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3rd INTERNATIONAL MAMMOTH CONFERENCE
FIELD GUIDE TO QUATERNARY RESEARCH IN THE KLONDIKE GOLDFIELDS

Edited by D.G. Froese and G.D. Zazula

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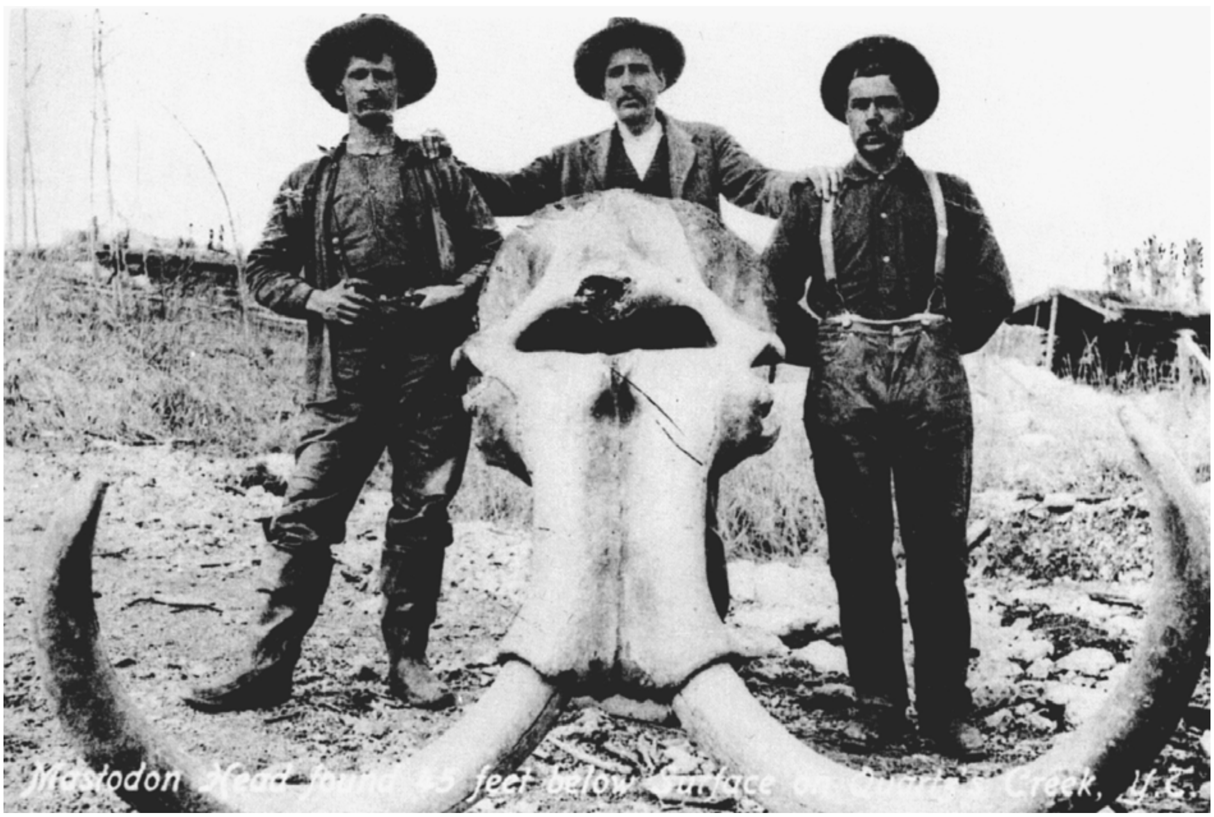
Field Guide to Quaternary Research in the Klondike Goldfields



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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 Territory. The exceptional exposure of Plio-Pleistocene deposits by Yukon placer miners in the Klondike, 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 the last 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. The trip from Whitehorse to Dawson crosses the McConnell (late Wisconsin), Reid (mid-Pleistocene) and pre-Reid (multiple late Pliocene-early Pleistocene) drift surfaces, entering into the unglaciated region of central Yukon Territory. The field excursion on the final day of the conference introduces the rich glacial record of the Klondike goldfields and the late Pleistocene environments associated with the Mammoth steppe in Yukon Territory.

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, Lee Olynyk, David Millar, Mike McDougall, Norm Ross, Stuart Schmidt, Frank Short, Wayne Tatlow, 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.

The summaries and research presented in this guide were made possible through grants from various agencies which have supported Quaternary studies in Yukon by the respective authors. In particular, we would like to acknowledge the Geological Survey of Canada, which has provided the main support for a host of researchers. Recent grants from NSERC and support from Yukon Geology Program and Heritage Branch of the Yukon Territorial Government have been critical to continuing this work. As well, many graduate students working in Yukon have received support from the Northern Science Training Program administered by the Department of Indian Affairs and Northern Development.

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 Saint 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 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.

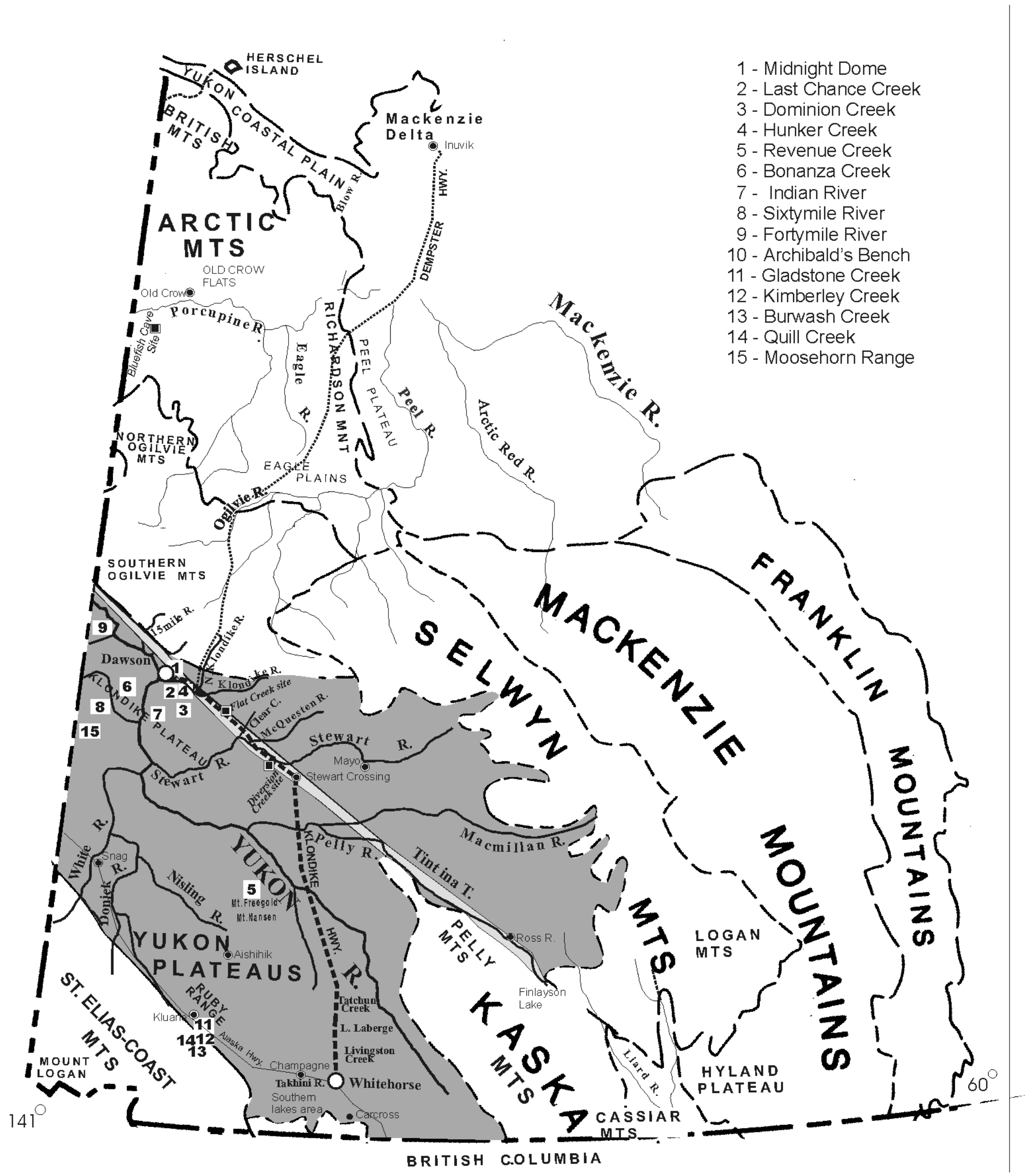

Figure 1. Physiography of Yukon and location of sites mentioned in text.

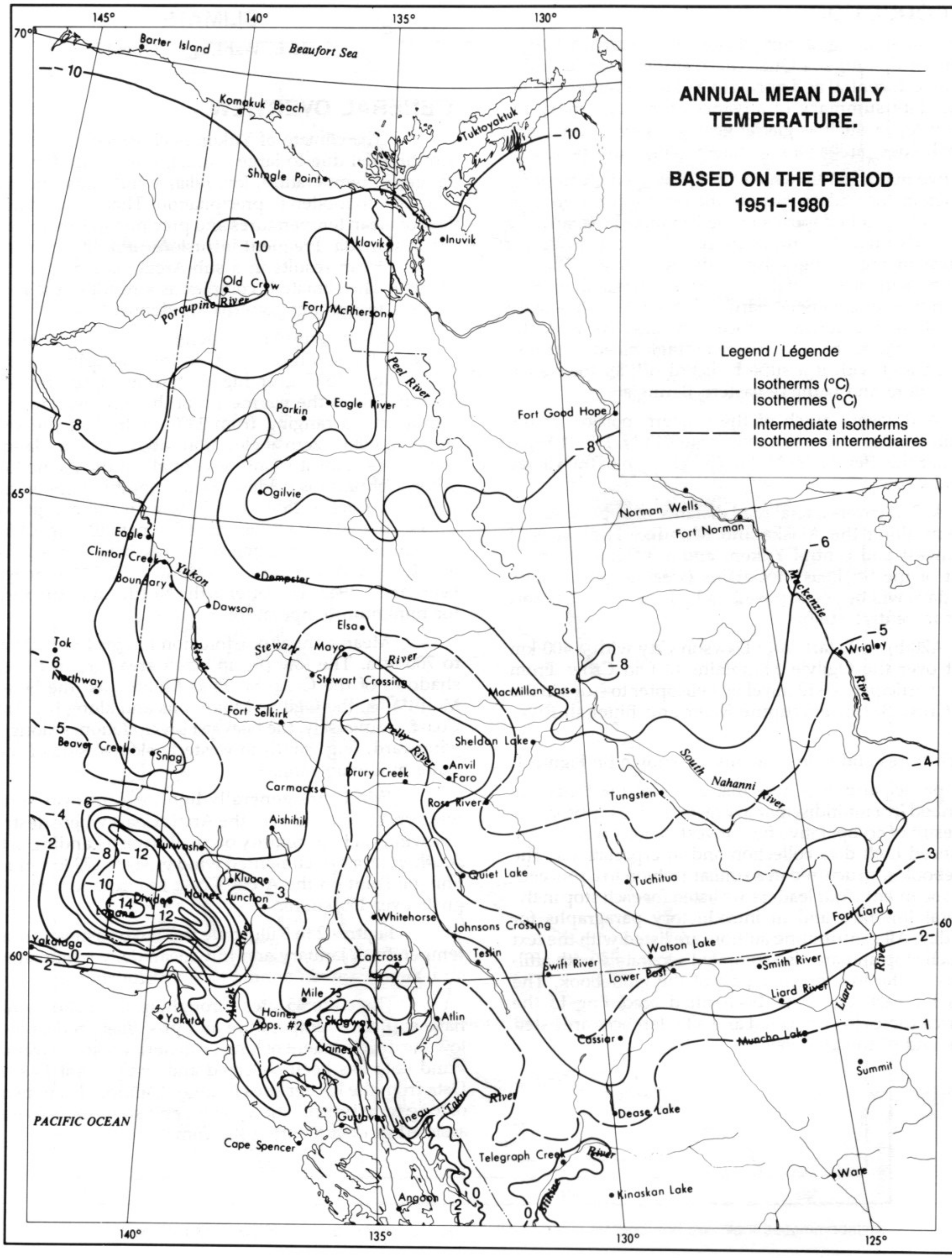

Figure 2. Annual mean daily temperature, Yukon Territory (from Wahl et al. 1987).

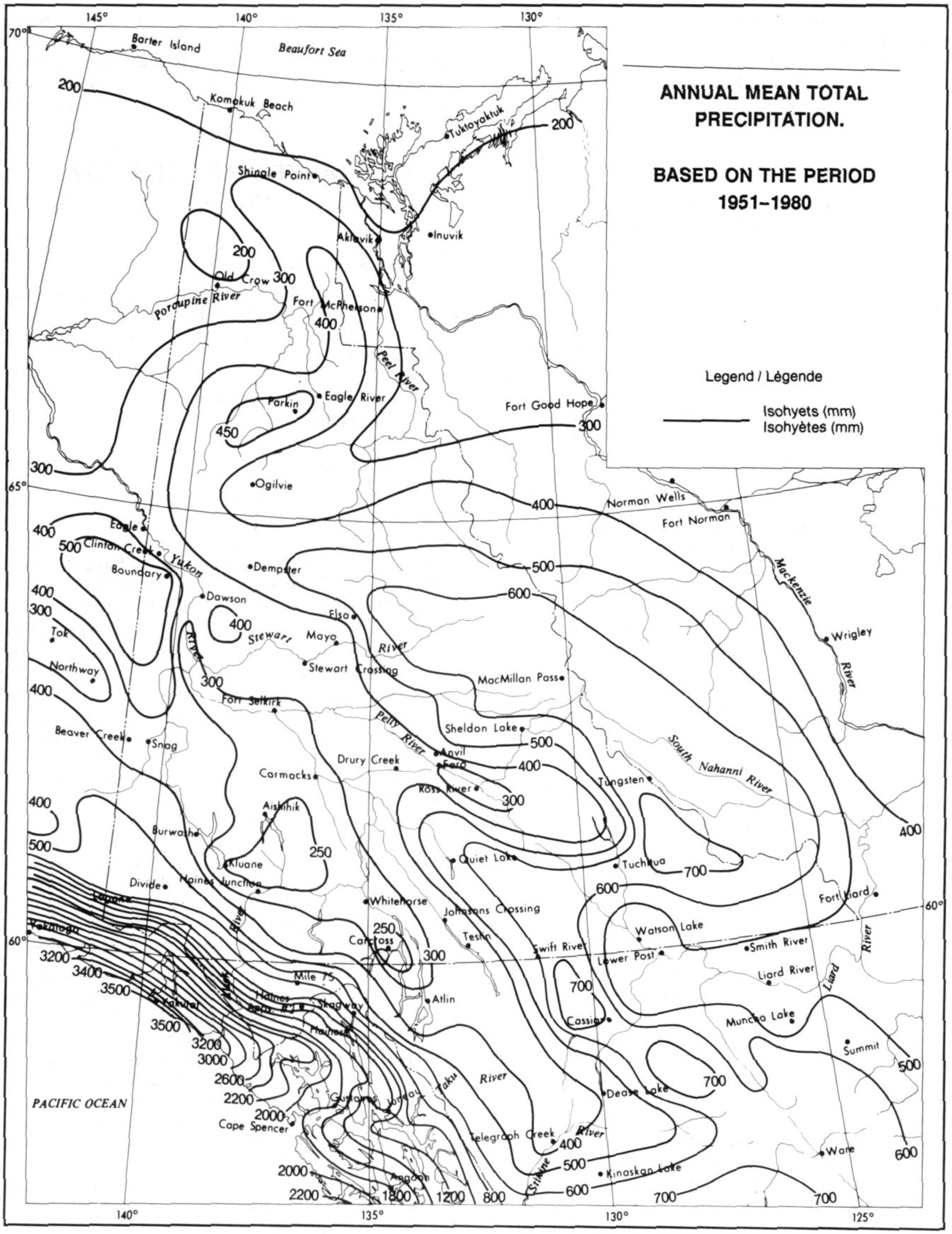

Figure 3. Annual mean total precipitation, Yukon Territory (from Wahl et al. 1987).

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 longwave 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 3. 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 increases 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 help 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, and be-

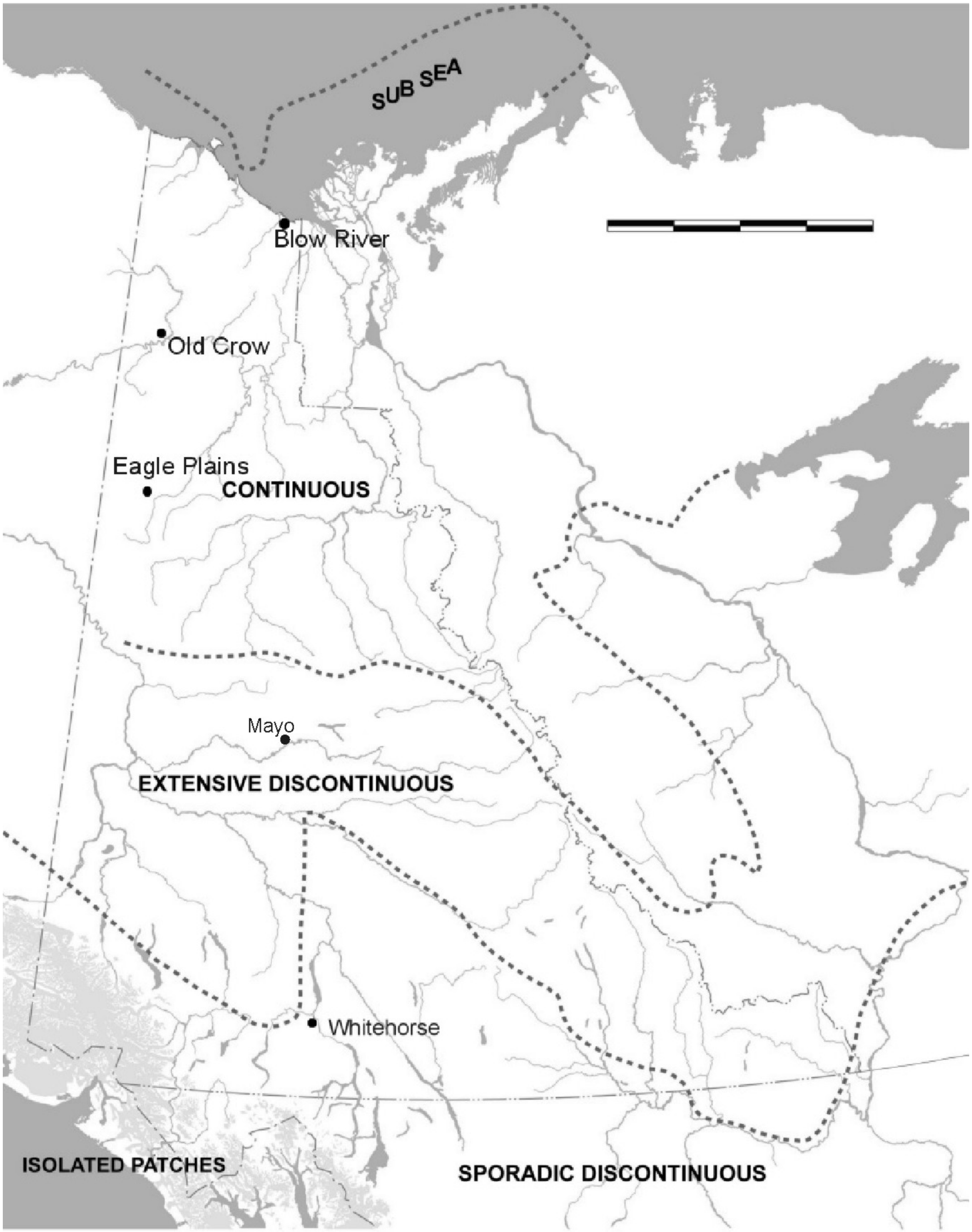

Figure 4. Permafrost distribution in Yukon, and locations of sites mentioned in text.

tween 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 exposed 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 forests. 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 wetlands 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 Armtrong 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. 2002), 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 Macauly Glaciation (Late Pleistocene) and Mirror Creek Glaciation (Middle Pleistocene).

The extent of glaciations in Yukon follows a gen-

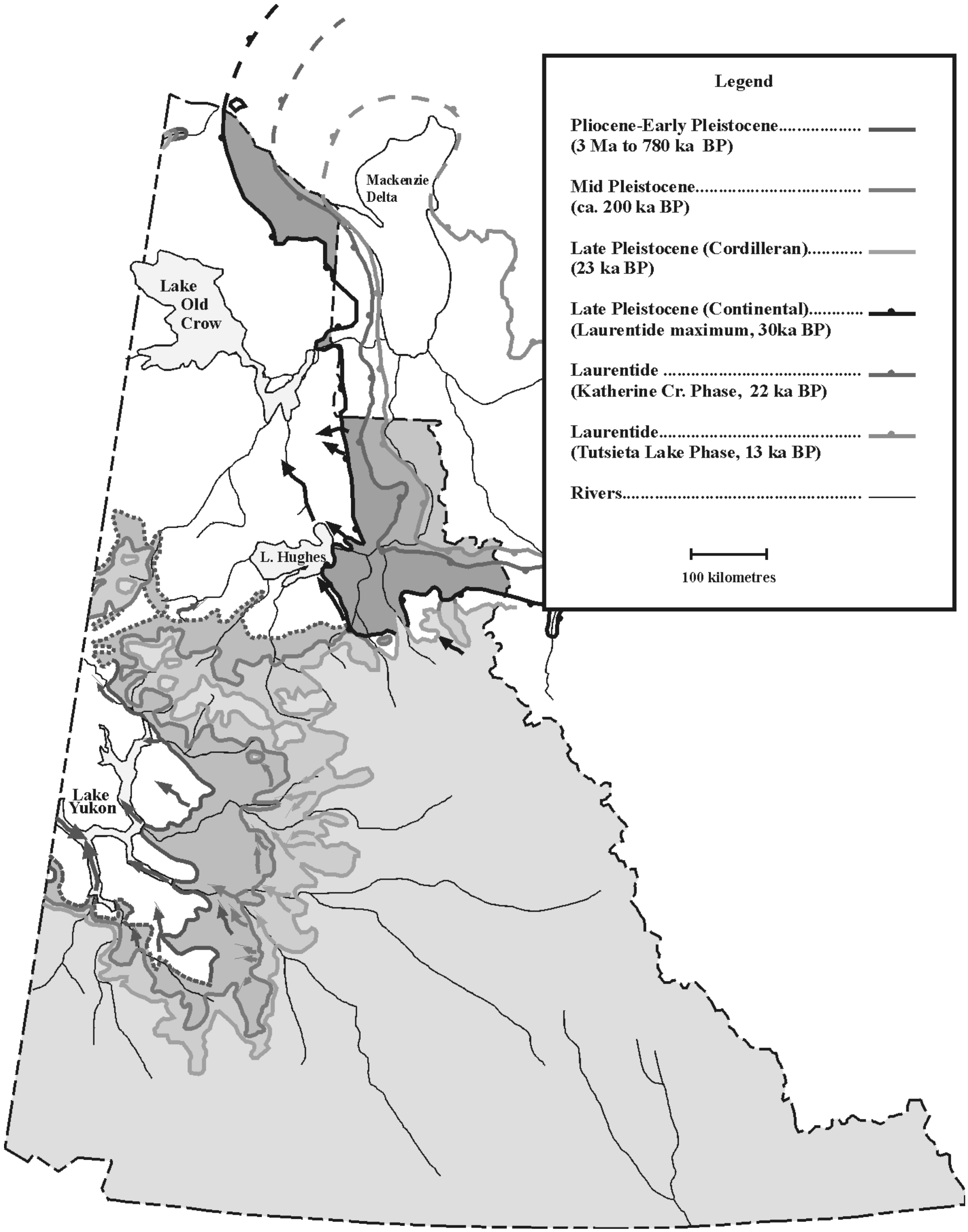

Figure 5. Extent of Plio-Pleistocene glaciations in Yukon.

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 cups 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 1991; Lemmen et. al. 1994), traceable from southern Mackenzie Mountain to the Coastal Plain; and 2) Tutsieta 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; Zazula et al., in review). 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 (paleo-Mackenzie River); Pacific Ocean (Bering Strait: Kwikpack River and Gulf of Alaska: paleo-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 paleosol 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. 2001). 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 paleosol (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

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 support 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.

Permafrost is widespread and discontinuous

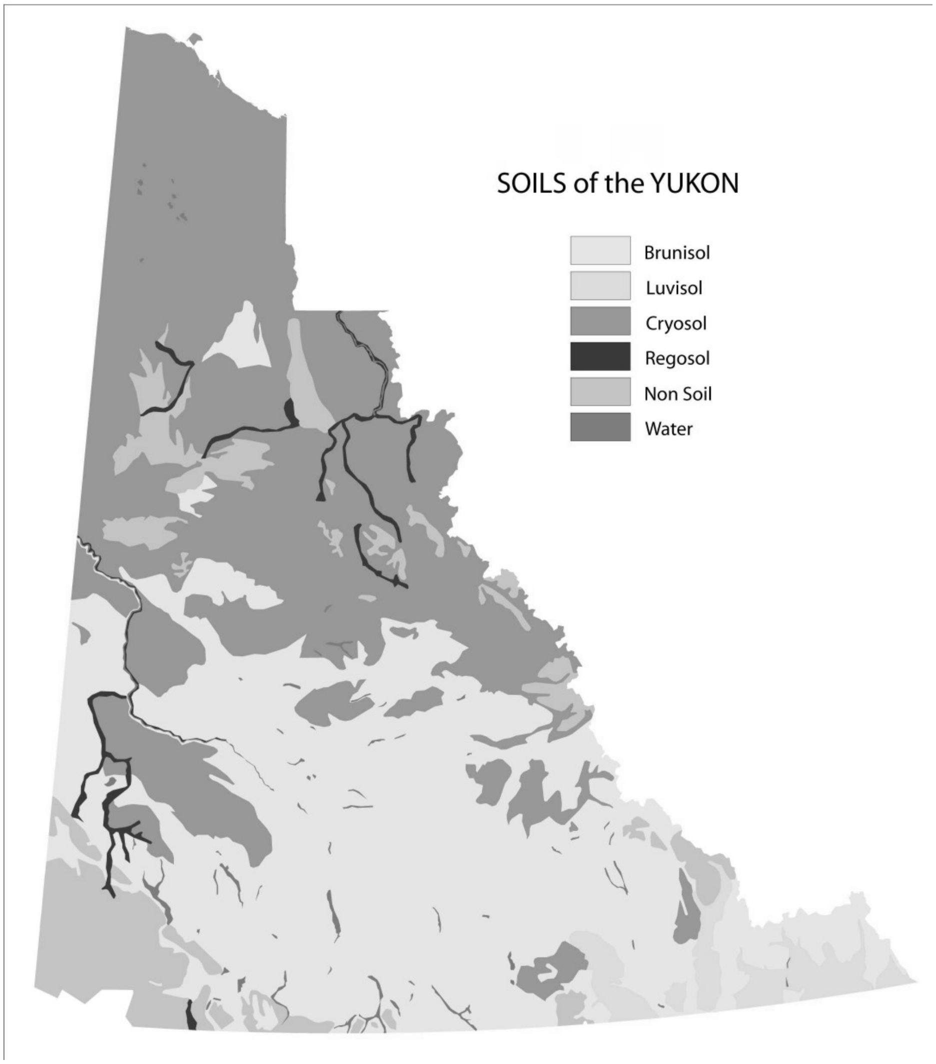

Figure 6: Distribution of major soil groups in Yukon.

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 aufeis 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 paleosols. These paleosols 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 paleosols (designated the Wounded Moose paleosol 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 Saint Elias Icefields contain vast areas without soil formation or vegetation cover. Here high-elevation icefields and rock summits dominate 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 centers 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. These exposed shelves formed the Bering Land Bridge, which connected western Alaska to ice free

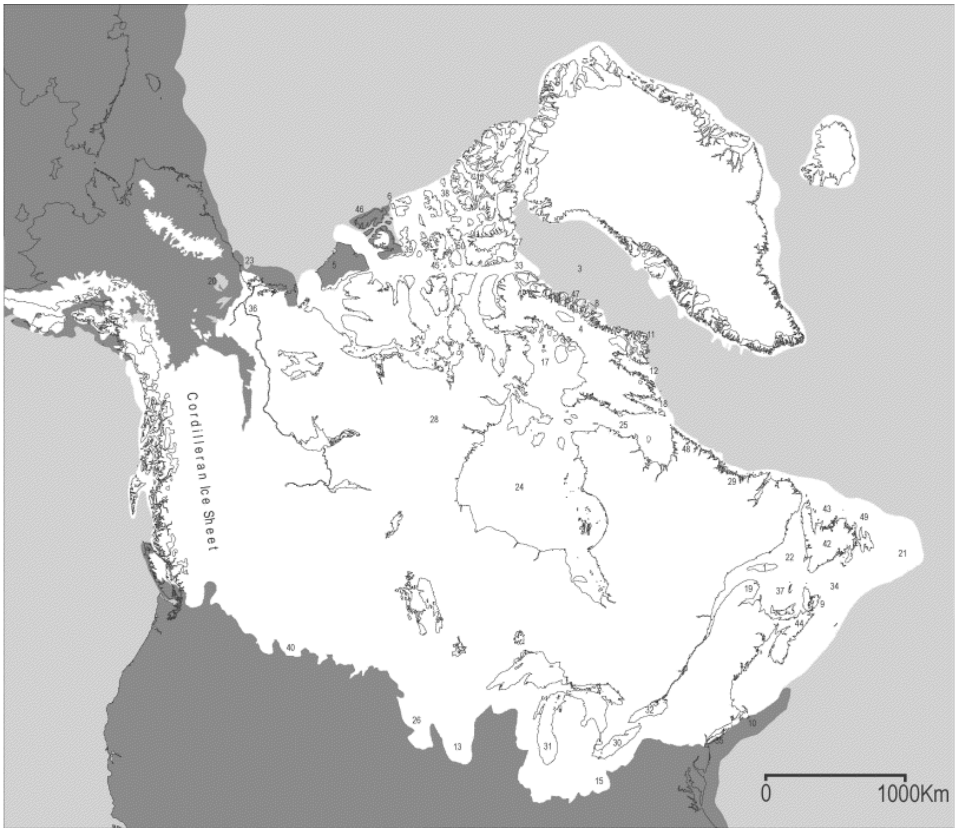

Figure 7. Extent of last (late Wisconsinan) glaciation in Canada (from Dyke et al., 2002).

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 1.8 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.

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 pro-

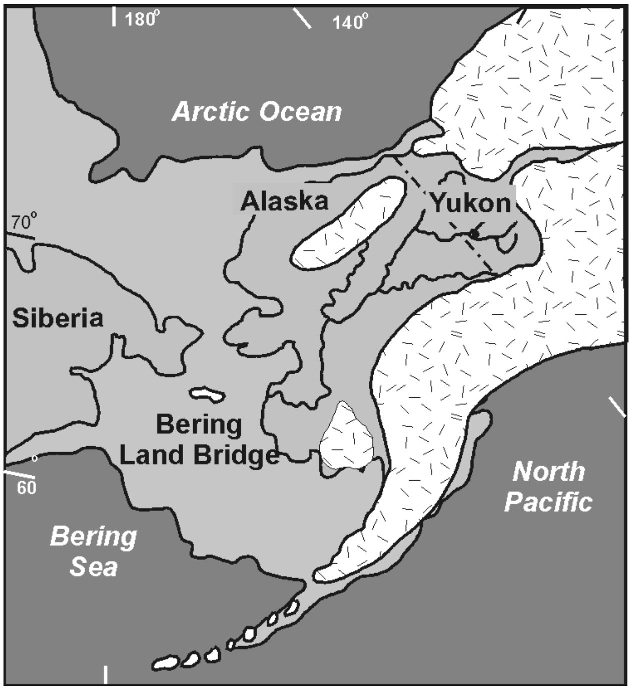

Figure 8. Beringia during the last glaciation (ca. 18 ka BP).

duced 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), *Artemisia* (sage) 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/cm^2/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 (Zazula, 2003) 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 refugia 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 ^{14}C 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 Klondike 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 (Table 2). 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 (Eocene; 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±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 quantity. 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 1988), 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.

Table 2. Radiocarbon ages of dated fossils from the Klondike area, Yukon Territory. Compiled from Morlan (2003b), and other sources.

Fossil	Locality	^{14}C yr B.P.	Lab number	Reference	Collection Number	Material
Klondike District						
Beaver (*Castor canadensis*)	Hunker Creek	6,920 ±105	Beta-28763	Harington, 2002	ETH-4760	cranium
Arctic ground squirrel	Dago Gulch, Hunker Creek	13,910 ±70	Beta-111606	Kotler and Burn, 2000	plant remains in nest	
(*Spermophilus parryi*)	Lower Sulphur Creek	13,350 ±265	QC-664	Guthrie, 1990		plant remains in nest
	Dominion Creek	12,200 ±100	GSC-2641	Harington, 1977, 2002; Pirozynski et al., 1984	NMC-21094	plant remains in nest
	Hester Creek, Hunker Creek	22,090 ±140	Beta-136365	Storer, 2002 pers. comm.	YG-76.5	mummified carcass
	Quartz Creek	24,280 ±130	Beta-161239	Froese et al., 2002		*Draba* seed capsules in rodent burrow
	Quartz Creek	23,990 ±130	Beta-161238	Froese et al., 2002		Grass stems from rodent burrow
Badger (*Taxidea taxus*)	Hunker Creek	15,240 ±130	Beta-81133	Harington, 1977	NMC-17260	cranium
	80 pup, Hunker Creek	37,990 ±750	Beta-83413	Harington, 1980	CRH-95-17	
Black-footed Ferret (*Mustela nigripes*)	Hunker Creek	30,370 ±560	Beta-23347	Youngman, 1993	NMC-44305	partial mummy, age on bone
Short-faced bear (*Arctodus simus*)	Ophir Creek, Indian River	20,250 ±110	Beta-79852	Harington, 2002	CRH-9593	cranium
	Hunker Creek	24,850 ±150	TO-3707	Harington, 2002, Matheus, 1995	NMC-50367	
	Gold Run Creek	26,040 ±270	TO-2696	Harington, 2002, Matheus, 1995	NMC-7438	cranium
	Hester Creek, Hunker Creek	26,720 ±290	OxA-9259	Barnes et al., 2002	CMN-49874	ulna
	80 pup, Hunker Creek	29,695 ±1200	I-11037	Harington, 1980, 1997; Matheus, 1995	NMC-37577	humerus
Brown bear (*Ursus arctos*)	Hunker Creek	41,085 ±1050	Beta-16159	Harington, 1977, 1989	NMC-35965	mandible
American lion (*Panthera leo*)	Thistle Creek	32,750 ±370	TO-7743	Storer, 2002 pers. comm.	YG-54.1	mandible
Wolf (*Canis lupus*)	Nugget Gulch, Eldorado Creek	27,920 ±650	Beta-33191	Harington, 2002	NMC-45574	
	Dominion Creek	>47,240	Beta-89988	Harington, 2002		rib
Bison (*Bison* sp.)	Upper Dominion Creek	45,200 ±2100	Beta-79857	Harington, 2002	NMC-33876	metatarsal
Bison (*Bison priscus*)	Hunker Creek	20,370 ±230	Beta-28762	Harington, 2002	NMC-45289	foot muscle tissue
	Gold Run Creek	22,200 ±1400	I-3570	Harington and Clulow, 1973	horncore	
	Flat Creek	24,900 ±1000	I-3575	Harington, 1977, 1980		
	Quartz Creek	30,300 ±1850	I-3571	Harington, 1977, 1980	horncore	

Bison (*Bison priscus*) cont'd	Nugget Gulch, Eldorado Creek	30,810 ±975	Beta-33192	Harington, 2002; Harington and Morlan, 2002	NMC-46320	radioulna[1]
	Thomas Gulch and Klondike River confluence	39,290 ±540	TO-5650	Froese, 1997	DF-95-98-115	right metacarpal
	80 pup, Hunker Creek	44,660 ±1900	Beta-79856	Harington, 1980	CAMS-19419	thoracic vertebra
Bison (*Bison alaskensis*)	Gold Run Creek	>39,900	I-5405	Harington and Clulow, 1973; Harington, 1977	NMC-13506	horncore
Bison (*Bison bison athabascae*)	Quartz Creek	1,350 ±95	I-5404	Harington, 1977,1980	NMC-17519	frontal and horncore[1]
	Hunker Creek	1,465 ±85	I-11051	Harington, 1980		tibia[1]
Horse (*Equus* sp.)	Dominion Creek	27,760 ±650	Beta-8865	Harington, 2002	CR-79-10	metatarsal
	Dominion Creek	14990 ±220	I-9316	Guthrie, 1990		
	Dominion Creek	16,270 ±230	I-9271	Guthrie, 1990		
Horse (*Equus lambei*)	15 pup, Last Chance Creek	26,280 ±210	Beta-67407	Harington and Eggleston-Stott, 1996	CAMS-9769	partial carcass, age on bone
	Dominion Creek	14,870 ±260	I-3659	Harington, 1977, 1980	metacarpal	
Caribou (*Rangifer tarandus*)	Quartz Creek	5,010 ±100	I-8642	Harington, 1977	NMC-26011	radioulna
	Hunker Creek	11,350 ±110	Beta-27512	Harington and Morlan, 1992	ETH-4582	antler [1]
Giant moose (*Alces latifrons*)	Friday Gulch, Sulphur Creek	40,230 ±1100	Beta-79862	Harington, 2002	CRH-95-13	cranium
	Dominion Creek	35,610 ±340	TO-3712	Harington, 2002	NMC-25969	antler beam
Mammoth (*Mammuthus* sp.)	Hunker Creek	34,290 ±580	Beta-79855	Harington, 2002	NMC-9927	molar
	Gold Run Creek	32,350 ±1750	I-4226	Harington, 1997; Harington and Clulow, 1973	thoracic vertebra	
Woolly Mammoth (*Mammuthus primigenius*)	80 pup, Hunker Creek	25,680 ±580	I-8583	Harington, 1980	NMC-25996	metatarsal IV
	Gold Run Creek	18,030 ±120	Beta-70099	Harington, 2002		
	Quartz Creek	37,220 ±830	LU-3010	Harington, 2002		
Columbian Mammoth (*Mammuthus columbi*)	Quartz Creek	29,190 ±400	I-10971	Harington, 1977, 1980	NMC-29270	mandible with right third molar
American mastodon (*Mammut americanum*)	Gold Run Creek	18,460 ±350	TO-7745	Storer, 2002	YG-29.199	tooth
Western camel (*Camelops hesternus*)	Hunker Creek	19,770 ±640	TO-7740	Storer, 2002	YG-50.1	metacarpal
Dall Sheep (*Ovis dalli*)	Bonanza Creek	23,000 ±600	I-4225	Harington, 1977		horncore
	Hunker Creek	23,900 ±470	I-8580	Harington, 1977; Harington and Morlan, 1992	NMC-25926	antler
	Thistle Creek, Edas gulch	19,710 ±380	TO-7744	Storer, 2002 pers. comm.	YG-47.1	cranium

Tundra muskox (*Ovibos moschatus*)	Brewer Creek	2,800 ±100	I-3568	Harington, 1980		cranial fragment
Woodland muskox (*Bootherium bombifrons*)	Dominion Creek	23,710 ±320	TO-7741	Storer, 2002 pers. comm.	YG-66.1	braincase
Sedge (Carex sp.)	Last Chance Creek	25,700 ±400	Beta-171748	Zazula et al., 2003		Sedge seeds in peat within gravels below mucks
Sixtymile District						
Snow goose (*Chen caerulescens*)	Sixtymile River	39,020 ±690	TO-2698	Harington, 1997	NMC-38229	
Arctic ground squirrel (*Spermophilus parryi*)	Glacier Creek	11,170 ±205	QC-662	Guthrie, 1990	N/A	plant remains in nest
	Sixtymile River	45,160 ±1200	TO-2704	Harington, 1997		
	Glacier Creek	47,500 ±1900	Beta-16157	Harington, 1989	NMC-37557	skin and hair
Black-footed ferret (*Mustela nigripes*)	Sixtymile River	39,560 ±490	TO-214	Harington, 1989	NMC-43786	fibula
Wolverine (*Gulo gulo*)	Sixtymile River	41,420 ±1100	TO-2701	Youngman, 1993; Harington, 1997	NMC-42492	
Short-faced bear (*Arctodus simus*)	Sixtymile River	44,240 ±930	TO-2699	Harington, 1997	NMC-42388	
Brown bear (*Ursus arctos*)	Sixtymile River	36,500 ±1150	Beta-16162	Harington, 1989	NMC-38279	humerus
	Bolinden Creek	>43,400	Beta-68923	Harington, 2002	NMC-42105	cranium
	Sixtymile River	35,970 ±660	CAMS-51808	Barnes et al., 2002, Leonard et al., 2000	CMN-42381	ulna
Scimitar cat (*Homotherium serum*)	Sixtymile River	43,430 ±1100	Beta-89989	Harington, 1997	NMC-29194	humerus
	Sixtymile River	>52,300	Beta-68926	Harington, 1997	NMC-46442	humerus
Horse *(Equus* cf. *E. varae*)	Sixtymile River	46,660 ±840	TO-2702	Harington, 1997	NMC-46507	metatarsal
Caribou (*Rangifer tarandus*)	Sixtymile River	45,470 ±1150	TO-2700	Harington, 1997	NMC-42431	
Moose (*Alces alces*)	Miller Creek	4,040 ±60	TO-2356	Harington, 1997	NMC-47691	antler
Tundra muskox (*Ovibos moschatus*)	Miller Creek	3,280 ±90	I-10985	Harington, 1997	NMC-36137	cranium
	Miller Creek	21,160 ±280	Beta-13869	Harington, 1989	NMC-42047	cranium
Helmeted muskox (*Symbos cavifrons*)	Sixtymile River	>43,200	Beta-68924	Harington, 1997	NMC-42517	mandible
American mastodon (*Mammut americanum*)	Miller Creek	24,980 ±1300	Beta-16163	Harington, 1989	N/A	molar
Western camel (*Camelops hesternus*)	Sixtymile River	23,320 ±640	Beta-8864	Harington, 1989	NMC-38227	humerus
	Sixtymile River	39,030 ±350	Beta-115207	Harington, 1997	NMC-42549	mandible
	Sixtymile River	43,270 ±510	Beta-115205	Harington, 1997	NMC-46728	radius
	Sixtymile River	43,620 ±1100	Beta-89985	Harington, 1997	NMC-29194	
Dall sheep (*Ovis dalli*)	Sixtymile River	>49,800	Beta-68925	Harington, 1997	NMC-44711	horncore
Spruce stump (*Picea* sp.)	Sixtymile River	26,080 ±300	Beta-13870	Harington, 1997		
	Sixtymile River	>38,000	GSC-4485	Harington, 1997		

Others						
Arctic ground squirrel (*Spermophilus parryi*)	Stirling Bend, Stewart River	11,460 ±80	TO-4875	Bond, 1997	N/A	
Horse (*Equus* sp.)	Dublin Gulch, Haggart Creek, McQuesten River	31,450 ±1300	I-10935	Harington, 1989	NMC-34964	metatarsal
Horse (*Equus lambei*)	Scottie Creek	20,660 ±100	Beta-70102	R.M. Gotthardt and C.R. Harington, p.c. 1998	CAMS-11244	mandible
Mammoth (*Mammuthus* sp.)	Stewart River	32,300 ±1050	BGS-1022	Harington, 2002	N/A	tusk and bone
Mammoth (*Mammuthus* sp.)	Scroggie Creek	16,200 ±130	GSC-1893	Harington, 1977	NMC-21301	
Proboscidea (*Mammuthus* ?)	Stewart River	>33,000	SRC-3606	Bond, 1997	tusk (Reid outwash)	
Spruce wood (*Picea* sp.)	Hungry Creek	36,900 ±300	GSC-2422	Hughes et al., 1981		Beaver chewed wood

[1] Culturally modified

Pleistocene Archaeology in Yukon

Richard E. Morlan

The Yukon interior is situated at the eastern end of Beringia, a land-mass stretching from the Kolyma River in Siberia to the Mackenzie basin of northwestern Canada that remains ice-free even during continental glaciations. When sea level is lowered during glacial advances, the floor of the Bering and Chukchi Seas are exposed to form the Bering Land Bridge at the heart of Beringia, and the land bridge has been exposed more often than not during the past 2.5 million years (Morlan 1997). The land bridge has provided an avenue for inter-continental dispersals of many organisms during the late Cenozoic. For example, mammoths and bison moved eastward into North America, and camels and horses moved westward into Eurasia (Kurtén and Anderson 1980). Beringia also provided a route to the western hemisphere for Homo sapiens, a species ultimately of African origin, and it is logical to suppose that the earliest evidence for human occupation in the Americas should be found in eastern Beringia (Morlan 1987). In fact, such evidence may already have been found, but it is difficult to interpret, and is often controversial.

Most recent conservative summaries of eastern Beringian archaeology present the Nenana complex, dated to 11.5-12.0 ka in Alaska, as the beginning of the regional archaeological record (e.g. Dixon 2001; Yesner 2001). The Nenana complex is known from a series of securely dated archaeological sites where assemblages of stone tools and a few assemblages of animal bones are interpreted as the campsites of hunter-gatherers. Nenana technology exhibits similarities to the Dyuktai culture of Upper Paleolithic Siberia and possibly also to the fluted point traditions that were widespread south of the ice sheets at the close of the late Wisconsinan stadial. Dyuktai, Nenana, and the fluted point cultures represent the beginning of the readily interpretable part of the archaeological record, but a growing body of evidence suggests that people arrived in eastern Beringia long before these cultures arose. The earlier evidence comes from the Bluefish Caves and various localities in Old Crow Basin, northern Yukon, and from a few localities in the Klondike district of west-central Yukon.

The Bluefish Caves site consists of three small limestone rockshelters just east of the Yukon-Alaska border, about 65 km southwest of the village of Old Crow, near the headwaters of Bluefish River. Artifacts and bones are buried in loess both within and in front of the rockshelters (Cinq-Mars 1979, 1990, n.d.). The loess consists of silt derived from the Bluefish Basin during drainage events of glacial Lake Old Crow, an enormous lake system that covered several interior basins and was filled and drained in two major phases during the late Wisconsinan (Zazula et al. 2003). A pollen sequence in Bluefish Cave 1 reveals the usual eastern Beringian pattern of herbaceous steppe-tundra during full glacial time, followed by a late glacial birch rise dated elsewhere to 13.5-14.0 ka, followed by the formation of open spruce woodland at the beginning of the Holocene (Ritchie et al. 1982). These environmental changes are also seen in a sequence of microtine rodent assemblages that reflect a shift from xeric herbaceous steppe-tundra to more mesic shrub tundra and formation of boreal forest (Morlan 1989). Evidence of intermittent human occupation is found throughout the late glacial sequence, but the site seems to have been abandoned by people during the Holocene.

The most obvious artifacts from Bluefish Caves are microblades, microblade cores, a core tablet, burins, and burin spalls. These are not directly dated, but one of the burins was closely associated with xerophilous microtine taxa, the collared lemming (*Dicrostonyx torquatus*) and the singing vole (*Microtus miurus*). These two species lived near the site only when xeric herbaceous steppe-tundra existed there, before 13.5 ka, therefore before the appearance of the Nenana complex in Alaska. A whittled and polished sheep (*Ovis dalli*) tibia resembling a broken fleshing tool was found in the lower loess at Cave 1 and was directly dated to 13.5 ka. A similar broken tool made on a caribou (*Rangifer tarandus*) tibia from basal loess at Cave 2 was dated to 24.2 ka. Just above the caribou tibia was a flake detached from a mammoth (*Mammuthus* sp.) limb bone by percussion flaking. Nearby was the mammoth limb bone core to which the flake can be refitted on its scar. Dates on the flake and core average 23.5 ka. Percussion flaking of mammoth bones has been shown to be a method of producing sharp-edged bone knives that are useful butchering tools (Stanford et al. 1981). A number of bones from caves 1 and 2 exhibit distinctive butchering scars inflicted with stone tools, and dates on these bones range from 24.0 to 12.0 ka (Cinq-Mars and Morlan 1999).

Even older evidence of human occupation has been found in Old Crow Basin, but most of the artifacts have been recovered from secondary deposits. Lacking an archaeological context to aid their interpretation, these artifacts can be interpreted only through taphonomic analysis that seeks to link specific morphologies with specific agents and processes. The artifacts consist of fresh-fractured proboscidean limb bones, percussion flaked bone cores, and flakes detached from such cores. Detailed descriptions are available elsewhere (Morlan 1980, 2003a). It is important to note that after more than 20 years of investigation into elephant bone taphonomy (e.g., Haynes 1991) no process except human technology has been found to produce such complexly modified specimens. Dates on these modified bones range from 39.5 ka to 25.2 ka, and the modifications could have been made only when the bones were fresh (Morlan 1984; Morlan et al. 1990). *Bison priscus* bones exhibiting butchering marks inflicted with stone tools have been found in situ at two Old Crow Basin localities, and these are dated to 42.0 ka and 36.5 ka, respectively. This body of evidence is interpreted to mean that people arrived in Old Crow Basin during marine isotope stage 3, the Boutellier Non-glacial Interval, and they lived there until the basin was flooded by glacial Lake Old Crow (Zazula et al. 2003), after which they inhabited uplands surrounding the northern Yukon basins, as shown at the Bluefish Caves.

Evidence from the Klondike district suggests that people were widespread in easternmost Beringia during isotope stages 2 and 3. Among the thousands of fossil bones recovered from the frozen organic silt ("muck") in the Klondike stream valleys, several dozen specimens exhibit cut marks that may reflect human butchering and distinctive fracture patterns consistent with those induced by humans during marrow retrieval and tool production. Two such specimens have been studied, dated, and reported so far. One is a *Bison priscus* radioulna from Nugget Gulch that exhibits a ring-crack on its anterior surface (Harington and Morlan 2002). The size and placement of the ring-crack suggest it was most likely produced with a hammerstone wielded by a person intending to open the marrow cavity, and it has been dated to 31.0 ka. The second specimen is a 12.8 cm segment of caribou (*Rangifer tarandus*) antler from Hunker Creek (Harington and Morlan 1992). A nearly flat but slightly keeled surface has been created at one end by grooving and snapping the antler beam, and the other end is shaped into a blunt point that terminates in cortical tissue, avoiding the spongy medulla. The artifact may have served as a punch for indirect percussion while knapping stone tools. Force applied by percussion on the butt would have been transmitted to the blunt point held in contact with a stone object such as a microblade core (e.g. Sanger 1968: Fig. 11). Dated to 11.35 ka, this punch could be a tool that corresponds with the Nenana and Denali complexes in nearby Alaska.

These glimpses of Pleistocene archaeology in Yukon suggest that people had already reached eastern Beringia long before the last major advances of Laurentide and Cordilleran glaciers. The people lived by hunting and scavenging the abundant big game of the region, including mammoths, bison, horses, sheep and caribou, continuing cultural practices long established during the Upper Paleolithic of Eurasia. These groups were in a position to move southeastward and colonize the rest of the hemisphere with little or no change in technology or subsistence practices. A few sites in both North and South America suggest that they took advantage of this opportunity, but every one of the sites is controversial for one reason or another, and this important aspect of ancient history remains unresolved.

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 $^{40}Ar/^{39}Ar$ (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 (AAAP) 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 AAAP (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): <u>White</u> <u>River</u> <u>Ash</u> (1-2 ka), Dawson (c. 24 ka), VT (0.13±0.03), Old Crow (0.14±0.01), Dominion Creek (0.17±0.02), <u>Sheep</u> <u>Creek</u> (0.19±0.02), Midnight Dome (1.09±0.18), <u>Flat</u> <u>Creek</u> (1.23±0.18) Mosquito Gulch (1.45±0.14), <u>Paradise</u> <u>Hill</u> (1.54±0.13), <u>Quartz</u> <u>Creek</u> (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.

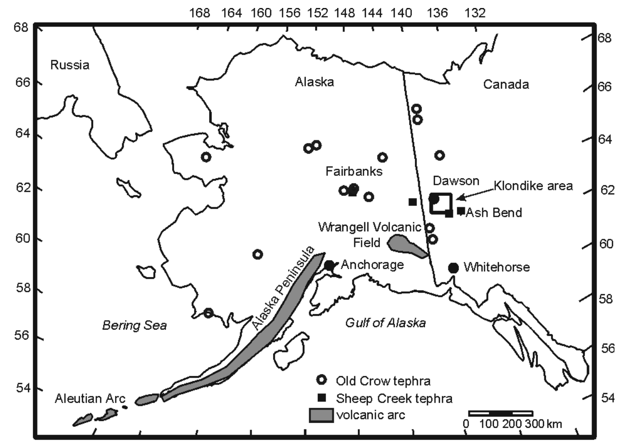

Figure 9. Location of the Klondike district in relation to the Aleutian - Alaska Peninsula arc, the Wrangell volcanic field, and the distribution of the widespread Old Crow and Sheep Creek tephra beds.

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

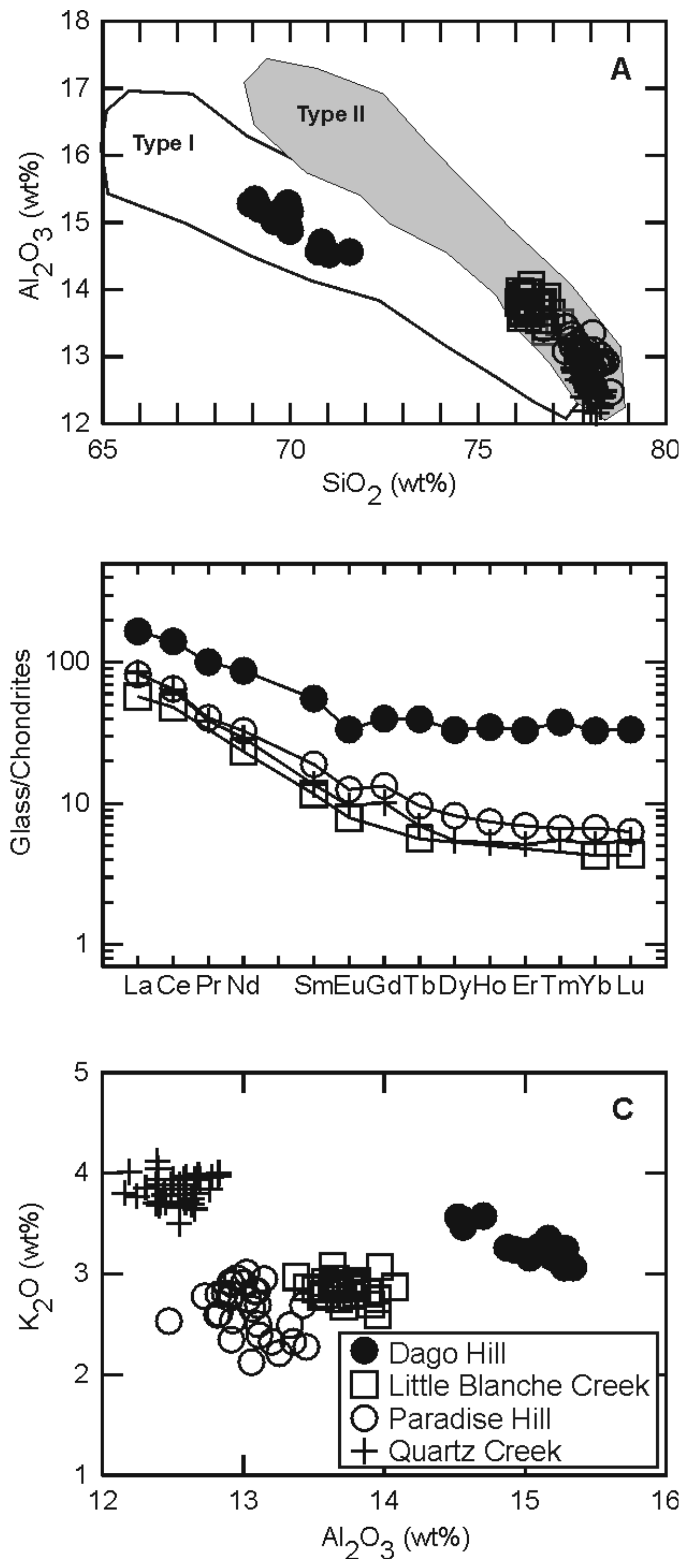

Figure 10. Major- and rare-earth-element composition of glass shards in Dago Hill, Paradise Hill, Little Blanche Creek, and Quartz Creek tephra beds. A: Classification of tephra beds into type I and type II groups of Preece et al. (1999). B: Rare-earth element profiles. C: Oxide variation plot showing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

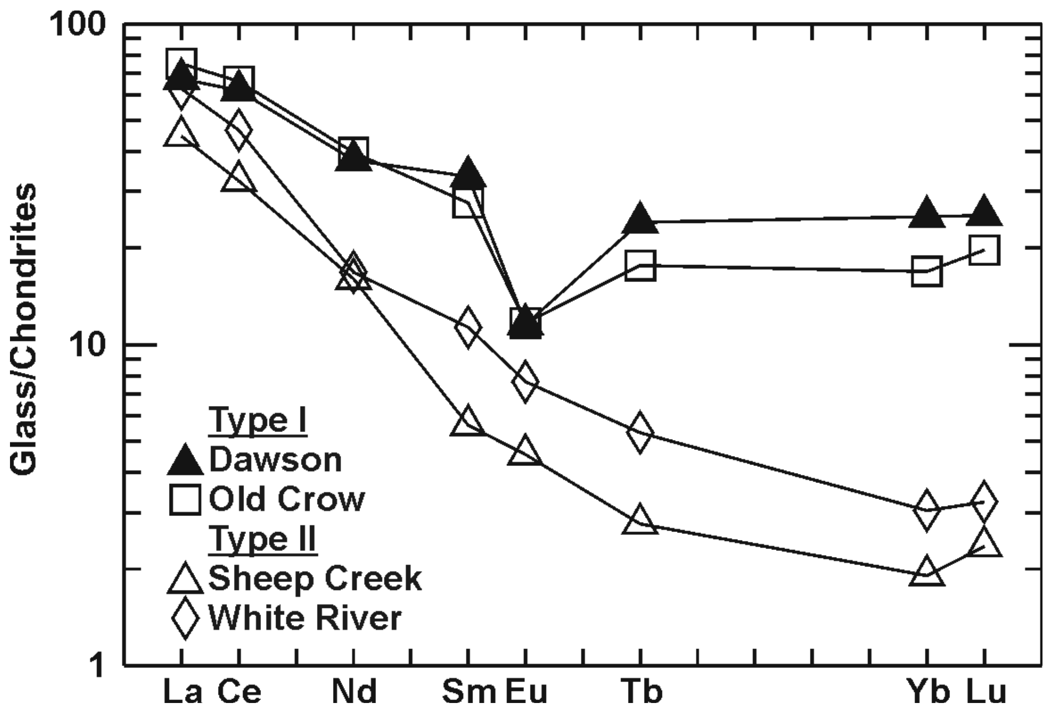

Figure 11. Chondrite-normalized rare-earth element profiles of tephra beds in the Klondike district, Yukon. Tephra beds have been grouped into two classes, type I and type II. Glass analyses by the instrumental neutron activation method.

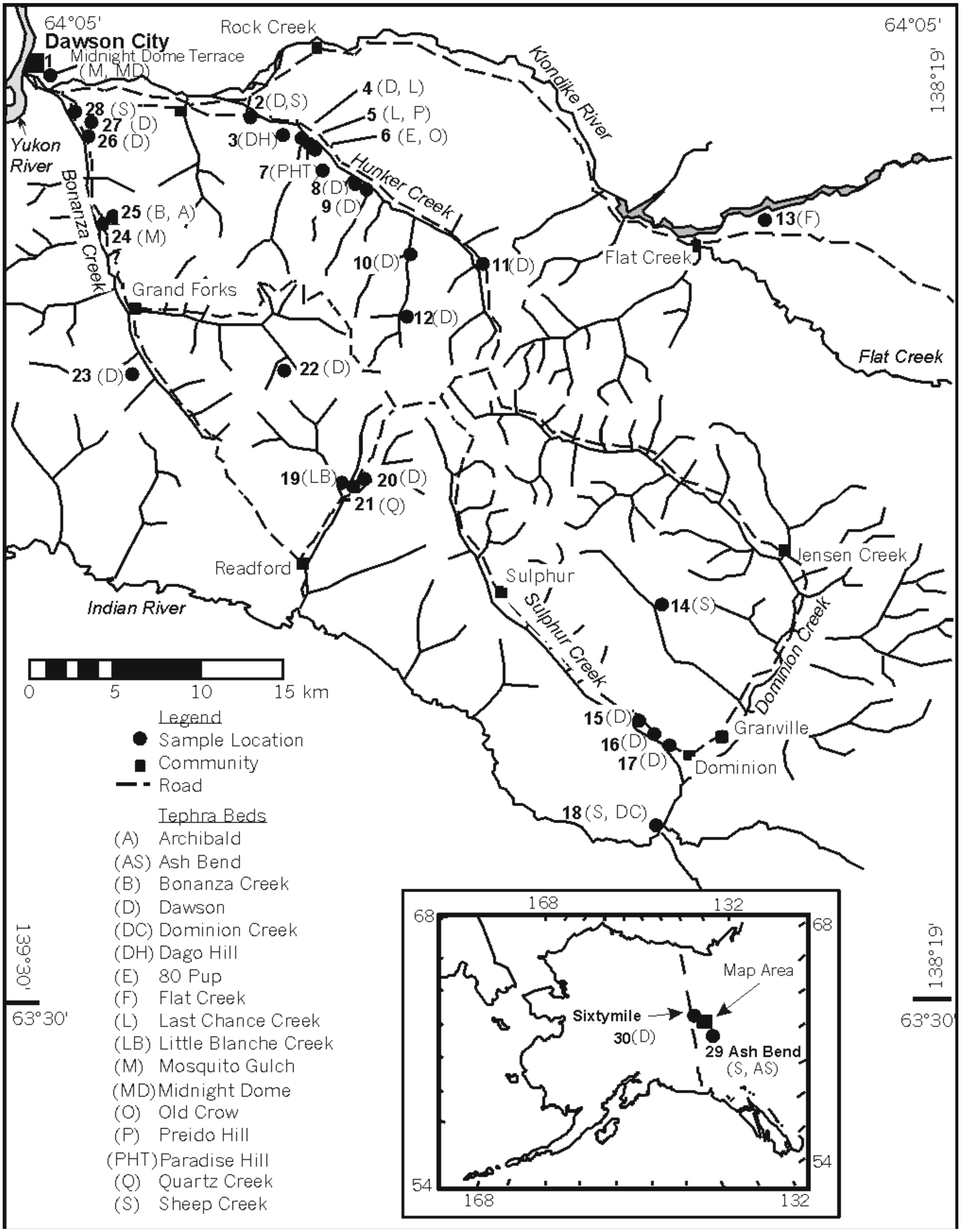

Figure 12. Location and identity of tephra samples collected in the Klondike district, Yukon.

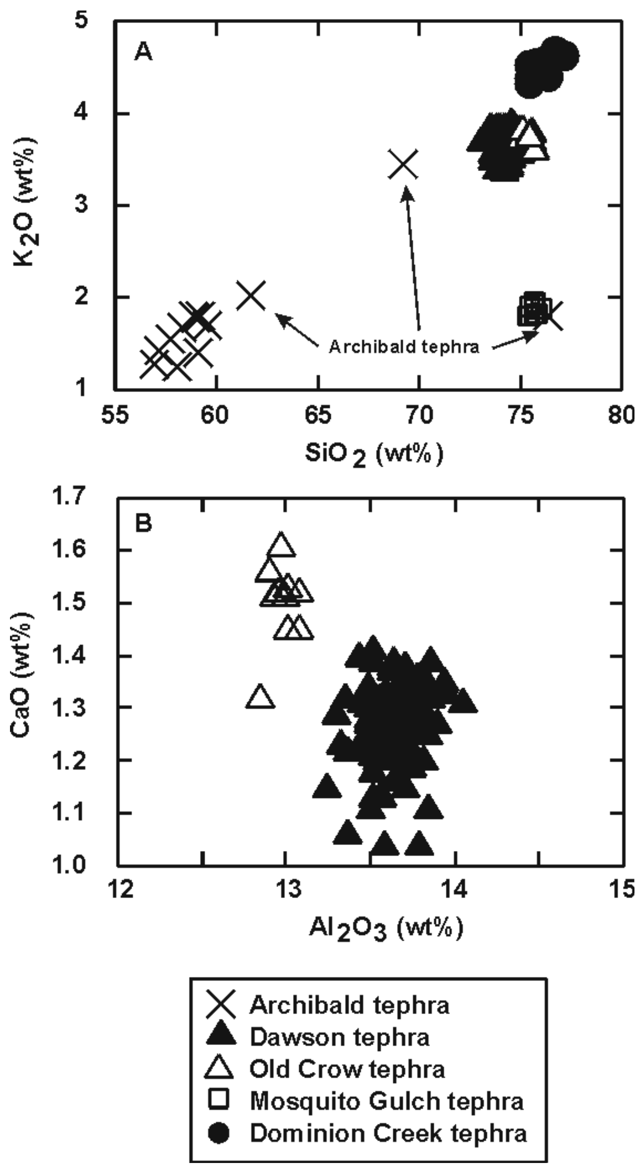

Figure 13. Oxide-variation diagrams of type I and Archibald tephra beds in the Klondike region showing distinctive chemical composition of glass shards in each bed.

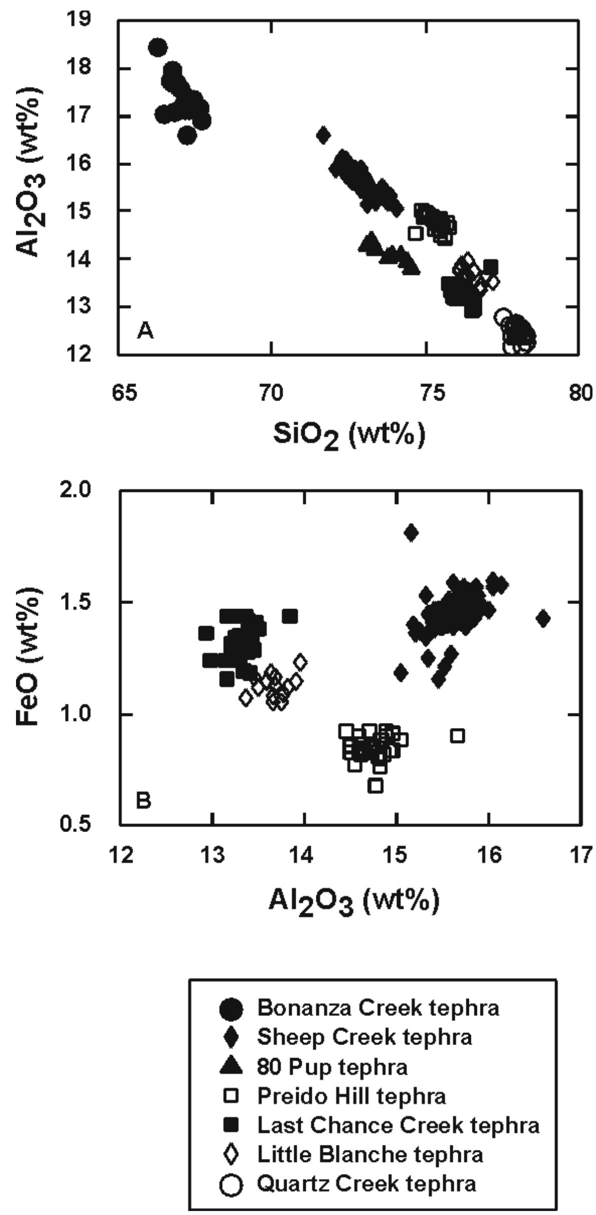

Figure 14. Oxide-variation diagrams of type II tephra beds in the Klondike region showing distinctive composition of glass shards in each bed.

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 areas, 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 outside 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 northwestward. 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

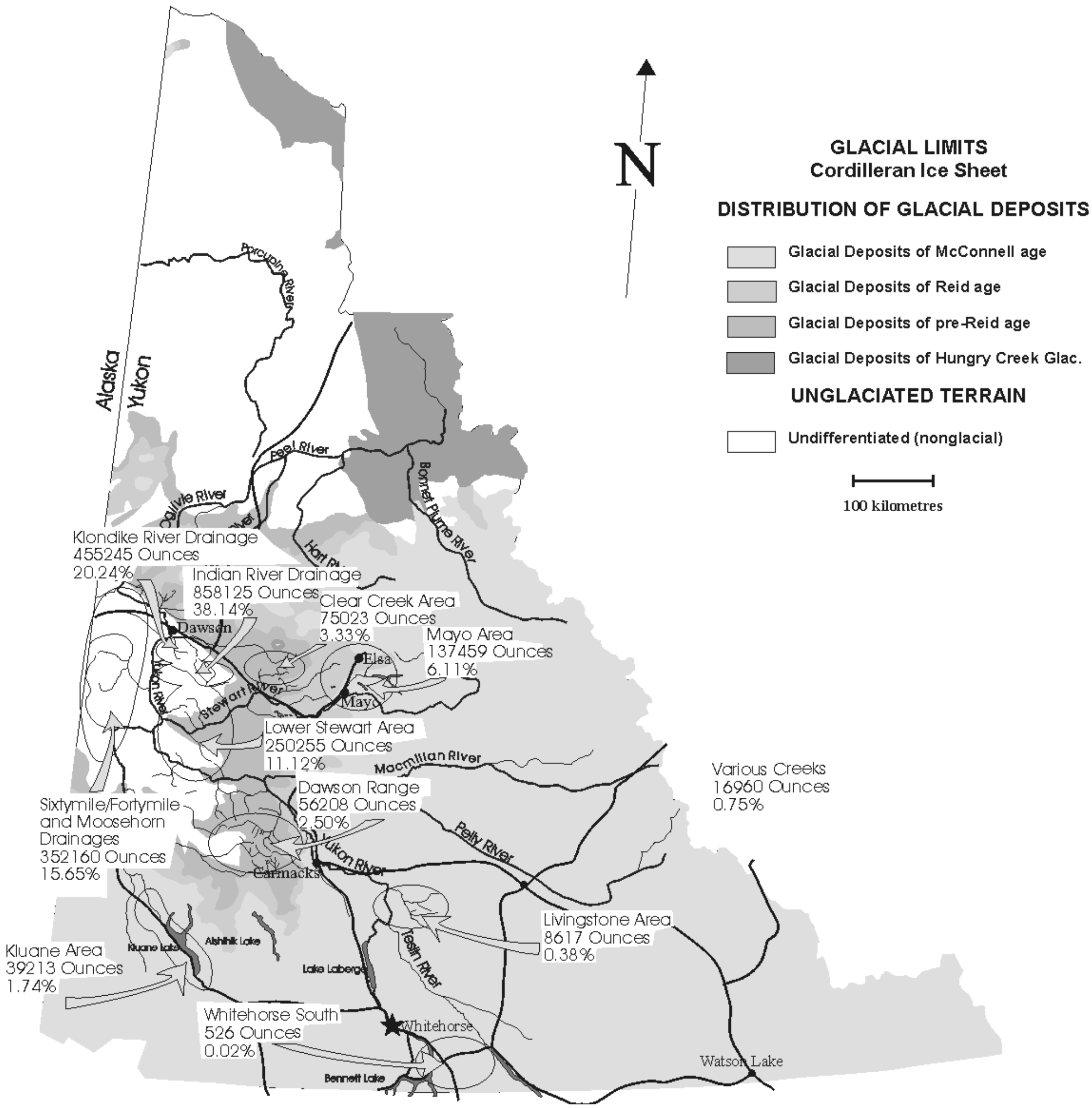

Figure 15. Yukon placer gold production by region 1978-2000.

Field Excursion Itinerary: Whitehorse to Dawson and Stewart River Valley, May 29, 2003

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

D.G. Froese and J.D. Bond

The route from Whitehorse to Five Finger Rapids is entirely within the limits of the last glaciation (McConnell or late Wisconsinan). At Five Finger Rapids we are near the McConnell limit of the Cordilleran Ice Sheet. Here, the Yukon River channel was diverted at the end of the last glaciation through a resistant unit of conglomerate and sandstone of the Laberge Group. Four rock islets divide the flow of the river into 5 channels.

Up until the early 1950s when the Yukon River was still the primary means of transportation between Dawson City and Whitehorse, these rapids posed a navigational hazard to the riverboats. Overpowered by the current, the boats had to be winched through the narrow channel at this point of the Yukon River. Large anchor points are still visible along the valley walls.

On the far side of the Yukon River the upper terrace 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 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 for 25 kilometres to the site of Minto. The highway then traverses Reid terrain en route to Pelly Crossing.

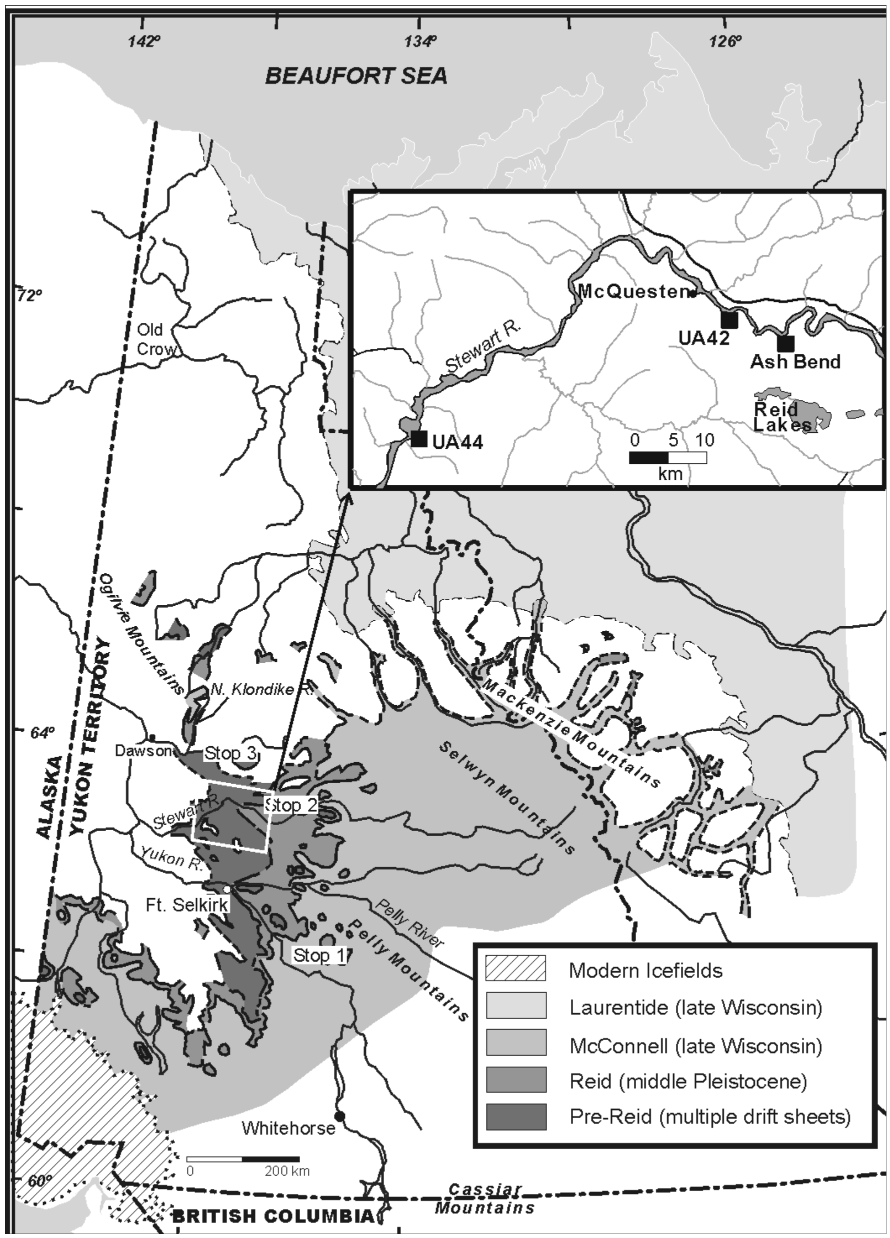

Figure 16: Location of stops in relation to limits of glaciation in Yukon Territory (adapted from Westgate et al., 2001).

STOP 2: Overview of Ash Bend Section, Stewart River: Type section for the Reid glaciation and early Beringian Bison fossils (63° 30' N, 137° 15' W)

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

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 debris-flow diamict (Figures 17 and 18; Bond 1997). The diamict can be separated into two diamicts, separated by eolian sand. Both contain red, elongate slabs of paleosol 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 18).

Incised into the Reid drift and exposed in a gully near the section is an interglacial organic silt (Figures 18, 19). The deposit consists of organic silts with transported spruce limbs and logs, abundant spruce pollen, and remains of bison, mammoth and moose (Table 3). The overlying silty beds contain less organic material, a reduced spruce pollen content, and the Sheep Creek tephra (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, Ash Bend tephra (UT1051) occurs in the lower silts, 90 cm above the Sheep Creek tephra.

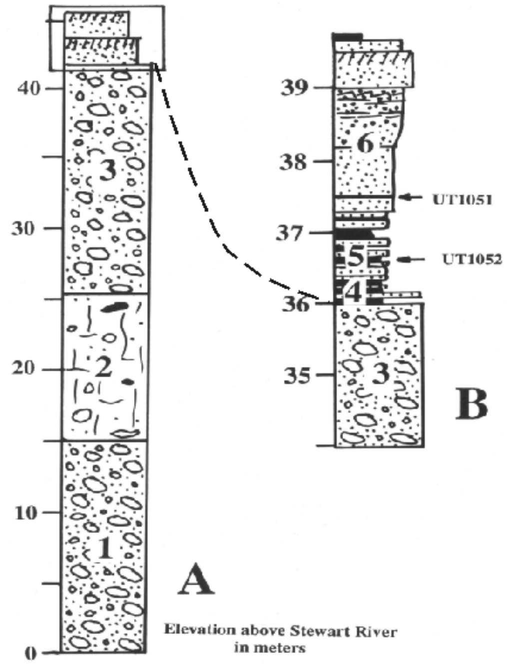

Figure 18: Ash Bend stratigraphy. Unit 1 is advance Reid outwash, Unit 2 is Reid basal till, Unit 3 is retreatal Reid outwash and Units 4 and 5 are a post-Reid organic silt (interglacial) which contains Sheep Creek tephra (UT1052) and Ash Bend tephra

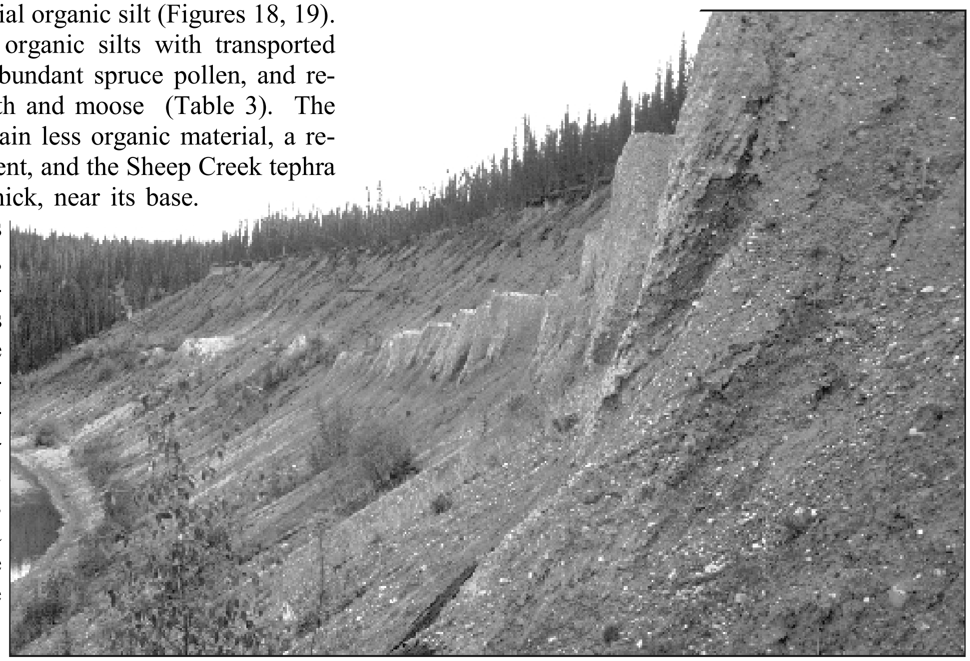

Figure 17: Ash Bend on the Stewart River in Tintina Trench. This is the type section for the Reid glaciation in central Yukon. The 60 m high section consists of (from bottom to top) advance outwash, Reid till, retreatal outwash and a debris-flow deposit. View is to the east.

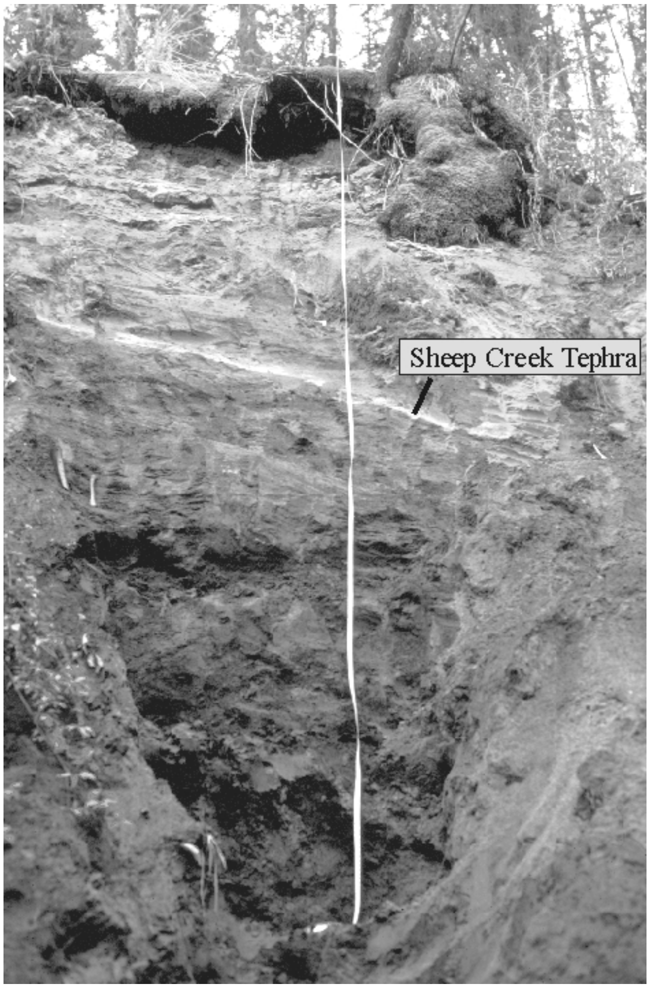

Figure 19: A swale carved into the Reid deposits at Ash Bend contains organic-rich silts, permafrost, vertebrate fossils and the Sheep Creek tephra. The Sheep Creek tephra is approximately 190 ka BP (Berger et. al 1996; Westgate et al., 2001).

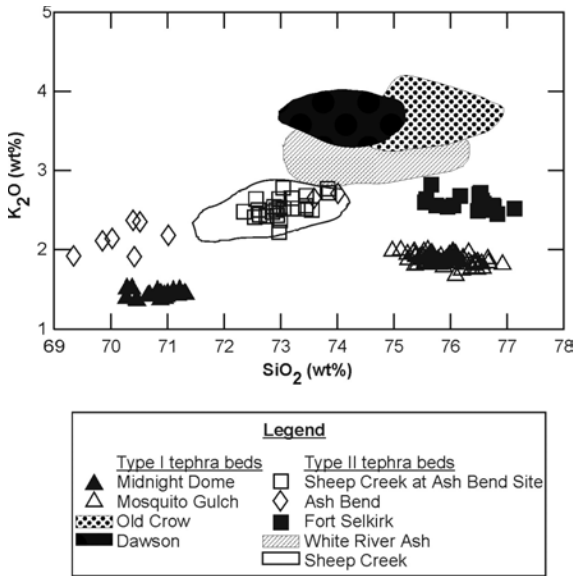

Figure 20. K_2O - SiO_2 scatter plot of glass shards in tephra beds with corresponding data on a few other widespread, well known tephra beds in the

The youngest deposits of this channel infill consist of massive sands with a paleosol, covered by another sandy bed and woody peat. The organic silt deposit contains permafrost.

Paleoenvironmental 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 21). 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 Oxygen Isotope Stage (OIS) 6 and 7, exactly the palaeoenvironmental setting recorded at Ash Bend (Figs. 20, 21). In other words, the dense boreal forest belongs to interglacial OIS 7 and the succeeding tundra environment to OIS 6. All these constraints strongly suggest an OIS 8 age for the Reid Glaciation.

A corollary issue is the extent of the Cordilleran Ice Sheet in central Yukon during OIS 6. Because the Reid Glaciation is older than OIS 6 and the next most extensive drift sheet belongs to the McConnell Glaciation of late Wisconsin age (Fig. 16), it follows that the advance of the Cordilleran Ice Sheet across central Yukon during OIS 6 was less extensive than during the later OIS 2. Glacial deposits of OIS 6 age must have been entirely covered by McConnell glacial drift. These observations deny the simplistic model of progressive restriction in the extent of the Cordilleran Ice Sheet in central Yukon with time.

Vertebrate Paleontology

The main interest in the Ash Bend site from a vertebrate paleontological perspective is that bison bones have been recovered *in situ* beneath Sheep Creek tephra, now dated to 190 ka BP (Berger et al., 1996; Westgate et al., 2001). These specimens indicate that bison were present by at least isotope stage 7

Table 3. Fauna recovered from Ash Bend (modified from Hughes et al., 1987).

Fossil	Location	Collector	Material (Collection Number)
Bison sp.	from slump in gully - "presumably eroded from muck overlying out-wash gravel	O.L. Hughes, 1965	first phalanx (NMC 10532)
Mammuthus sp.	from slump about 6 m above stream level	O.L. Hughes and C.R. Harington, June 21, 1966	left ulna, proximal end (NMC 11686)
Bison priscus	from slump about 4.6 m above stream level	O.L. Hughes and C.R. Harington, June 21, 1966	left horncore and part of left frontal and orbit (NMC 11687) immature?
**Bison* sp.	in place in gray silt 1.1 m below Sheep Creek tephra	O.L. Hughes and C.R. Harington, June 21, 1966	right metacarpal (NMC 11688)
Bison sp.	from base of slump	O.L. Hughes and C.R. Harington, June 21, 1966	?right metacarpal, distal epiphy-sis (NMC 11689) - immature
**Bison* sp.	in place in gray silt 1.1 m below Sheep Creek tephra	O.L. Hughes and C.R. Harington, June 21, 1966	right pelvic bone (acetabulum) (NMC 11690)
**Bison* sp.	in place in grey silt 0.8 m below Sheep Creek tephra	O.L. Hughes and C.R. Harington, June 21, 1966	first phalanx (NMC 11691)
Bison sp.	from base of slump	O.L. Hughes and C.R. Harington, June 21, 1966	first phalanx (NMC 11692)
Bison sp.	from slumped material	O.L. Hughes and C.R. Harington, June 21, 1966	first phalanx (NMC 11693)
Bison sp.	from base of slump	O.L. Hughes and C.R. Harington, June 21, 1966	right astragalus (NMC 11694)
Bison sp.	from slump	O.L. Hughes and C.R. Harington, June 21, 1966	left humerus, distal end (NMC 11695)
Bison sp.	from slump	O.L. Hughes and C.R. Harington, June 21, 1966	lumbar vertebra fragment (NMC 11696)
Bison sp.	from slump about 0.3 m above stream level	C.R. Harington, June 21, 1967	first phalanx (NMC 17336)
Bison sp.	from slump about 12.4 m above stream level	C.R. Harington, June 21, 1967	vertebra, epiphyseal plate
**Bison* sp.	in place in black silty clay 1.3 m below Sheep Creek tephra	C.R. Harington, June 21, 1967	right pelvic fragment (NMC 17335)
Alces sp.	from slump (Stn. 4, 39.5 m), greyish sand	O.L. Hughes, 1969	right tibia lacking proximal end (NMC 35841)
cf. *Mammuthus* sp.	from slump	O.L. Hughes, 1969	right scapula, proximal fragment (NMC 35842)
Bison sp.	not in place ("float")	O.L. Hughes, 1969	left radio-ulna, distal half (NMC 35843)
Bison sp.	not in place ("float")	O.L. Hughes, 1969	right radio-ulna, distal end (NMC 35844), smaller size indicates different individual than NMC 35843

Bison sp.	from slump in gully	O.L. Hughes, 1969	second phalanx (NMC 35845)
Mammuthus sp.	in stream	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	right basioccipital region (NMC 29297)
**Bison* sp.	in place 1.2 m below Sheep Creek tephra and just above organic silts representing former forest floor	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	right mandible fragment, ascending ramus (NMC 29296)
**Bison* sp.	in place 1.2 m below Sheep Creek tephra and just above organic silts representing former forest floor	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	approximately 6th thoracic vertebra (NMC 29301)
**Bison* sp.	in place 1.2 m below Sheep Creek tephra and just above organic silts representing former forest floor	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	approximately 7th thoracic vertebra (NMC 29302)
**Bison* sp.	in place 1.2 m below Sheep Creek tephra and just above organic silts representing former forest floor	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	approximately 8th thoracic vertebra (NMC 29300)
+Cervidae?	from slump	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	long bone fragment with spiral fracture (Archaeological Survey of Canada collection, Kj Vd-1)
+*Mammuthus* sp.	from slump	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	rib fragment with chewed area near fracture (ASC collection, Kj Vd-1)
+*Bison*?	from slump	R.E. Morlan and R. Bonnichsen, June 4, 5, 1975	long bone fragment with spiral? fracture (Archaeological Survey of Canada collection, Kj Vd-1)
**Bison* sp.	in place in organic silt at or just below Sheep Creek tephra NMC 43897, 43898 and 43899 articulate and are part of an individual	J.V. Matthews, Jr. and C.E. Schweger, June 23, 1983	left astragalus (NMC 43897)
**Bison* sp.	in place in organic silt at or just below Sheep Creek tephra	J.V. Matthews, Jr. and C.E. Schweger, June 23, 1983	left tibia, distal end (NMC 43898)
**Bison* sp.	in place in organic silt at or just below Sheep Creek tephra	J.V. Matthews, Jr. and C.E. Schweger, June 23, 1983	left calcaneum, proximal half (NMC 43899)
Mammuthus	On toe of colluvial apron, likely weathered from glaciofluvial gravel	J.D. Bond, 1994	tusk (>33,000 14C yr B.P., SRC-3606)

* fossils found in place

\+ specimens identified by R.E. Morlan, remainder identified by C.R. Harington

ASH BEND, STEWART RIVER, YUKON: RELATIVE PERCENT DIAGRAM

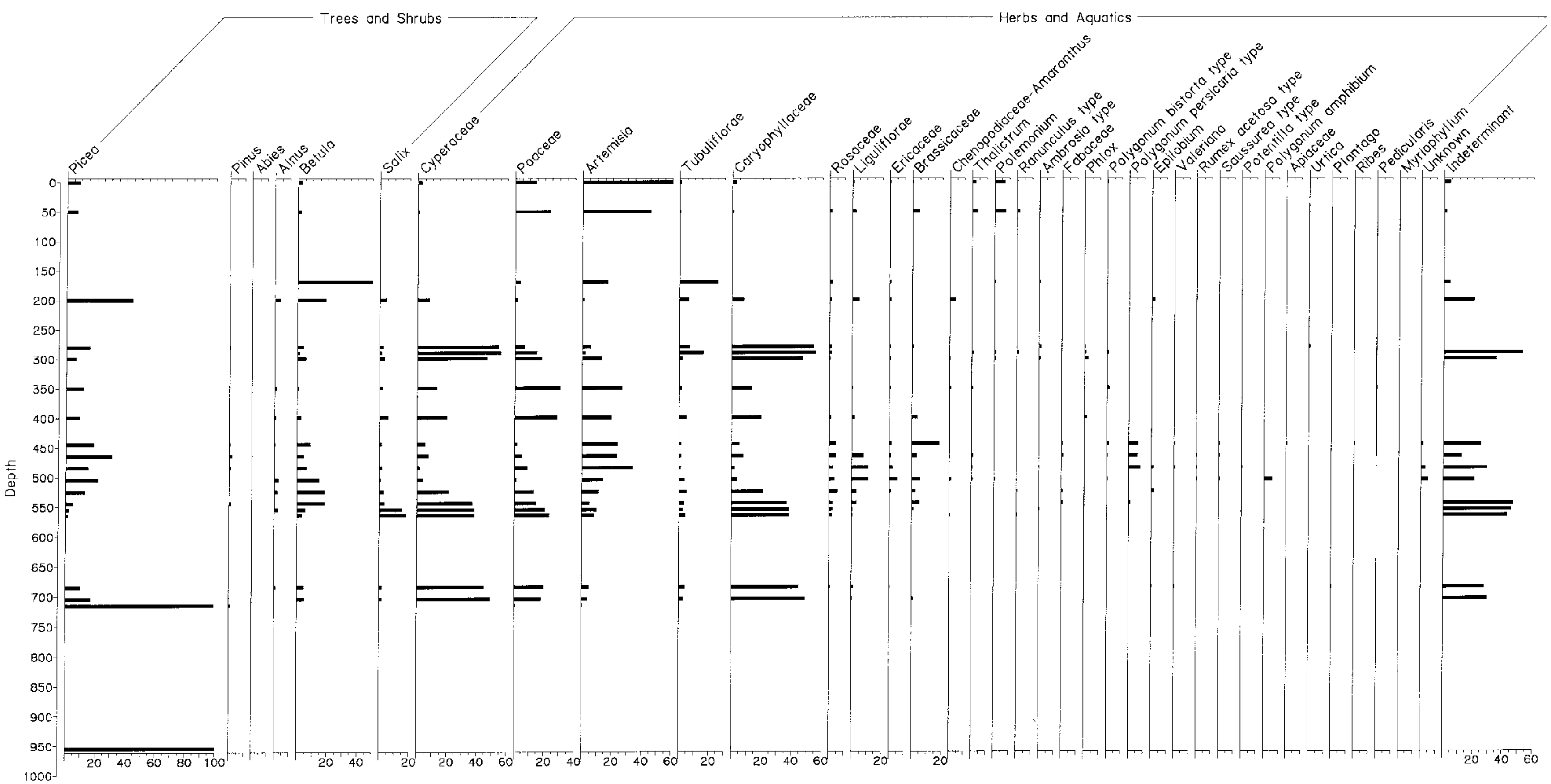

Figure 21. Fossil pollen diagram of organic sediments at Ash Bend section, Stewart River.

(late Middle Pleistocene), thus are the oldest dated bison in eastern Beringia with clear stratigraphic and temporal context. Bison, mammoth and moose remains have been collected at the site by geologists, paleobiologists and archaeologists since the discovery of the site in the summer of 1965. Although some bones have been found in place, most have been recovered from slumped material in a gully where the tephra is exposed.

Bison have long been considered to have immigrated to North America from Eurasia during the Rancholabrean land mammal age, now often defined as beginning at the base of the Bruhnes paleomagnetic chronozone (0.78 Ma, Graham, 1998). Others have narrowed the timing of their first arrival in the continental United States during late Illinoian (stage 6 or 7) or early Sangamon (stage 5e) time (Guthrie, 1990). However, the timing of bison colonization in North America is controversial and unresolved because specimens recovered from sediments that are beyond the range of radiocarbon dating often do not have well established ages. Within central Alaska, some bison specimens were speculated to be of middle Pleistocene age (Péwé, 1975), but these earlier reports are largely based on stratigraphic correlation of lithologic units, most of which lacked independent chronology (Péwé, 1975; *c.f.* Westgate et al. 1990; Preece et al., 1999).

STOP 3: Tintina Trench Lookout: Brief geology of the Tintina Trench and Ogilive 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 22).

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 below the unconformity. These beds are composed of northward tilted conglomerate, sandstone, silt and peat. Pliocene to middle Pleistocene fluvial and glacial conformable strata are above 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 overlie the tilted mid Miocene beds. 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.

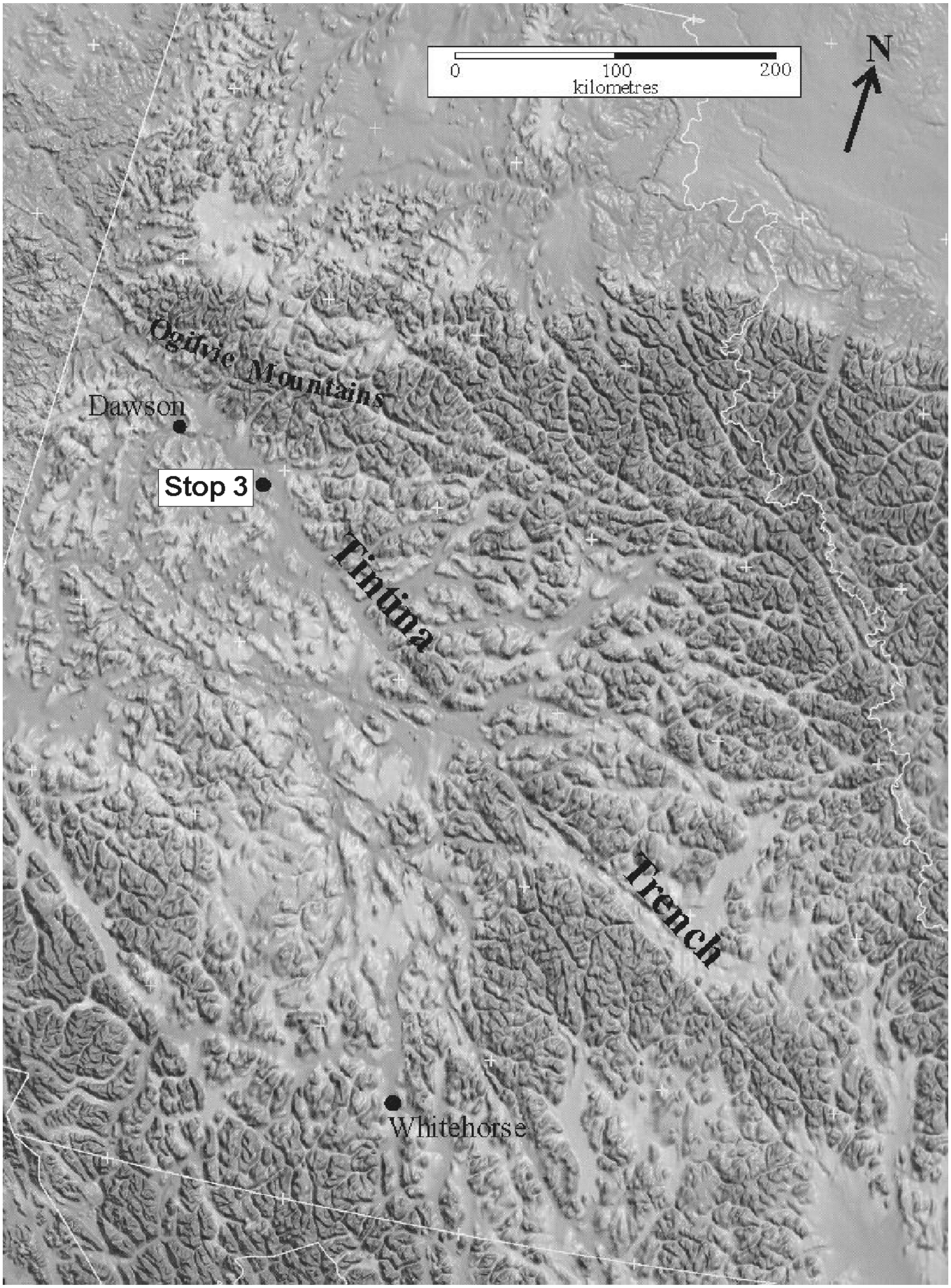

Figure 22. The Tintina Trench forms a distinct lineation in the physiography of central and southern Yukon.

Field Excursion: Klondike Valley and Northern Klondike Goldfields, May 29, 2003

The post-meeting excursion will be spent looking at 'muck' exposures in the Klondike goldfields (Figure 23). We will do an orientation on the Midnight Dome summit above Dawson City, followed by trips to sites that illustrate the taphonomic and paleoecologic setting of the Pleistocene fauna.

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 paleomagnetic timescale is shown in Figure 24.

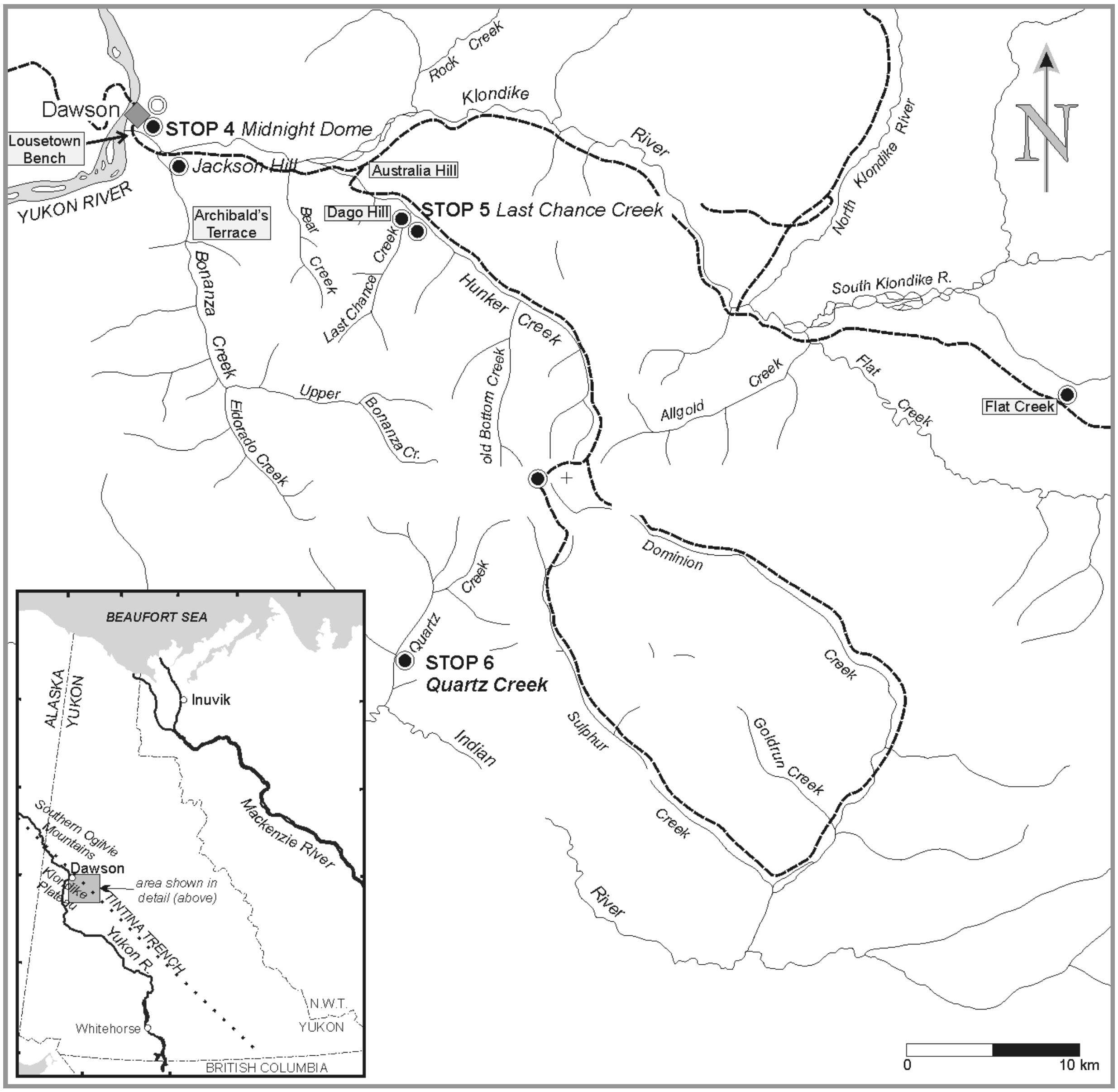

Figure 23. Map of Klondike Placer District with sites mentioned in text.

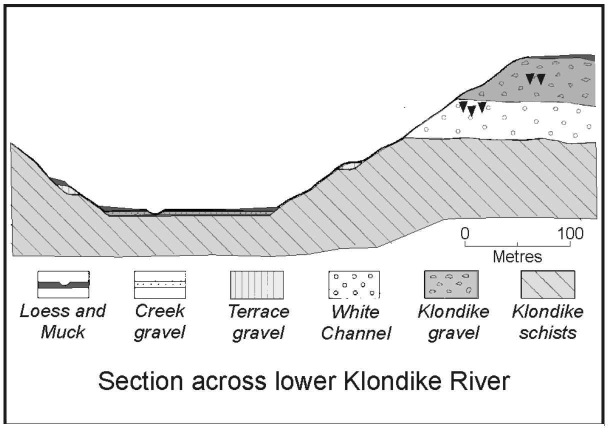

Figure 24 Generalized cross-section of terrace gravels across Klondike valley (after McConnell 1907).

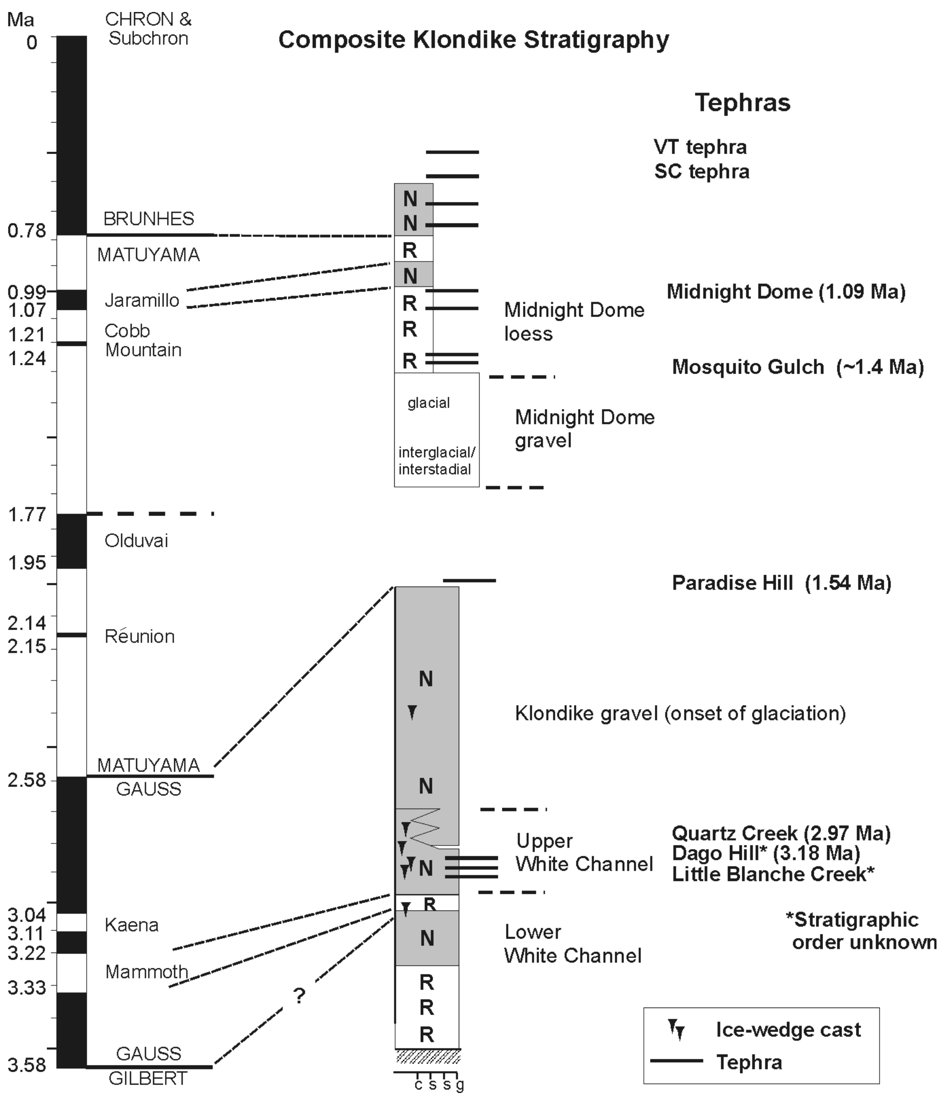

Figure 25. Correlation of Klondike stratigraphic units to the geomagnetic polarity timescale.

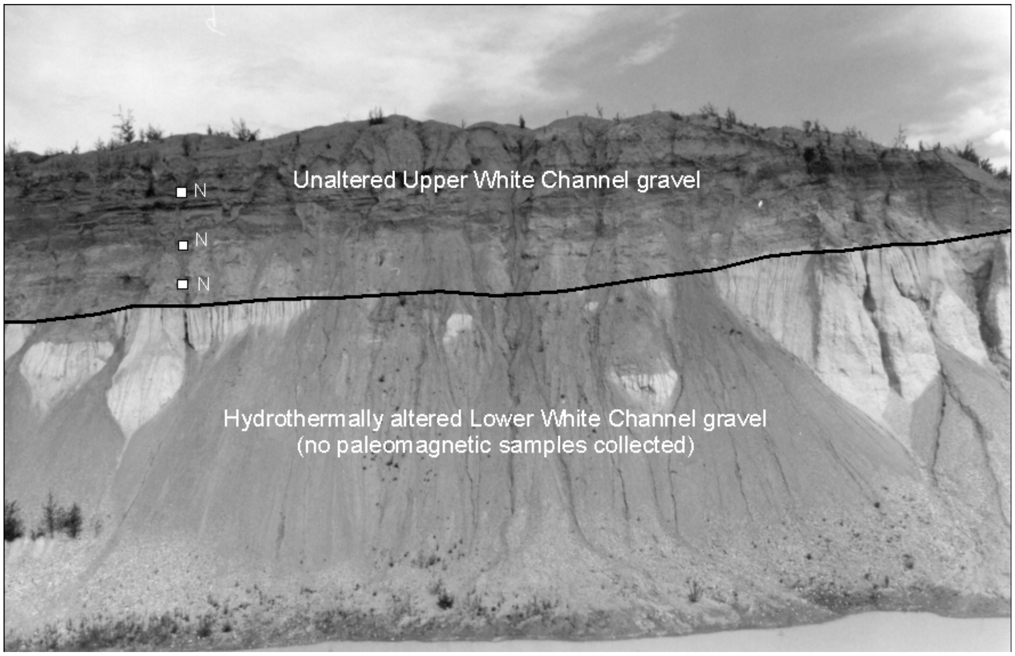

Figure 26. White Channel gravels on Dago Hill.

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 26, 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 24), 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 25; 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 23) 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. 2001) 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.

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 because they preserve abundant faunal remains and provide an opportunity for palaeoecological investigations, as you will see on the trip.

Stop 4 Midnight Dome Overview (64º 03' N, 139º 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

Rivers draining the north slope of the St. Elias Mountains and adjacent Alaska Range of Yukon and Alaska, the headwaters of the Yukon River, make a circuitous route to the sea. While the water falls near the Pacific coast, it makes a nearly 2500 km journey through western Yukon and Alaska before joining Pacific waters in the Bering Strait. This anomaly of drainage coupled with relatively young, late Tertiary mountain ranges forming the drainage divide, has led to speculation that rivers of interior Yukon and Alaska drained across the region of the St. Elias and Alaska ranges prior to, and through, mountain building (Brooks, 1902; McConnell, 1907; Cairnes 1915; Tempelman-Kluit, 1980; Duk-Rodkin *et al*., 2001) (Fig. 26). Extending this logic, Tempelman-Kluit (1980)

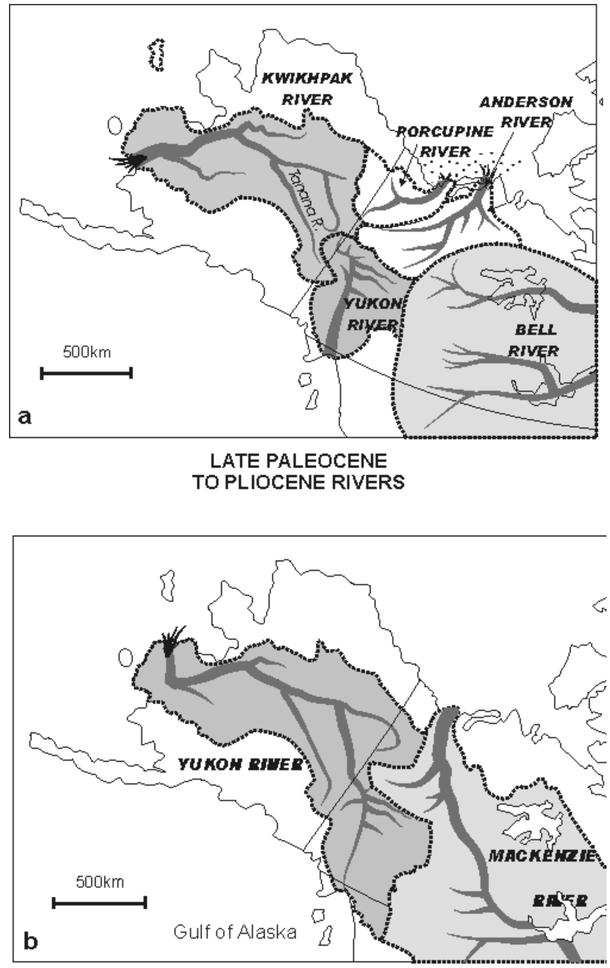

Figure 27. A. Pre-glacial drainage systems of northwestern North America. B. Modern drainage systems resulting from extensive Plio-Pleistocene glaciation.

observed a series of 'barbed' drainages along the middle reaches of the Yukon River in west-central Yukon, which he suggested may be explained by the presence of a pre-glacial drainage divide between a south flowing Yukon River and a westerly flowing Alaskan River system about 50 km east of the Yukon-Alaska border. This he argued, was consistent with reconstructions of Mesozoic clastic deposits and interpreted Tertiary erosional surfaces sloping south opposite the present drainage. Duk-Rodkin et al., (2001) mapped a series of southerly-sloping terraces and their lithologies from the hypothesized paleo-continental divide to Dawson City, opposite to the present river direction, consistent with the reversal hypothesis (Figs. 27, 28).

However, to date the evidence for a south-flowing Yukon River system has been equivocal in that southerly-dipping terraces, 'barbed' drainages and other anomalies of drainage may simply be reflecting local tectonism (Hughes et al., 1972).

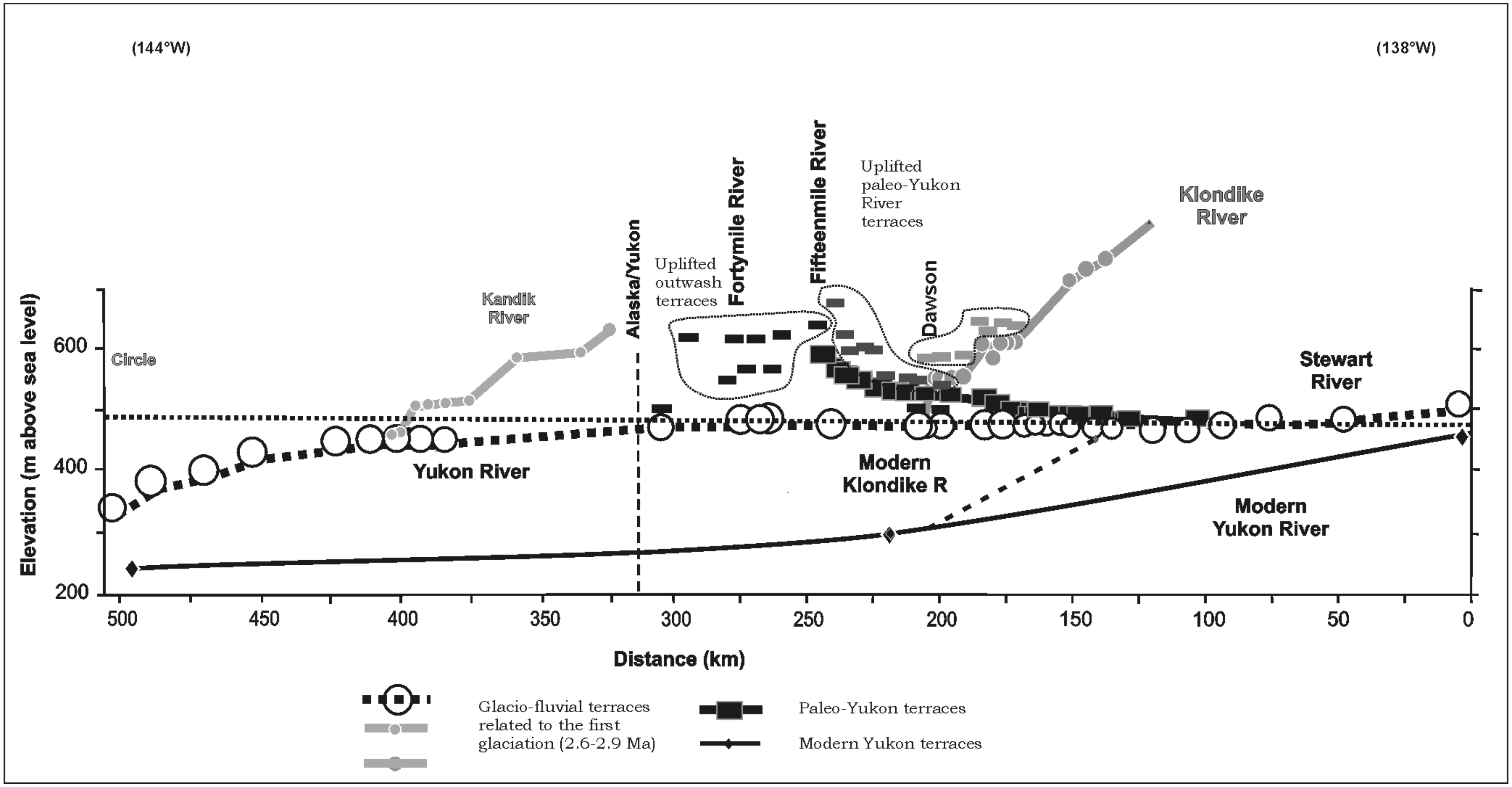

Figure 28. Terrace profiles along Yukon River (from Duk-Rodkin et al., 2001).

Recently, placer-mining exposures on the Lousetown Bench immediately south of Dawson City at the Klondike-Yukon river confluence (Fig.23) have exposed early-mid Pliocene gravel with flow directions to the south (Froese et al., 2001). Gravel lithologies include cherts derived from the Ogilvie Mountains to the north and clasts from the Klondike goldfields, indicating that the Lousetown bench was the confluence of the pre-glacial Klondike and a southerly-flowing paleo-Yukon River. In addition to the pebble fabric data, placer gold at the site extends to the southern limit of the Lousetown bench and is similar in morphology and fineness to that derived from lower Bonanza Creek. This surface marks the location of a southerly-flowing paleo-Klondike River across the Lousetown bench in the early-mid Pliocene. Subsequently, Lionel Jackson of the Geological Survey of Canada has shown that southerly-flowing pebble-fabric at the confluence of the Sixtymile and Yukon rivers south of Dawson.

Regional glacial history indicates that the most extensive glaciation of the Yukon was also likely the earliest in the late Pliocene (Froese et al., 2000; Duk-Rodkin et al., 2001). In the Klondike area, this early glaciation is marked by the Klondike gravel (Fig. 24, 25). The formation of this early Cordilleran Ice Sheet would have blocked the southerly-flowing Yukon River, forming a large proglacial lake between the ice sheet and paleodivide, named glacial Lake Dawson .

Pebble fabric from the Klondike gravel indicates that the Yukon River was westerly-flowing by the time of its deposition, suggesting the Yukon River had been established in its present direction by at least this time. Evidence for the diversion of the Yukon River into central Alaska is found in the Yukon Flats Sedimentary Basin (Fig. 26). In the Yukon Flats, a large lake existed through the Miocene and early Pliocene, and records an abrupt transition from silt and clay deposition to gravel after 3 Ma, consistent with the timing of the pre-glacial to glacial transition in central Yukon Territory (Froese et al., 2001).

Modern Yukon River

The Yukon River is the fifth largest in North America, twentieth globally, and has an average discharge > 6000 m^3/s at its confluence with the Bering Sea (Brabets et al. 2000). At Dawson, the Yukon drains an area of about 250 000 km^2 with an average discharge of ~3000 m^3/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 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 (Brabetts 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 Paleo-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 5: Last Chance Creek: Klondike Muck cryostratigraphy (63º 59' N., 139º 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 remains 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 subdivisions 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 29).

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

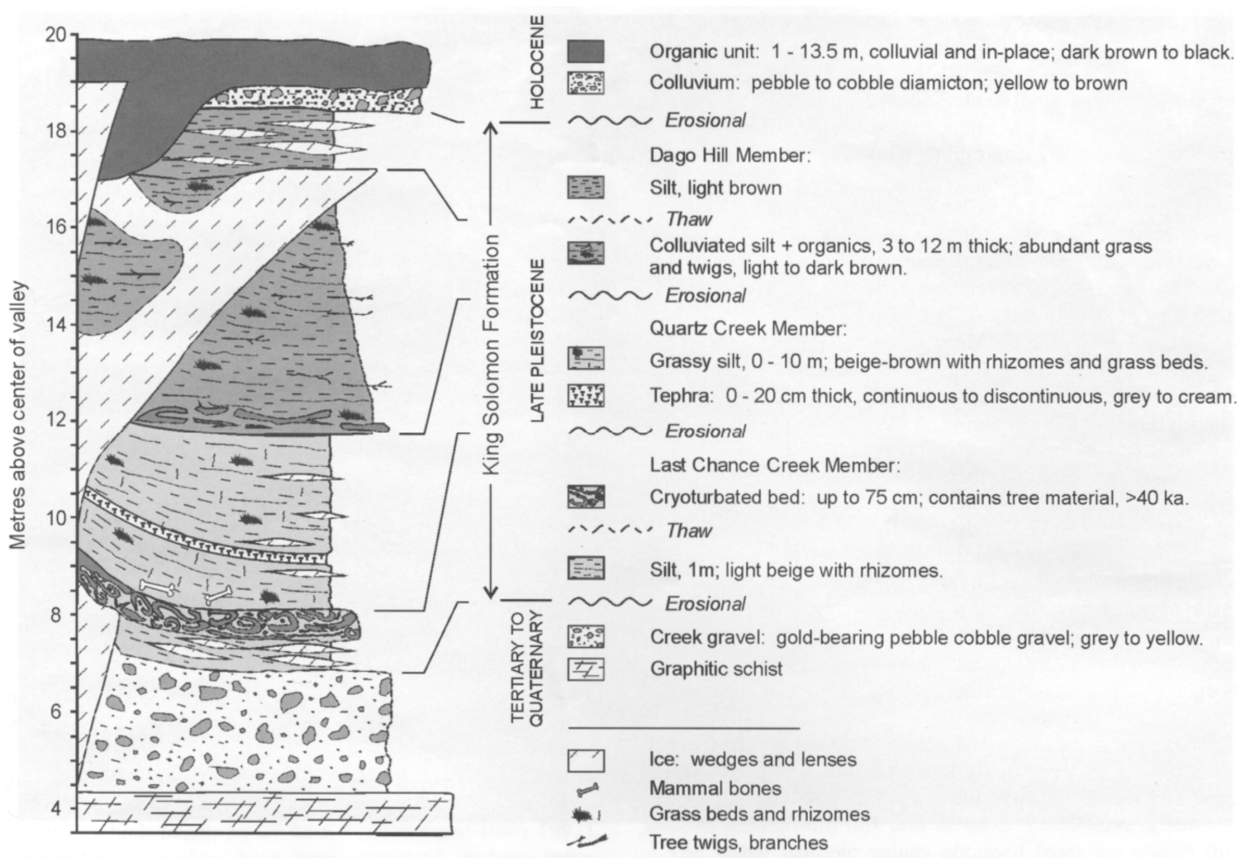

Figure 29. Generalized cryostratigraphy of frozen, unconsolidated sediments in the Klondike area, YT. This figure is a representative section compiled from the 21 sections examined by E. Kotler in 1997 and 1998. It combines the sedimentological and geocryological characteristics of the deposits and summarizes the stratigraphic position of the various cryostratigraphic units described in the text.

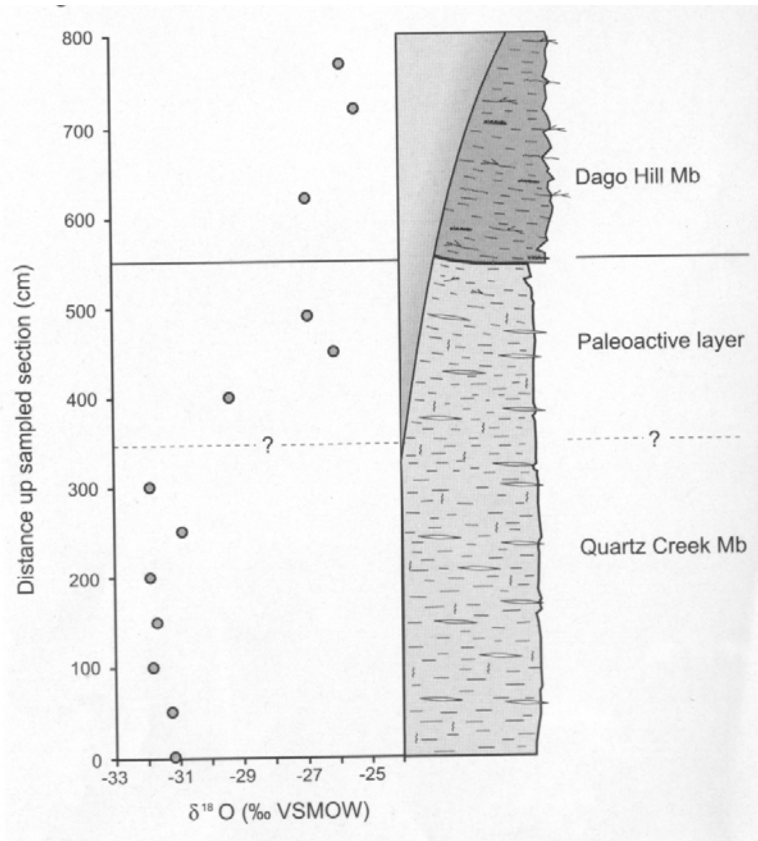

Figure 30. ^{18}O profile across the contact between Quartz Creek Member and Dago Hill Member at Quartz Creek. The data were obtained from water which was released upon thawing of the sediment samples. (from Kotler and Burn 2000).

Quartz Creek Member

This member comprises the McConnell (late Wisconsin) 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; Froese et al., 2002). 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

A 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 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 30). 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.

Middle Pleistocene Muck

Initial studies of Klondike muck deposits by Fraser and Burn (1997) concluded that all of the ice-rich silt deposits dated to the last glaciation, and that earlier units were likely eroded by subsequent fluvial activity. Kotler and Burn (2000) showed that some muck deposits dated to the limit of the radicarbon method. Recently, Froese and Westgate (unpub. data) have shown that relict ice wedges and ground ice are overlain by tephra dating to the middle Pleistocene. This is in contrast to work in central Alaska that suggests permafrost completely melted out during the last interglacial (Pewe et al., 1997).

Paleoecology of Late Pleistocene "muck" deposits in the Klondike area

G.D. Zazula and D.G. Froese

Placer mine exposures in the Klondike goldfields of west-central Yukon Territory are among Canada's most prolific sources of Pleistocene vertebrates (Harington and Clulow, 1973; Harington 1977, 1980, 1989, 2002). Most fossils are recovered near the base of frozen "muck" deposits, the unconsolidated, organic-rich silt found in valley bottoms, overlying gold bearing gravel (Fraser and Burn, 1997). "Muck" deposits similar to those from the Klondike are also described for interior Alaska (Guthrie, 1990; Thorson and Guthrie 1992) and have been the focus of paleontology and stratigraphy. These eastern Beringian deposits may be analogous to the northeast Siberian Duvanny Yar-aged *yedoma* silt that has been studied both paleoecologically and paleontologically (Gitterman et al. 1982; Tomirdiaro, 1982; Sher 1997). Other than the paleoecological implications made from the presence of vertebrate fossils, little systematic paleoecological investigation has been conducted at Klondike "muck" exposures. These Pleistocene fossils are dominated by large mammal grazers, including mammoth (*Mammuthus primigenius*), horse (*Equus lambei*), and bison (*Bison priscus*) suggesting the predominance of herbaceous grassland or steppe vegetation (Harington 1980; Guthrie 1990).

The first paleoecological study in the Klondike was conducted on Dominion and Hunker Creeks (McAtee 1979). McAtee described 5 stratigraphic sections and studied pollen from the silt with interbedded peat and gravel. These sediments were found to yield abundant pollen and spores when a heavy liquid separation technique was used. The use of this method enabled large samples to be processed to obtain sufficient pollen and spores for statistical analyses. However, McAtee noted that many of the samples were not polleniferous. Pollen and spore data suggest a Pleistocene warm interval dominated by boreal forest, followed by a cold tundra period, and then a return to warmer forested conditions (McAtee 1979). Radiocarbon ages and pollen were interpreted by McAtee as representative of the last glacial cycle, spanning the Middle Wisconsinan interstadial (isotope stage 3) to the Holocene. Middle Wisconsinan assemblages indicate a closed spruce forest with possibly some shrub birch and junipers. Full-glacial aged spectra are typical for eastern Beringia, dominated by grasses, sedges, willows, composites, chenopods, *Artemisia* and other herbs, indicative of steppe-tundra with some scattered trees. Abrupt increased frequencies of spruce, ericads and some pine (*Pinus*) mark the reappearance of Holocene closed boreal forest, perhaps with climate warmer and drier than that of today. Although McAtee notes the presence of abundant insects, wood and other plant macrofossils in the "muck" peat beds, these were not analyzed.

The presence of pine (*Pinus*) pollen in McAtee's diagrams is problematic in light of recent palynological investigations that suggest pine was absent from Yukon during pre-Holocene interglacials and interstadials and did not become re-established until ca. 1500-2000 years ago (Schweger and Froese 2001). McAtee's pine pollen may thus be the result of reworking from Tertiary sources because there are also high quantities of pre-Pleistocene spores and indeterminable conifer pollen. Further, the examination of McAtee's pollen diagrams indicates significant radiocarbon age reversals that make their chronologies problematic and correlation with recent work from the Klondike uncertain. McAtee describes at least two different tephras in his stratigraphic sections, yet the age or identification was not determined.

Recent work has described the Klondike "muck" stratigraphy in greater detail, suggesting a consistent chronostratigraphy at numerous localities (Fraser and Burn 1997; Kotler and Burn 2000). The 24,000 year old Dawson tephra provides a regional stratigraphic marker for studying the paleoecology of the last glaciation and correlating data from various sites (Froese et al., 2002). At Quartz Creek, Dawson tephra is exposed immediately above the muck-gravel interface. A second, younger tephra is exposed near the top of the section. Peat within gravel underlying Dawson tephra contains spruce (*Picea*) cones, probably representing open-spruce forests during the latter part of the stage 3 interstadial. Massive and bedded loess overlies Dawson tephra, being deposited during cold-arid conditions of the full-glacial after ca. 24,000 yr BP. The upper tephra is associated with a thaw unconformity and large epigenetic and syngenetic ice wedges that signal the transition to present interglacial conditions. Soil associated with this tephra contains spruce needles, suggesting it was deposited during a period of transition from open herbaceous vegetation to closed boreal forest with shallow active layers.

Pleistocene Rodent Burrows

Frozen nests within microtine rodent burrows can be observed melting out of "muck" sediments. Radiocarbon ages from nests in west-central Yukon span the last glacial interval (isotope stages 3 and 2) (Harington 1989; 1997; Pirozynski et al. 1984). The burrows contain rodent droppings, frozen plant material that was used to construct nests and seeds caches used for winter food storage (Krog, 1953). At Quartz Creek, a nest is inset approximately 10 cm into the surface of Dawson tephra and yielded ages of 24,280 ±130 yr BP (Beta-161239) and 23,990 ±130 yr BP (Beta-161238) (Froese et al., 2002). This nest contained abundant seeds and florets of Poaceae, mostly *Poa* type, grass leaf blades and leaves of *Artemisia frigida*. The seed cache is dominated by sedge (*Carex*) seeds, probably totalling 95% of the entire assemblage. Diverse herbs are represented by seeds of rush (*Juncus/Luzula*), mustard (*Draba*), cinquefoil (*Potentilla)*, buttercup (*Ranunculus)*, sedge (*Carex)*, pepperwort (*Lepidium)*, chickweed (*Cerastium)*, campion (*Silene* cf. *involucrata)*, dandelion (*Taraxacum)*, and fairy candelabra (*Androsace septentrionalis)*. Pollen analysis of a frozen arctic ground squirrel (*Spermophilus parryi*) nest and droppings was conducted by McAtee (1979) at Hunker Creek. The pollen and spore assemblage is dominated by grasses (Poaceae, 73.5%) and sage (*Artemisia*, 14.6%) with smaller quantities of goosefoot (Chenopodiaceae), willow (*Salix)*, asters (Asteraceae) birch (*Betula)* and mosses. Together, these data suggest that late Pleistocene microtine rodents were foraging within a xeric grassland or steppe environment. Several nests and seed caches have been sampled by the authors from various sites in the Klondike to develop a unique paleoecological record in eastern Beringia.

Rodent burrows from the "mucks" have some important paleoecological implications. Firstly, permafrost active layer depths during the last glaciation were deeper than at present to enable small mammals to burrow into silty, well-drained soil on north-facing slopes (see Guthrie, 1990; Schweger, 1997; Plug, 1992). Secondly, these small mammals foraged and gathered plant remains from a limited radius around the burrows, thus paleobotanical remains from them reflect local vegetation. Arctic ground squirrels will eat almost any vegetation growing in its local habitat, thus the plant remains within burrows are not biased for dietary preferences and likely reflect the most readily available plants to the animal (Mayer 1953). Like packrats in the American southwest, Yukon ground squirrels were little botanists who collected plants that one day could be analyzed by paleoecologists. In 1967, seeds of the arctic lupine (*Lupinus arcticus*) recovered from a frozen late Pleistocene collared lemming (*Dicrostonyx torquatus*) nest on Dominion Creek were germinated in a laboratory, exemplifying the exceptional preservation of some paleobotanical remains (Porsild et al., 1967).

New chronostratigraphic information enables systematic paleoecological research now being conducted by the authors. Investigations are focussed at sites with Dawson tephra to obtain samples of "muck" silt, peat, paleosols, rodent burrows and gravel, that span the interval between ca. 24,000 –12,000 yr BP. The "mucks" contain a detailed, multi-proxy record of local environments. Pollen from the retransported silt in the "mucks" is likely local in origin because primary loess has low overall pollen concentrations (Edwards 1997). The "muck" pollen spectra probably reflect vegetation that inhabited local hillslopes that became entrained within silts by slopewash during spring runoff or spring/summer rainfall and deposited into the valley bottoms. Active loess deposition and transport downslope implies an open herbaceous pattern of vegetation cover with exposed mineral soils. However, the retransported, colluvial nature of these "muck" deposits may hinder the interpretation of these pollen records and the uncertainty introduced by sediment and pollen reworking must be acknowledged (Edwards 1997). These problems were demonstrated in McAtee's pollen study. Thus, it is crucial to work on sites that have consistent stratigraphy and reliable chronology and conduct sampling with caution. Analyses of plant macrofossils and silica-phytoliths can refine the interpretations based on the pollen data, confirming the local presence of particular taxa and taxonomically expanding the interpretation of past vegetation. The presence of in situ vegetated surfaces and soils have also been briefly described and dated (Fraser and Burn 1997; Kotler and Burn 2000). Such sites have preserved the remains of rooted vegetation that inhabited the fluves and interfluves during the late Pleistocene. Together, pollen, plant and insect macrofossils, phytoliths, paleosols and mammalian fossil remains have the potential to yield detailed and unique paleoecological data. The "mucks" may provide crucial paleoecological data for interpreting the nature of the full-glacial refugia and presenting new data in the long-standing debate on Beringian environments.

Dawson Tephra: A large magnitude eruption during the late Pleistocene of Beringia and its stratigraphic significance

D.G. Froese, J.A. Westgate, S.J. Preece and J. Storer

Dawson tephra is the most prominent tephra in the Pleistocene deposits of the Klondike area, and likely the largest Quaternary volcanic eruption in Yukon or Alaska (Froese et al. 2002). Dawson tephra consists mainly of thin, bubble-wall glass shards of rhyolitic composition (Westgate et al., 2000). The crystal component consists of orthopyroxene, plagioclase, magnetite and ilmenite with minor amounts of clinopyroxene, apatite and zircon. Its chemistry and mineral content are similar to the Old Crow tephra, a Type I bed in the terminology of Preece et al. (1999), indicating a source area in the Aleutian arc-Alaska Peninsula region of southwestern Alaska. This composition is distinct from the Type II beds derived from Wrangell-sourced eruptions, the latter having highly inflated pumice and abundant hornblende such as the Sheep Creek and White River tephra (Preece, et al. 1999; 2000).

Dawson tephra has been identified at twenty sites in the Klondike and Sixtymile placer districts of western Yukon (Fig.31). It occurs near the base of the 'muck' deposits, marking the transition from to active loess deposition in the Klondike near the start of stage 2 (Fraser and Burn, 1997). In the Klondike area, most muck deposits are of stage 2 age dating from 28 000 to 12 000 ^{14}C yr BP (Fraser and Burn, 1997; Kotler and Burn, 2000). Dawson tephra is typically 15 to 30 cm thick, making it the most prominent tephra bed in the region (Fig. 31, 32).

Age

A single fission-track age of less than 52 000 years old was obtained for Dawson tephra by Naeser et al. (1982). An accurate age for the Dawson tephra eluded Westgate et al. (2000) due to a lack of concordance between bracketing radiocarbon ages associated with the tephra, and they concluded its age was likely younger than 22 000 and older than 11 600 ^{14}C yr BP. A more precise age is provided by a frozen rodent burrow inset approximately 10 cm into the surface of the Dawson tephra (Fig. 32), and in turn overlain by 12 m of massive and crudely bedded frozen silt (muck). The burrow was chosen for dating because it was constructed from contemporaneous plant material, eliminating potential reworking of 'old' organic material above the tephra bed. In addition, individual components of the burrow (seed capsules and grass stems) could be dated separately to verify the age determination.

One sample, obtained from *Draba* sp. seed capsule casings have an age of 24 280 ± 130 ^{14}C yr BP (Beta 161239). Grass stems from the burrow yielded an age of 23 990 ± 130 ^{14}C yr BP (Beta 161238). These samples overlap at two standard deviations giving an estimated age for the burrow of 24 020 - 24 250 ^{14}C yr BP, or approximately 27,000 cal yr BP using the calibration curve of Kitagawa and van der Plicht (1998). We think this closely approximates the actual tephra age because of the proximity of the burrow to the tephra (Fig.3) and evidence for rapid burial in the early period of muck deposition (Fraser and Burn, 1997).

A maximum age for the tephra is provided by concordant ages from two sites. On Quartz Creek a peat 30 cm below the tephra returned an age of 25 240 ± 140 ^{14}C yr BP (Beta 133410), which is consistent with the minimum ages. This age is further supported by a conventional radiocarbon age from an *in-situ* grass bed immediately below the tephra on Quartz Creek of 24 025 ± 550 ^{14}C yr BP (BGS 1775, Table 1; Fraser and Burn, 1997). A third age on Last Chance Creek from a peat bed within fluvial gravel 4 m below the tephra of 25 700 ± 400 yrs BP (BETA-171748) is consistent with the Quartz Creek ages (Zazula et al., 2003).

Probable distribution and significance

The chemistry and petrography of Dawson tephra indicate a Type I bed sourced in the Aleutian arc- Alaska Peninsula of southwestern Alaska (Preece et al., 2000; Westgate et al., 2000). Glass shards are dominated by fractured bubble walls with minor pumice fragments similar to the Old Crow tephra. Given the nearest vent of the Aleutian arc-Alaska Peninsula is 700 km southwest of the Klondike region, the eruption must have been a very large one. Recent work by the US Geological Survey at Emmons Lake on the Alaska Peninsula, 1500 km southwest of Dawson, suggests that one of the two main caldera-forming events, comparable in size to the Old Crow event (>50 km^3), is the likely source for the Dawson tephra (Waythomas et al., 2001). An Ar-Ar age at Emmons Lake on the second caldera event of 16,000 ± 10,000 yr BP indicates the complex was active during the time that Dawson tephra was deposited (Drake et al., 2001).

The probable minimum distribution of Dawson tephra shown on Figure 1 is based on the source area of Waythomas et al. (2001). We assume that west-central Yukon lies along the axis of maximum thickness of the lobe, which is a conservative estimate. We have extended the plume 150 km to the northwest and southeast to derive the distribution in Figure 1. We chose this value because the tephra bed has not been found in the Fairbanks area. However, it should be noted that much of the tephrochronological work in this region (Naeser et al., 1982; Westgate et al., 1990; Preece et al., 1999; Begét, 2001) has focussed on the pre-late Wisconsinan record, and Dawson tephra may well be present in this area. And further, its distribution could easily extend much further north than that shown in Figure 31.

Dawson tephra is a valuable stratigraphic marker. It is presently being used to correlate late Pleistocene faunal remains throughout the Klondike area (Storer, 2001), and marks the beginning of 'muck' deposition in the region (Fraser and Burn, 1997; Kotler and Burn, 2000). Furthermore, it is likely that Dawson tephra is locally preserved in sediments in the Gulf of Alaska, offering the possibility of an effective linkage between marine and continental sedimentary sequences of eastern Beringia during the late Pleistocene.

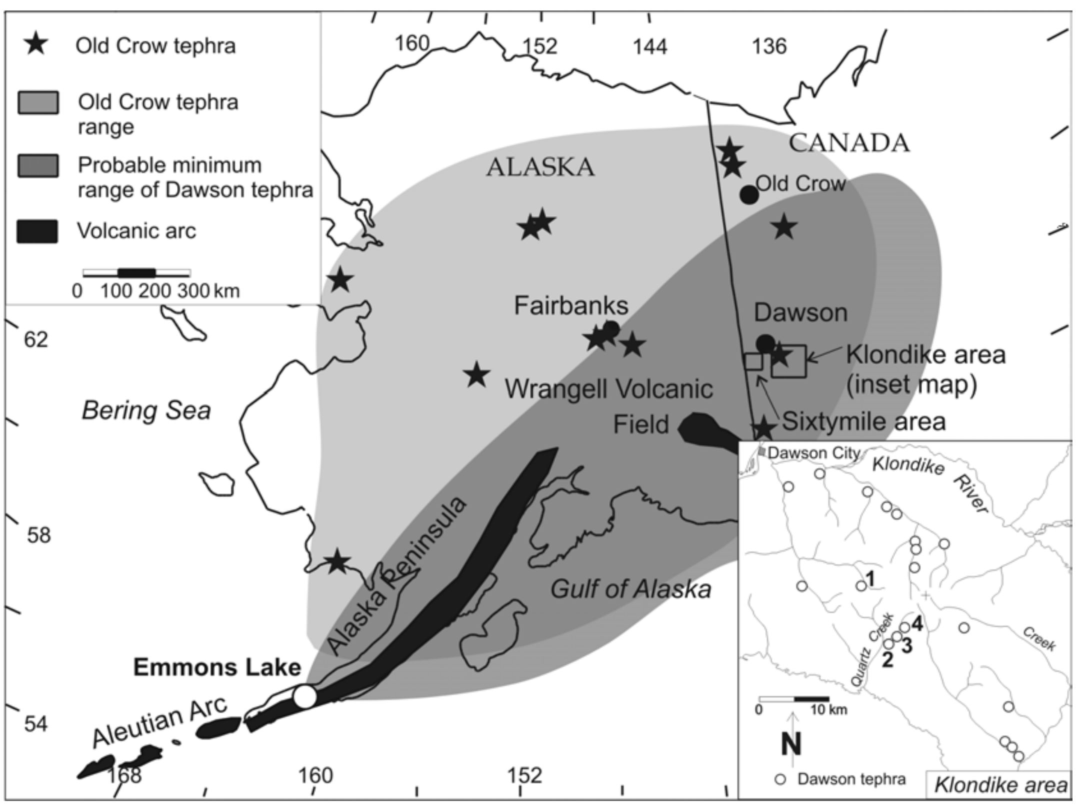

Figure 31. Probable distribution of Dawson tephra (dark pattern), based on its source area in southwestern Alaska in comparison to the widespread Old Crow tephra (stars and grey pattern).

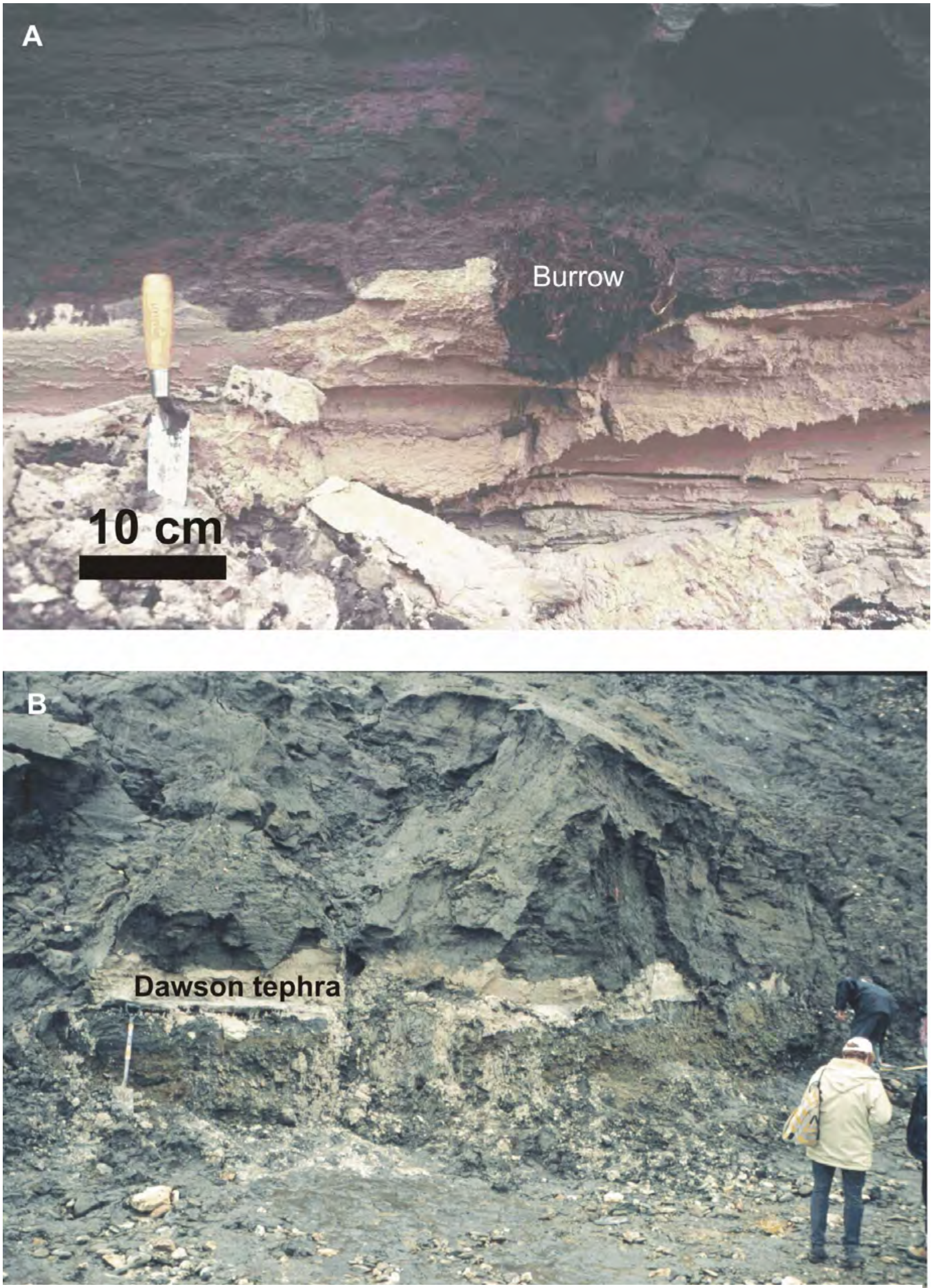

Figure 32. Dawson tephra with dated rodent burrow on Quartz Creek (A), and Dawson tephra at base of mucks on Quartz Creek (B).

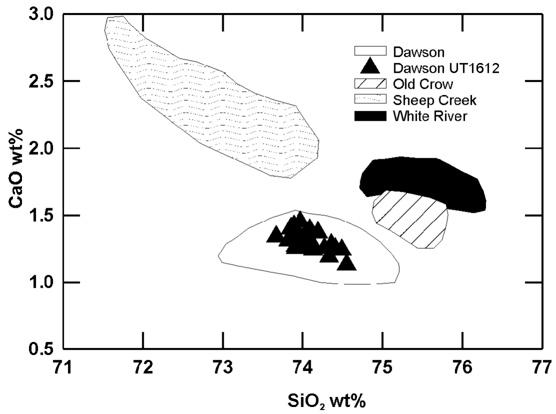

Figure 33. Oxide variation diagram showing differences in glass composition between Dawson tephra and other well-known Quaternary tephra beds in Yukon. UT1612, the tephra sample with the ^{14}C control at Quartz Creek, is highlighted in the field of all other glass analyses of Dawson tephra.

Stop 6: Quartz Creek: Pliocene ice wedges/ Quartz Creek tephra (63º 46' N. ,139º 03' W.)

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The Quartz Creek site is located 35 km southeast of Dawson City near the confluence of Little Blanche and Quartz creeks (Figures 12, 23). 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 34, 35. 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 $^{40}Ar/^{39}Ar$ 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 paleomagnetic measurements (Froese et al. 2000) calibrated with fission-track ages that place its range between 2.6 - 3.3 Ma (Figure 34).

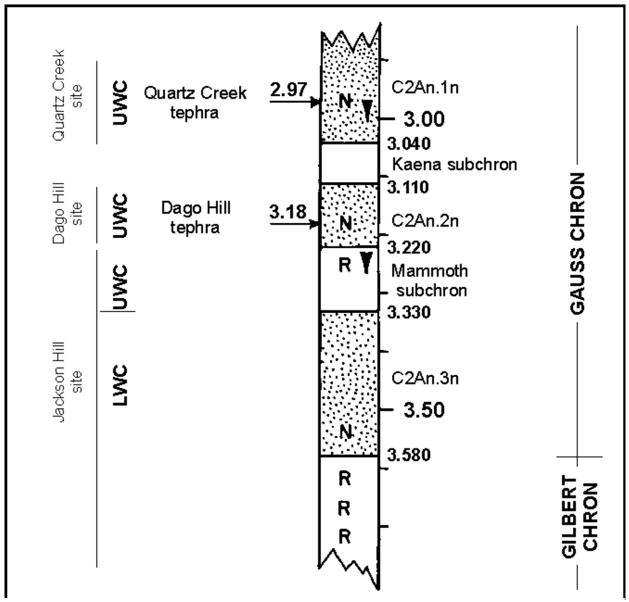

Figure 34. Proposed correlation of the Upper and Lower White Channel gravels with the geomagnetic polarity timescale of Cande and Kent (1995). Ages in Ma. (from Westgate et al., 2003)

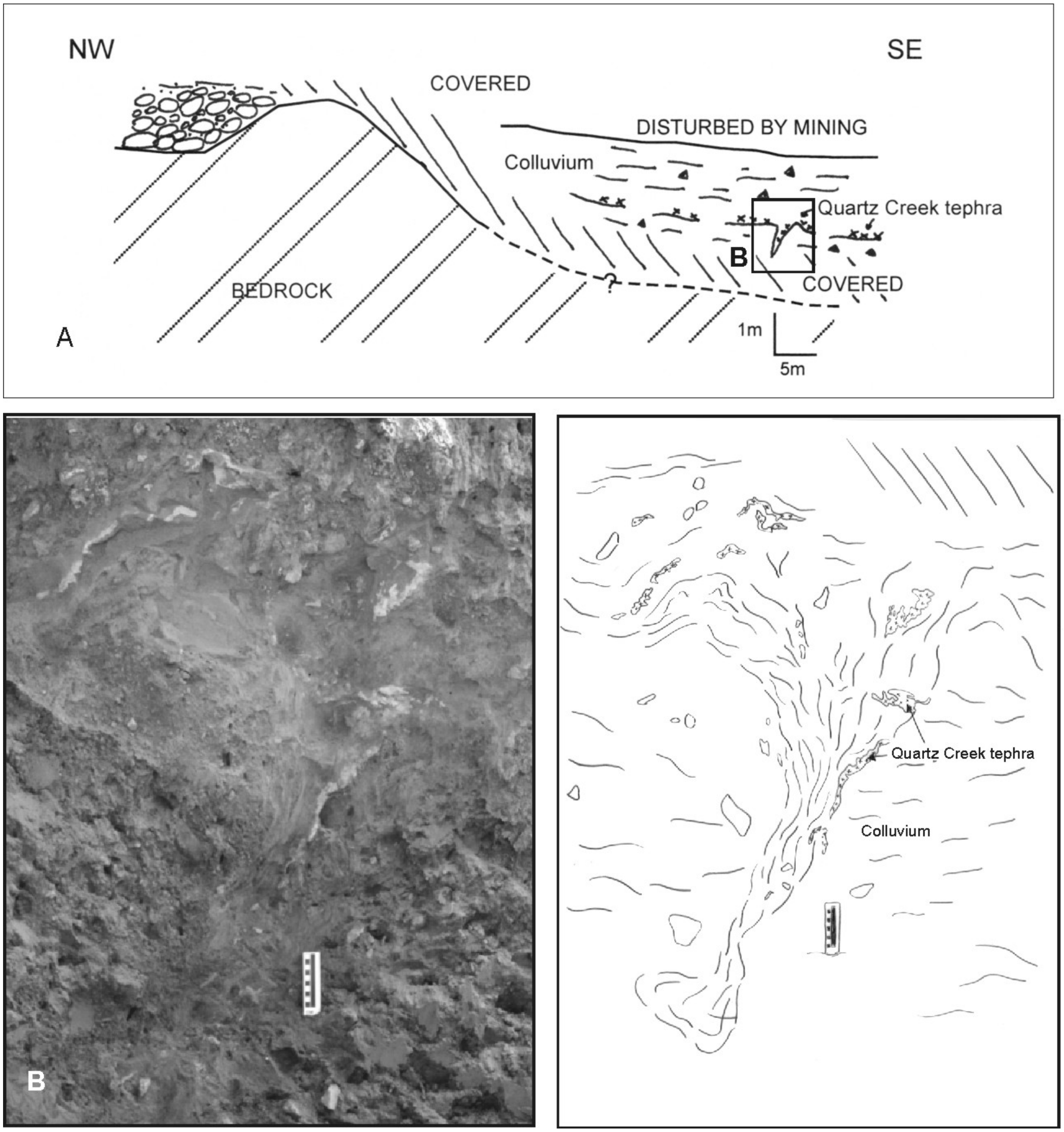

Figure 35. Generalized stratigraphic relation between Quartz Creek tephra, colluvial sediments, and White Channel gravel at site along Quartz Creek (Figure 27). B. Photograph and sketch of Quartz Creek tephra. The tephra has a diameter-corrected fission-track (DCFT) age of 3.00±0.33 Ma and an isothermal 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.

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