, Paradise Hill, Little Blanche Creek, and Quartz Creek tephra beds. A: Classification of tephra beds into type I and type II groups of Preece et al. (1999). B: Rare-earth element profiles. C: Oxide variation plot showing good separation of the four tephra beds on the basis of their K and Al contents.

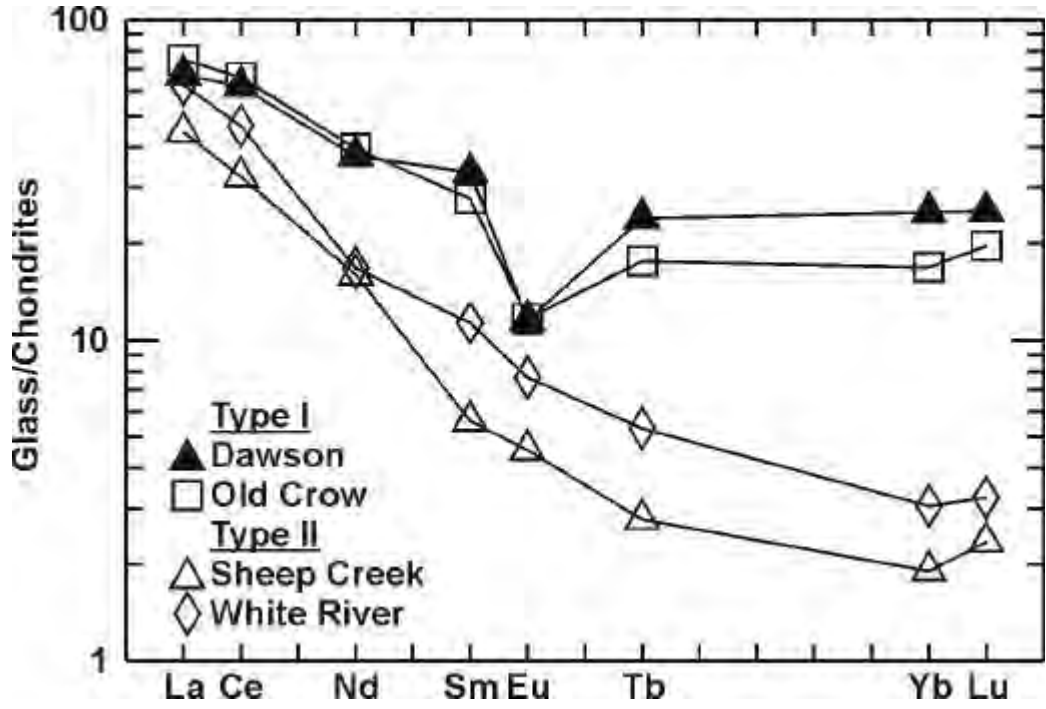


Figure 11. Chondrite-normalized rare-earth element profiles of tephra beds in the Klondike district, Yukon. Tephra beds have been grouped into two classes, type I and type II. Glass analyses by the instrumental neutron activation method.

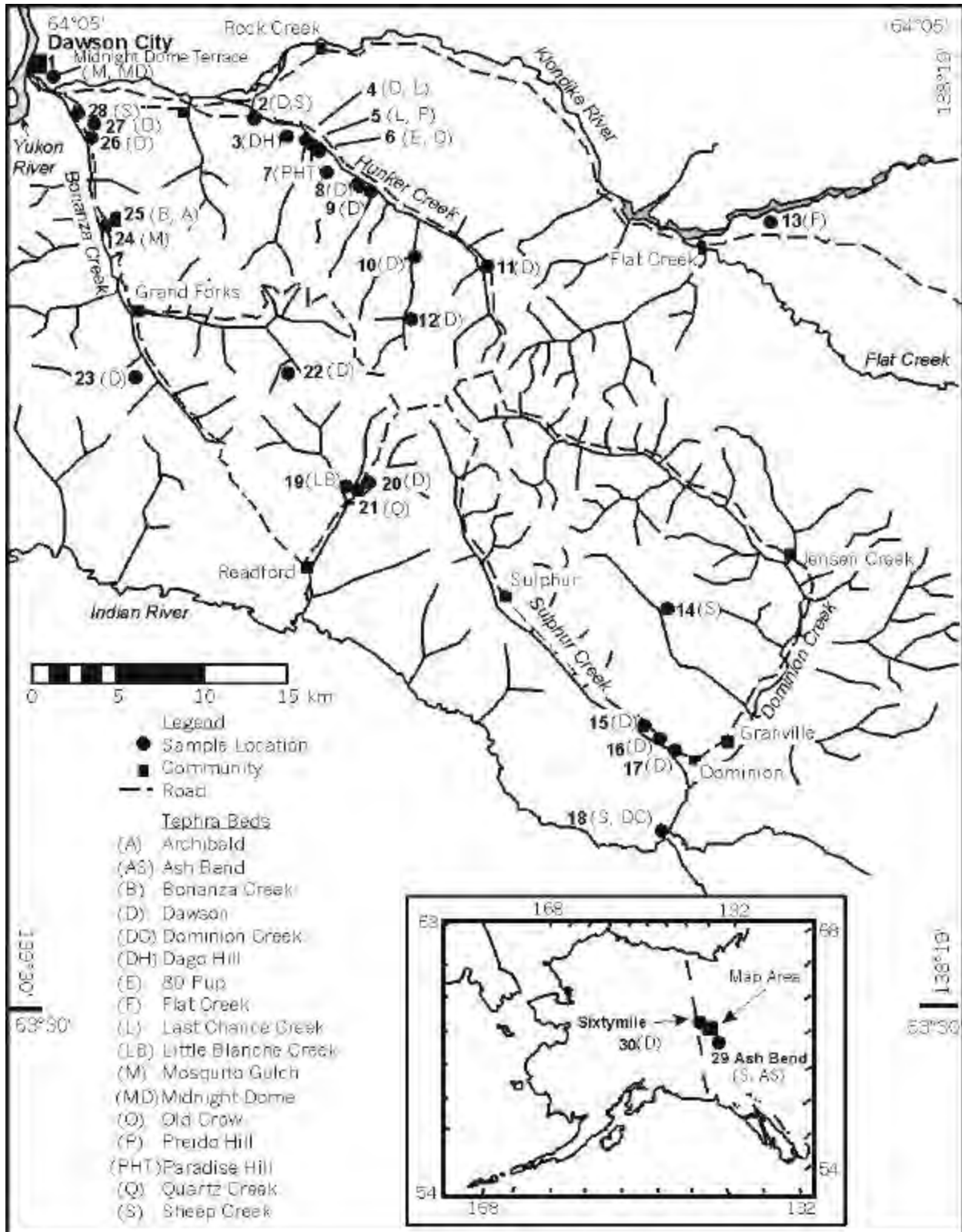


Figure 12. Location and identity of tephra samples collected in the Klondike district, Yukon.

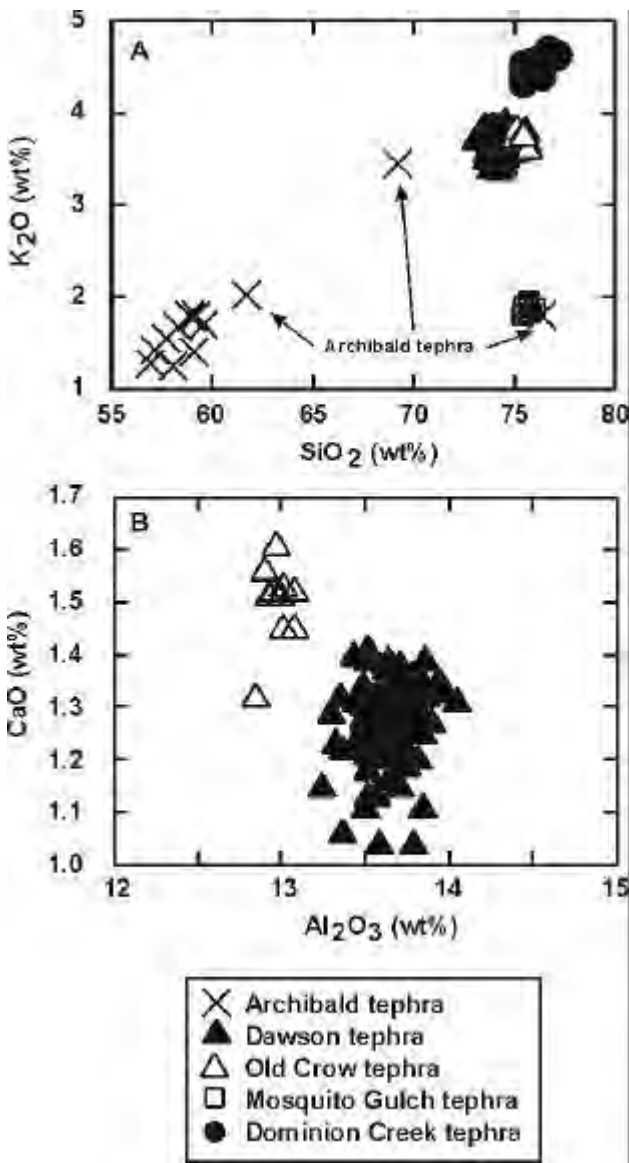


Figure 13. Oxide-variation diagrams of type I and Archibald tephra beds in the Klondike region showing distinctive chemical composition of glass shards in each bed.

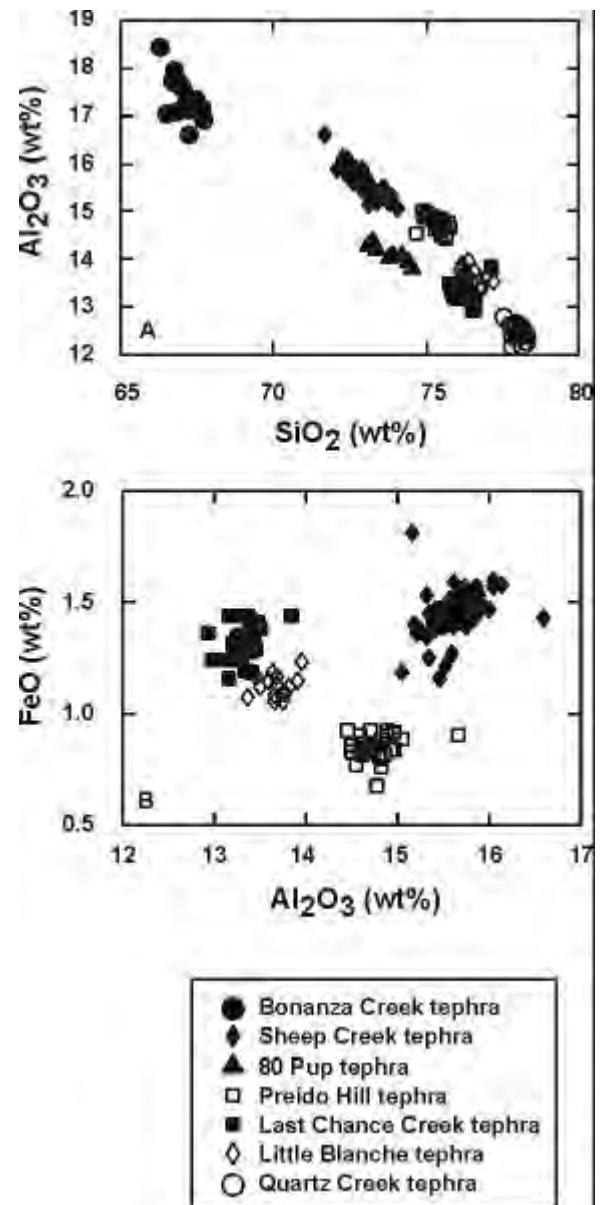


Figure 14. Oxide-variation diagrams of type II tephra beds in the Klondike region showing distinctive chemical composition of glass shards in each bed.

eas, with the remainder coming from glaciated areas (Figure 15). This relationship is slowly changing as reserves in traditional areas are being depleted and new reserves are found in glaciated areas.

Placer Deposits in Unglaciated Areas

Klondike/West Yukon: Placer deposits in unglaciated areas share a number of characteristics including glacial history, type of lode gold source, overall weathering history and the main geomorphic setting of the placers. They occur as low-level valley fill and gravel terraces in valley-bottoms, alluvial fans, and gulches (e.g., Moosehorn Range, Bonanza and Hunker creeks, Klondike, Indian, Sixtymile and Fortymile rivers), as intermediate-level gravel terraces (e.g., Midnight Dome and Archibald's Bench), and as high-level gravel terraces such as the White Channel and Klondike gravels (Lowey 1998). Since these areas are unglaciated, alluvial sediments have undergone extensive weathering and fluvial reworking since the Tertiary Period. Consequently, a continuing cycle of uplift and erosion has concentrated placers into rich pay streaks in valley bottoms, valley side alluvial fans and alluvial terraces. The most likely lode gold sources in these areas are auriferous mesothermal quartz veins which dissect local schist and gneiss bedrock.

Along the unglaciated reach of the Stewart River, placer gold occurs on active point and channel bars along the current course of the river, and along abandoned channels and oxbows. This placer gold is transported primarily during flood events from a number of dispersed gold sources, from tributaries and from glaciofluvial sediments on adjacent bedrock terraces.

Placer Deposits in Glaciated Areas

Placer deposits within glaciated terrain are much more complex than those in unglaciated regions reflecting drainage diversions, deep burial by glacial deposits and the range of glacial sedimentary processes which may result in economic heavy mineral concentrations. As a result, placer deposits formed in glaciated terrain in Yukon can be divided both geographically in relation to ice-flow directions and time since glaciation, and sedimentologically, in terms of the mechanisms responsible for heavy mineral concentration.

Within Pre-Reid Limits

Clear Creek: The Clear Creek area lies just out-

side the limit of the Reid valley glaciation, but includes areas which may have been subject to alpine glaciers during the Reid episode (Morison 1985). Surficial deposits include Tertiary (Pliocene?) gravels similar to the White Channel deposits, pre-Reid glacial drift which has covered the Tertiary gravels, Reid alpine drift, Quaternary valley-bottom and buried placers, and colluvial deposits. Felsic intrusions dissected by gold-quartz veins are the likely lode gold source.

Mt. Freegold/Mt. Nansen: Placer gold occurs in pre-Reid glacial till and glaciofluvial gravels, as well as in non-glacial gravels which were deposited after and on top of pre-Reid glacial and glaciofluvial deposits (LeBarge 1995). Gold was preserved in the glacial material because of the limited dispersion during the alpine pre-Reid glaciation. The gold is likely derived from lode sources within numerous felsic intrusions and related vein systems in the area (LeBarge 1995).

Within Reid Limits

Mayo Area: These placers lie at the margins of both the Reid and the McConnell glaciations, and are known to occur in a wide variety of geomorphic settings, including alluvial fans, fan-deltas, gulches, valley-bottoms (alluvial plains), and bedrock terraces which have been variably buried and reworked by glaciofluvial processes (Hein and LeBarge 1997). Placer gold is also known to occur in glacial till and glaciofluvial gravel, especially where these sediment types have encountered pre-existing placers and have reconcentrated gold in a zone close to bedrock. The bedrock source of gold in the area is likely related to intrusions and quartz veins which cut the local Paleozoic schist and quartzite.

Within McConnell Limits (Multiple Glaciations)

Livingstone/South Big Salmon Area: Auriferous interglacial gravels formed between the Reid and the McConnell glaciations occupy east-west trending valleys which are transverse to the direction of ice movement (Levson 1992). These placers were buried by several metres of glacial drift, which protected them from the erosive action of the ice which later scoured the ridges as the ice sheet moved northward. The gravels were later re-exposed by a large amount of fluvial downcutting at the end of the glaciation and during a period of post-glacial fluvial reworking. The lode source of gold in the Livingstone area is likely auriferous quartz veins hosted by graphitic schist (Stroink and Friedrich 1992).

Whitehorse South The placer gold-bearing gravels the Whitehorse South area (i.e., Moosebrook, Pennycook, Sidney, and Iron creeks) are poorly understood as little scientific work has been done in the area. They may be similar in genesis to the placer deposits on Livingstone Creek, where auriferous interglacial gravels formed during the long period of fluvial action between the Reid and the McConnell glaciations.

Kluane area Placer deposits generally occur in two physiographic settings which are geographically divided by Kluane Lake. West of Kluane Lake, drainages including Burwash Creek, Quill Creek and Kimberley Creek contain placer gold in modern stream gravels and low terraces. Placer gold distribution there is mainly controlled by variations in stream gradient and valley shape throughout numerous canyons and bedrock-confined alluvial plains. On the east side of Kluane Lake (e.g., Gladstone Creek) placer deposits consist of auriferous glaciofluvial and recent stream gravels which have reconcentrated gold above bedrock on top of glacial till and glacial lake sediments.

Yukon Placer Gold Production 1978-2000

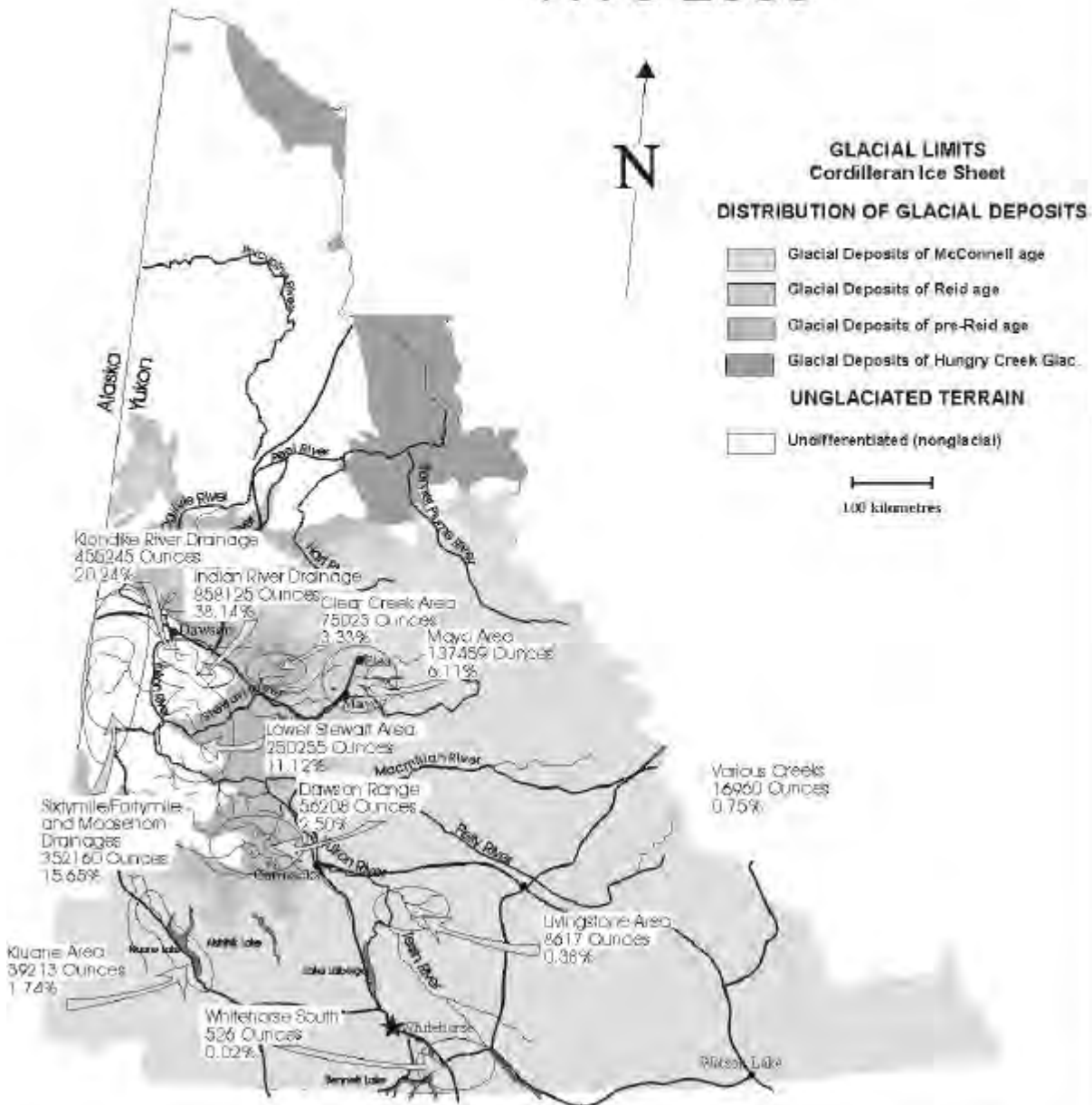


Figure 15. Yukon placer gold production by region 1978-2000.

Field Excursion Itinerary, Day 1: Whitehorse to Dawson and Stewart River Valley, August 25th, 2001

For stop locations see Figure 16 (in map pocket)

STOP 1: Five Finger Rapids: Yukon River rapids and overview of McConnell glaciofluvial terrace (62° 15'N, 136° 25'W)

J.D. Bond

These rapids posed a navigational hazard to the riverboat skippers on their journey up river from the Klondike gold fields. Overpowered by the current, the boats had to be winched through the narrow channel at this point of the Yukon River.

The glaciofluvial terrace observed on the far side of the Yukon River is composed of McConnell outwash. At this locality we are within 15 kilometres of the McConnell glacial limit, which lies downriver and to the west of the viewpoint. Approximately 10 kilometres north on the highway we will pass a braided stretch of the Yukon River. This is very proximal to the McConnell glacial limit and the highway climbs over the actual limit at this locality. The highway then drops onto a McConnell glaciofluvial terrace, which it follows

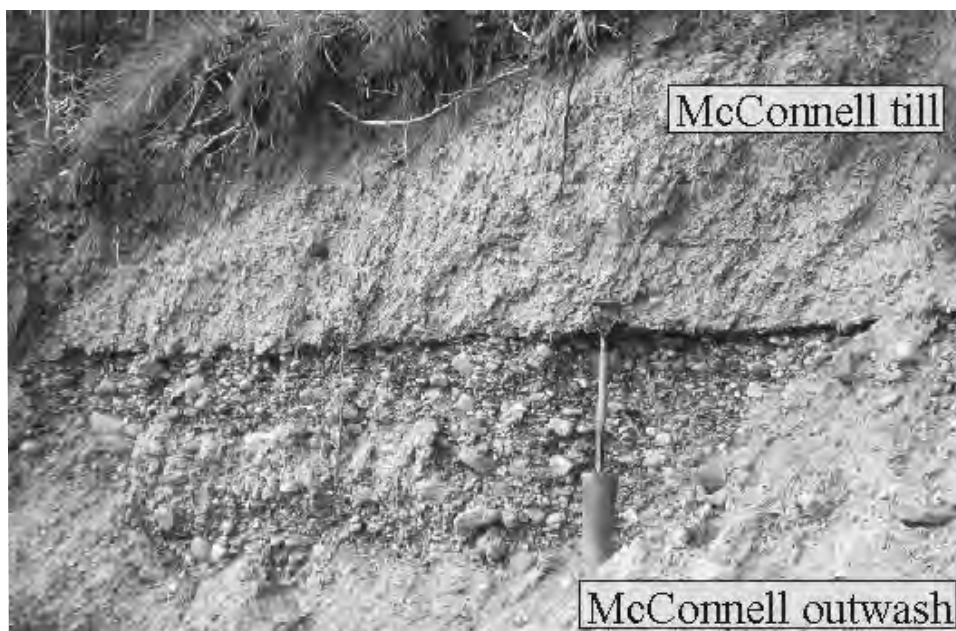


Figure 17: Contact between McConnell till and advance outwash at the McConnell end moraine section, Stewart River.

for 25 kilometres to the site of Minto. The highway then traverses Reid terrain en route to Pelly Crossing.

STOP 2: McConnell End Moraine/Stewart Neosol, Stewart River Valley: Type locality for the last glaciation in Yukon. Locality for modern soil development in central Yukon (63° 31'N, 136° 14'W)

J.D. Bond

In 1900 R.G. McConnell of the Geological Survey of Canada published notes on the glaciation of the Stewart River valley. This is his description of the end moraine: "Below Mayo river (on the Stewart river) a wide ridge 200 feet in height crosses the valley. The ridge is several miles in width and is built of silt, sand and gravel alternating with and often capped by bands of boulder-clay."

The McConnell glaciation is estimated to have reached this location, its maximum extent, between 26.3 ka BP and 12.5 ka BP (Jackson and Harington 1991; Ward 1989). The McConnell valley glacier that flowed down Stewart River was part of the leading edge of the Cordilleran ice sheet in central Yukon (Figure 17). The accumulation zone for the ice sheet, in this part of Yukon, originated to the east in the Selwyn Mountains. Local alpine glaciers also developed in isolated highlands near the leading edge of the ice sheet. Meltwater drainage occupied the current course of the Stewart River, flowing into Tintina Trench and eventually southwest towards the Yukon River across the Klondike Plateau.

Holocene soil development on McConnell surfaces is classified as the Stewart soil in central Yukon (Table 1; Smith et al. 1986). The Stewart soil is a weakly developed Orthic Eutric Brunisol, characterized by a shallow solum (30 cm) and lack of clay accumulation in the

IIBm horizon (Smith et al. 1986; Tarnocai 1987a). There is no chemical weathering or disintegration in rock materials (Smith et al. 1986). The type locality for this soil is on the McConnell end moraine in Stewart River valley.

STOP 3: Overview of Ash Bend Section, Stewart River: Type section for the Reid glaciation and contains last interglacial organics, vertebrate remains and the Sheep Creek tephra (63° 30' N, 137° 15' W)

J.D. Bond, J.A. Westgate, C.E. Schweger and D.G. Froese

Ash Bend section is an 800 metre long exposure of Reid drift plus an interglacial deposit on the Stewart River in central Yukon (Figure 16). The section is located a few kilometres within the outer limit of the Reid glaciation. The exposure consists of (from river level) 15 m of partially covered glaciofluvial gravel (advance outwash), 10 m of till (Reid glacial maximum), 15 m of glaciofluvial gravel (retreat outwash) and 2.2 m of post-glacial debris-flow diamict (Figures 18 and 19; Bond 1997). The diamict can be separated into two diamicts, separated by aeolian sand. Both contain red, elongate slabs of palaeosol material, sand lenses sheared out in the direction of slope, and fractured pebbles, some of which are dismembered with their fragments being strewn out in the direction of slope. Sand wedges with their characteristic primary lineation penetrate the lower diamict and are overturned in the direction of slope, and the entire sequence is blanketed with thin, massive loess (Figure 19).

Incised into the Reid drift and exposed in a gully near the section is a deposit of organic silt from a previous interglacial period (Figure 20). The deposit consists of organic silts with transported spruce limbs and logs, abundant spruce pollen, and remains of bison, mammoth and moose occur at the base of the channel (Hughes et al. 1987). The overlying silty beds contain less organic material, a reduced spruce pollen content, and a conspicuous, white tephra bed (UT1052), up to 3 cm thick, near its base. Above these deposits there is a 2 m thick, coarsening-upward sequence of inorganic silts and sands with sparse and

poorly preserved pollen of mostly Cyperaceae, Poaceae and *Artemisia*. A wispy, white, discontinuous tephra bed (UT1051) occurs in the lower silts, 90 cm above the lower tephra bed. The youngest deposits of this channel infill consist of massive sands with a palaeosol, covered by another sandy bed and woody peat. The organic silt deposit contains permafrost.

Reid Glaciation

The Reid glaciation reached its terminus in the Tintina Trench prior to 190 ka BP, according to the thermoluminescence date on loess bracketing the Sheep Creek tephra at Fairbanks, Alaska (Berger et al. 1996). The timing of the Reid glaciation, other than the pre-190 ka BP date, remains unresolved. According to the North Atlantic oxygen isotope scale and the pre-190 ka date, the most recent cold period the Reid glaciation could be assigned to is MIS 8 (approximately 250 – 300 ka BP). This is also supported by recent evidence from Huscroft et al. (2001) that indicates a maximum age for the Reid glaciation at 311 ka BP. This suggests that the Reid glaciation does not correlate with the middle Pleistocene penultimate advance (Illinois glaciation, MIS 6) elsewhere in Canada. Geomorphic evidence for a penultimate glaciation may not exist because the late Wisconsinan McConnell glaciation was more extensive. No evidence in the Yukon stratigraphic record has documented this hypothesis.

These observations also deny the simplistic model of the Cordilleran

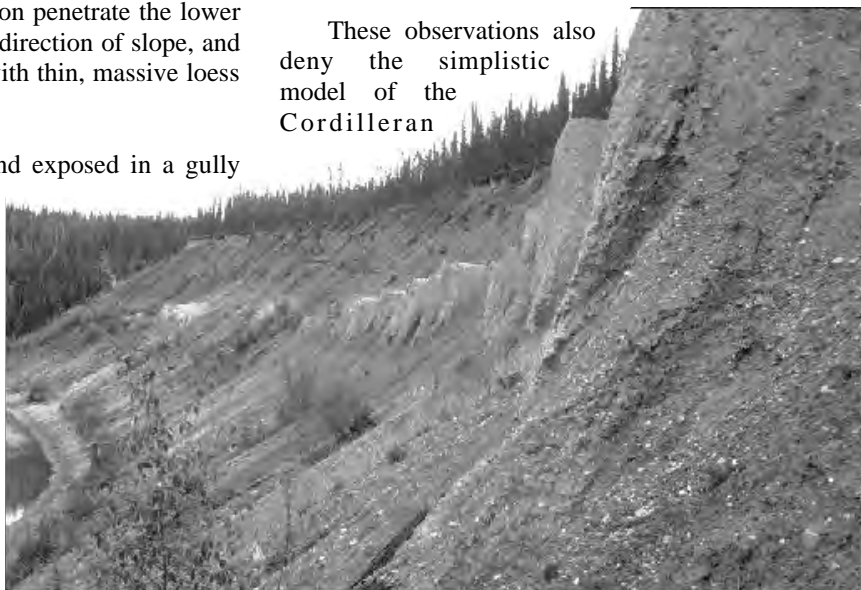


Figure 18: Ash Bend on the Stewart River in Tintina Trench. This is the type section for the Reid glaciation in central Yukon. The 60 m high section consists of (from bottom to top) advance outwash, Reid till, retreatal outwash and a debris-flow deposit. View is to the east.

ice sheet becoming progressively restricted with time in central Yukon. They also suggest that the Diversion Creek palaeosol formed over three climatic periods (MIS 5, 6 and 7).

Post-Reid Interglacial

Tephra Beds

Sheep Creek Tephra

The lower tephra bed (UT1052) is crystal-rich with abundant hornblende, plagioclase, and FeTi oxides and minor amounts of hypersthene and apatite. Its glass is mostly in the form of highly inflated pumice, although some chunky glass of low vesicularity is present. The glass has a rhyolitic composition that clearly separates it from the other tephra beds described in this study and confidently identifies it as belonging to Sheep Creek tephra (Figure 21). These characteristics indicate a type II tephra, derived from a source in the Wrangell volcanic

field. Thermoluminescence (TL) dating studies point to an age of about 190 ka for Sheep Creek tephra (Berger et al. 1996). This result agrees with a mean glass-ft age for Old Crow tephra of 140 ± 10 ka because Old Crow tephra is found 3 m above Sheep Creek tephra at Upper Eva Creek, Fairbanks, Alaska. A glass-ft age of 170 ± 20 ka for Dominion Creek tephra, which lies 30 cm above Sheep Creek tephra in the Klondike goldfields, likewise supports the TL age of Sheep Creek tephra.

Ash Bend Tephra

The younger tephra bed (UT1051) is the Ash Bend tephra. Its glass is in the form of frothy pumice and its mineral component consists of plagioclase, hornblende and FeTi oxides with rare amounts of basaltic hornblende and apatite. Its glass composition is similar to that of Sheep Creek tephra (Figure 21) but is distinguishable by higher amounts of Al, Fe and Ca. Compositional trends in Sheep Creek tephra are colinear with those of Ash Bend tephra and suggest a co-magmatic relationship. A time interval of a few thousand years probably separates these two volcanic eruptions given the pollen content and environmental interpretation of the sediments in this channel fill.

Palaeoenvironmental Setting of Sheep Creek Tephra

The palaeoenvironmental setting of Sheep Creek tephra at Ash Bend is important in assessing the accuracy of the TL age determination. Environmental conditions during deposition of the channel fill sequence changed from dense boreal forest to a transitional environment, and then to cold, dry tundra conditions (Figure 22). Sheep Creek tephra occurs near or at the base of the transitional zone, and separates a lower, dark organic-rich silt from a light-coloured silt with sparse organic matter. The TL age of 190 ± 20 ka for Sheep Creek tephra puts this tephra bed at the boundary between MIS 6 and 7, exactly the palaeoenvironmental setting recorded at Ash Bend. In other words, the dense boreal forest belongs to interglacial MIS 7 and the succeeding tundra environment to MIS 6.

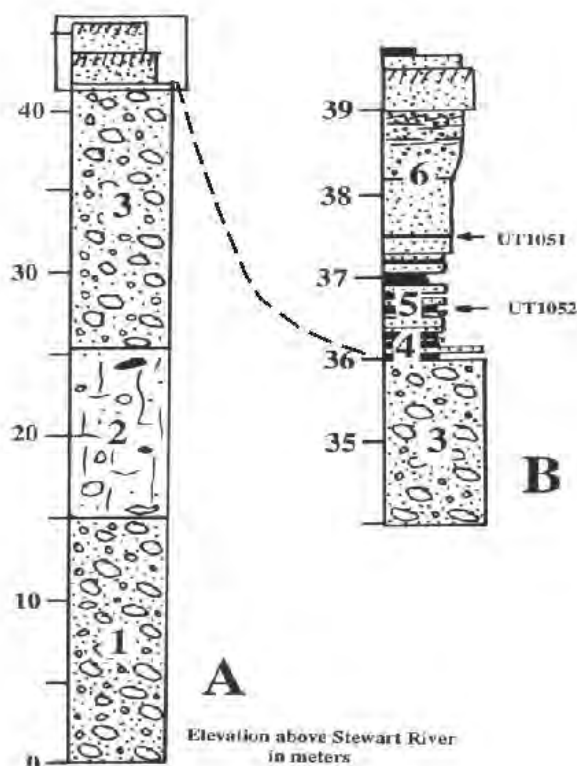


Figure 19: Ash Bend stratigraphy. Unit 1 is advance Reid outwash, Unit 2 is Reid basal till, Unit 3 is retreatal Reid outwash and Units 4 and 5 are a post-Reid organic silt (interglacial) which contains Sheep Creek tephra (UT1052) and Ash Bend tephra (UT1051). Unit 6 is a MIS 6 aeolian deposit.



Figure 20: A swale carved into the Reid deposits at Ash Bend contains organic-rich silts, permafrost, vertebrate fossils and the Sheep Creek tephra. The Sheep Creek tephra is approximately 190 ka BP (Berger et al. 1996).

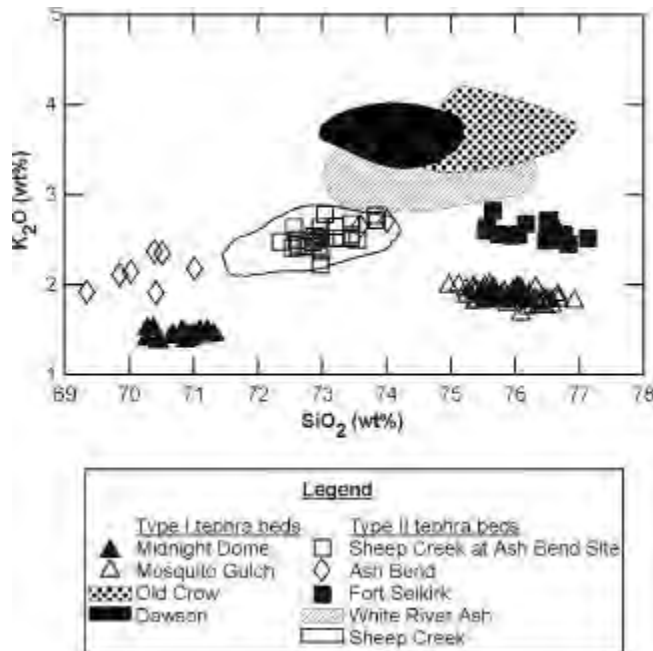


Figure 21. K_2O - SiO_2 scatter plot of glass shards in tephra beds with corresponding data on a few other widespread, well known tephra beds in the Yukon Territory.

STOP 4: Diversion Creek Palaeosol in Tintina Trench: Mid-Pleistocene palaeosol preserved beyond the McConnell limit on Reid (mid-Pleistocene) glacial surfaces ($63^{\circ} 34' N, 137^{\circ} 26' W$)

J.D. Bond and C.A.S. Smith

The Diversion Creek palaeosol is the remnant soil from the last interglacial (and possibly MIS 7) in central Yukon. It is developed on loess, till and glaciofluvial deposits resulting from the Reid glaciation (Bond 1997; Smith et al. 1986, 1987; Tarnocai 1987a). Reid loess and the A-horizons from the palaeosol are rarely preserved due to erosive katabatic winds during the McConnell glaciation. What is observed is that part of the soil is preserved in the resistant till and glaciofluvial deposits. At these localities the palaeosol has moderately well developed B horizons, weakly developed grain argillans and leached sola in the uppermost paleo B horizons (Smith et al. 1986; Table 1). Palaeo-sand wedges and ice wedges, oriented stones, frost shattering and ventifacts are regularly observed in these palaeosols. These cryo-features are remnants of the McConnell glaciation.

At a section exposed near the confluence of Vancouver Creek and the McQuesten River a full intact Di-

version Creek palaeosol was observed (Figures 23 and 24; Bond 1997). The Reid loess and A-horizon had been protected from McConnell katabatic winds in a small depression formed on Reid outwash. McConnell periglacial conditions further contributed to the preserving of the palaeosol by initiating colluviation into the depression, which buried the palaeosol. The intact palaeosol measures 130 cm thick from the A-horizon to the lowest B-horizon.

The Diversion Creek palaeosol can be used as a palaeoclimatic indicator for the last interglacial in central Yukon. Luvisols are the soils most commonly associated with the fine to medium-textured tills of the boreal forest zone of western Canada. Contemporary Luvisols in Yukon are found only in the Watson Lake area. While not quite the same morphologically as the Diversion Creek palaeosol, the contemporary Luvisols exhibit the same features of clay illuviation. They are associated with till and unlike the Diversion Creek palaeoluvisol, are generally absent in glaciofluvial parent materials.

STOP 5: Wounded Moose Palaeosol and Pre-Reid Glacial Deposits in Tintina Trench: Pre-Reid interglacial palaeosol preserved beyond the limit of the Reid glaciation near Clear Creek ($63^{\circ} 38' N, 137^{\circ} 37' W$)

J.D. Bond and C.A.S. Smith

The Wounded Moose palaeosol was originally postulated to be a remnant palaeosol from an interglacial period that preceded the Reid glaciation (Smith et al. 1986). It developed on till and glaciofluvial sediment originating from one of the pre-Reid glaciations (Tarnocai 1987b; Bond 1997). Pre-Reid glaciations remain undifferentiated due to their poor geomorphologic expression (Bond 1997). From investigations in Tintina Trench near Dawson a total of 6 pre-Reid glaciations can be accounted for the in middle to early Pleistocene stratigraphy (Duk-Rodkin et al. in press).

The Wounded Moose palaeosol, as it is characterized at the surface in Tintina Trench, represents a climatic period that followed the last pre-Reid glaciation. It is uncertain how long the climatic period(s) lasted and it is also unclear how well the palaeosol's character represents the climatic norm during that time. Regardless, this surface palaeosol provides us with a glimpse into a period of early Pleistocene climate of central Yukon.

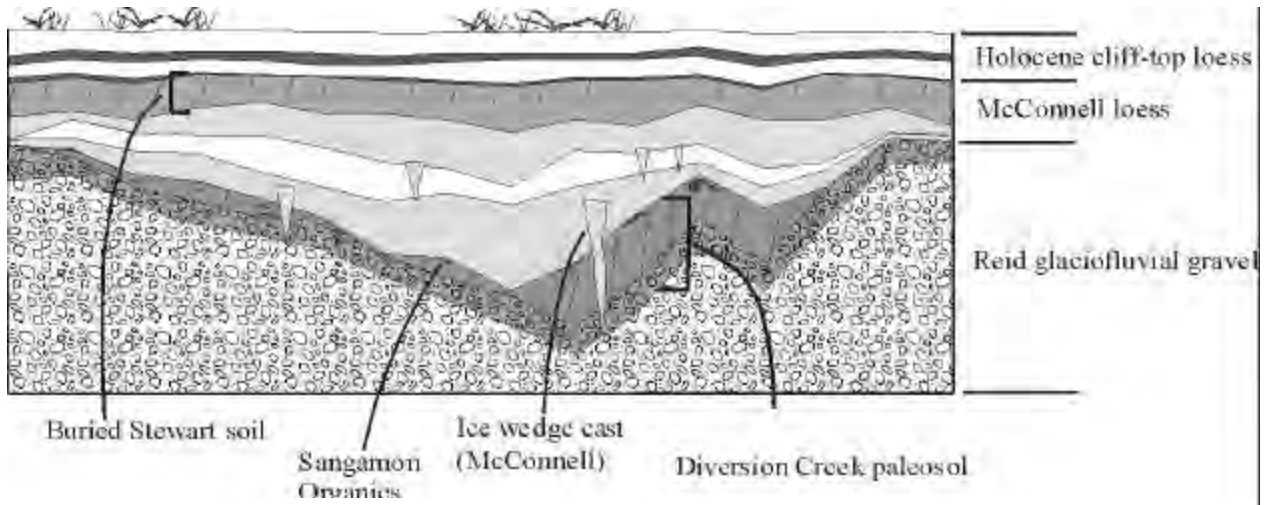


Figure 23: Vancouver Creek section showing a buried Diversion Creek palaeosol on a Reid outwash terrace. The Diversion Creek palaeosol is buried by periglacial sediment from the McConnell glaciation. Recent cliff-top loess has been deposited over the Stewart soil during the Holocene.



Figure 24: Photograph showing the Diversion Creek palaeosol in Reid outwash and loess (1), McConnell loess (2), and Holocene cliff-top loess (3).

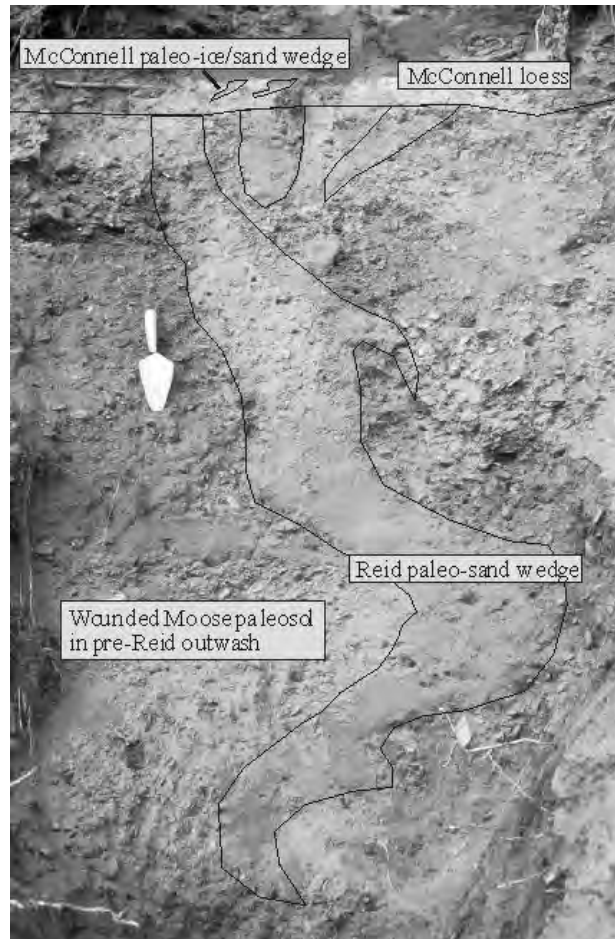


Figure 25: The Wounded Moose palaeosol preserved in pre-Reid outwash near Clear Creek in Tintina Trench. Reid and McConnell palaeo-thermal contraction cracks are visible in the soil profile. McConnell loess, ventifacts and polished stones are also present at the loess/outwash contact.

The Wounded Moose palaeosol attains solum thicknesses of up to 2 m on pre-Reid outwash (Figure 25; Tarnocai et al. 1985). Soil properties characteristically have strongly weathered, rubified, palaeoargillic Bt horizons with thick clay skins or argillans (Table 1; Smith et al. 1987). The argillic horizons found within the Wounded Moose palaeosol are characteristically more developed compared to similar horizons found in the Diversion Creek palaeosol. Smith et al. (1986) concluded that there is a direct relationship between the age and clay content of the palaeosols (Table 1; Smith et al. 1986). Evidence of cryoturbation, ventifacts and the presence of fossil sand and ice wedge casts indicate exposure to subsequent cold, glacial periods (Figure 25).

The climatic conditions necessary for the development of the Wounded Moose palaeosol required a temperature increase of 10°C above current averages for this area. Precipitation would also have to be higher

(Rutter et al. 1978; Smith et al. 1987). The clay mineralogy and clay chemistry suggested that the Wounded Moose palaeosol was initially subjected to a warm and subhumid climate with grassland-shrub vegetation (Kodama et al. 1976; Strakhov 1967; Rutter et al. 1978). This type of environment was favourable for the development of montmorillonite. Later on, the climate became more temperate and humid, which led to the degradation of montmorillonite to kaolinite. This climate was also responsible for the Luvisolic soil with the thick red palaeo-B horizon.

STOP 6: Tintina Trench Lookout: Brief geology of the Tintina Trench, Brewery Creek mine and Ogilvie Mountains (63° 56' N, 138° 24' W)

J.D. Bond and A. Duk-Rodkin

The Tintina Trench is a major structural valley (graben) that developed in the late Miocene along a continental-sized fault that displaced about 450 km during the late Cretaceous and early Tertiary. This major structural feature is similar to the Teslin fault which cuts across the Yukon from southeast to northwest, and the Denali fault which has a displacement of approximately 425 km (Roddick 1967; Tempelman-Kluit 1980) (Figure 26).

The stratigraphy of this part of the trench is formed by two groups of beds that are separated by an unconformity. Middle Miocene alluvial fans are at the top of the unconformity. These beds are composed of northward tilted conglomerate, sandstone, silt and peat. Pliocene to middle Pleistocene fluvial and glacial conformable strata are at the top of the unconformity. Exposures along the north and south sides of the trench show a steeper tilting than exposures along the north side of the trench. Generally north-dipping Tertiary beds were observed by Green (1972) and Hughes and Long (1980). Dip directions show that a major pull-apart basin along the Tintina Fault was established by Late Miocene northward systematic tilting of the middle Miocene alluvial fan deposits (Duk-Rodkin et al. 2001).

Additionally, there are indications of subsequent phases of tectonic activity, since younger fluvial and glaciofluvial terraces were also uplifted in this general region. Features such as sag ponds, numerous depressions and diverted drainages are masked by the later deposition of fluvial and glacial sediments that filled the depressions and lowered the ridges. Fluvial and glacial strata conformably overlain by the tilted mid Miocene beds are exposed along landslide scars and slopes along faults. Pre-glacial fluvial and as many as six glacial beds - tills, outwash, and loess, many of which are separated by palaeosols - have been preserved in exposures along the trench.

Brewery Creek gold mine is a bulk tonnage, heap leach operation, visible in the distance at this site. Gold mineralization is hosted by intrusions of the mid-Cretaceous Tombstone Plutonic Suite, and Silurian to Carboniferous clastic metasedimentary rocks (Burke 2001). Reserves as of December 31, 1999 were 3.1 million tonnes grading 1.59 g/t Au.

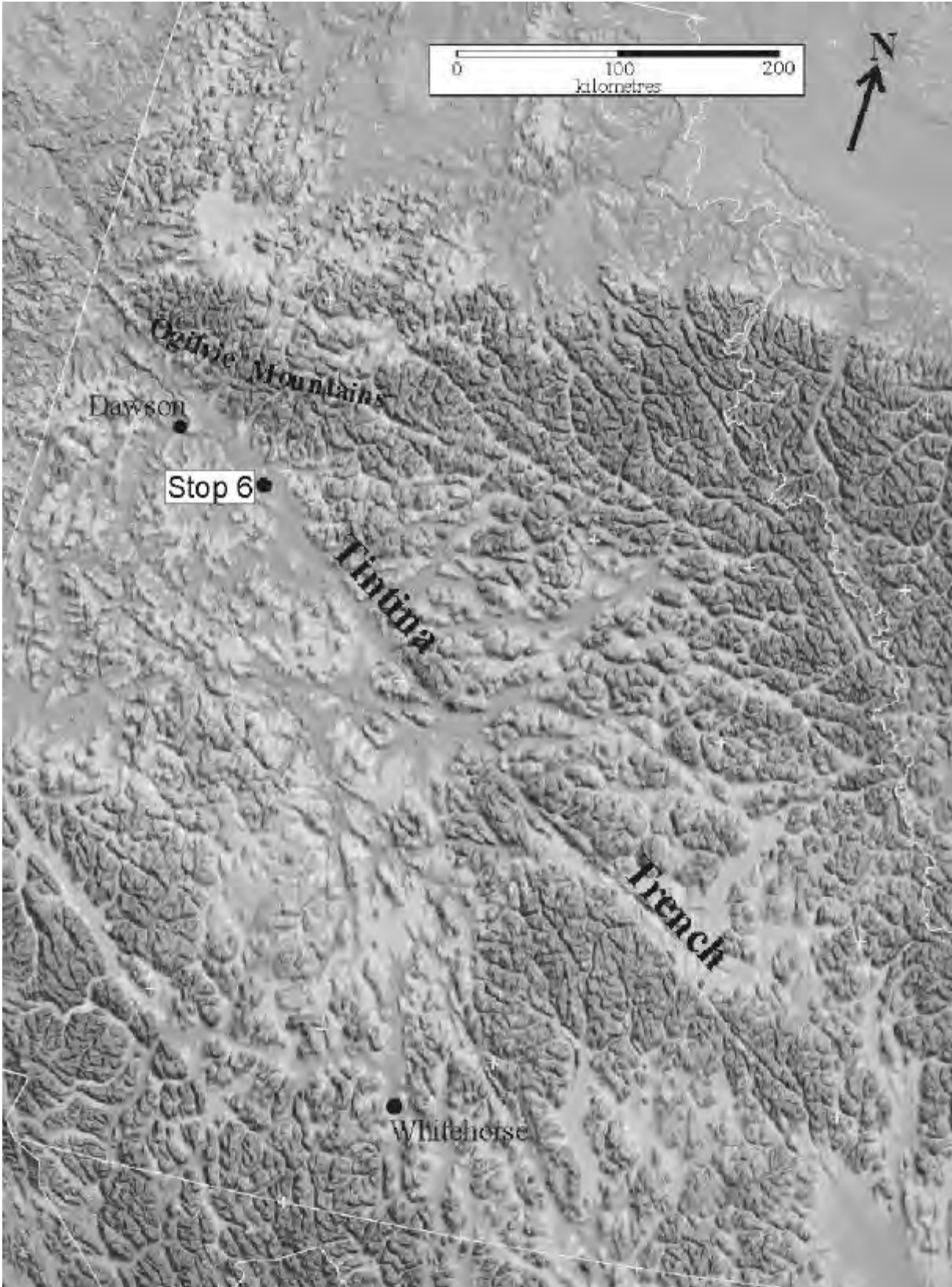


Figure 26: The Tintina Trench forms a distinct lineation in the physiography of central and southern Yukon.

Field Excursion Itinerary, Day 2: Klondike Valley and Northern Klondike Goldfields, August 26th, 2001

The second day of the field trip will be spent in the northern portion of the Klondike goldfields (Figure 27). We will do a quick orientation on the Midnight Dome summit above Dawson City, followed by a series of sites that illustrate the overall stratigraphic framework of the Klondike District.

Klondike Goldfields Stratigraphy

D.G. Froese, J.A. Westgate

McConnell (1905, 1907) divided the gravel of the Klondike goldfields into stream and gulch gravel, terrace deposits, and high-level terraces which included the White Channel gravel and a high-level river gravel known as the Klondike gravel (Hughes et al. 1972; Morison and Hein 1987). These relations are shown in Figure 28 and their correlation to the palaeomagnetic timescale is shown in Figure 29.

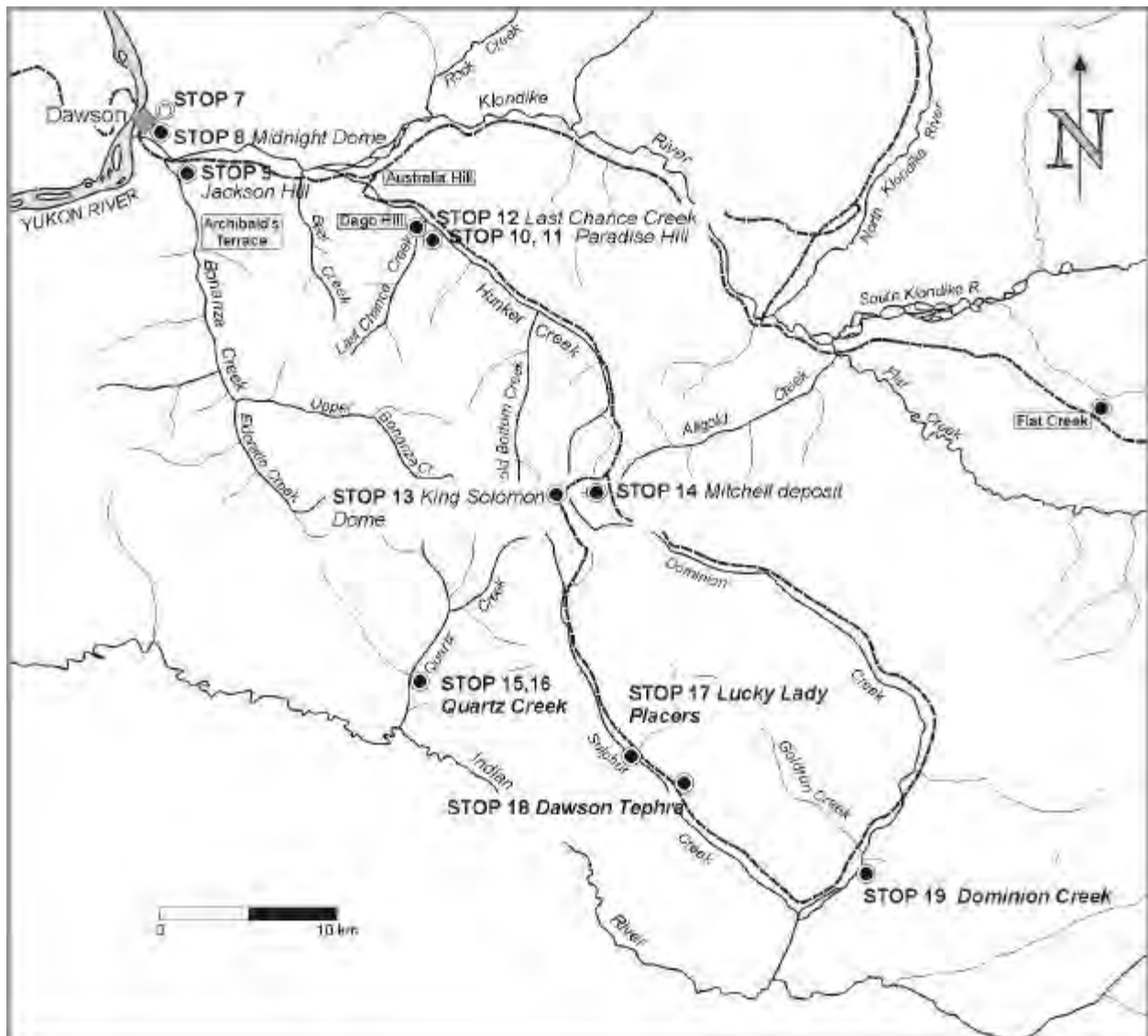


Figure 27. Map of Klondike Placer District with sites mentioned in text.

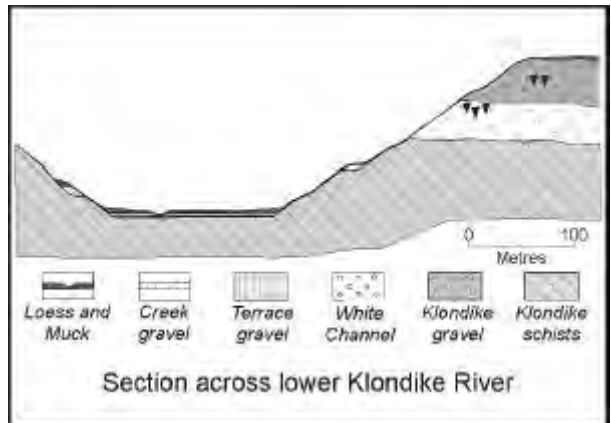


Figure 28 Generalized cross-section of terrace gravels across Klondike valley (after McConnell 1907).

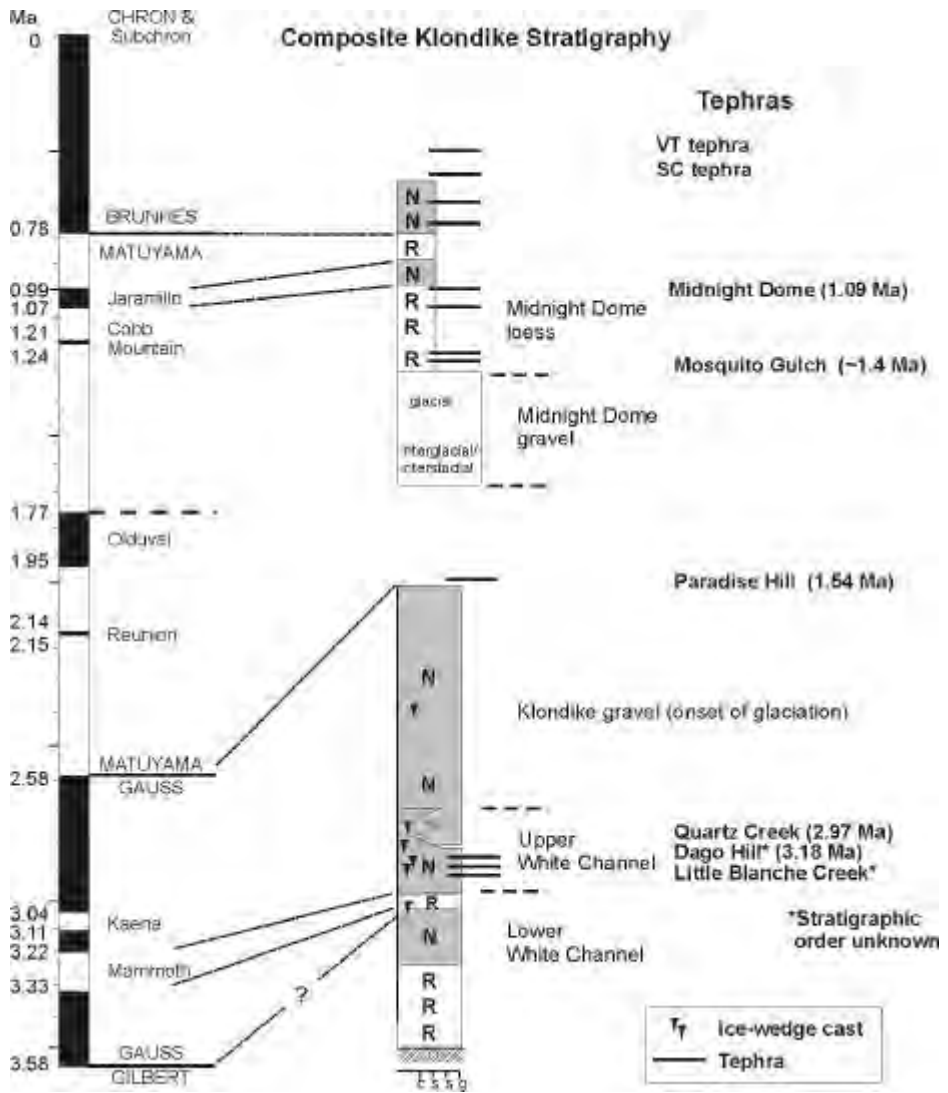


Figure 29. Correlation of Klondike stratigraphic units to the geomagnetic polarity timescale.



Figure 30. White Channel gravels on Dago Hill.

Lower White Channel Gravel

The Lower White Channel gravel (LWC) is a characteristically light grey to white quartz-rich gravel. Its age ranges from earliest Pliocene at the base (ca. 5 Ma) to mid Pliocene at its upper contact with the Upper White Channel (UWC) (ca. 3.2 Ma). At some sites, a distinct alteration is noticeable in the LWC which has been suggested as evidence of hydrothermal fluid flow and linked to the rich placers (Tempelman-Kluit 1982; Dufresne et al. 1986). Lithologies consist of white quartz and quartz-muscovite-chlorite schist, volcanic and meta-volcanic rocks derived entirely from the local drainage basin. The deposit is found on terraces cut into bedrock along Bonanza and Hunker creeks and is interpreted as a braided-river deposit derived from weathering and transport of material from the local drainage basin (Morison and Hein 1987). The LWC is generally considered to represent the time in which the majority of gold was released from bedrock sources in the Klondike.

Upper White Channel Gravel

Similar to the LWC, the UWC is derived entirely from local lithologies. Clasts are unleached and better preserved than those found in the Lower White Channel. Ice-wedge casts and involutions in the sediments are the diagnostic criteria of the UWC, and indicate the first evidence of permafrost in the Klondike area.

Sedimentologically, the Upper White Channel gravel is interpreted as a braided stream deposit that aggraded under periglacial conditions. At sites along Hunker Creek, there is a prominent unconformity between the Upper White Channel and altered Lower White Channel gravel, marked by up to 12 m of aggradation (Figure 30, Dago Hill). It represents the last aggradation of the White Channel during late Pliocene cooling, and the top of the unit interbeds in the lower valleys of Bonanza and Hunker creeks with Klondike gravel (Hughes et al. 1972; Froese et al. 2000).

Klondike Gravel

Klondike gravel is lithologically distinct from the locally-derived White Channel gravel. The presence of chert, various quartzites and slate clasts indicates a source in Selwyn continental-margin sedimentary rocks, found north of Tintina Trench, and quartz-feldspar porphyry from the Tintina Trench. These rocks were transported by glacial meltwater into Klondike River valley, providing the first evidence of Cordilleran Ice Sheet glaciation in the region (Froese et al. 2000). The areal extent of this gravel spreads eastward to the Flat Creek beds (McConnell 1905; Figure 27), and westwards along the Yukon River to at least the Alaska border (Duk-Rodkin 1996). The deposit is normally magnetized, and on the basis of its magnetism and the age of the interbedded UWC gravel, is correlated to the late Gauss Chron (>2.6 Ma) (Figure 29; Froese et al. 2000)

Intermediate Terraces

Terraces intermediate in elevation between the White Channel/Klondike gravel level and present valley bottom are poorly preserved in the Klondike area. Milner (1977) and Naeser et al. (1982) investigated Archibald's Terrace (Figure 27) which includes the Mosquito Gulch tephra on a terrace incised at least 60 m into the White Channel level. The fission track age of 1.45 ± 0.14 (Westgate et al. in press) clearly indicates a pre-Reid age for the surface. One intermediate terrace on the Midnight Dome, incised 100 m into the White Channel-Klondike gravel, is described at Stop 8

Valley Bottom Creek and Muck Deposits

Valley bottom deposits consist of organic-rich silts ("muck", derived from retransported and primary loess deposits) which cover valley-bottom gravels in much of the Klondike district. The majority of these deposits are late Pleistocene through Holocene age (Fraser and Burn 1997; Kotler and Burn 2000), but tephrochronology indicates that remnants date back to at least the middle Pleistocene (Preece et al. 2000). These deposits are of great significance from an economic point of view (in that they host the richest placers in the Klondike District), but they also preserve abundant megafaunal remains and provide an opportunity for palaeoecological investigations, as you will see on the trip.

STOP 7: Midnight Dome Overview (64° 03' N, 139° 29' W)

D.G. Froese and A. Duk-Rodkin

The top of Midnight Dome is an excellent location for an orientation to the Klondike region. There are a few features to note that are described below.

Origin of Yukon River

A complex history of tectonism, regional denudation and glaciation was involved in the evolution of the Yukon River, evidence of which is found in west-central Yukon. Today this river follows a highly anomalous course, having its headwaters in northwestern B.C., crossing over divides in the Dawson Range, passing through the Yukon flats, and draining into the Bering Sea. Changes in the course of the Yukon River occurred as a result of major movements along faults, and the uplift of the St. Elias Mountains in the late Tertiary. As recently as the Late Miocene to Early Pliocene, the Yukon River was south-flowing,

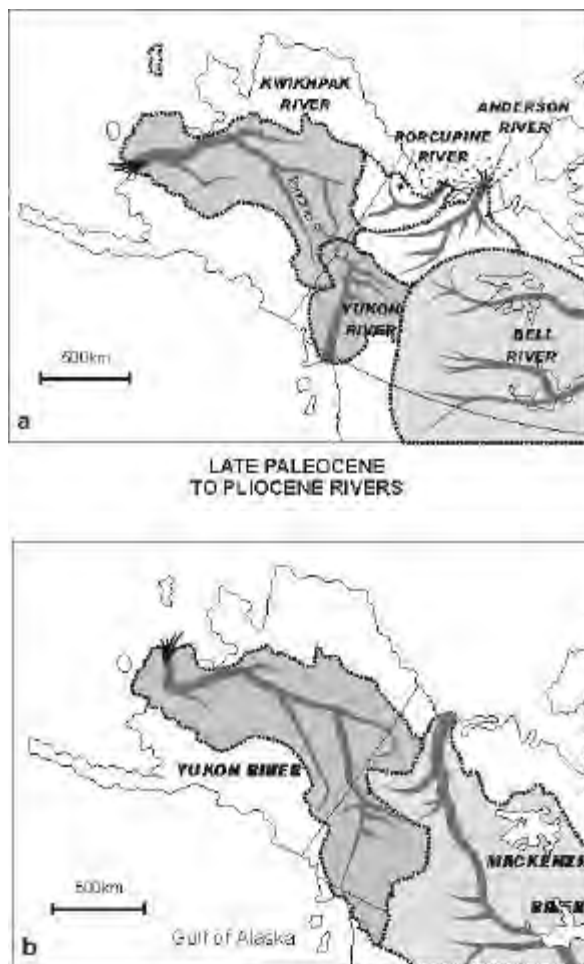


Figure 31. A. Pre-glacial drainage systems of northwestern North America. B. Modern drainage systems resulting from extensive Plio-Pleistocene glaciation.

having its headwaters in the Ogilvie Mountains to the north. Following the first Cordilleran glaciation in the Late Pliocene, the Yukon was diverted northwestward and joined the Kwikhpak River, now collectively referred to as the Yukon River (Figure 31).

Much of the history of the geologic evolution of this region is obtained from long stratigraphic records extending from the late Cretaceous to the mid Pleistocene, which are exposed in landslide scars along the walls of the Tintina Trench. In general, this stratigraphy can be divided into two groups separated by an angular unconformity: lower tilted fluvial and alluvial fan deposits of Tertiary age, and conformable fluvial and glacial beds of late Pliocene and Pleistocene age. The lower group of sediments, which is reversely magnetized, contains a pollen assemblage dominated by middle Miocene Pinaceae, suggesting a warm temperate environment. Dip directions of the sediments

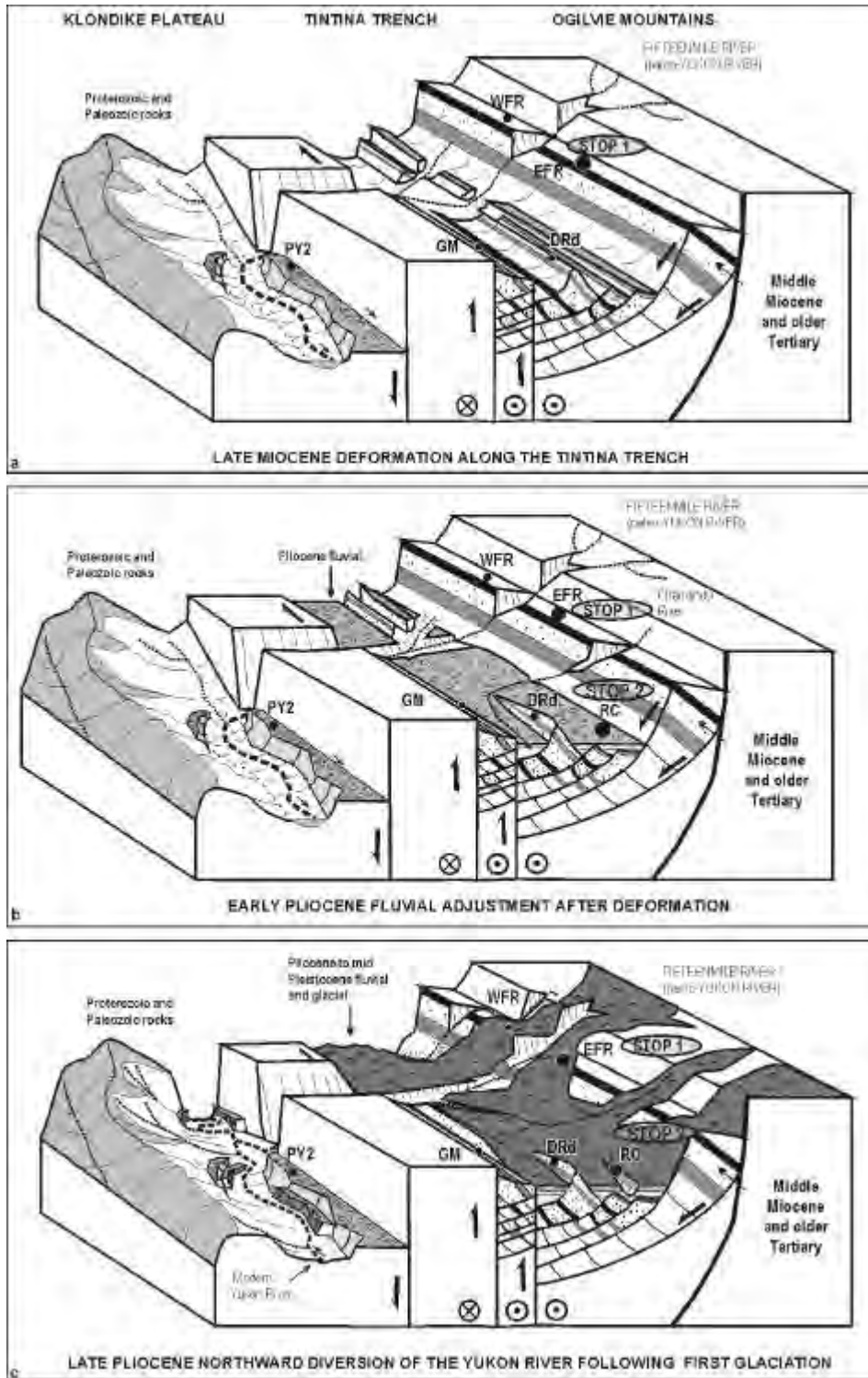


Figure 32. Evolution of Tintina Trench and Yukon River drainage through the Cenozoic (from Duk-Rodkin et al. in press).

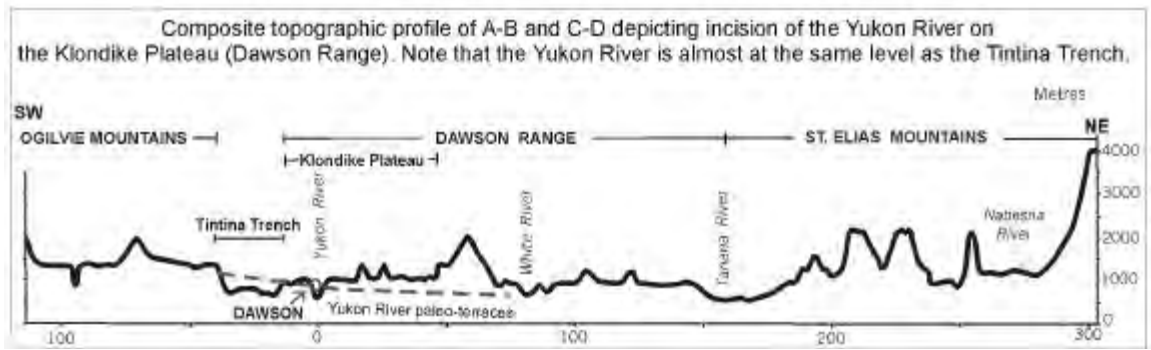
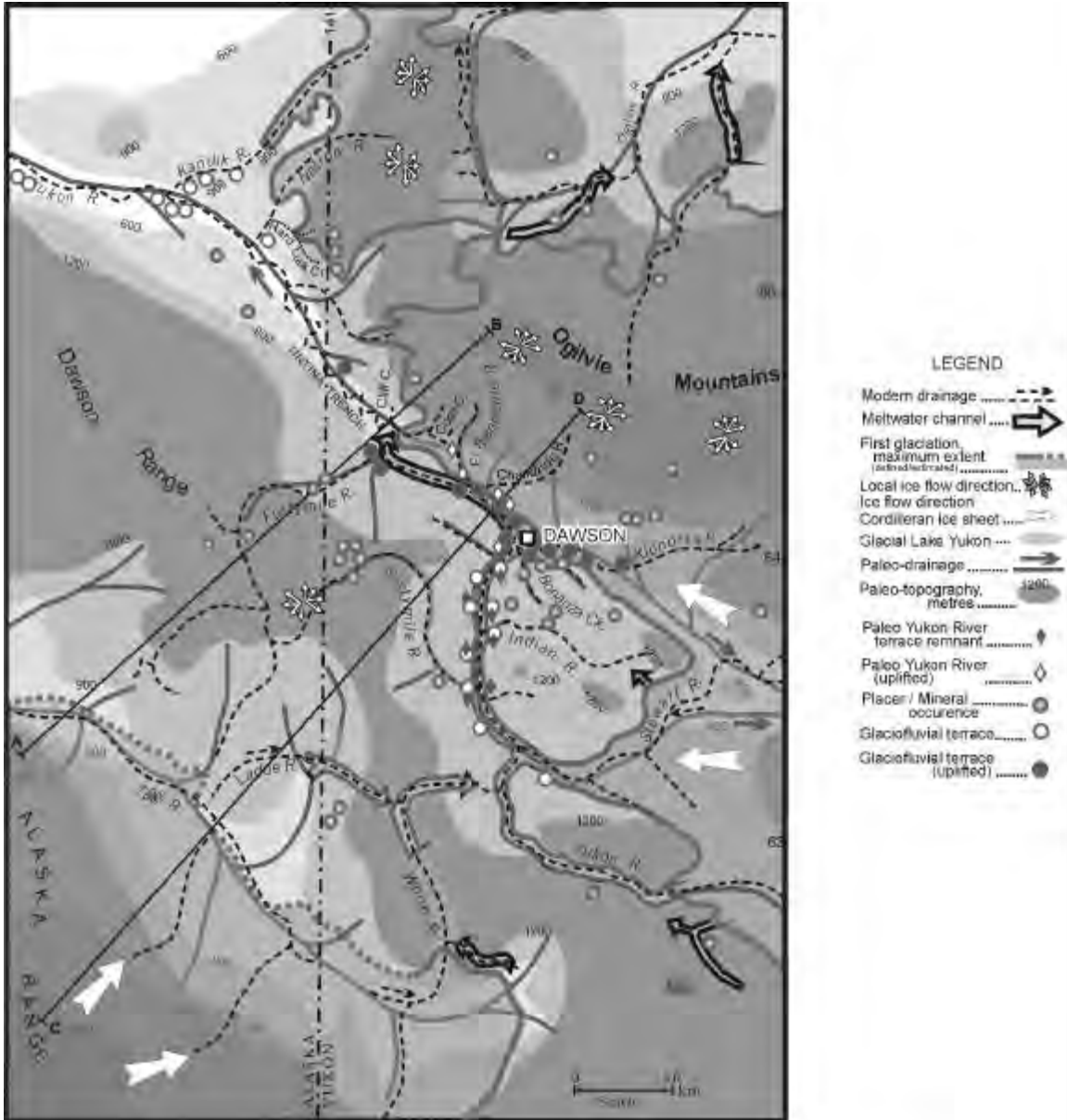


Figure 33. Generalized topography of west-central Yukon and adjacent east-central Alaska. Map shows limits of glaciation in the late Pliocene, location of meltwater channels, and palaeo-Yukon River terraces.

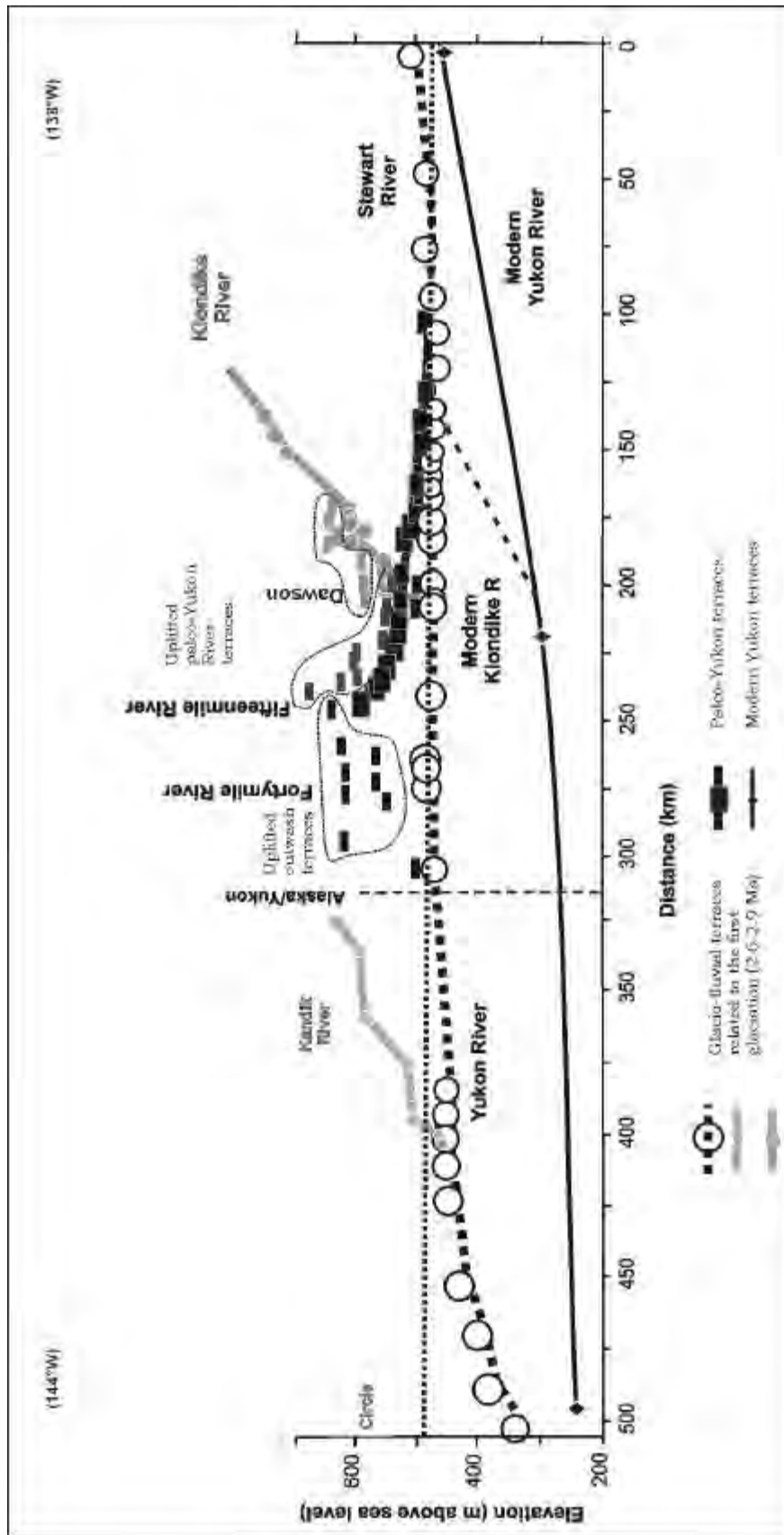


Figure 34. Terrace profiles along Yukon River (from Duk-Rodkin et al. in press).

suggest an extensional “pull-apart” motion in the trench during the middle Miocene while uplifted Pliocene fluvial and glaciofluvial sediments in the region suggest that uplift occurred subsequent to mid Miocene tectonism (Figure 32). The undeformed Pliocene sediments above the unconformity occur as fluvial terraces along the palaeo-Yukon River (some of which are seen here across the Yukon and the Klondike rivers) and as alluvial fan gravels in the Tintina Trench, exposed in landslide scars. The chert and argillite-rich gravels of these palaeo-Yukon terraces are imbricated to the south and clearly point to a northern provenance for the pebble lithologies and a south flowing river at that time (Figures 33 and 34). The considerable thickness of these gravels and the gentle gradients of the terraces suggest a prolonged period of aggradation interrupted by only short periods of downcutting. The trench sediments contain *Picea* and *Pinus*-dominated pollen assemblages and an abundance (21%) of *Polemonium*, suggesting a Pliocene or younger age. The absence of *Artemisia* suggests an age older than Pleistocene. The Pliocene sediments from both the palaeo-Yukon terraces and the trench sites yielded normal magnetization, with the exception of West Fifteenmile River section, which was reversed. Based on the presence of Pliocene pollen, and the fact that these gravels are preglacial, the polarities are assigned to the Gauss Normal Polarity Chron and Gilbert Reversed subchron, respectively. Glaciofluvial terraces associated with the first glaciation, as well as the preglacial palaeo-Yukon terraces, were uplifted by as much as 150 metres in places, suggesting that further tectonism occurred after the first glaciation of the area, but before subsequent glaciations since the younger fluvio-glacial terraces were unaffected by uplift.

The most significant effect of glaciation in the region was the diversion of the Yukon River and its tributaries to the Alaskan Kwikhpak River Basin, increasing its drainage area by some 20%. Iceflow into the trench resulted in the formation of a large glacial lake which formed an outlet to the west of the mouth of Fifteenmile River and established the location of the modern Yukon River in a glacial meltwater channel that was located south of the Tintina Trench, across a low divide, and thus at an elevation that is substantially higher than the trench floor. Evidence for the diversion of the Yukon River by the first glaciation is also seen near Circle, Alaska where Pliocene pre-glacial fluvial deposits (Ager et al. 1994) which are normally magnetized (Figure 19), are overlain by outwash gravels. The lower fluvial gravels there are devoid of argillites of eastern (Ogilvie Mountains) provenance, while the overlying outwash gravels con-

tain approximately 8 % argillites.

The tectonic evolution of the Yukon River in the early Pliocene, associated with major extensional faulting in the Tintina Trench (prior to the first glaciation), forms the geologic setting in which gold became aggregated in west-central Yukon. Gold occurs in various sediment sources, ranging from modern alluvium and colluvium, to interglacial and preglacial gravels, however, its largest occurrences relate particularly to the evolution of the Yukon River drainage basin.

Modern Yukon River

The Yukon River is the fifth largest in North America, twentieth globally, and has an average discharge > 6000 m³/s at its confluence with the Bering Sea (Brabets et al. 2000). At Dawson the Yukon drains an area of about 250 000 km² with an average discharge of ~3000 m³/s with a width of about 400 m and maximum depth of 12 m. Discharge is characterized by a nival regime which usually begins with a mechanical ice break-up in early to mid-May followed by a spring freshet in mid to late June. The modal break-up day for the Yukon is May 7, and a century of flood data indicates that all flood events in the Dawson area have been ice-related (Gerard et al. 1992). Morphologically, the Yukon River is largely braided with some lower reaches exhibiting a wandering planform upstream of the Yukon Flats in central Alaska.

The murky colour of the Yukon makes the low sediment load of the Klondike stand out with its large freshwater ‘hammer-head’ plume that extends along the bank at Dawson City. Upstream from the White River, south of Dawson, the Yukon River has a low suspended sediment load and the gravel cobbles of the bed are clearly seen. However, the addition of the abundant sediment load of the White River, originating in the glacierized St. Elias Range in southwestern Yukon quickly changes its colour and clarity. Local legend suggests that the Yukon was clear until the eruption of the late Holocene White River tephra. While the White River eruption certainly would have added sediment to the White River, the murky water is due to the roughly 30 MT/yr of suspended sediment which the Yukon River carries. Nearly 90% of this sediment load is derived from the glacial meltwater carried by the White River (Brabets et al. 2000).

Top of the World Highway

To the west, the Yukon River continues through the northern Yukon-Tanana Terrane and is largely a

bedrock-confined valley with a thin gravel valley-fill over bedrock. It passes through typical unglaciated country of the Top of the World Highway to the west dividing a portion of the Yukon Tanana Terrane from the Klondike goldfields.

Jackson Hill and Palaeo-Bonanza Creek

To the southeast is the large mining exposure known as Jackson Cut or Jackson Hill. The site consists of 40 m of pre-glacial White Channel gravel overlain by up to 55 m of Klondike gravel marking the onset of glaciation in the region. The White Channel gravel at the site extends to the south and marks the former position of Bonanza Creek prior to its being shifted to the west at the onset of glaciation. This shifting of Bonanza Creek at the onset of the Ice Age is responsible for the preservation of the extensive White Channel gravel at Jackson Hill and the adjacent Trail and Lovett hills immediately up Bonanza Creek. Also note the strath (bedrock) terrace level that the White Channel overlies, which extends up Bonanza Creek marking the former valley position in the Pliocene.

STOP 8: Midnight Dome Gravel and Loess: 1.5 Ma record of multiple tephras, soils and interglacials (64° 03' N, 139°25' W)

D.G. Froese, C.E. Schweger, J.A. Westgate, S.J. Preece, C.A.S. Smith and R.W. Barendregt, and J. Storer

The Midnight Dome site is located on an intermediate terrace, approximately 100 m above the modern valley-floor, 2 km east of Dawson City along the Midnight Dome road (Figure 27). The site consists of fluvial and glaciofluvial gravel overlying bedrock, covered by 10 -14 m of loess, retransported loess and colluvial deposits (Figure 35). Sedimentology of the gravel deposits is reported in Froese and Hein (1996) and the palaeomagnetism of the overlying loess is reported in Froese et al. (2000). Pollen from three interglacial organic silts within the loess was studied by C. Schweger and is reported as part of a larger study of late Cenozoic palaeoecology of the Klondike area (Schweger et al. n.d.). Fission track ages and geochemistry of two tephras at the site, including the basal tephra, and its implications for the age of the Midnight Dome glaciofluvial gravel are reported in Westgate et al. (in press). The geochemistry and correlation of the remaining tephras at the site are the subject of a manuscript being prepared by S. Preece and J. Westgate.

Midnight Dome Gravel

The Midnight Dome gravel consists of two successive depositional environments. Overlying bedrock is an interglacial or at least interstadial "wandering gravel bed" river deposit (Desloges and Church 1987) which is overlain by a glacial proximal-braided outwash deposit. The lower, wandering gravel deposit is dominated by low-angle lateral accretion and channel facies representing migration of gravelly point-bars and pools of the palaeo-Klondike River (Froese and Hein 1996). This style of sedimentation is consistent with the present (Holocene interglacial) Klondike River in the valley today, suggesting that the wandering gravel bed assemblage is similarly of interglacial (or at least interstadial) character. This unit is overlain by 2 - 5 m of coarse, imbricate, cobble-gravel interpreted as a proximal braided deposit (Figure 35). This aggradation represents an outwash deposit as a result of a pre-Reid glaciation in the Southern Ogilvie Mountains to the north (Froese and Hein 1996; Westgate et al. in press).

Midnight Dome Loess

The MDL consists of 10-14 m of loess, retransported loess, colluvial deposits and interbedded organic material spanning an early-mid-Pleistocene interval from ~1.5 Ma to perhaps 0.5 Ma (at present there is no upper minimum age besides normal polarity). A typical lithostratigraphic log through the MDL is described below and shown graphically with associated tephras and palaeomagnetism in Figure 36.

Unit 1: 0-0.8 m: Aeolian coversands

The topography of the upper outwash gravel deposit is infilled by well-sorted, cross-bedded medium sand. Unit 1 shows frosting on grain surfaces and exhibits compound cross-bedding (climbing translucant stratification) (Figure 37A) and ranges from 0.5-1.5 m thick. These sands are interpreted as an aeolian coversand, deposited as low-relief dunes on the braidplain of the palaeo-Klondike River.

Unit 2: 0.8-1.0 m: Organic-rich silt (Interglacial 1) with discontinuous tephra pods

The coversands are sharply overlain by an organic-rich silt ranging from 0.2-0.3 m thick, which preserves an interglacial pollen assemblage (see below). Important to note, however, is that this first interglacial organic unit has an erosional contact at its base, and bedding within the organic unit and associated silts suggest an alluvial/colluvial origin for the

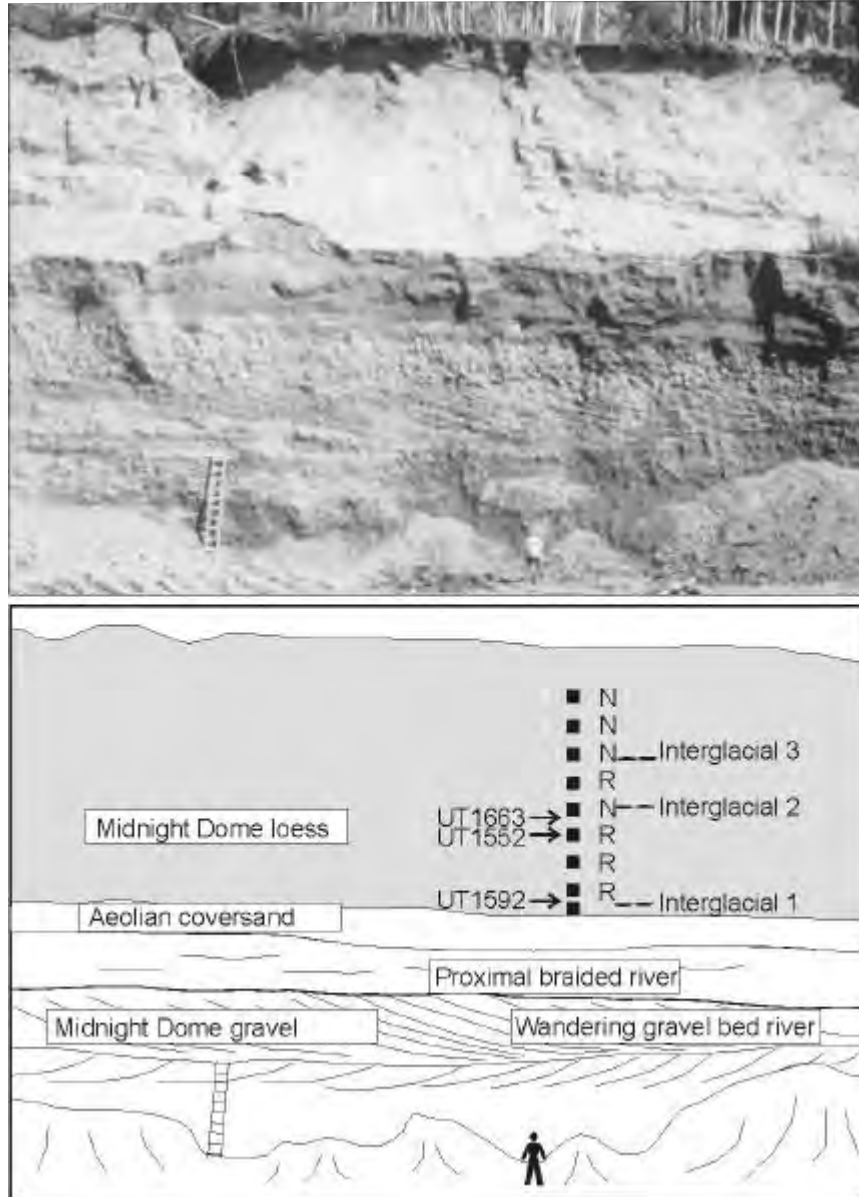


Figure 35. Generalized stratigraphic setting of the Midnight Dome Terrace site in the lower Klondike Valley with exposure documented in 1996. Arrows indicate tephra beds, as follows: lowermost is UT1592, Mosquito Gulch tephra; middle is UT1552, Midnight Dome tephra; uppermost is UT1633, reworked Midnight Dome tephra. The primary position of the Midnight Dome tephra is in the reversely magnetized loess a few cm below UT1633.

facies. Unit 2 also includes the Mosquito Gulch tephra.

Unit 3: 1.0-1.2 m: Pebble diamict

On the north side of the exposure a discontinuous pebble diamict (up to 20 cm thick) is present. The

unit is interpreted as a solifluction deposit during a periglacial period.

Unit 4: 1.2-5.8 m: Massive silts with rare horizontal laminations, and ice-wedge cast with a prominent Type II tephra

Massive silts overly the diamict sediment (1.2-5.8

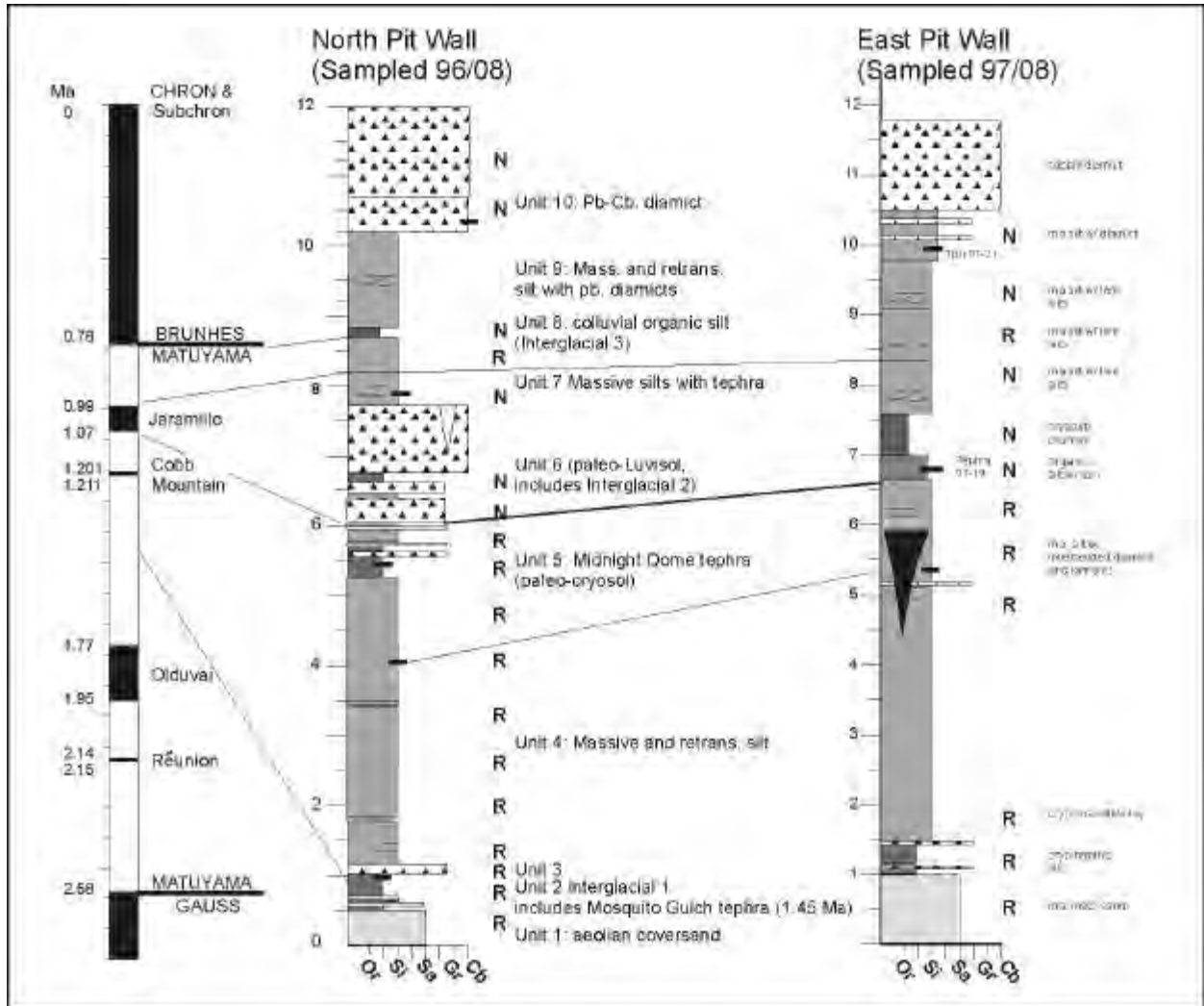


Figure 36. Midnite Dome loess lithostratigraphic logs with units mentioned in text.

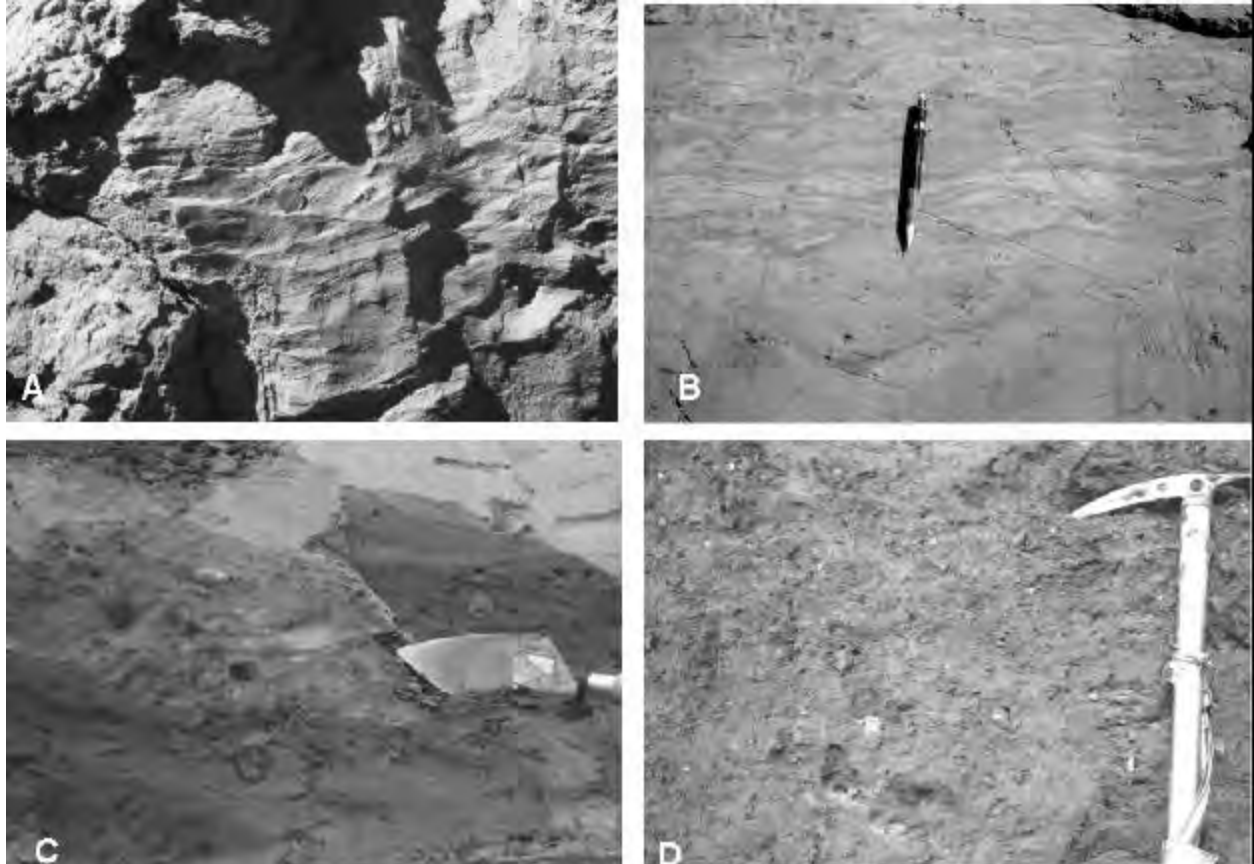


Figure 37. A. Aeolian coversands at the base of the Midnight Dome loess. Unit ranges from 0.3-2.0 m in thickness and infills the topography of the underlying braidplain surface. Photo shows climbing-translucent stratification. B. Pebble-to-gravel diamict with organic pods. C. Midnight Dome tephra (1.09 Ma) occurs within organic rich sediments in distinctive load-casts, indicating the eruption most likely occurred during a periglacial episode in which an uneven tundra landscape was present at Midnight Dome. Polarity evidence for the Midnight Dome tephra includes one site in which the polarity was magnetically reversed, and two sites in which the polarity was normal. Given the mean age (1.09 Ma) and that the deposit straddles the Matuyama-Jaramillo boundary, we argue that the eruption likely occurred during MIS 32 which agrees with the mean IPFT age (Figure 39). D. Pebble-gravel diamict with discontinuous organic pods. Unit interpreted as a remobilized palaeo-luvisol. Pollen analysis from organic pods within the diamict indicates an interglacial assemblage.

m). This facies is interpreted as direct air fall loess during glacial conditions. In some areas these silts are weakly laminated, suggesting that reworking and re-transportation of silts is also an important process during deposition. Near the contact with Unit 5, a prominent anti-syngenetic ice-wedge cast was present in the east wall of the exposure (Figure 38). The cast had a maximum width and depth of nearly 3 m and included the tephra.

Unit 5: 5.8-6.1 m: Organic silt with load casts and Midnight Dome tephra (1.09 Ma)

Overlying Unit 4 is a distinctive organic silt unit with load casts which can be traced laterally around

the section for over 500 m (Figure 37). The unit has a wavy relief, and despite a very high organic silt content, returned no pollen. This unit is interpreted as a palaeo-cryosol, and was likely the palaeo-surface during ice-wedge formation.

Unit 6: 6.1-7.3 m: Pebble-Gravel diamict with discontinuous organic pods (Interglacial 2)

Immediately overlying the organic silt unit is a discontinuous pebble-gravel diamict unit. These diamicts are prominent and can be traced laterally across the section recording a period of hillslope instability. Within this unit, organic silt pods occur characterized by interglacial pollen, and the unit includes

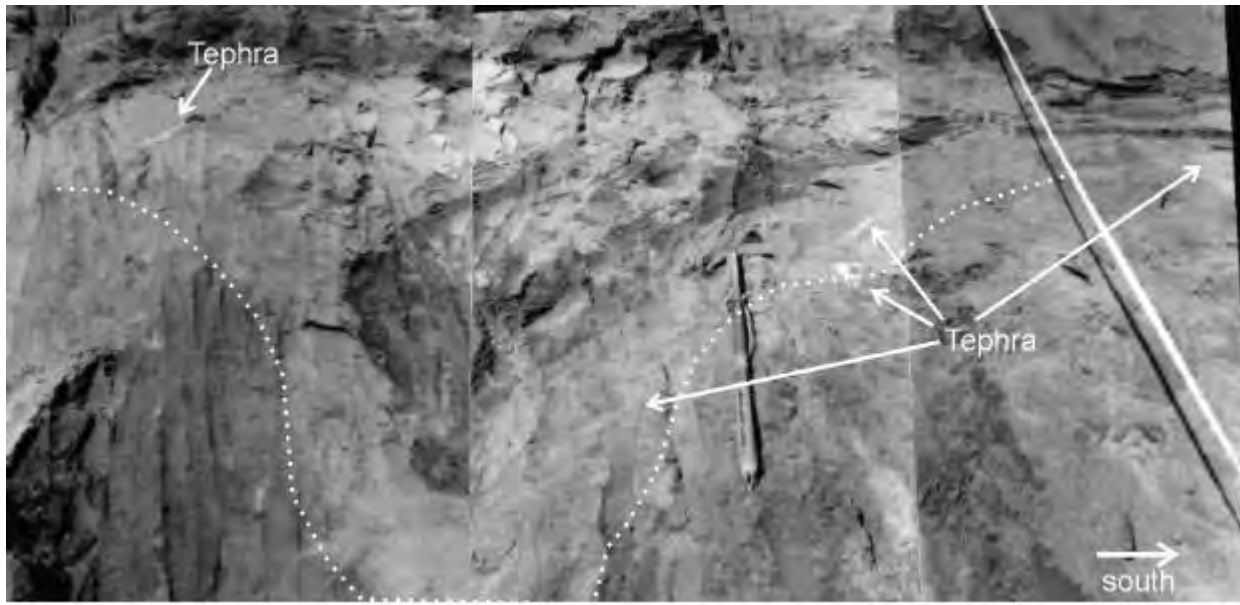


Figure 38. Ice-wedge pseudomorph complex developed in immediately pre-Jaramillo sediments (probably about 1.2 Ma) on the east wall of the Midnight Dome site (photographed in 1996). The complex is interpreted as an anti-syngenetic ice-wedge cast which developed on a degrading (incising) hillslope during a glacial. Such unconformities are likely common within the loess deposits, but rarely are seen due to the lack of textural heterogeneity of the deposits.

10 YR B horizon material suggesting it formed due to remobilization of a palaeo-Luvisol (Figure 39). The organic material is interpreted as portions of remobilized A-horizon.

Unit 7: 7.3-7.8 m: Massive silts

These massive silts overlie the diamict unit (Unit 6) with normal polarity and include massive silts interpreted as direct air-fall loess.

Unit 8: 7.8-8.0 m: Discontinuous organic silt with pebble-diamict (Interglacial 3)

Unit 8 occurs discontinuously across the section as a reworked zone of organic silts interbedded with diamict. Interglacial pollen was collected from this unit.

Unit 9: 8.0-10 m: Massive silts

This unit overlies Interglacial 3 (Unit 8) and consists of massive silts interpreted as direct air-fall loess.

Unit 10: 10-12 m: Pebble-gravel diamict

The uppermost diamict is present in all exposures of the MDL and contains at least one tephra bed.

Palaeoecology

Pollen analysis was undertaken on samples collected from 3 horizons displaying very fine organic material, and typical of loess, most samples proved sterile. Midnight Dome Organic Horizon 1 (Unit 2) is known from only one sample but has been recollected for further pollen and macrofossil analysis. Midnight Dome 2, collected from Unit 6, is known from 6 polleniferous samples and Midnight Dome 3 (Unit 8) from a single sample. Spruce pollen dominates the 3 horizons ranging from 70 to 92%, followed by fir at 1.3-11%, with pine reaching only 3.4% and alder 4.3%. Dense boreal forests were present at the time the 3 horizons were deposited but because of the larger amounts of fir and the smaller amounts of pine and alder these forests were unlike Holocene forests of the Yukon. Warm, mesic conditions with less fire and the absence of permafrost may account for these differences.

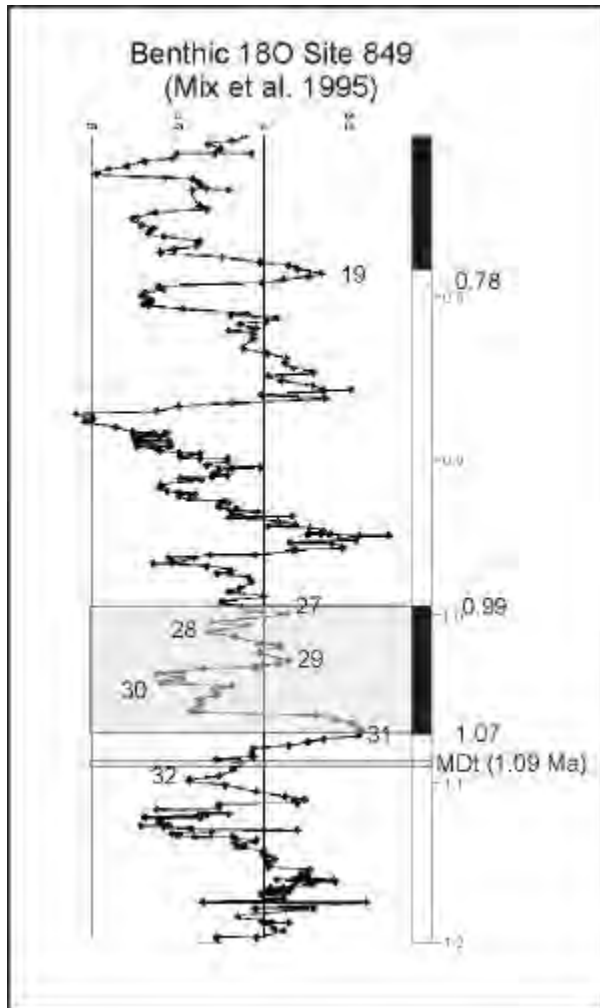


Figure 39. Astronomically-tuned benthic foraminifera record from equatorial Pacific Site 849 with reference to palaeomagnetic timescale (data from Mix et al. 1995). The Midnight Dome tephra, occurring within a palaeo-cryosol immediately below the Jaramillo unit, most likely erupted during MIS 31 and was mobilized downslope during MIS 30.

Palaeosols

The section was examined for evidence of soil development associated with the presence of organic sediments at the base of the loess sequence and at several locations associated with tephra beds within the upper portion of the exposure. In all instances materials appeared to be retransported and mixed. However there is good evidence for a relatively continuous 50 cm thick, brown (10YR 5/4), soliflucted soil B horizon and associated disrupted and discontinuous humus-rich soil A horizon. The soil development is like that expected to form under well drained upland forested conditions. The horizons preserved the suban-

gular blocky structure associated with such a soil but had been retransported locally (soliflucted?) so that the original positions and thicknesses of horizons are lost. As such the palaeosol features in the section (Figure 36) suggest initial interglacial formation (corroborated by interglacial pollen associated with the humus-rich horizons) and subsequent movement downslope as the result of environmental changes during the Jaramillo.

Tephrochronology

Several thin and discontinuous tephra beds occur in the Midnight Dome loess, of which three have been studied so far. Thin pods of reworked Mosquito Gulch tephra (UT1592) are present in the lowermost interglacial silt, close to the base of the loess. Pods of reworked tephra (UT1633) occur within a palaeo-cryosol, as well as in the immediately underlying loess (UT1552) (Figure 35). UT1552 and UT1633 belong to the same tephra bed, which we have named the Midnight Dome tephra.

Mosquito Gulch and Midnight Dome tephra beds belong to the type I class of Preece et al. (1999), their source vents being located in the Aleutian arc - Alaska Peninsula region. They have a low crystal content, abundant bubble-wall glass shards, pumice with large vesicles, and very small amounts of brown glass. Feldspar, orthopyroxene, and clinopyroxene are abundant with minor amounts of amphibole, magnetite, ilmenite, apatite and zircon. Glass shards of the Mosquito Gulch tephra have a rhyolitic composition but those of the Midnight Dome tephra vary from a dacitic to rhyolitic composition. The major-element composition of the glass in each tephra bed is very distinctive with the Midnight Dome tephra containing more Ti, Al, Fe, and Ca but less K and Si than the Mosquito Gulch tephra (Figure 40).

Midnight Dome tephra is known only from the Midnight Dome Terrace site but Mosquito Gulch tephra has been identified at three sites: Archibald's Bench (UT19; Naeser et al. 1982; Preece et al. 2000), Midnight Dome Terrace (UT1592), and the Gold Hill site at Fairbanks, Alaska (NT tephra bed, UT1104; Preece et al. 1999).

Mosquito Gulch tephra offers an opportunity to date the oldest glacial - interglacial sequence at the Midnight Dome Terrace site, namely the outwash gravels, aeolian coversand (glacial) and the overlying, thin, organic silt with Mosquito Gulch tephra (interglacial). The glass-fission-track age of Mosquito Gulch tephra is 1.45 ± 0.14 Ma. The enclosing silt is reversely magnetized (Froese et al. 2000), in agree-

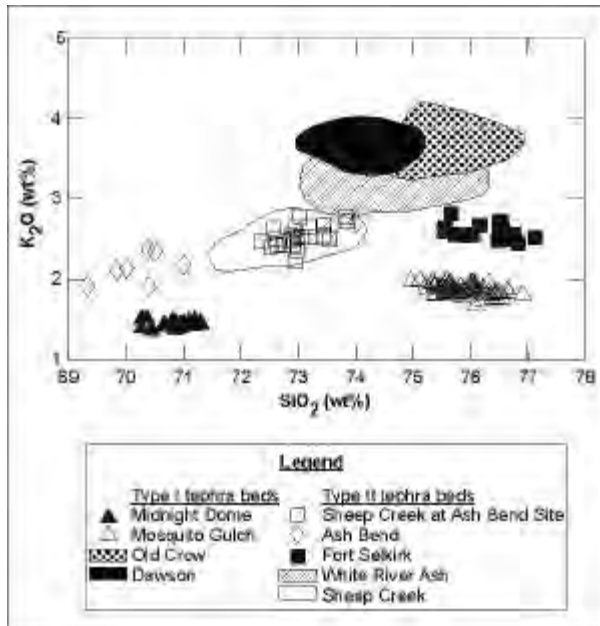


Fig. 40. K₂O - SiO₂ scatter plot of glass shards in tephra beds with corresponding data on a few other widespread, well known tephra beds in the Yukon Territory.

ment with the geomagnetic polarity timescale of Cande and Kent (1995).

Midnight Dome tephra (UT1552) is 8 m above Mosquito Gulch tephra and occurs just below sediments with a normal magnetic polarity, interpreted by Froese et al. (2000) as belonging to the Jaramillo Subchron. Its glass-fission-track age of 1.09 ± 0.18 Ma confirms their favoured age scenario.

Early Pleistocene alpine glaciation affected the Ogilvie Mountains about 1.5 Ma, given the age of Mosquito Gulch tephra, which was deposited during the succeeding interglacial interval. It follows that the Klondike outwash gravels, whose geomorphic and stratigraphic setting indicate a much older age than the sediments at the Midnight Dome Terrace site, must be related to a much earlier Cordilleran glaciation, thought to be of late Pliocene age by Froese et al. (2000).

STOP 9: Jackson Hill: Composition of White Channel/Klondike gravel terrace (64° 01' N, 139° 22' W)

Duane Froese, John Westgate and James White

Jackson Hill is located on the south side of the lower Klondike River valley, above the canyon mouth of Bonanza Creek (Figure 27). Mining at the site focused initially on exposing bedrock (and the gold) by hydraulic mining of the gravel. In large part this produced the exposure at the site and the substantial tailings fan that Callison (the local industrial park) is built upon. More recent mining at the site through the 1970s and 1980s has gone underground from adits exploiting the deep permafrost at the site to maintain the integrity of the gravels.

A century of mining at the site has exposed approximately 40 m of White Channel gravel overlain by 55 m of Klondike gravel (Figure 41). The contact between the White Channel and Klondike gravels at the site is sharp, but if one proceeds up Bonanza Creek to a more marginal braidplain position of the Klondike gravel, the two units interbed, suggesting contemporaneous deposition.

Sedimentology

The lower White Channel gravel at Jackson Hill is slightly altered by fluids, but well enough preserved for sedimentologic characterization (Figure 42). In the western side of the exposure, 3.5 m of basalmost gravel are exposed, consisting of coarse cobbly massive gravel with crude normal grading in horizontally stratified facies. Facies between 12 and 40 m above bedrock, overlying a covered interval, consist largely of planar-tabular cross-bed sets of gravel with interbedded massive gravel and infrequent sand facies characteristic of a medial to distal braided stream environment with rare hyper-concentrated flows.

There is no clear break in sedimentation between Lower and Upper White Channel gravel at the Jackson Hill site, but ice-wedge casts have been observed within a few metres of the Klondike gravel contact (Figure 42). The site represents deposition by the palaeo-Bonanza Creek through the Pliocene in an area of flow-expansion where it met the palaeo-Klondike valley.

Ice wedge casts are present in the UWC gravel near the Klondike gravel contact, and midway through the Klondike gravel (Figure 42). These casts, or pseudomorphs, are epigenetic casts of limited width (30-50 cm) but may extend to depths of 2 m in fluvial gravel. There is no evidence for extended exposure of the surfaces (i.e., soil development), and similar ice-wedge casts have been observed along the modern Stewart River near Mayo (Chris Burn, pers. comm. 2000). They are therefore interpreted as short-term (100s of years) exposure of terraces due to topographic differentiation of the braidplain (during short-term channel

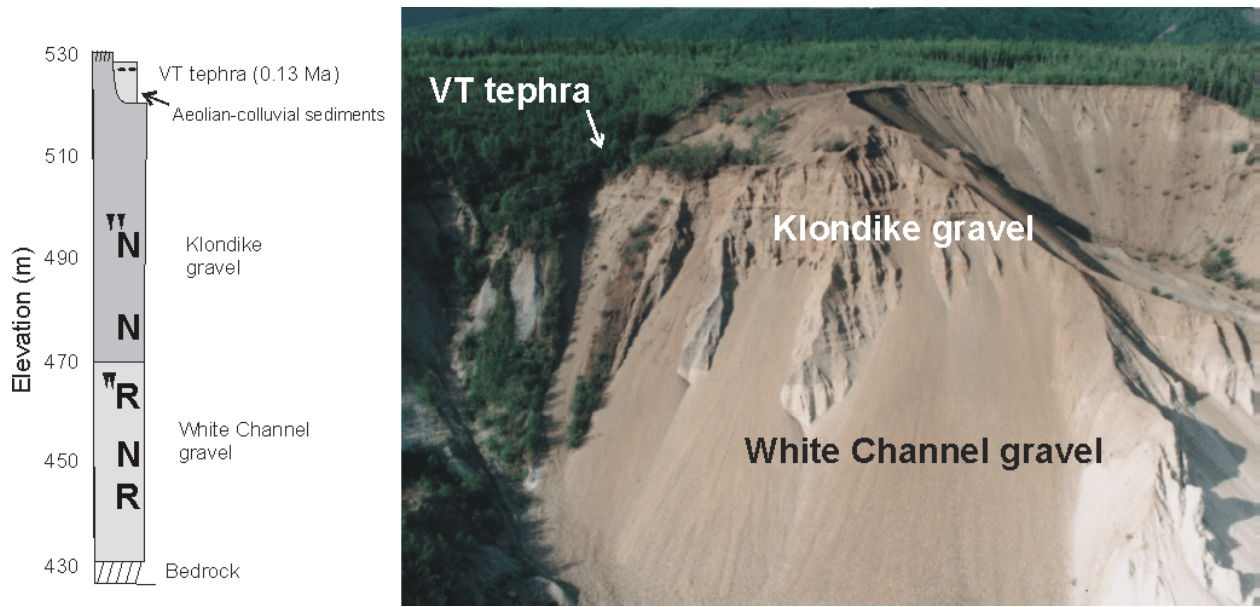


Figure 41. Jackson Hill site with generalized stratigraphic log. Bold N and R indicate Normal and Reverse polarity, respectively. VT tephra occurs at the site within an aeolian-colluvial fill incised into the Klondike gravel.

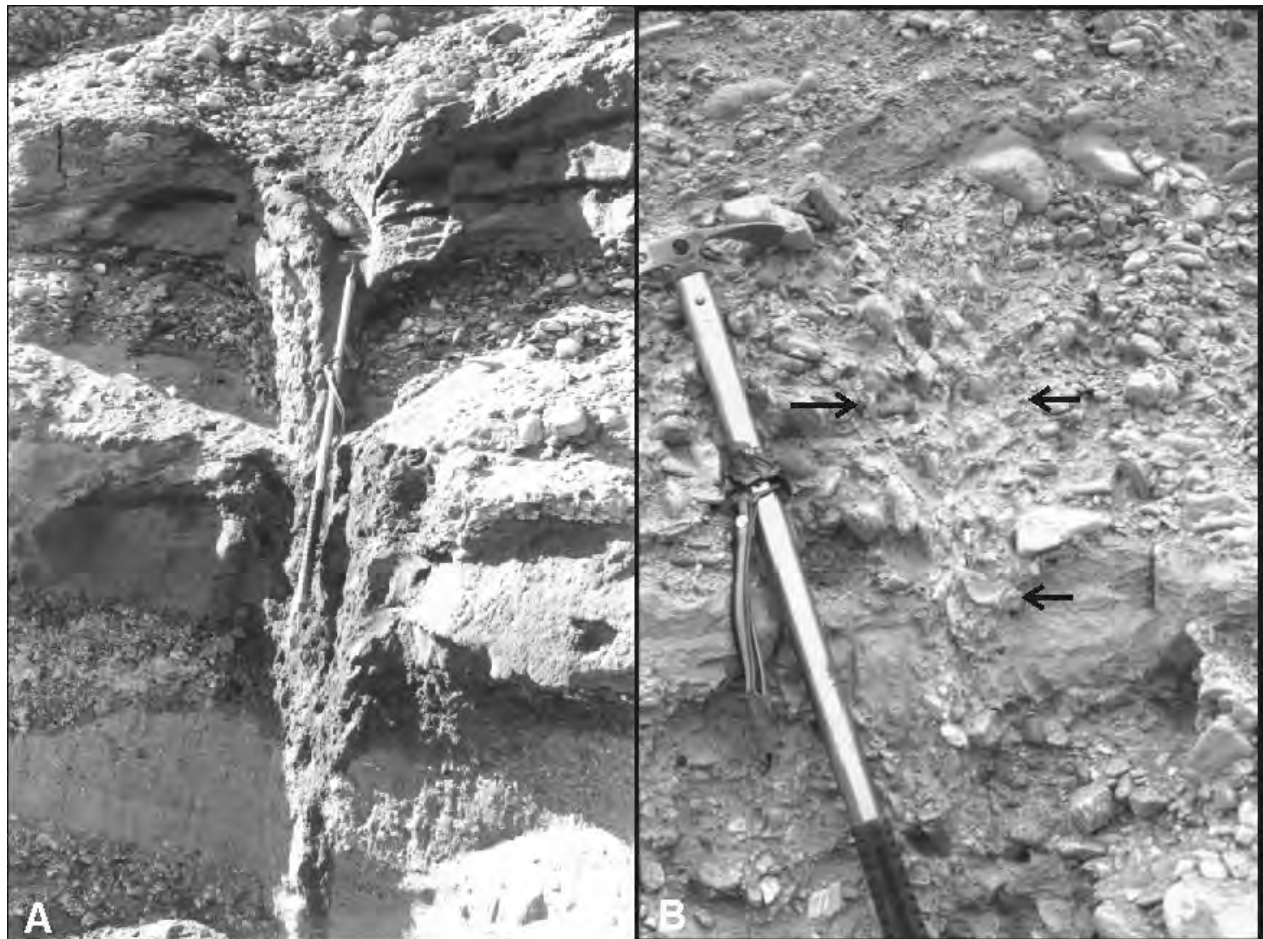


Figure 42. Ice wedge casts at Jackson Hill. A. Ice wedge cast located midway through Klondike gravel— indicating short-term braidplain abandonment and subsequent aggradation. B. Ice wedge cast developed in Upper White Channel gravel near the contact with Klondike gravel.

incision). Mean air temperature is likely to have been no warmer than present during formation and may extend to the continuous permafrost zone (mean air temperature -6°C).

Klondike Gravel

Klondike gravel is lithologically distinct from the locally-derived White Channel gravel. The presence of chert, various quartzites and slate clasts indicate a source in Selwyn continental-margin sedimentary rocks, found north of Tintina Trench, and quartz-feldspar porphyry found within Tintina Trench, and volcanics from south of the Tintina Trench. These rocks were transported by glacial meltwater into Klondike River valley, providing the first evidence of Cordilleran Ice Sheet glaciation in the region (Froese et al. 2000). The areal extent of this gravel spreads eastward to the Flat Creek beds (McConnell 1905; Figures 27 and 43), and westward along the Yukon River to at least the Alaska border (Duk-Rodkin 1996).

Klondike gravel consists of braided stream deposits which are largely massive and imbricate at Flat Creek, and show a greater frequency of planar-tabular crossbeds in the more westerly Klondike valley sites at Trail, Jackson and Australia hills (Figure 27), suggesting a more distal position from the palaeo-ice margin (Smith 1985). Palaeoflow measurements at Flat Creek indicate a source from the southeast, suggesting the Klondike gravel is derived from the first Cordilleran Ice Sheet advance into western Yukon, rather than local glaciation of the Ogilvie Mountains to the north (Figure 43).

Froese et al. (2000) document a reverse-normal-reverse polarity sequence through the White Channel gravel with reversed polarity above the ice wedge cast shown in Figure 42. The Klondike gravel is normally magnetized and correlated to the late Gauss chron (>2.6 Ma). The Jackson Hill tephra (0.13 Ma) occurs in a loess-filled gully incised into the Klondike gravel surface (Sandhu et al. 2001).

STOP 10: Paradise Hill: Upper White Channel stratigraphy and chronology (63° 59' N, 139° 04' W)

D.G. Froese and J. Westgate

Paradise Hill is located on the left limit of Hunker Creek between Hester and Last Chance creeks (Figure 2-1). Extensive open pit mining at the site has opened an approximately $\frac{1}{2}$ km wide exposure from bedrock. The site is important for a number of reasons: 1) it documents the relationship between alteration of the

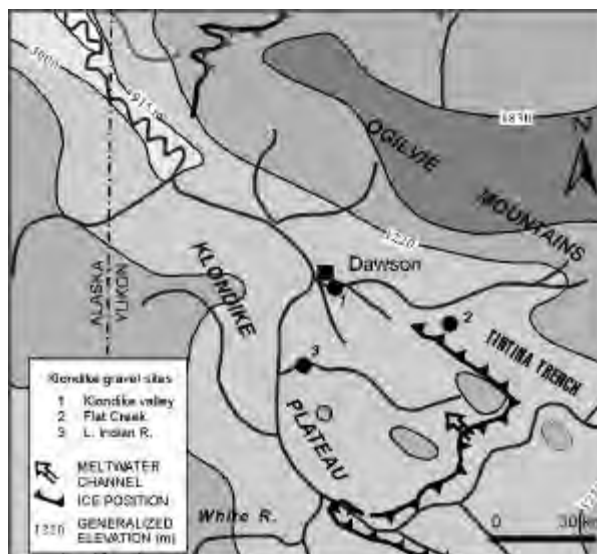


Figure 43. Distribution of Klondike gravel sites in the west-central Yukon. At each site the gravel is normally magnetized. The site on lower Indian River was deposited when the Cordilleran Ice Sheet overtopped a divide with the Stewart River, while palaeoflow indicators at the Flat Creek site indicate a source from the southeast (rather than the north if it had been local ice as Hughes et al. 1972 suggest).

gravel and neotectonics; 2) the Upper White Channel gravel at the site contains at least 2 tephra, including the Little Blanche tephra within the UWC gravel, and Paradise Hill tephra immediately overlying the UWC gravel; and 3) ice-wedge casts within the UWC are present at the site.

Strong alteration at the site makes the unconformity between Lower and Upper White Channel gravel especially pronounced (Figure 44). In the northeast end of the pit, at least two major bedrock thrust structures were measured emplaced into the Lower White Channel gravel up to 6 m, and extending at least 200 m (Figure 45).

A transition occurs at the 10 m level where fluvial sands (containing a tephra) are sharply overlain by a 10 cm organic silt. Polarity of both units is normal, and the organic silt contains an interglacial pollen assemblage consisting boreal-type spruce forest without pine.

Chronology

Two tephra beds are exposed at Paradise Hill. Little Blanche Creek tephra (UT1623) is preserved as pods, up to 5 cm thick, in a sand lens about 4 m below



Figure 44. The Upper White Channel gravel exposed in mining cut at Paradise Hill in 1996. Note unconformity between altered Lower White Channel and unaltered Upper White Channel gravels. Person for scale on lower right of photo near contact.



Figure 45. Bedrock structure emplaced into LWC gravel at Paradise Hill. Some sites at Paradise Hill show strong alteration of the gravel (see Figure 44), and others show little evidence of disturbance. UWC gravel overlying the bedrock structure is undisturbed by the intrusion, while the lower gravel is highly disturbed indicating that emplacement occurred pre-UWC (>3.2 Ma). Note person for scale on bedrock ridge.

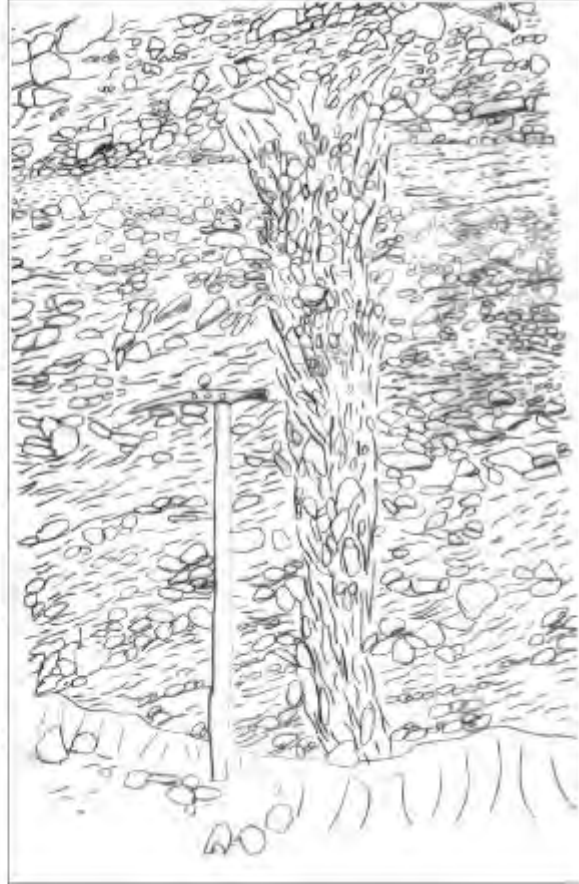
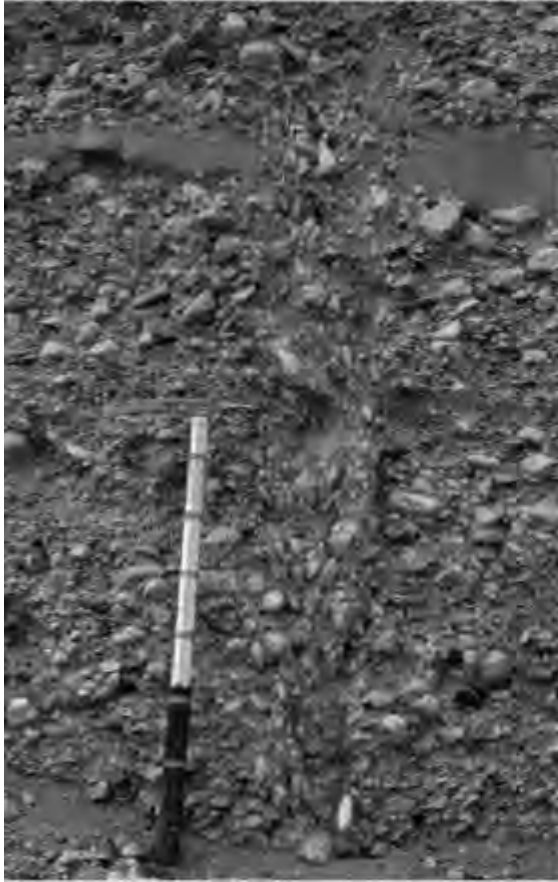


Figure 46. Ice wedge cast developed in Upper White Channel gravel at Paradise Hill. Ice axe in photo is 80 cm long.

the top of the Upper White Channel gravel, and Paradise Hill tephra (UT1624) has been reworked into multiple, thin, discontinuous beds over a stratigraphic interval of 20 cm in sands and silts about 50 cm above the top of the Upper White Channel gravel. Colluvial silt with organic-rich beds overlies Paradise Hill tephra. At Dago Hill, the mode of occurrence of Dago Hill tephra (UT1553) is very similar to that of Little Blanche Creek tephra, except that in this case only a wispy lens was preserved in the Upper White Channel gravel. These two tephra beds offer the potential of improving age controls on the Upper White Channel gravel, with the other, Paradise Hill tephra, likely giving a close minimum age for this unit.

Little Blanche Creek and Paradise Hill tephra beds are crystal-rich with abundant hornblende, plagioclase, and FeTi oxides and minor amounts of hypersthene, apatite and zircon. Their rhyolitic glass is mostly in the form of highly inflated pumice, although some chunky glass of low vesicularity is present but very rare in Paradise Hill tephra. These two tephra beds can be readily distinguished from one another on

the basis of their glass composition, each bed possessing a homogeneous population (Figure 10). Their compositional characteristics, including the weak Eu anomaly in the rare-earth-element (REE) profiles (Figure 10), demonstrate that they belong to the type II class of Preece et al. (1999) and come from vents in the Wrangell volcanic field of Alaska (Figure 9).

Dago Hill tephra is very different. It has abundant bubble-wall glass shards, mostly of a brownish hue. The small amount of material available prevented a representative description of its mineralogy. However, being a type I tephra bed (Figure 10), crystals would be sparse and consist of plagioclase and pyroxene with minor amounts of amphibole, magnetite, ilmenite, apatite and zircon. Glass shards have a rhyolitic composition. The silica content -- at a given Al_2O_3 concentration -- is low relative to the type II tephra beds and there is a distinct Eu anomaly in the REE profile (Figure 10); both indicate a type I identity (Preece et al. 1999), the source vent being located in the Alaska Peninsula - Aleutian arc region (Figure 9).

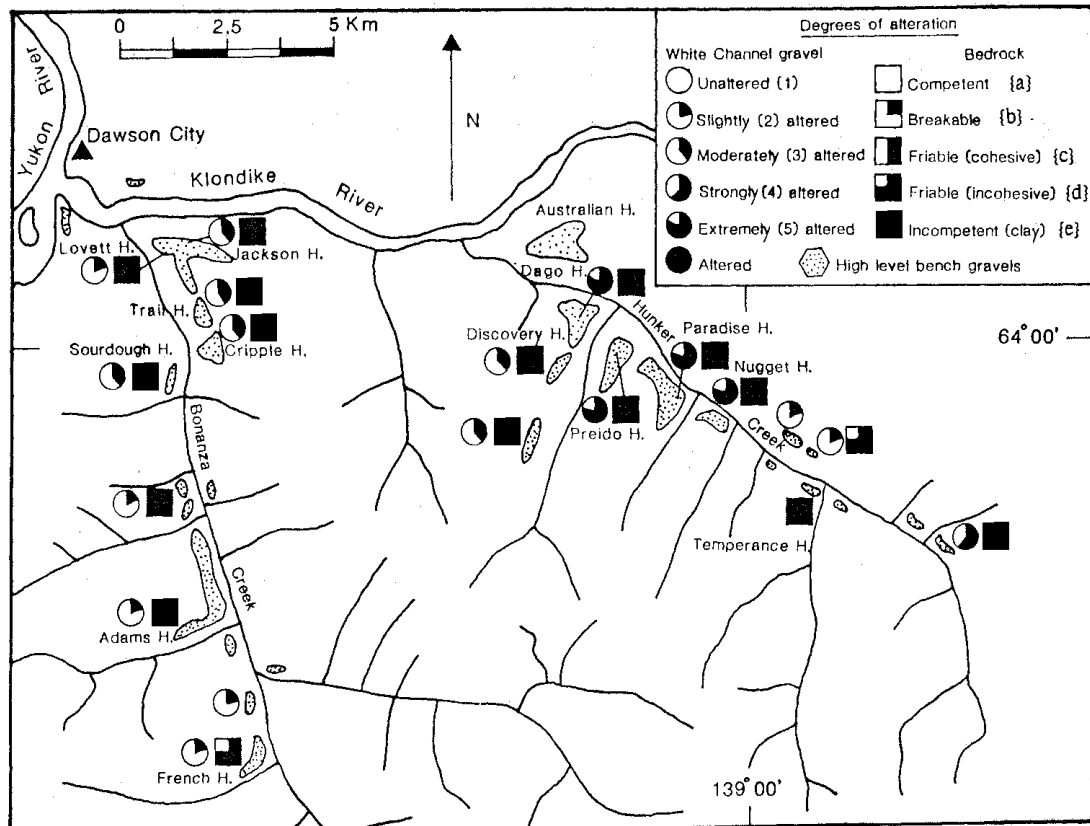


Figure 47. Distribution of alteration in the northern Klondike, (from Dufresne 1987).

Dago Hill tephra (UT1553) has a glass-ft age of 3.18 ± 0.41 Ma, which agrees with the normal polarity of sediments just above it. Therefore, the Upper White Channel gravel is of Late Pliocene age. Little Blanche Creek tephra at Paradise Hill is not datable by glass-ft methods because of its highly pumiceous glass, but the abundance of hornblende suggests that it might be possible to determine its age by the $^{40}\text{Ar}/^{39}\text{Ar}$ method. Paradise Hill tephra (UT1624) has a glass-ft age of 1.54 ± 0.13 Ma, which is considerably younger than expected, given its stratigraphic position immediately above the Upper White Channel gravel. The top of the Upper White Channel gravel at this locality must be defined by a significant unconformity.

STOP 11: White Channel Alteration: Paradise Hill

G.W. Lowey

This stop provides an opportunity to observe several of the more spectacular effects of alteration in the White Channel Gravel and to discuss the process, or processes, responsible for the alteration.

McConnell (1905, p. 83) described the White Channel gravel as consisting of "a compact matrix of small, clear, little-worn and often sharply angular grains of quartz and scales of sericite thickly packed with rounded quartz pebbles and rounded to sub-angular and wedge-shaped pebbles of sericite schist". He further noted (p. 83) that the "schist pebbles are usually decomposed and crumble rapidly when thawed out", and that the color of the White Channel gravel is "characteristically white or light gray due to the preponderance of the quartz constituents and the leaching out of the greater part of the iron". In a later report, McConnell (1907, p. 220) observed that the "bedrock underlying the White Channel gravel is more decomposed than in the creek". Regarding the distribution of gold, McConnell (1907, p. 219) concluded that the "greater part of the gold in the hill and creek gravels occurs on or near bedrock, either in the lower four to six feet of gravel or sunk for some distance in the bedrock itself". He noted also (p. 221) several exceptions to this general rule, such as at Dago Hill where a "considerable enrichment takes place at a point sixty feet above bedrock", and at Paradise Hill where the "main gold zone here is in many places not found on bedrock, but at elevations of from three to

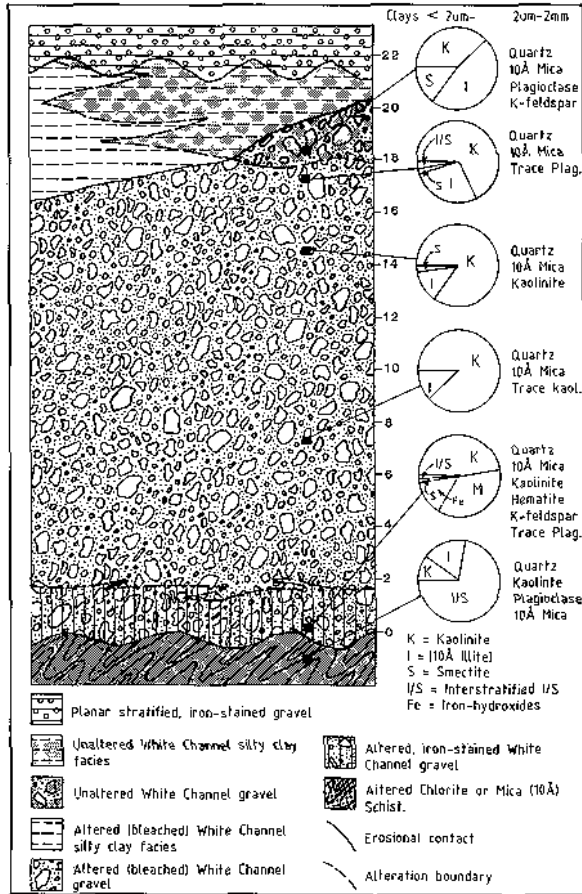


Figure 48. Alteration of Lower White Channel gravel at Dago Hill (from Dufresne et al. 1986).

twelve feet or more above it". McConnell (1905, p. 105) concluded that there is "little doubt that the Klondike gold, or the greater part of it, at least, is detrital in origin, and has been largely derived from the auriferous quartz veins cutting the older schists". However, he suggested (p. 106) that a "small percentage may have been precipitated from water carrying gold in solution", citing as evidence a boulder from the Sixtymile River area, the upper surface of which was partially covered with dendrites of gold.

Tempelman-Kluit (1982) thought that the leached matrix and decomposed pebbles in the White Channel Gravel were due to alteration from groundwater, and (p. 78) proposed that the "gold was deposited from the same groundwater that altered the rocks, and that this gold is a near-surface, low-temperature deposit". This prompted a study of the alteration and distribution of gold in the White Channel Gravel by Dufresne (1987). Dufresne (1987) recognized three alteration zones (i. e., bleached, iron and footwall) and concluded that, based on particle size analysis, XRD analysis, major

and minor trace element compositions, crystallinity of clay minerals, and fluid inclusion data of the gravel and/or underlying bedrock, the alteration and gold in the White Channel gravel were due to low temperature (i.e., 120-130° C), low salinity (i.e., 1-2 wt% NaCl) aqueous fluids characteristic of epithermal mineralization (Figures 47 and 48). However, Knight, Mortensen and Morison (1999a, b), using major and trace element compositions of gold particles from placer and lode deposits, demonstrated that the gold was in fact detrital in origin and mainly derived from mesothermal quartz veins, although they did not discount Dufresne's (1987) interpretation that the leaching and decomposition of pebbles in the White Channel Gravel were products of hydrothermal alteration.

If the gold in the gravel is no longer considered hydrothermal in origin, then why is there alteration? Note that alteration (including new mineral assemblages) produced by hydrothermal solutions (referred to also as wall rock alteration) may be very similar to that produced by weathering, diagenesis or metamorphism. Is the alteration of the White Channel gravel due to hydrothermal solutions, weathering, diagenesis, metamorphism, or a combination of some or all of these processes? How can we tell? And what was the dominant alteration process?

STOP 12: Last Chance Creek: Klondike Muck cryostratigraphy (63° 59' N, 139° 08' W)

E. Kotler and D. Froese

"Muck" is the term used in the mining districts of Yukon and Alaska for the unconsolidated fine-grained, ice-rich deposits that are found in valley-bottom sites overlying gold-bearing gravels. They have a distinctive odour, and are perhaps best known for their role in preserving ice age mammal bones, and in rare cases, articulated skeletons of extinct mammals (Storer this volume; Guthrie 1990). Despite their clear link to Ice Age environments, studies of these deposits and what they might reveal in terms of ice age climate in western Yukon only began in earnest a few years ago.

"Muck" is found on valley sides facing north and northeast and in the bottom of narrow valleys. In general, the thickness of the deposits increases southward from 1 to 10 m on lower Bonanza and Hunker Creeks, to over 20 m on Quartz and Dominion creeks (Figure 27). Kotler and Burn (2000) subdivided the "muck" into four sedimentary units: an organic unit, two sub-

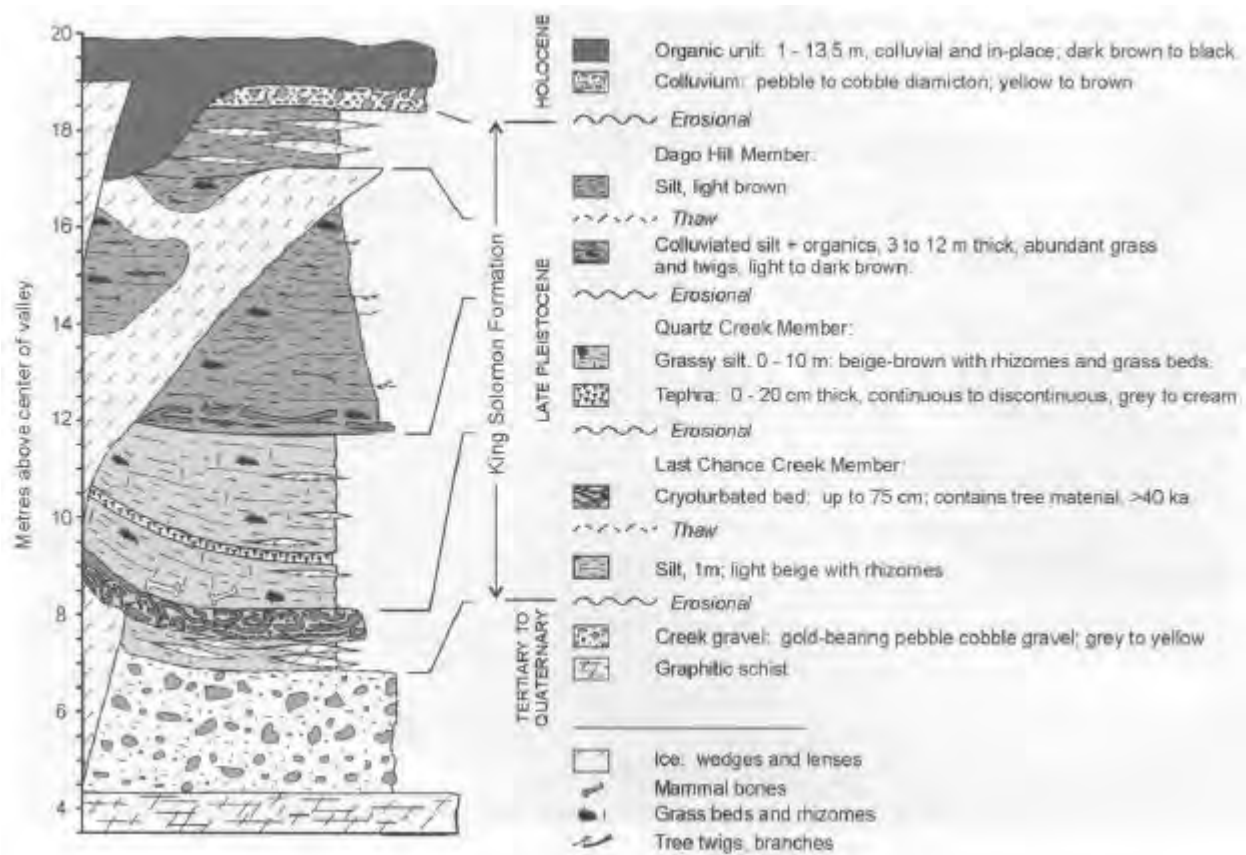


Figure 49. Generalized cryostratigraphy of frozen, unconsolidated sediments in the Klondike area, YT. This figure is a representative section compiled from the 21 sections examined by E. Kotler in 1997 and 1998. It combines the sedimentological and geocryological characteristics of the deposits and summarizes the stratigraphic position of the various cryostratigraphic units described in the text.

divisions of the Pleistocene silt of Fraser and Burn (1997), and an additional unit beneath it. Cryostratigraphically, the three lower units are considered distinct members of the King Solomon Formation (Figure 49).

Last Chance Creek Member

Pre-McConnell loess (probably mid-Wisconsin) was described from sections on Last Chance Creek. Radiocarbon ages on rhizomes from the base of the unit range from 45-40 ka and may be considered >40 ka. The unit contains discontinuous, planar ice lenses, up to 20 cm long, and narrow syngenetic or epigenetic ice wedges up to 4 m in height extend beyond the base of the unit and penetrate the underlying gravel

Quartz Creek Member

This member comprises the McConnell (late Wis-

consin) Pleistocene silt described by Fraser and Burn (1997). Radiocarbon ages range from 27 ka at the base of the unit to 13.9 ka near the top of the unit. The unit commonly overlies gravel, and is up to 10 m thick. The sediments consist of massive-to-bedded loess, and commonly contain the late Pleistocene Dawson tephra (Westgate et al. 2000). The relative absence of massive ice bodies and tree and shrub material distinguish the Quartz Creek Member from the underlying Last Chance Creek Member. Despite the absence of visible ice, the Quartz Creek Member is ice rich with a mean volumetric ice content of 65%.

Dago Hill Member

The redeposited (upper) portion of the silt unit of Fraser and Burn (1997) was observed throughout the study area to comprise loess interbedded with dark organic material such as peat, rhizomes, twigs and small branches making the sediments very dark brown

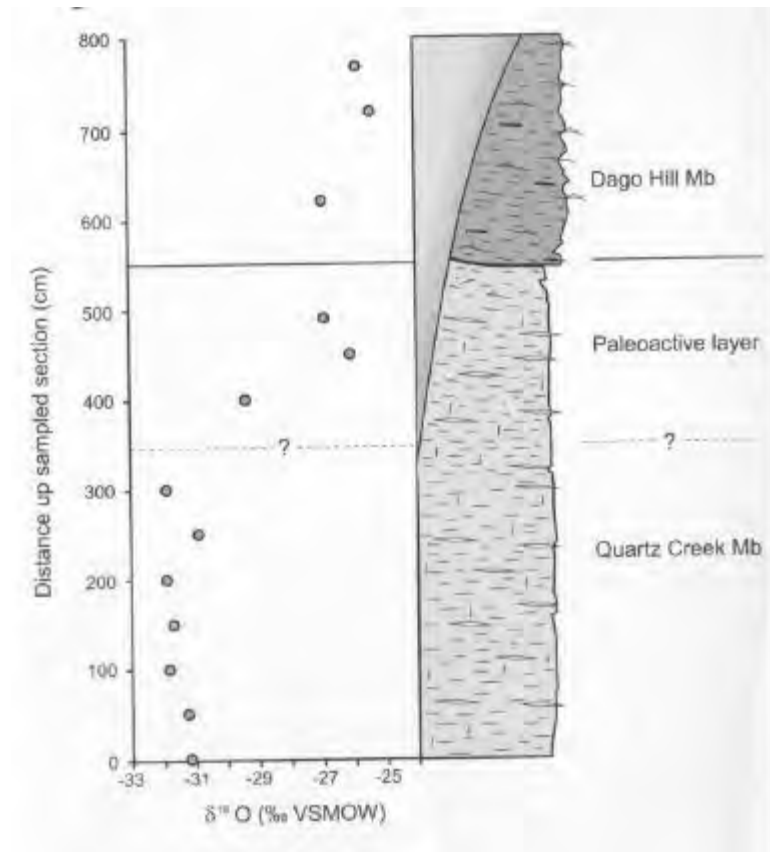


Figure 50. ^{18}O profile across the contact between Quartz Creek Member and Dago Hill Member at Quartz Creek. The data were obtained from water which was released upon thawing of the sediment samples. (from Kotler and Burn 2000).

to black. The sediments are up to 12 m thick and have a sharp contact with the Quartz Creek Member. Radiocarbon ages from the base of the unit range from 11.6 to 10 ka. The sediments are ice rich, and 10-20 cm long ice lenses parallel to bedding are common. The unit has a parallel-wavy layered cryostructure. Ice wedges up to 8 m in height are abundant within the member. Syngenetic and epigenetic ice wedges occur in valley bottoms, whereas very broad ice wedges, joined at their tops, are found on valley sides (Figure 50). These wedges, with upturned sediments at their margins, and vertical to nearly horizontal foliation, resemble anti-syngenetic wedges described by Mackay (1990). Organic and mineral material, as well as pool ice, have been incorporated into most of these anti-syngenetic wedges through erosion along their troughs. Bubbles typically occupy 30% of the ice by volume, are spherical to tabular, and commonly have orientations unrelated to foliation. The mean volumetric ice content of 43 samples collected from the Dago Hill Member was 70%.

Organic Unit

Disorganized organic material such as branches and peat makes up this surficial unit with a maximum thickness of 13.5 m (Fraser and Burn 1997). The lower contact with the King Solomon Formation is unconformable, irregular and sharp. Narrow syngenetic ice wedges occur at various levels within the unit, and can be found rejuvenating truncated ice wedges in the Dago Hill Member. At several locations, the tops of these wedges have been deformed downslope. The lower cryostratigraphic contact is a thaw unconformity. Ice wedges, originating at the base of the present active layer, are neither as large nor as abundant as in the Dago Hill Member.

Field Excursion Itinerary, Day 3: Southern Klondike Goldfields, August 27th, 2001.

STOP 13: King Solomon Dome

C. Schweger and D.G. Froese

Overview stop of radial drainage into Klondike River and south into Indian River.

Northern Limit of Lodgepole Pine

Pine (*Pinus*) was a major component of the pre-glacial (late Tertiary) forests of northern North America, but seems to have been largely absent during subsequent interglacials of the last 2.6 Ma (Schweger et al. n.d.). From palynological studies, the Holocene seems to be anomalous relative to previous interglacials in that Pine (*Pinus contorta* ssp. *latifolia* Cody 1996) has become a significant component of the boreal forest of Yukon today. Pollen evidence indicates that pine migrated 2200 km northward from south of the continental ice sheets to its present limit in central Yukon during the past 12,000 years, only crossing the border between 1500 and 2000 years ago. Pine reaches its northwestern limit in south-central Yukon (near Reid Lakes on the Klondike Highway) where it reproduces and spreads successfully following fire. A single tree is known in the Klondike at the summit of King Solomon Dome immediately south of the parking lot of the Ridge Road Heritage Trail (Figure 51).

STOP 14: Lode Deposits- Mitchell Deposit

G.W. Lowey

Approximately 311 Mt (13 million ounces) of placer gold has been extracted from the Klondike goldfields (Lowey 1998). The gold ranges from fist size nuggets (weighing up to 72 ounces) to minute flakes capable of floating on water, with a fineness ranging from about 750 to 800. This stop provides an opportunity to observe what is thought to have been one of the primary sources of placer gold in the Klondike goldfields, the Mitchell deposit.

There are only four types of lode gold occurrences in the Klondike goldfields (Knight et al. 1999a, b; Mortensen 1996; Rushton 1991): 1) Early Cretaceous mesothermal quartz veins (discordant type) in schist; 2) Syngenetic (i.e., Paleozoic) gold in pyritic schist; 3) Tertiary epithermal quartz fluorite veins in



Figure 51. A single lodgepole pine is known in the Klondike District near the parking lot of the Ridge Road Heritage Trail.

porphyry dikes and sills; and 4) Quaternary (?) low temperature epithermal quartz veins in schist (Knight et al. 1999b). According to Knight et al. (1999b, p. 662), “most if not all” of the placer gold was derived from the discordant mesothermal quartz veins, with minor contributions from the pyritic schist and none from the epithermal veins.

Typical discordant mesothermal quartz veins are exposed at the Lone Star and Mitchell properties. The Lone Star property, located near the headwaters of No. 7 Pup on Victoria Gulch, produced the only lode gold from the Klondike goldfields (approximately 40 kg of gold were extracted from underground workings in the early 1900s, Yukon MINFILE 1997) (Figure 52). Due to poor road conditions, access to the Lone Star property is generally difficult; however, similar features can be seen at the much more accessible Mitchell property.

The Mitchell property is located on a ridge north

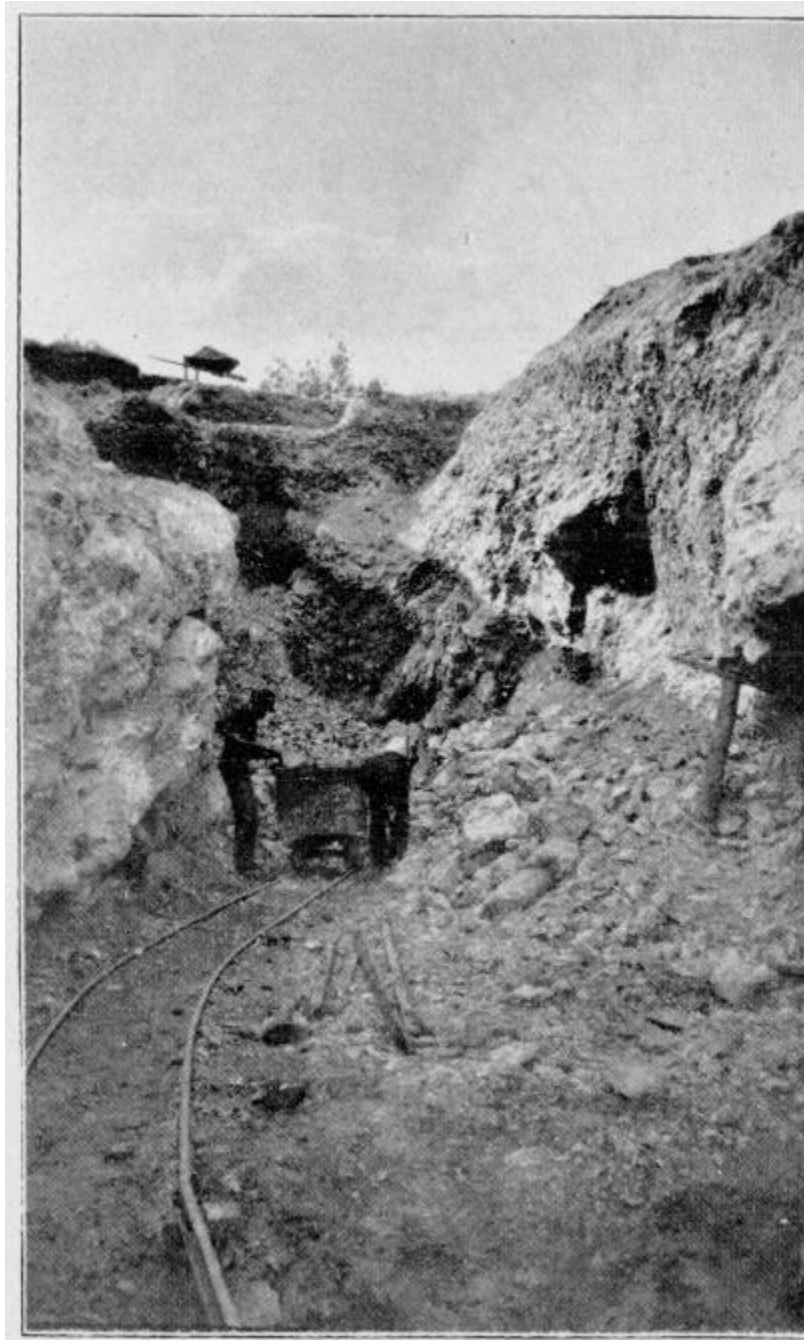


Figure 52. Photo of Lone Star workings in the early 1900s.

of King Solomon Dome. The following description is taken from the Yukon MINFILE (1997, 115O 068):

At the Mitchell showing, spectacular samples of free gold were reportedly found on the surface in the early days. The main showing consists of two parallel quartz veins striking 060° which cut the chlorite schist and impure quartzite, and have been traced for a length of 1 km. One vein is 1.2 to 2 m wide and barren, while the other is 10 to 45 cm wide and contains small pockets of visible gold along with rutile and coarse pyrite which has replaced euhedral magnetite porphyroblasts. The mineralized vein is surrounded by a pyritic alteration envelope.

Other minerals present in this and similar veins on the property include bornite, galena, sphalerite, tetrahedrite and arsenopyrite. Reported grades range from 1.4 to 48 g/t Au (Yukon MINFILE 1997).

Rushton (1991) concluded that the discordant quartz veins are earliest Cretaceous in age and formed from moderate temperature (i.e., 200-350°C), low salinity (i.e., <6 eq. wt% NaCl) aqueous fluids characteristic of mesothermal mineralization. He suggested that they were emplaced during uplift of the Klondike area in the Late Jurassic-Early-Middle Cretaceous.

STOP 15: Quartz Creek: Pliocene ice wedges/Quartz Creek tephra (63° 46' N, 139° 03' W)

J. Westgate and D.G. Froese

The Quartz Creek site is located 35 km southeast of Dawson City near the confluence of Little Blanche and Quartz creeks (Figures 12 and 27). It is an important site because of the presence of a tephra bed within an ice-wedge cast that penetrates the Upper White Channel gravel, which, locally, is covered by a colluviated facies of the Upper White Channel gravel (Figure 53). This association offers constraints on the age of the Upper White Channel gravel and the inception of permafrost conditions in the area.

This tephra bed has been named Quartz Creek tephra (UT1001). It is of the type II variety and comes from a volcano in the Wrangell volcanic field (Figure 9). This white tephra is very pumiceous and rich in phenocrysts of feldspar, amphibole, pyroxene and iron-titanium oxides. Glass shards have a homogeneous, rhyolitic composition, being enriched in K

relative to Little Blanche Creek and Paradise Hill tephra beds (Figure 10).

Quartz Creek tephra (UT1001, UT1634) at Quartz Creek (Figure 12) has a diameter-corrected fission-track (DCFT) age of 3.00±0.33 Ma and an isothermal plateau fission-track (ITPFT) age of 2.93±0.36 Ma, giving a weighted mean age of 2.97±0.24 Ma. This age estimate is compatible with its normal magnetic polarity (Gauss Chron) of the enclosing sediments (Froese et al. 2000). It is also in agreement with the ⁴⁰Ar/³⁹Ar age determination by Kunk (1995), which gives a minimum age of 2.71 Ma, a total gas (maximum) age of 3.01 Ma, and an isochron age of 2.64±0.24 Ma. Hence, a Late Pliocene age for this tephra bed and its enclosing sediment can be considered secure as is the age of the Upper White Channel gravel with its palaeomagnetic measurements (Froese et al. 2000) calibrated with fission-track ages that place its range between 2.6 - 3.3 Ma (Figure 54).

STOP 16: Dawson Tephra at Quartz Creek

J. Westgate

A 13-cm thick, creamy white tephra bed occurs in organic-rich silt that is exposed in the upper reaches of Quartz Creek (Figure 12). This is the Dawson tephra, the most prominent tephra bed in the late Cenozoic deposits of the Klondike and Sixtymile areas of the Yukon Territory. It has been identified at 16 sites and has a maximum thickness of 30 cm, but most values are in the range 15 to 30 cm (Figure 55), which suggests that its primary thickness is probably in this range. The roadcut along Sulphur Creek, just above its junction with Dominion Creek (Figure 55), is considered the reference area. Exposures in this area are very accessible and have survived for over 30 years.

Dawson tephra consists mostly of thin, bubble-wall shards. Orthopyroxene, plagioclase, magnetite and ilmenite are relatively abundant in the small crystal component, which also contains minor amounts of clinopyroxene, apatite and zircon. The glass has a rhyolitic composition and is very different to that of Sheep Creek tephra and White River Ash, which have both been recognized in the Yukon, but is more similar to that of Old Crow tephra, although the latter has a distinctly higher SiO₂ content (Figure 56).

Dawson tephra belongs to the type I group and so is derived from a volcano in the Aleutian - Alaska Peninsula region. Its close resemblance to Old Crow tephra (Figures 11 and 56) suggests a common source.

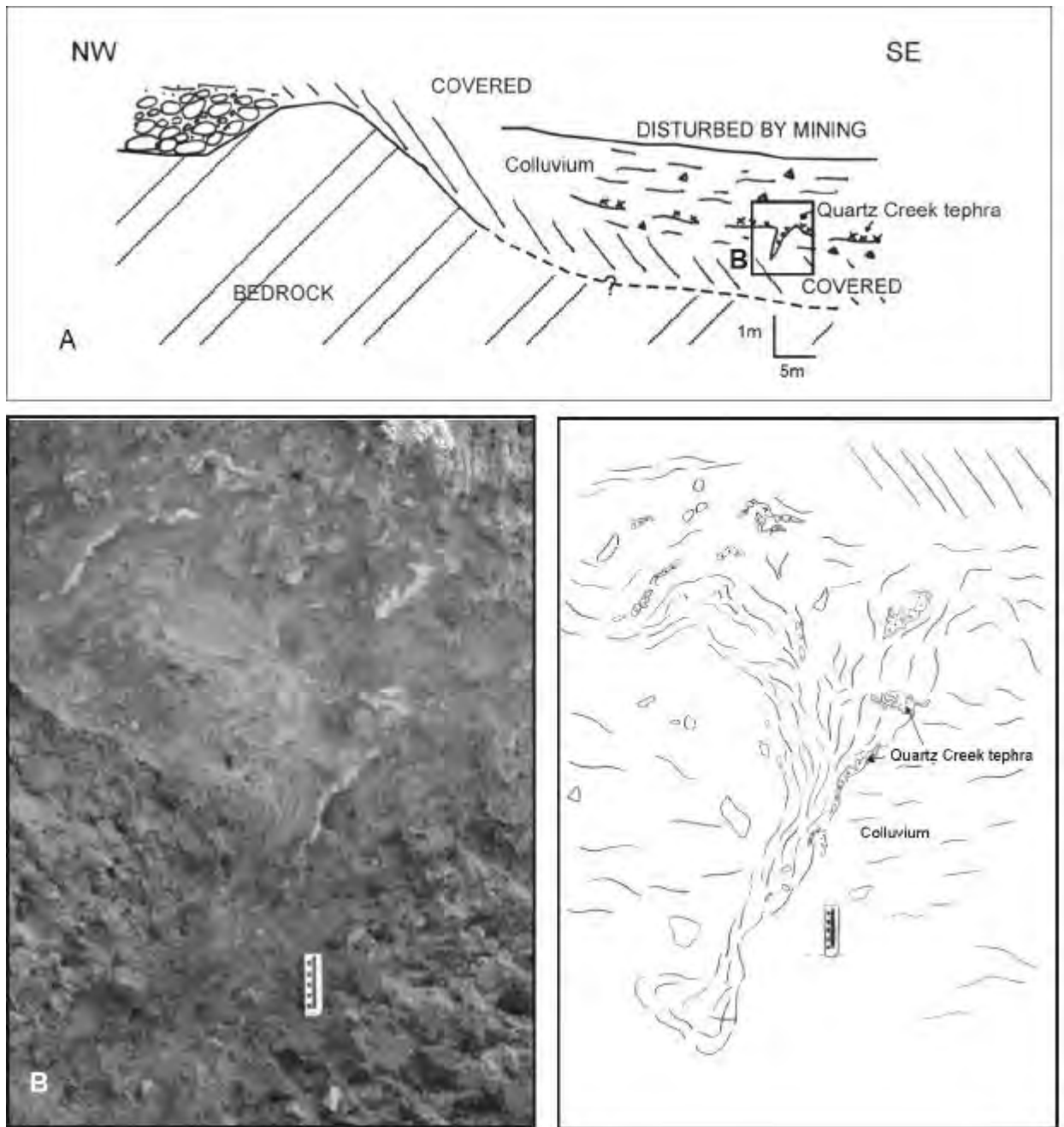


Figure 53. Generalized stratigraphic relation between Quartz Creek tephra, colluvial sediments, and White Channel gravel at site along Quartz Creek (Figure 27). B. Photograph and sketch of Quartz Creek tephra. The tephra has a diameter-corrected fission-track (DCFT) age of 3.00 ± 0.33 Ma and an isothermal plateau fission-track (ITPFT) age of 2.93 ± 0.36 Ma, giving a weighted mean age of 2.97 ± 0.24 Ma. These ages provide a minimum age for the White Channel gravel in the Klondike region and secure a late Pliocene age for the first occurrence of permafrost in the Yukon.

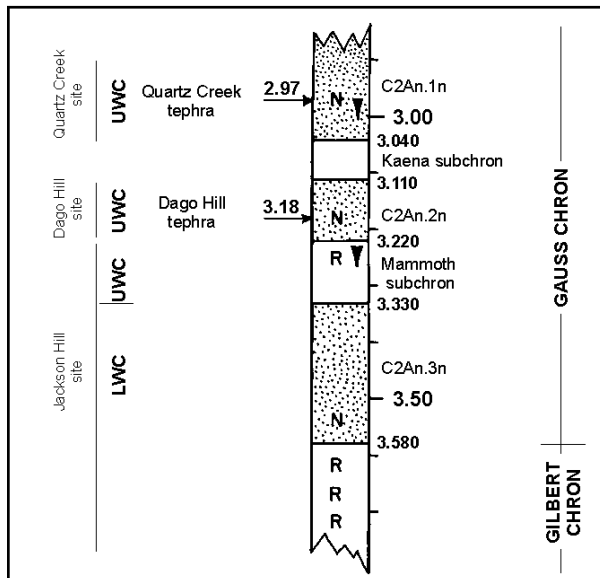


Figure 54. Proposed correlation of the Upper and Lower White Channel gravels with the geomagnetic polarity timescale of Cande and Kent (1995). Ages in Ma.

Given that the nearest vent in the Alaska Peninsula is 700 km away from the Klondike district and that the primary thickness of the Dawson tephra in this region is likely in the range of 15 - 30 cm, it follows that this explosive eruption must have been of great magnitude, distributing tephra over southern and central Alaska and across the Yukon.

A single fission-track age determination suggests that Dawson tephra is younger than 52,000 years (Naeser et al. 1982) but more recent stratigraphic studies by Kotler (1998) provide a more precise age estimate. Dawson tephra occurs within loess deposited between 28 ka and 13 ka, demonstrating a Late Wisconsinan age. At the Ready Bullion Gulch site, where UT1612 was collected (Figure 55), two AMS ^{14}C dates on small twigs and grass bracket the tephra. Organics immediately below and a few cm above the tephra have uncalibrated ^{14}C ages of $22,300 \pm 190$ (TO-6967) and $23,520 \pm 210$ years BP (TO-6968), respectively. Stratigraphic reversal of these ^{14}C dates is probably due to colluvial processes transporting older material downslope to overlie younger organics. Hence, Dawson tephra is younger than 22,300 years but older than 11,620 years, the age of organic material overlying Late Wisconsinan loess (Kotler 1998).

STOP 17: Lucky Lady Placers, Sulphur Creek: (63° 44' N, 138° 51' W): Mid-Wisconsin environments of the Klondike and central Yukon

D.G. Froese, C.A. Schweger, C.A.S. Smith, S.J. Preece and J.A. Westgate

The Lucky Lady mine is located at the confluence of Brimstone Gulch with Sulphur Creek and extends south to the opposite side of the Sulphur Creek valley (Figure 57). Mining at the site (observed between 1998 and 2000) has exposed an area immediately below the Sulphur Creek Road. The site consists of 3 m of rusty-brown Dominion Creek gravel overlain by 3-6 m of loess and retransported loessal sediments with organic horizons which includes the Lucky Lady tephra (LLt) and a syngenetic ice wedge complex (Figures 58 and 59).

The sediments are divided into 7 units:

Unit 1: 0-2.5 m, Massive gravel, with weak planar-tabular cross-bedding. The upper 30 cm of gravel contains detrital wood, and retransported peats derived from the local hillslope.

Unit 2: 2.5-2.8 m, Fibrous, organic-rich silt, weak horizontal stratification with rare interbedded clasts derived from the hillslope. Unit interpreted as retransported forest-floor organics with loess.

Unit 3: 2.8-3.4 m, Tan-olive disorganized silt with organic silts, and laminae of tephra marbled through the upper surface. This unit is interpreted as a result of cryoturbation (frost-churning) and solifluction of organic material under continuous permafrost conditions with ice-veinlet infillings of remobilized tephra. (Figure 60).

Unit 4: 3.40-3.45 m, Lucky Lady tephra. LLt is up to 15 cm thick (in retransported strata), however its primary thickness was likely closer to 3 cm. Tephra is laterally continuous across the section with a wavy relief of up to 40 cm reflecting the palaeo-surface variability which the LLt fell upon. The tephra is draped within an ice-wedge cast, suggesting melting at least part of the ice-wedge shortly after the tephra eruption.

Unit 5: 3.45-3.47 m, Continuous fibrous organic silt. Immediately overlies the LLt across the exposure and is interpreted as a former tundra surface (palaeo-cryosol).

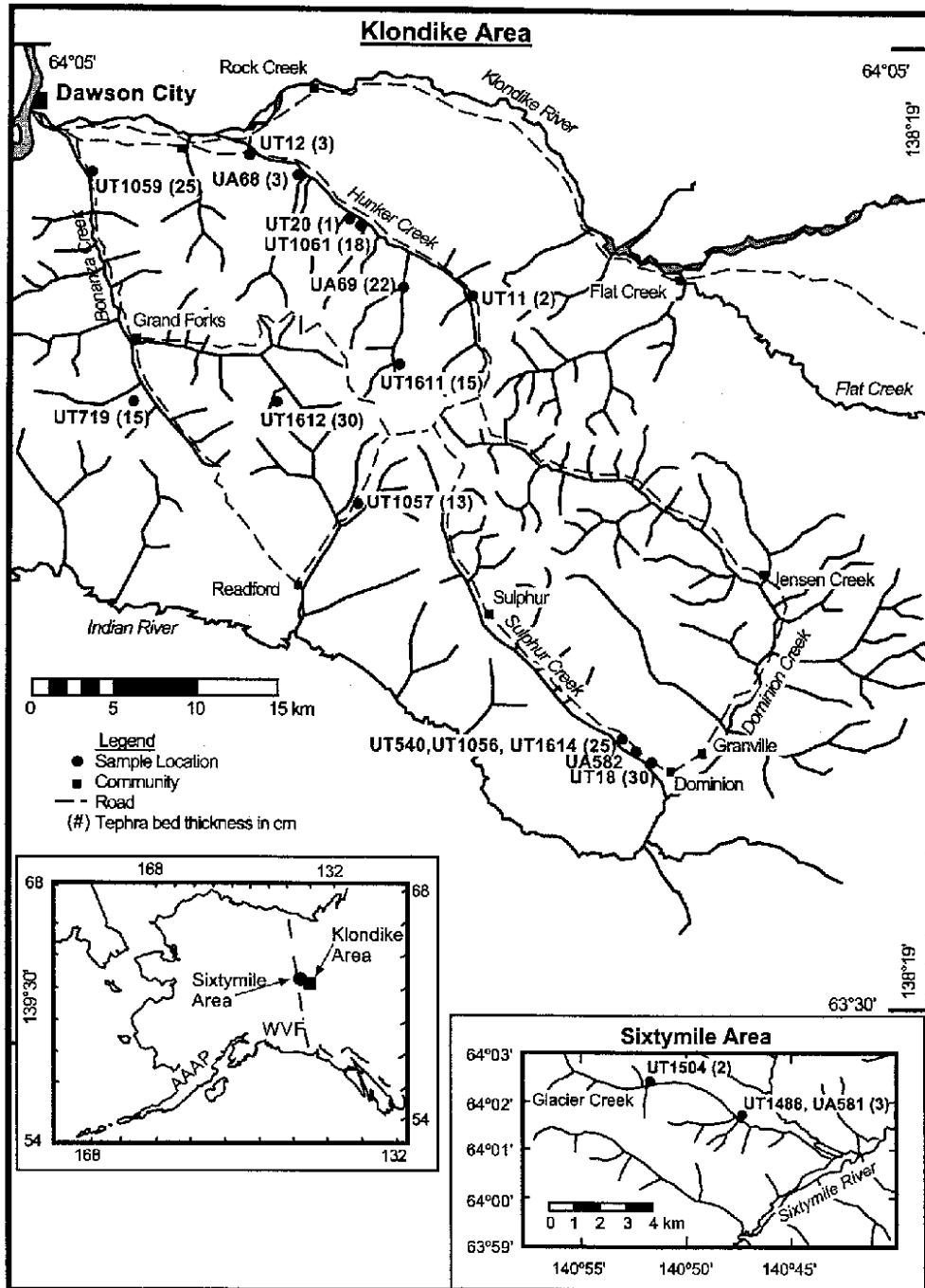


Figure 55. Location of samples and thickness of Dawson tephra in the Klondike and Sixtymile areas, Yukon. AAAP is Aleutian arc - Alaska peninsula region; WVF is the Wrangell volcanic field.

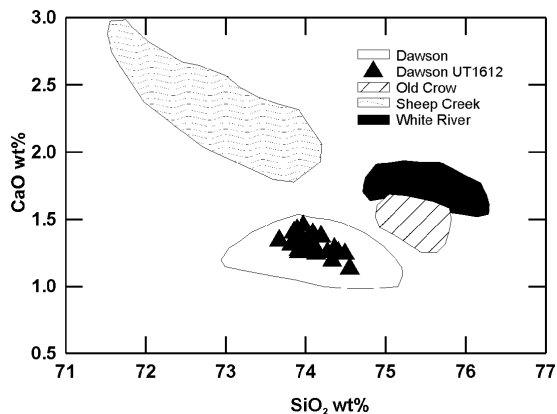


Figure 56. Oxide variation diagram showing differences in glass composition between Dawson tephra and other well-known Quaternary tephra beds in Yukon. UT1612, the tephra sample with the ^{14}C control at Quartz Creek, is highlighted in the field of all other glass analyses of Dawson tephra.

Unit 6: 3.47-4.5 m, Olive-grey, crudely-stratified silt with rare organic pods. The silts are very dense and the organic pods (3-5 cm thick) are distributed throughout, but form no coherent horizon. The sediments are strongly reduced (gleyed horizons) and cut by syngenetic ice-wedge casts. Interpreted as a surface which aggraded under continuous permafrost conditions with active solifluction and primary loess deposition.

Unit 7: 4.6 – 6 m, Massive and retransported loess. Upper surface disturbed by mining.

Palaeoecology

Nine organic samples were collected through units 1, 2, 3 and 4, of which only 3 returned a countable sums. Sample DF99-59 collected from an organic pod within the upper 30 cm of the gravel unit was the most productive. The sample contained 34% *Picea* with nearly 40 % composed of *Artemisia* and grasses.

Chronology

Two radiocarbon ages were determined from the site. A piece of wood from the upper 30 cm of Unit 1 had AMS age of $49\,390 \pm 1350$ (TO-7943), and a sample immediately above the LLt at the site (Beta-128237) had an AMS age on wood of $48\,370 \pm 1400$. Given that these ages are at about the ^{14}C dating maxi-

mum they must be treated with some caution since they may be infinite with some younger carbon contamination. However, given the similarity of the ages (samples overlap at 1σ) and that they come from two separate labs (Isotracer and Beta) we have no convincing reason to reject the age of the sediments at ca. 50 ka. However, the composition of the LLt, with its similarity to Sheep Creek tephra (190 ka) suggests a much greater antiquity for the LLt that the radiocarbon ages suggest.

At present, the LLt has not been found elsewhere in Yukon or Alaska and thus cannot be used to supply chronology yet. Two samples of tephra were collected from small retransported pods above the LLt within Unit 6 in 1999 that are still awaiting analyses. We hope that one of these samples will correlate with a previously recognized tephra from either Yukon or Alaska and either corroborate the ^{14}C ages presented here, or provide new chronology. If we accept the ages on face value, this mid-Wisconsin pollen sequence is the oldest finite sequence in eastern Beringia (Anderson and Lozhkin 2001).

Summary

The Lucky Lady sequence is interpreted as a hill-slope record of a cooling cycle (Dansgaard-Oeschger event?) during the mid-Wisconsin with the following sequence of events:

1. Deposition of Dominion Creek gravel and concentration of placer gold derived from Brimstone Gulch and Sulphur Creek during a 'warm' interval of the mid-Wisconsin (likely prior to 50 ka BP) (Unit 1).
2. Transportation of forest-floor organics and hillslope deposits downslope to valley-bottom in response to cooling conditions (Unit 2).
3. Development of continuous permafrost conditions, deposition of Unit 3, development of ice-wedge.
4. Frost churning of Unit 3, deposition of Lucky Lady tephra (Unit 4), development of Arctic soil (Unit 5).
5. Melting of ice-wedge (at least partly), disruption of tephra and associated sediment to form an epigenetic ice wedge cast.
6. Increased sedimentation to Brimstone loess-fan under continuous permafrost condition, with continued syngenetic ice-wedge formation.

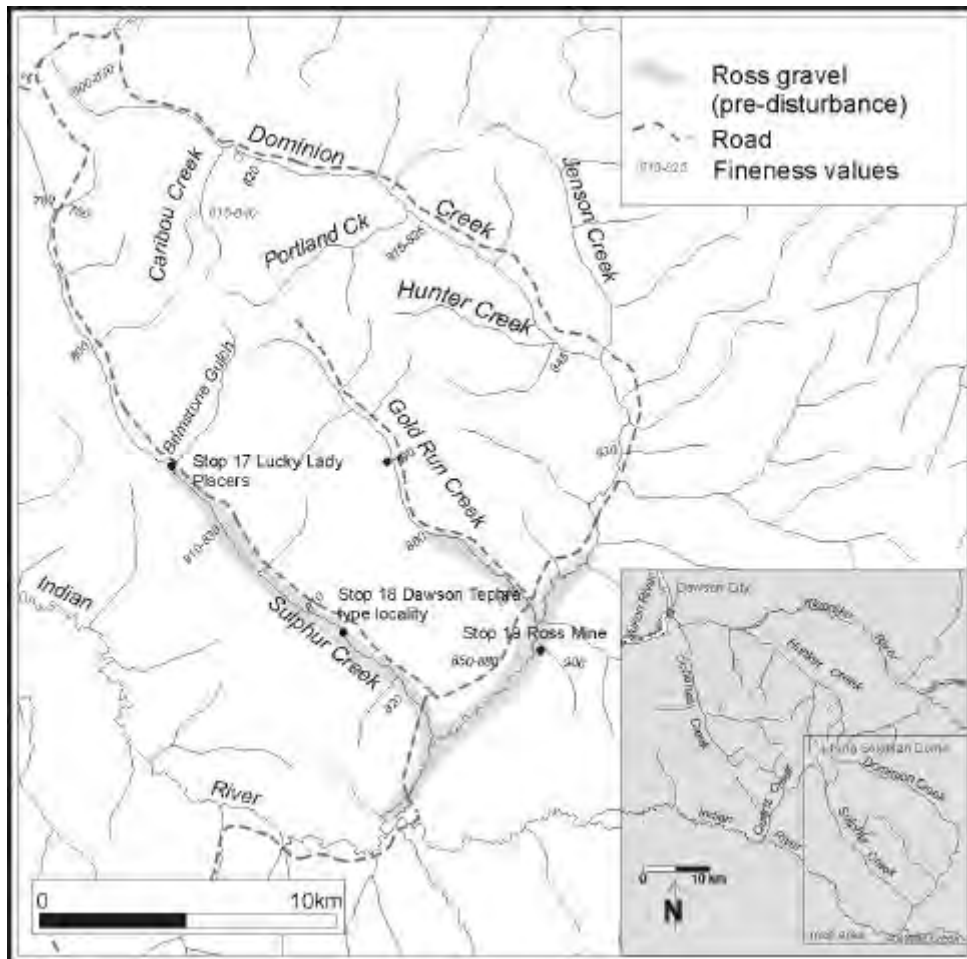


Figure 57. Map of Dominion Creek area. Shaded area marks distribution of the early-Pleistocene Ross gravel in the Dominion Creek area, and fineness values of placer gold recovered.



Figure 58. Fine-grained (loess and retransported loess) sediments overlying fluvial gravel at the Lucky Lady site. Lucky Lady tephra forms a prominent horizon across the exposure and is disrupted into an ice-wedge cast (Figure 59). Radiocarbon ages on wood from the upper gravel and above the tephra are finite (ca. 50 ka BP).

STOP 18: Dawson Tephra Type Locality (Westgate) (63° 41' N, 138° 43' W)

Immediately south of the Lucky Lady mine is the type locality of Dawson tephra (see above description). The roadcut along Sulphur Creek, just above its junction with Dominion Creek (Figures 27 and 57), is very accessible and has survived for over 30 years.

STOP 19: Dominion Creek: Ross Mine (63° 41' N, 138° 34' W)

Duane Froese, J. Westgate

Dominion Creek is the largest tributary of Indian River, and forms the southeastern boundary of the Klondike Placer District (Figure 57). Dominion Creek and its tributaries (principally Sulphur and Gold Run creeks) reported production of roughly 450 000 ounces of raw gold between 1978 and 1997 (Mineral Inspections Division 1998) making it the largest placer producing area in Canada over that period. This value, however, represents a small fraction of the area's total production since discovery in 1896, which is likely close to 3 million ounces. The following description is taken from Froese et al. (2001a).

Gravel Stratigraphy

The stratigraphy of Dominion Creek fluvial deposits is similar to the sequence of terraces known in the north Klondike. The gravels are divided into 1) Pliocene White Channel gravel, 2) Pleistocene terraces, 3) incised-valley-fill gravel (Ross gravel), 4) Dominion Creek gravel, and 5) gulch and stream deposits. These relations are shown in Figure 61, and photos of the deposits in Figure 62.

The major difference from the south Klondike is that the valley-bottom on Dominion Creek, on the basis of a reversed magnetization of the Ross gravel (indicating an age of >780 ka), incised to present level by at least 780 ka ago. In contrast, valley-bottoms of the north Klondike were perhaps 50-100 m above their present level suggesting greater uplift of the north Klondike relative to the south over the last ca. 1 Ma.

Placer Settings and Gold Character on Dominion Creek

The Dominion Creek basin is located within the Yukon-Tanana Terrane and consists largely of meta-sedimentary and meta-volcanic rocks at chlorite-

biotite to garnet metamorphic grade (Mortensen, 1990, 1996). Lode gold occurrences are associated with metavolcanic rocks of the Klondike Schist and mesothermal quartz veins (Mortensen et al. 1992). The erosion of mesothermal quartz veins appears to be the main source of the Klondike placer deposits based upon elemental similarities (microprobe geochemistry) between placer and lode gold (Knight et al. 1999b). Erosion of bedrock sources and transport by fluvial processes is supported on Dominion Creek by hydraulic equivalence data amongst gravelly depositional unit grain size and size/weight of gold grains recovered from placer gravel (Christie 1996).

Fineness values on Dominion Creek (plotted from Mining Inspection Division 1998; Figure 57) show considerable similarity on each of Sulphur (750-830), Gold Run (790-850) and main Dominion creeks (800-900) and generally increasing down-valley as has been noted previously in the Klondike region (Hester 1970; Knight et al. 1999b). The increase in down-valley fineness likely reflects prolonged mechanical weathering of gold grains increasing high-fineness rims. Gold morphology data, presented by Knight et al. (1999a) suggests that flat, well-rounded gold, like the majority of that recovered on Dominion and Sulphur creeks, was transported 10-15 km, indicating a major source in the area of King Solomon Dome. A high fineness lode source is well known on King Solomon Dome (McConnell 1905; Milner 1977, Knight et al. 1999b).

The majority of gold produced on Dominion, Gold Run and Sulphur creeks in the last century has been from Ross gravel. On Dominion Creek, Ross gravel is at least 800 000 years old, suggesting little gold has been eroded or concentrated in the last 800 000 years in this area. This contrasts with the majority of gold produced on Bonanza and Hunker creeks in the Klondike drainage where deposits are largely of late Pleistocene and Holocene age (valley-bottom gravel/muck ages reported in Fraser and Burn 1997 and Froese 1997).



Figure 59. Ice wedge cast complex associated with Lucky Lady tephra. The tephra is draped within the ice wedge cast suggesting an initial ice wedge developed which subsequently melted depositing the tephra within an epigenetic cast. Subsequent aggradation of the surface resulted in the development of a syngenetic cast with decreasing organic material.

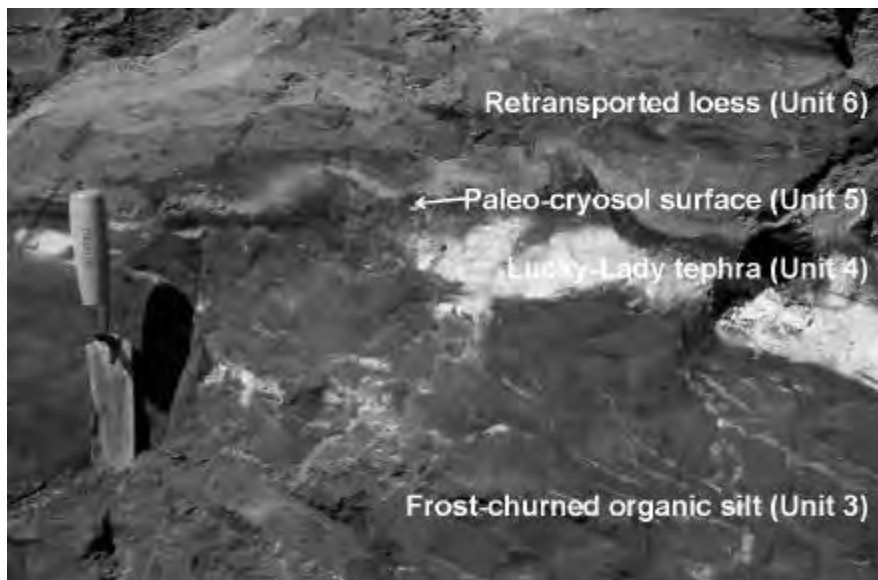


Figure 60. Close-up view of Lucky Lady tephra. Unit 3 is a tan-olive disorganized silt with laminae of tephra marbled through the upper surface. This unit is interpreted as a result of cryoturbation (frost-churning) under continuous permafrost conditions with ice-veinlet infillings of remobilized tephra. The upper surface of the tephra has a thin (1-4 cm) fibrous organic horizon (Unit 5) interpreted as a palaeo-cryosol which was the active surface during frost-churning of the tephra. Trowel in photo is 15 cm long.

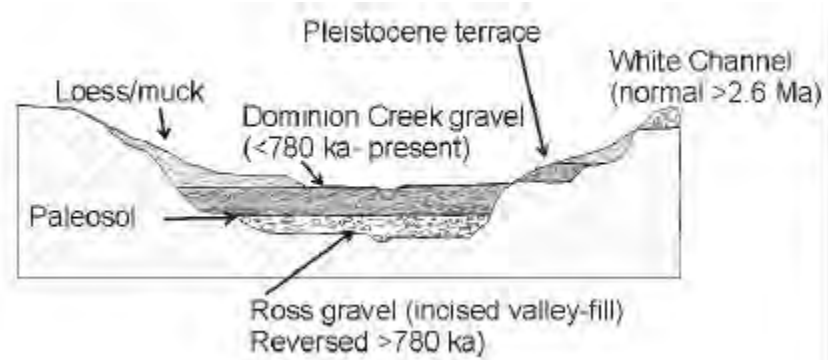


Figure 61. Schematic cross-section of gravel stratigraphic units on Dominion, lower Sulphur and Gold Run creeks.

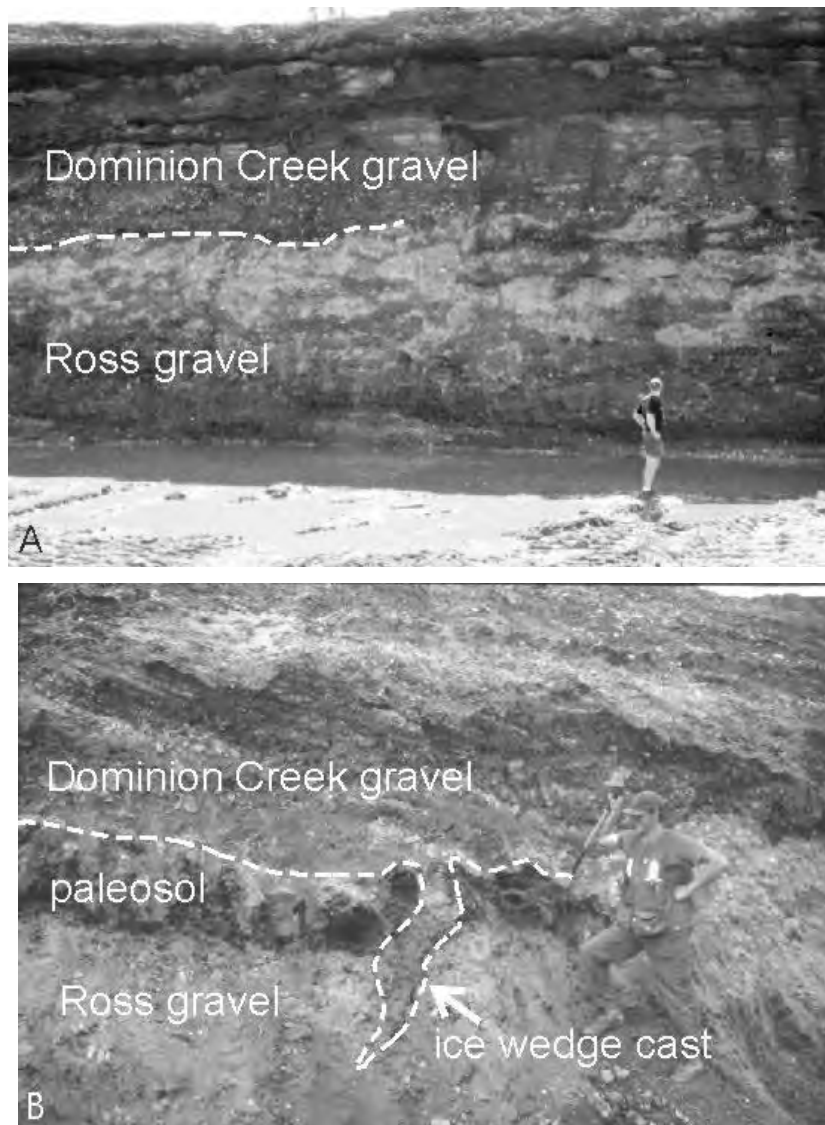


Figure 62. A and B: Exposures of Ross and Dominion Creek gravel units. The lower Ross gravel is magnetically reversed and shows evidence of a later alteration event from a normal overprint isolated from its palaeosol. Its magnetization (reverse) indicates deposition prior to 780 ka ago. The Dominion Creek gravel has abundant iron-staining and shows an unambiguous normal magnetization.

Field Excursion Itinerary, Day 4: Fifteenmile River, Rock Creek and Tombstone, August 28th, 2001

We will be starting in Dawson, flying (by helicopter) along the Yukon River to view a palaeo-Yukon River terrace from the air, and then continue northward to the East side of Fifteenmile River exposure (EFR). Walk from the landing site to the west to view exposure of Tertiary and Quaternary stratigraphy.

Continue the trip, flying along the Tintina Trench to view other landslide scars exposing Tertiary and Quaternary strata on the way to the Rock Creek Site (RC). At the RC Site we will visit only the 2b part of the exposure. Site 2a will be seen from the air on the way back to Dawson in the evening.

STOPS 20 and 21: Tintina Trench Sites (EFR and RC): *The exposures contain a middle Miocene to middle Pleistocene stratigraphic record composed of fluvial preglacial sediments, multiple tills, outwash, loess, and palaeosol sequences.*

A. Duk-Rodkin, R.W. Barendregt, J. White, S. Smith, and C. Schweger.

A long stratigraphic record extending from late Cretaceous to mid Pleistocene has been preserved along the walls of the Tintina Trench, and is exposed in modern landslide scars or along slopes. The sites are located on both sides of the trench along a stretch of approximately 100 kilometres starting about 30 kilometres from the Yukon/Alaska border (Figure 63). Stratigraphic reconstruction was carried out at many localities within the trench area. Sites with most relevant information are at Fifteenmile River (the west side, from hereon referred to as WFR, and the east side referred to as EFR), Grouse Mountain (GM), and Ditch Road (DRd; Figure 64). Each locality contains many sites and transects which have been arranged into composite stratigraphic sections shown in Figure 64. These sections show a partial or complete stratigraphy of two groups of beds. These two groups are separated by an unconformity containing tilted fluvial beds of Tertiary age below an unconformity, and conformable fluvial and glacial strata above the unconformity. We will visit the two sites (EFR and RC in Figure 65) which contain the most complete record. Stratigraphic work at these sections was done following Quaternary mapping of the Dawson area in 1996 by A. Duk-Rodkin. No previous stratigraphic

work had been carried out at these cuts. Older Eocene coal-bearing fluvial strata have been described along the Tintina Trench by Green (1972) and Hughes and Long (1980). Our studies of these sections included sedimentology, palaeomagnetism, palynology and palaeosols.



Figure 63. Field trip and location map.

Importance of the sites

The most important aspect of these sections is that they record the stratigraphy of the drainage evolution system in west-central Yukon. This evolution relates to the formation of major late Miocene extensional faulting in the Tintina Fault area which triggered early Pliocene incision-aggradation cycles of drainage and gold placer formation in the palaeo-Yukon River basin. The sections also record the first glaciation in west-central Yukon as well as another six glacial/interglacial events, making this the most complete terrestrial record of glaciations and interglaciations in North America.

STOP 20: EFR (Fifteenmile River east site; Figure 63) ($64^{\circ} 23' N, 139^{\circ} 46' W$)

Stratigraphy

This section shows well exposed beds below and above the unconformity. The lower sequence has been grouped into two parts: a lower recessive sequence of claystone and very coarse grained pebbly sandstone with coal beds, and an upper sequence of interbedded

conglomerate, sandstone, organics (pond deposits), and in most places indurated silt beds at the top. The beds analyzed here relate to the upper sequence since no detailed work was carried out on the lower one. The sequence above the unconformity is represented by Pliocene fluvial preglacial beds at the base which are overlain by glacial strata of tills, outwash and loess separated by palaeosols.

Stratigraphy Below Unconformity

Unit 1: Fluvial: Alluvial fan deposits composed of interbedded conglomerates, sandstones, and peats, make up Unit 1, with conglomerate beds being dominant in the basal unit of this section. They appear tilted by various amounts. The deposits depict facies ranging from proximal, to midfan, to distal, depending on their location in the trench.

Proximal deposits are represented at WFR site (Figure 66), where they consist of coarse, massive conglomerate beds with thin interbedded sandstones (1-2 cm thick). Proximal to midfan deposits are found at EFR, and consist of massive conglomerate beds intercalated with rare 1 metre thick sandstones (Figure 67). Distal deposits are found at the GM and DRd sites where sandstone beds are interbedded with small sized conglomerates and peat beds. The source of these sediments was in the Ogilvie Mountains to the north, about 9 to 50 kilometres distant. The total thickness of the Tertiary sequence occurring below the unconformity has been estimated to be about 900 m by Prindle (1913) and 700 m by Hughes and Long (1980). Prindle and Hughes identified two sequences in the Tertiary package, a lower recessive sequence represented by fine grained clastics (sandstone, clay, shale and lignite) and an upper sequence of conglomerate and sandstone. The upper sequence has a total

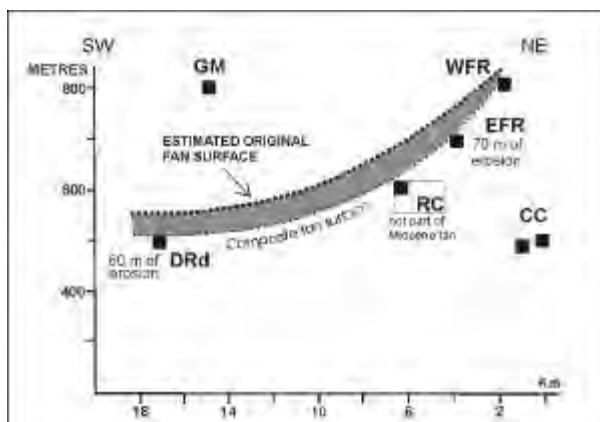


Figure 66. Reconstruction of Middle Miocene fan profile in the Tintina Trench.

thickness of 250 m at the Fifteenmile River sites.

At the EFR section below the unconformity, an assemblage containing abundant Pinaceae, plus *Tsuga* sp. cf. *T. canadensis*-type, *Sciadopitys*, *Alnus*, *Carya*, *Pterocarya*, *Osmunda* and *Cicatricosisporites* spores, the latter presumably recycled. A poorly productive sample (C-248520) yielded only Pinaceae and *Alnus*. The presence of *Carya*, *Pterocarya* and *Sciadopitys* argues for an age no younger than the Cyperaceae subzone, in the latest Miocene (White et al. 1999). *Pinus koraiensis* -type pollen was also recovered from a correlative site (WFR), and this pollen has been identified as being middle Miocene (White et al. 1999).

At this site a reversed polarity was obtained from below the unconformity (see Figure 64 for correlative site). Reversed polarity in the Middle Miocene occurs over relatively short intervals. The sampled units were most probably deposited during a reversal near the top of the Middle Miocene which occurred between 10.6 and 12.4 Ma. The probability that these deposits would be assigned to reversed periods at the beginning of the Middle Miocene, i.e., 15.25 -16.2 Ma, is less likely because, stratigraphically, this site represents the upper part of the 250 m Miocene sequence.

Stratigraphy Above Unconformity (Figure 64)

This section has revealed a normal-reverse-normal-reverse-normal palaeomagnetic sequence. Starting with **Unit 2: pre-glacial gravels:** a normal (Gauss) polarity for the basal pre-glacial interbedded sand and gravel beds is recorded. Three metres of highly oxidized interstratified sands and gravels are exposed, containing abundant broken pollen, much of which is difficult to identify. The deposits contain largely hematite which could not be demagnetized, but they nevertheless gave a crude normal direction probably corresponding to the Gauss Magnetic Polarity Chron.

Unit 3: First till and/or outwash gravel and sand: At this site a till overlies Miocene gravels, and in places Pliocene gravels. These gravels are overlain by a few metres of highly compacted till which mark the first glacial event in west-central Yukon. The till has a high percentage of clasts, some of which are striated. Along the eastern part of the exposure interbedded till and outwash can be seen. Both till and outwash yield a normal polarity, assigned to the Gauss Chron. The contact in both cases is contorted indicating deformation of the underlying pre-glacial fluvial beds. The till is stony, with 40% of the clasts being approximately a metre in diameter. They are subrounded, and many are weathered throughout. The matrix is a silty sand with minor clay.

STOP 20

EFR (eastern side of exposure)

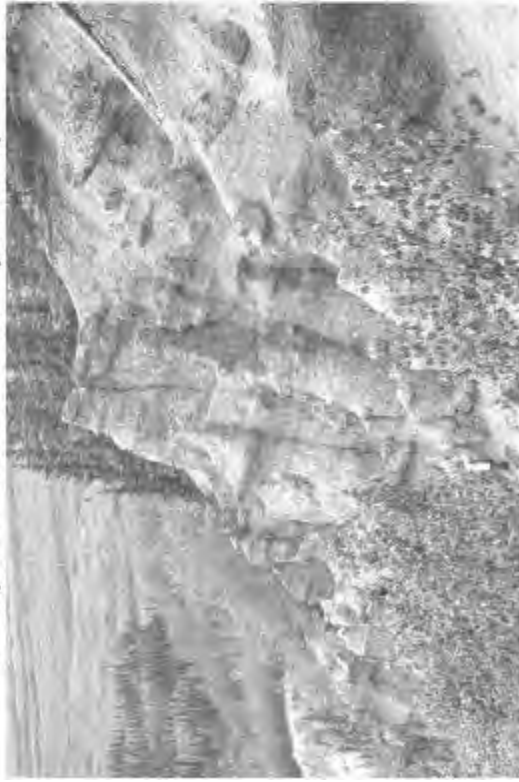


STOP ONE



East Fifteemile Site (top three hills)

EFR (western end of exposure)



EFR (central-west part of exposure)



Figure 67. Photos of East Fifteen mile River Tertiary-Quaternary exposures.

A groundwater affected palaeosol occurs at the surface of the till (Figure 67).

Unit 4: Till (Figure 67, photo 2): A magnetically reversed till overlies the previous till. It is highly compacted and contains a high percentage of clasts (approximately 40%). It contains a well developed luvisol at its surface. Both soil and parent material are reversely magnetized.

Unit 5: Till: This till is also very compact and it has a palaeosol at the upper contact that developed under wet conditions (ferragleyisol). Normal polarity was obtained from the base of the till and from the soil which has been assigned to the Olduvai Subchron of the Matuyama Chron.

Unit 6: Till: This deposit is in turn overlain by a reversed till which is less compact and has a higher sand content than the underlying deposits. This till is capped by a palaeosol (gleysol) which has been affected by ground water. This deposit is assigned a post Olduvai age.

Unit 7: Loess: Overlying the till of unit 6 is a reversely magnetized loess which has been affected by ground water.

Unit 8: Loess: A normal loess follows with no obvious contact between it and the reversed loess below. Groundwater has affected both units 7 and 8 but not the overlying unit 9. The loess of unit 8 did however, yield a small number of pollen grains of *Picea* (3.9%), *Pinus* (19.6%), *Alnus* (5.9%), *Betula* (29.4%), *Salix* (5.9%), *Ambrosia* type (33%), and *Botrychium* (2%). Pine pollen appears present in only small percentages (< 1%) in the Jaramillo and (3.4%) in the Matuyama/Brunhes boundary loess deposits at the Midnight Dome Site (Schweger, White and Froese in prep.). Pine appears not to have been present in the Yukon after 3 Ma.

Units 7 and 8 are not different from each other. They represent one deposit with no breaks. A reversed to normal polarity was obtained and is interpreted as being from the latest Matuyama to the earliest Brunhes, that is, it appears to straddle the Matuyama/Brunhes boundary. The age for these units correlates with units described and dated by Froese in the Midnight Dome Site (Figure 68). Also, at WFR Site equivalent loess deposits yielded a N-R-N polarity, extending from the Jaramillo to the Matuyama/ Brunhes boundary.

Unit 9: Till: About 200 metres to the southeast of the top of unit 8 and stratigraphically above, is a till of middle Pleistocene age (Reid equivalent). This unit is a

thin till veneer that conforms to the underlying topography. It contains a brunisol at the upper contact. Regionally, it has been correlated with deposits of Reid age (Bostock 1966; Hughes et al. 1969; Duk-Rodkin 1996; see attached surficial geology map). It is normally magnetized and is assigned to the Brunhes Chron.

Unit 10: Loess: A discontinuous thin loess cover of younger age covers the landscape.

STOP 21: RC (Rock Creek site) (64° 13' N, 139° 05' W)

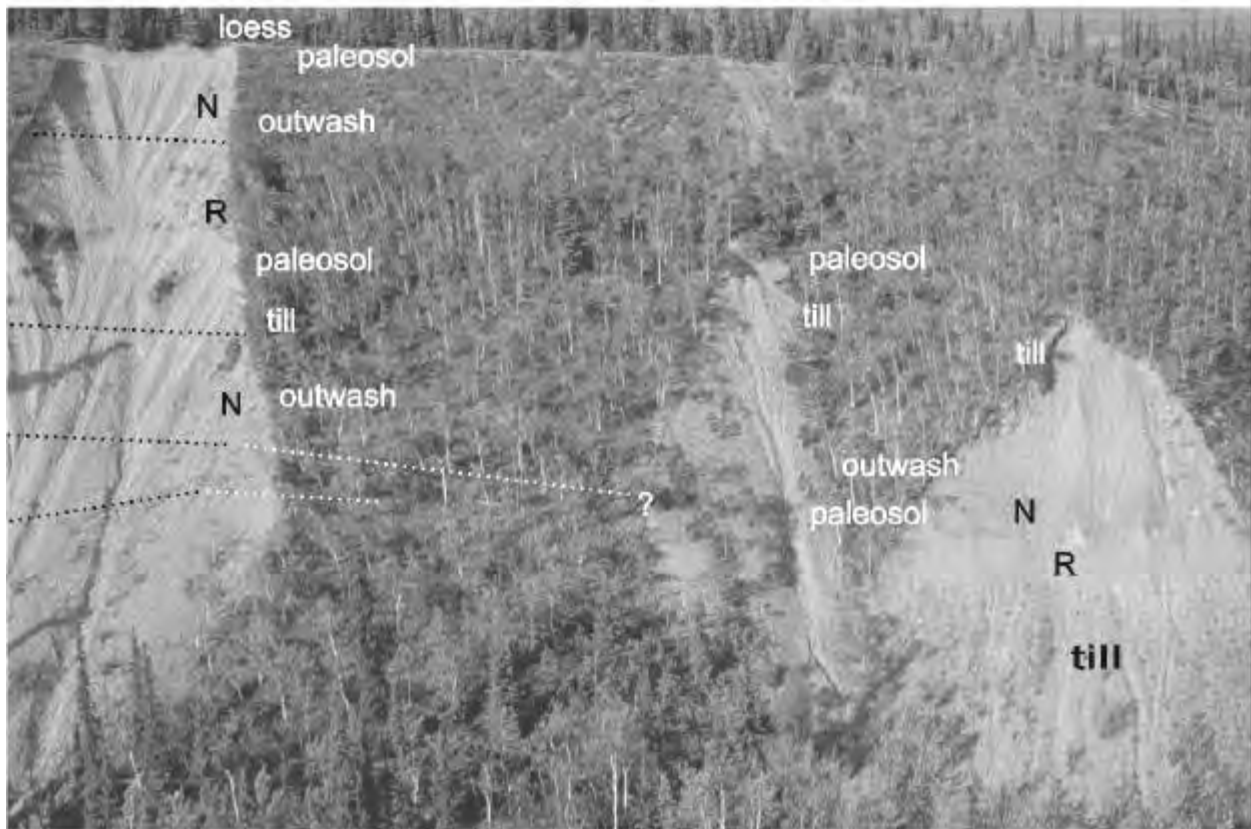
This site is located 37.5 km southeast of the EFR site. It is on the north side of the Tintina Trench and is exposed along a southwest facing slope of a small western tributary to Rock Creek (Figure 69). At RC, approximately 16 metres are exposed at the east end of the bluff (Stop 21a), of which 8 metres are partially covered. The western side exposes the rest of the sequence (Stop 21b, Figure 69a).

Stop 21a

Unit 1: fluvial gravel and lacustrine silts: At RC the base of the sequence exposes pre-glacial gravel deposits (mudflow) capped by a thin bed of lacustrine sediments. Both units are stratigraphically above the unconformity, and based on the Palynomorph assemblages they contain, are Pliocene or younger in age. RC section samples have the tree taxa *Pinus*, *Picea*, *Larix/Pseudotsuga*, cf. *Tsuga*, *Alnus* and *Betula*, and an abundant shrub-herb assemblage, including pollen of cf. Cyperaceae, Poaceae, *Ambrosia*-type, other Tubuliflorae, *Corylus*-type, Onagraceae and Ericales and *Polemonium*. An uncertain identification of *Paraalnipollenites* may suggest recycling from Paleocene or Eocene rocks.

Unit 2: Outwash: An outwash exposed for over 20 metres, overlies with a sharp contact the thin bed of preglacial lacustrine (silty clay) and mudflow deposits of Pliocene age (Figure 69). The outwash gravels are crudely stratified, interbedded with discontinuous silty clay and minor sand beds that vary in thickness from 0.5 to 1.0 m. The outwash is clast supported, with clasts up to 0.2 m in size, subrounded to rounded, some having striae.

Samples for palaeomagnetic analysis were collected from the pre-glacial lacustrine bed and from the silty clay and minor sand beds in the outwash. The two group of samples yielded a normal polarity (N). The pre-glacial mudflow deposits contain the same



Rock Creek Site (west end)



Rock Creek Site (east end)

Figure 69. Top photo is of Stop 21a. Bottom photo is of Stop 21b.

pollen as the overlying lacustrine bed. The lacustrine bed yielded the same normal polarity as the overlying outwash gravels. These preglacial undeformed sediments are all of Pliocene age, based on the pollen assemblage. The normal directions of magnetization for this site are assigned to the early Pliocene Gauss Magnetic Polarity Chron. Corroborating these assignments are extensive palaeomagnetic data obtained from the White Channel gravels on the Klondike Plateau (Froese et al. 2000).

Stop 21b (Figure 69b)

Unit 3: Till: This unit is exposed about one kilometre west of Stop 21a on the same slope. It is very compact with a high pebble content, and has a luvisol developed at its upper contact. It yielded a reversed polarity that has been assigned to the early Matuyama Chron.

Unit 4: Mudflow? Stratigraphically above unit 3, this highly compact deposit has a high clast content >75% in a silty/clay/gritty matrix that holds the clasts together. It is capped by a soil of undetermined depth. No pollen or palaeomagnetic record was obtained from this deposit.

Unit 5: Outwash/till: This unit is composed of two sub-units, an outwash gravel interbedded with a few thin sandy beds. A normal polarity was obtained from the fine grained beds. It is overlain by a coarse till which is capped by a palaeosol/weathering horizon. No polarity was obtained from the till. Because of the absence of a palaeosol between the two subunits they are assigned to the same glacial event, that is, to the Olduvai subchron.

Unit 6: Outwash: This unit is composed of interbedded gravel with sand beds and minor silt. A reversed polarity was obtained from the base of the unit, assigned here to the early upper Matuyama. The unit ends at an erosional contact, and near the top, sandy pockets reveal a reversed magnetization overprinted by a normal.

Unit 7: Outwash: This unit is similar to the underlying unit. It yielded normal polarity that can be assigned to a cold/glacial period (probably stage 30) of the Jaramillo subchron, and thus most likely correlates with the base of the loess sequence at the WFR Site. Unit 7 is capped by a Wounded Moose palaeosol (luvisol) which in turn is covered by a younger loess.

Comments

Magnetostratigraphy alone constrains the age for the oldest glacial deposits in the Tintina Trench. Suitable samples for palaeomagnetic analysis were obtained

from fine-grained beds within till and/or outwash sediments. All the sites yield normal polarity at the base. Thus the N-R-N-R-N glacial sequence in the trench sites must fall within the Matuyama and its subchrons, and the underlying normally magnetized preglacial beds which contain Pliocene pollen must belong to the Gauss Chron. Pliocene pollen defining pre-glacial Pliocene conditions are: *Pinus*, *Picea*, *Larix/Pseudotsuga*, cf. *Tsuga*, *Alnus* and *Betula*, and an abundant shrub-herb assemblage, including pollen of cf. Cyperaceae, Poaceae, *Ambrosia*-type, other Tubuliflorae, *Corylus*-type, Onagraceae and Ericales and *Polemonium*, *Polygonum persicaria*-type and ? Cyperaceae.

At two sections, three productive samples taken from above the unconformity have yielded simple pollen assemblages, namely an arboreal component dominated by Pinaceae and a good representation of shrubs and herbs, which is consistent with a Plio-Pleistocene age. In particular, the herbaceous species, *Polemonium*, is characteristic of the Plio-Pleistocene Poaceae and *Artemisia* zones in the interior of Yukon and Alaska. It has not been observed in Miocene strata. The absence of *Artemisia* pollen in the assemblages above the unconformity suggest an age older than Pleistocene (White et al. 1999). The environment appears to have not reached the level of aridity, conducive to the growth of *Artemisia*, which occurred in the Pleistocene.

Day 4: Glaciation of the Southern Ogilvie Mountains via the Dempster Highway

For those not taking the helicopter portion of the Day 4 tour, there is a vehicle trip up the southern Dempster Highway to look at the moraine stratigraphy of the southern Ogilvie Mountains. The original work on the Quaternary history of the southern Ogilvie Mountains was completed by Vernon and Hughes (1966) who outlined a record of at least three glaciations which they named Oldest, Intermediate and Last. Subsequent mapping by A. Duk-Rodkin (1996) correlated, more or less, these limits of glaciation with the pre-Reid, Reid, and McConnell advances of the Cordilleran Ice Sheet which we examined on Day 1 (Figure 70). More recently, B. Beierle has questioned the age of the mapped Reid limit in the vicinity of Chapman Lake on the basis of the age of lake sediments within that lake and kettle ponds in the region. Beierle, as will see at Stop 26, argues that the lake is part of a Late Wisconsin deglacial sequence and thus the moraine is similarly of Late Wisconsinan age.

STOP 22: Reid Outwash Surface (64° 00' N, 138° 44' W)

D.G. Froese

Vernon and Hughes (1966) mapped a prominent moraine set which the Dempster Highway crosses near Stop 23 (Figure 70). South of this position is a broad outwash plain that is considered correlative with the moraine and thus also of, probably, middle Pleistocene age (Duk-Rodkin 1996). There is little in the area that can be used for independent chronology associated with the moraine, but based on soil development, Tarnocai et al. (1985) considered soils associated with the outwash to be Diversion Creek soils and thus correlative with the Reid glaciation. An occurrence of the Old Crow tephra (ca. 140 ka) on an extension of this outwash surface to the southwest suggests an age of at least marine isotope stage 6 (ca. 150 ka; Froese unpublished data). Soil development at the site we will stop at shows downward movement of clays to at least 60 cm depth, suggesting a truncated Diversion Creek soil.

STOP 23: Reid Ice-Marginal Deposits (63° 4' N, 138° 32' W)

D.G. Froese

Along the west side of the Dempster Highway, a gravel pit exposes a section of the Reid moraine (Figures 71 and 72). The site consists of poorly-sorted boulder-gravel and steeply-bedded cobble-diamict. The boulder-gravel complex is interpreted as a proximal glaciofluvial complex, while the cobble-diamict units are interpreted as a meltout till associated with ice in the immediate region of the north Klondike valley.

STOP 24: Tombstone Valley Outlook

D.G. Froese

One of the most spectacular, and often photographed, views in Yukon is the Tombstone Valley look-out immediately north of the campground (Stop 24, Figure 70). The view looks up the North Klondike River to Tombstone Mountain (Figure 73). Tombstone Mountain is part of a middle Cretaceous syenite intrusion. Tombstone Mountain is part of a broad arc of lithologically similar intrusions distributed in a broad arc northeast of the Tintina Fault. These intrusions include the Tombstone Suite (92 Ma) and the Selwyn Suite (97-112 Ma; Mortensen unpublished, cited in Bremner 1994). Interest in the mineral potential of these rocks has been

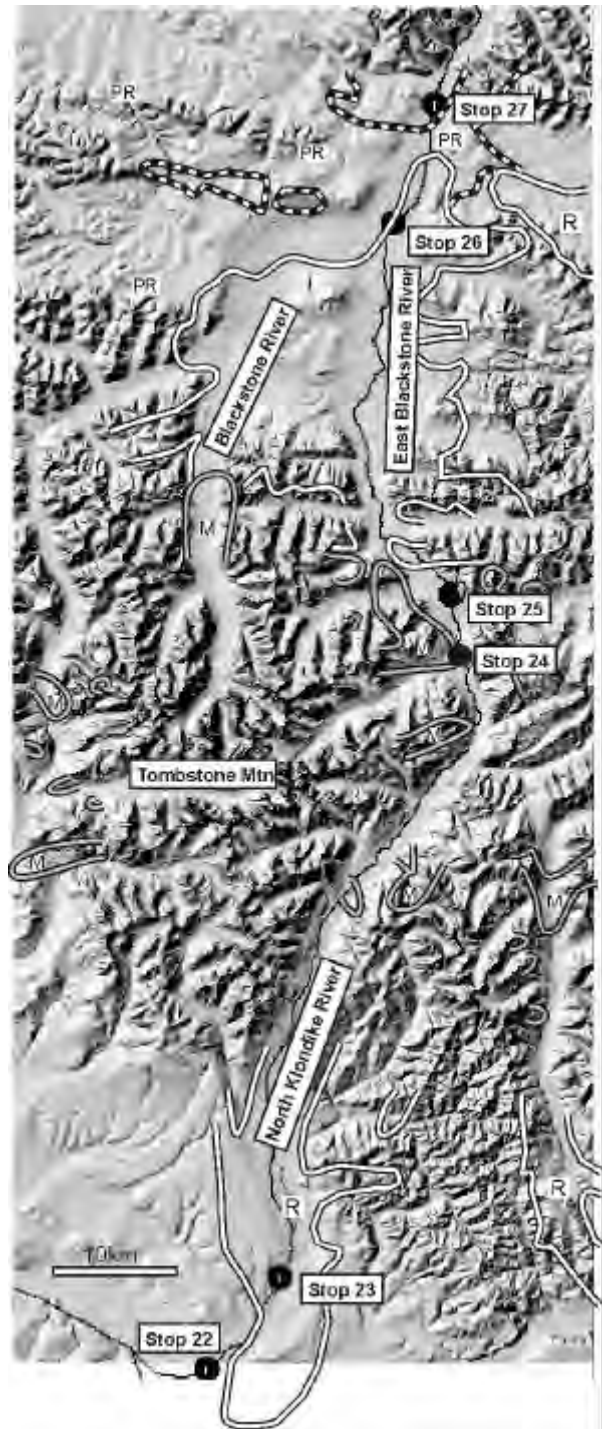


Figure 70. Shaded relief image of southern Dempster Highway and stops on Day 4. PR– Pre-Reid drift, R– Reid Drift, M– McConnell (late Wisconsinan Drift surfaces); glacial limits after Duk-Rodkin 1996.

driven most recently by recognition of the similarity to the Fort Knox porphyry deposit in the Fairbanks area which would occur adjacent to these intrusions



Figure 71 Diversion Creek soil developed on Reid (?) surface at Stop 22.

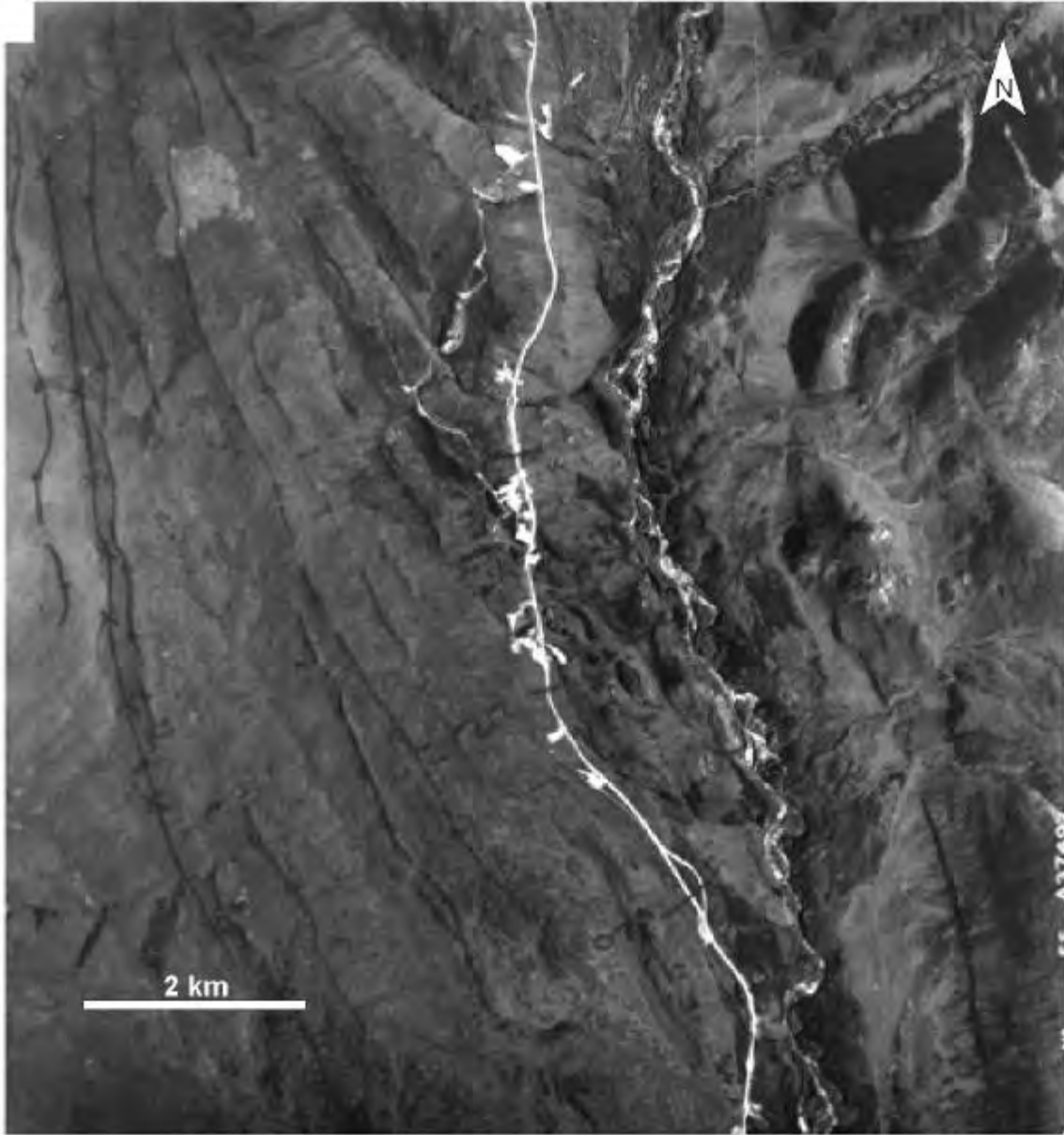


Figure 72. Aerial photograph of the Reid moraine (Duk-Rodkin 1996) along the southern Dempster Highway in the North Klondike Valley. The moraine has a subdued topography and lacks kettle lakes. Surface soils in the area of the moraine and outwash are similar to Diversion Creek-type soils associated with Reid surfaces.



Figure 73 View from the Tombstone valley lookout toward Tombstone Mountain.



Figure 74. North Klondike canyon formed by the displacement of an unnamed tributary of the Blackstone, prior to the McConnell Glaciation, into the North Klondike at the time of deposition of the North Fork Pass moraines.

with restoration of the 450 km offset of the Tintina Fault.

The spectacular topography of the mountain and adjoining peaks (e.g., Mt. Monolith) is a result of the resistance to erosion of the intrusions relative to the sedimentary rocks in the area, well developed vertical joints in the intrusion and repeated cycles of glaciation. Throughout the late Pliocene and Pleistocene, the Tombstone Mountain area, which contains the highest peaks of the Southern Ogilvie Mountains, was the center of the accumulation zone for glaciers that radiated out along the North Klondike, Blackstone and Tombstone/Chandindu valleys toward the Tintina Trench and Northern Ogilvie Mountains.

In the area of the lookout parking lot are numerous syenite boulders which were transported by glaciers from the Tombstone Mountain area. Vernon and Hughes (1966) and Duk-Rodkin (1996) placed the limit of their 'Last' glaciation (McConnell- late Wisconsinan) at about the lookout parking lot, and remnants of the moraine can be seen along the valley walls to the west.

Immediately to the north is a deep canyon which the northernmost tributary of the North Klondike River has cut to join the North Klondike River (located between Stops 24 and 25; Figure 74). The sharpness of the feature and its proximity to the North Fork Pass moraine suggest this drainage was tributary to the East Blackstone River prior to the advance of the North Fork Pass glacier (Stop 25). At this time, during the McConnell, the tributary was likely cut off from the Blackstone drainage and diverted into the North Klondike valley.

STOP 25: North Fork Pass moraine (63° 34' N, 138° 14' W)

D.G. Froese

At approximately Stop 25 and for a few km to the north, the Dempster Highway travels along the margin of the North Fork Pass moraine (Figures 70 and 75). The moraine has a sharp margin and contains numerous boulders (>2 m across) and kettle lakes. The sharp morphology of the moraine and the lack of recessional moraine features outside the limit suggest the North Fork Pass moraines are an end moraine system associated with the Late Pleistocene advance (McConnell) in the Ogilvie Mountains (Vernon and Hughes 1966; Duk-Rodkin 1996).

STOP 26: Chapman Lake, Kilometre 117, Dempster Hwy. (64° 50' N, 138° 21' W)

Brandon Beierle

Chapman Lake is the largest of a group of kettle lakes in what has been previously mapped as a Reid aged terminal moraine (Figure 76). Sediment cores and radiocarbon dates obtained from the lake, coupled with local geomorphology, suggest that Chapman Lake is considerably younger than previously thought, and that the moraine was deposited during the McConnell glaciation.

Chapman Lake lies near the confluence of the Blackstone and East Blackstone Rivers, which have incised into the terminal moraine complex, leaving the surface of Chapman and other lakes on the moraine perched ~10 m above the current Blackstone River. The scarp along the edge of the Blackstone River (Figure 77) is composed of laminated glaciolacustrine sands and silts contacting till at or below the water level in the river.

The surface of the moraine is comprised of hummocky moraine, with numerous small kettle lakes and marshes in low-lying areas. Superposed on the surface of the moraine are several large palaeo-meanders (Figure 77) from a time when the Blackstone River flowed across the surface of the moraine. An outlet channel exists at the north end of Chapman Lake (Figure 77) and flows through several of the smaller kettle lakes before it is truncated by the incision of the Blackstone River into the moraine. Flow was diverted away from Chapman Lake when the Blackstone River was captured by the East Blackstone River, resulting in the modern drainage configuration. This relationship indicates that the diversion of the Blackstone River away from Chapman Lake must have happened prior to incision of the moraine. If this were not the case, the outlet channel would be graded to the current level of the Blackstone River, which would have resulted in the drainage of Chapman Lake due to the ~10 m elevation difference between the river and the lake.

Three sediment cores taken from Chapman Lake record the isolation of the lake as an abrupt decrease in grain size (Figure 78) which allows the diversion of the Blackstone River to be radiocarbon dated at ~12,500 BP. This age also constrains the timing of incision of the moraine to the past ~12,500 years, which would require a 200,000 year greater delay in incision of the moraine if it were of Reid age. This

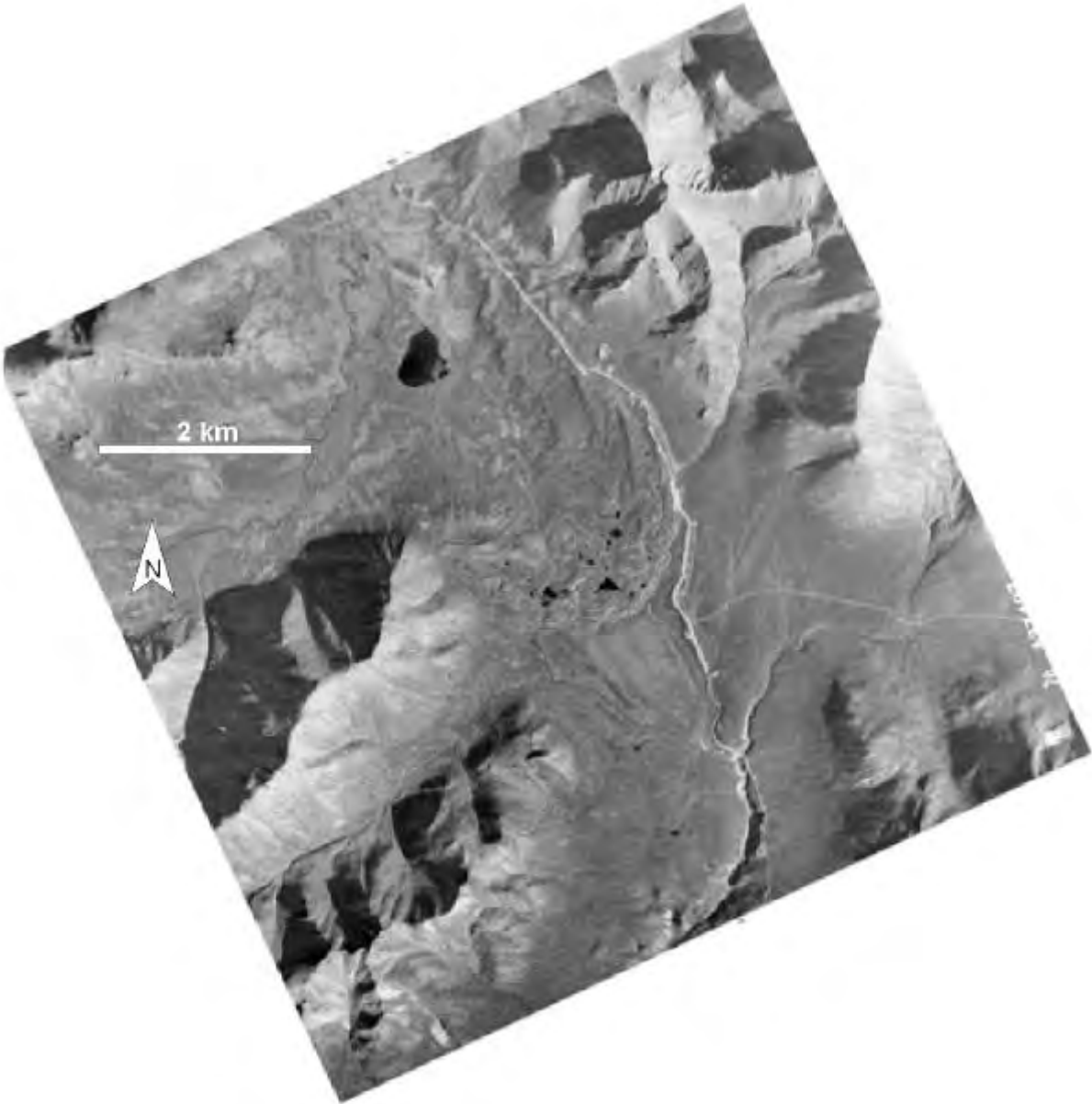


Figure 75. Aerial photograph of the North Fork Pass moraine.

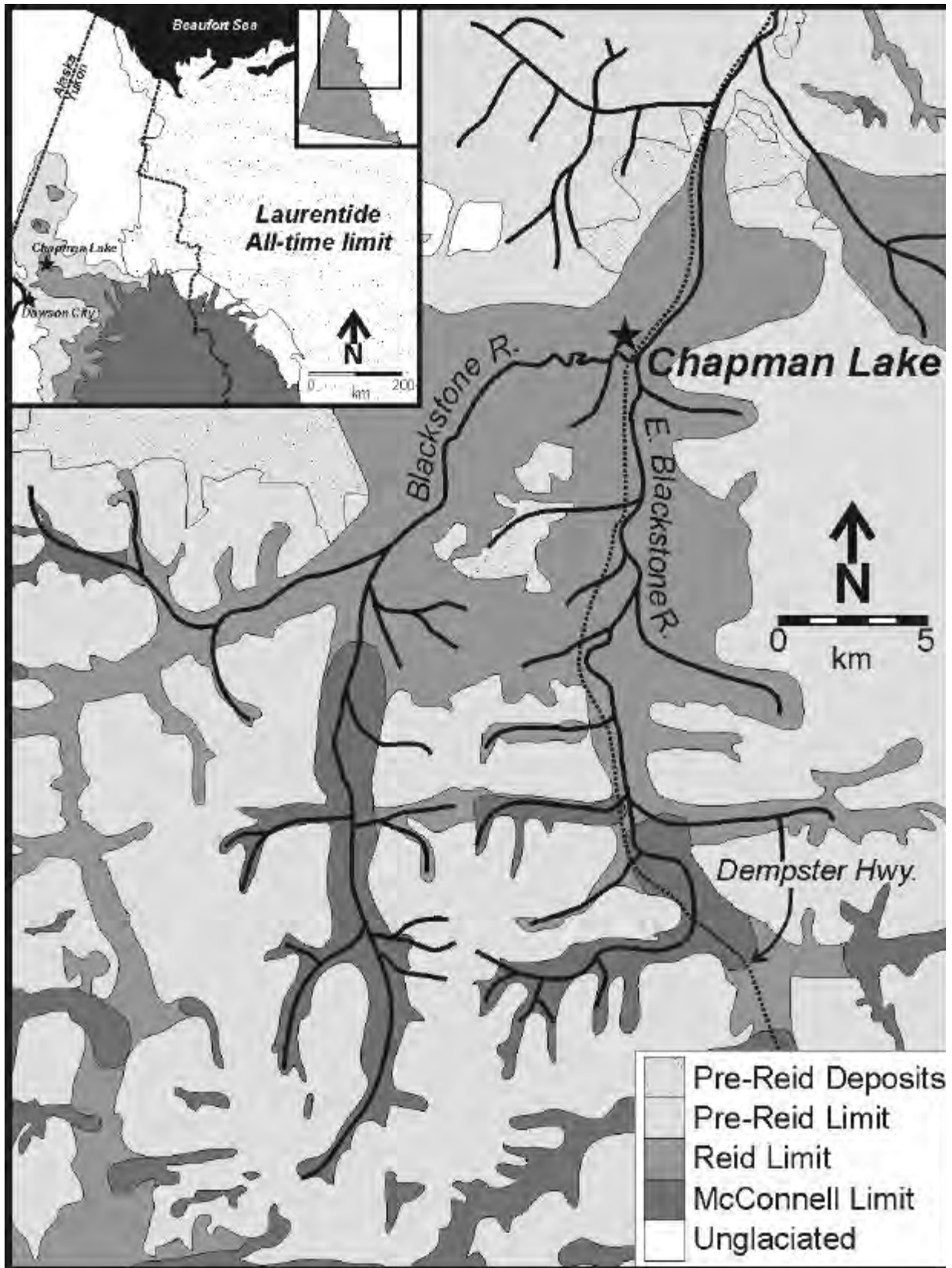


Figure 76. Glacial limits in the Ogilvie Mountains relative to Chapman Lake (after Duk-Rodkin 1996).

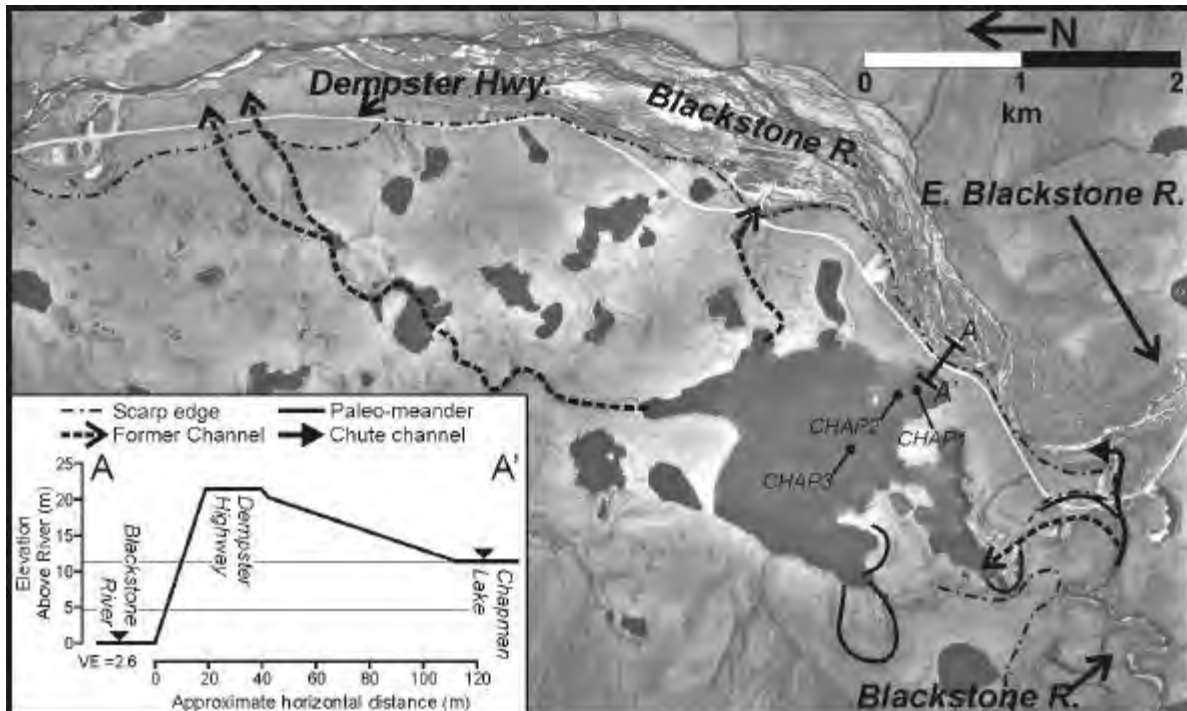


Figure 77. Aerial photograph of Chapman Lake with geomorphic features and locations of cores.

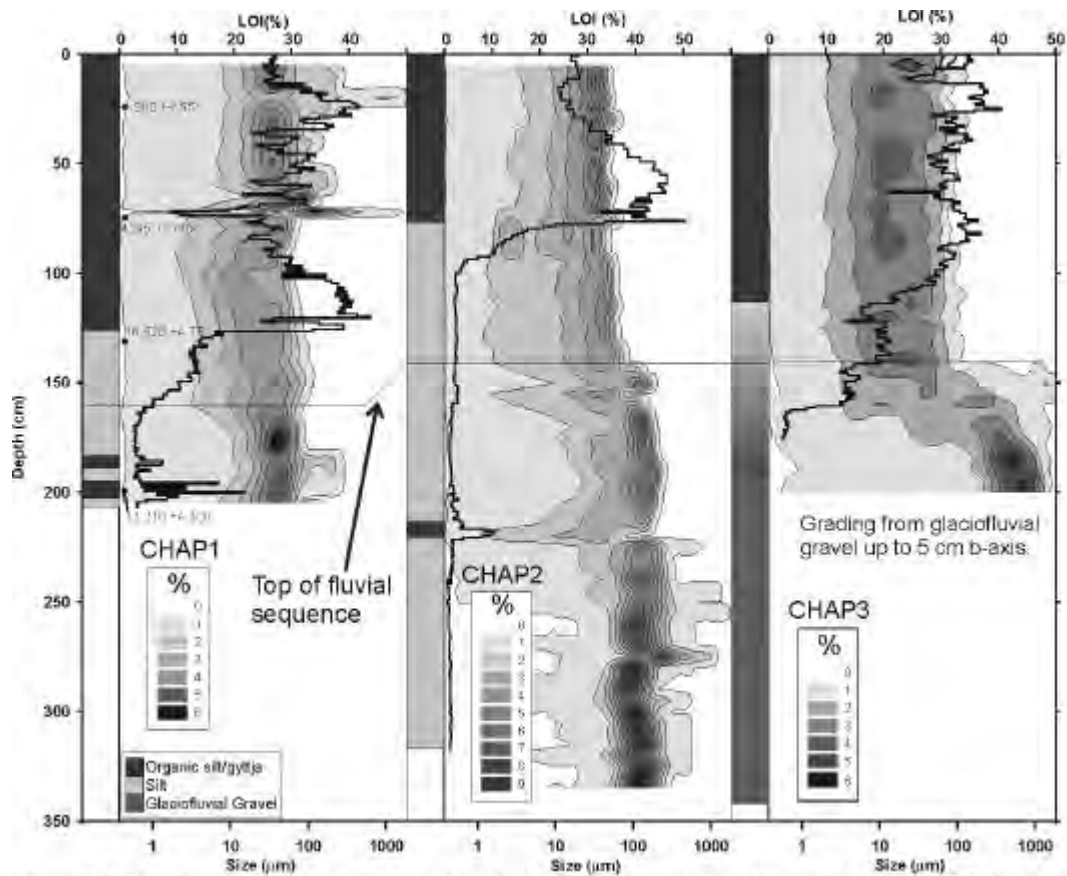


Figure 78. Lithostratigraphy, chronology, grain size and organic carbon from three cores from Chapman Lake (see Figure 76 for location).

incision chronology would, however, be consistent with deposition of the moraine during the McConnell. This argument is supported by a fining upwards sequence present in all three cores (Figure 78), which likely represents sedimentation by the Blackstone River into Chapman Lake. The most proximal core (CHAP3) contains glaciofluvial gravels at the base of this sequence, which fine upwards into coarse sand (Figure 78), abruptly ending at ~12,500 BP. The succession from gravel to sand in the proximal core is suggested to represent the recession of glacial ice from the Chapman Lake area, shortly after deposition of the moraine. The top of this fining sequence spans a radiocarbon date of 13,200 BP, suggesting that it was deposited after the McConnell glaciation.

STOP 27: Unnamed Felsenmeer Ridge with Tors and pre-Reid glacial erratics (64° 56' N, 138° 16' W- pull off location along Highway-ridge 2 km west)

From Stop 27, with weather permitting, we will hike from the pullout located on the east side of the Dempster Highway, to the west across the Tussock surface onto a heavily eroded limestone ridge. The round trip takes 2 to 3 hours. As we approach the top of the ridge we will see a number of Tors and well-developed Felsenmeer surfaces developed from the limestone. These are often an indication of landscape antiquity and have been associated with unglaciated regions (e.g., Pewe 1975). However, fine-grained and medium-grained volcanic rocks from the upper valleys of Blackstone River are present, particularly preserved on flat slopes, indicating these surfaces have been glaciated in the past (Duk-Rodkin 1996). The timing of this glaciation is unclear, but probably dates to early Pleistocene based on the extent of glaciers which occurred along the southern margin of the Southern Ogilvie Mountains.

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