GEOLOGICAL SURVEY OF CANADA
OPEN FILE 5487

Geology of Quiet Lake and Finlayson Lake map areas, south-central Yukon – An early interpretation of bedrock stratigraphy and structure

D.J. Tempelman-Kluit

2012

Natural Resources Canada
Ressources naturelles Canada

Canada

GEOLOGICAL SURVEY OF CANADA
OPEN FILE 5487

Geology of Quiet Lake and Finlayson Lake map areas, south-central Yukon – An early interpretation of bedrock stratigraphy and structure

D.J. Tempelman-Kluit

2012

doi:10.4095/291931

This publication is available for free download through GEOSCAN (http://geoscan.ess.nrcan.gc.ca/).

Recommended citation

Tempelman-Kluit, D.J., 2012. Geology of Quiet Lake and Finlayson Lake map areas, south-central Yukon – An early interpretation of bedrock stratigraphy and structure; Geological Survey of Canada, Open File 5487, doi:10.4095/291931

Publications in this series have not been edited; they are released as submitted by the author.

Preparation of this volume jointly funded by the Yukon Geological Survey and the Geological Survey of Canada

PREFACE

Quiet Lake and Finlayson Lake map areas include the heart of the Pelly Mountains. They provide excellent bedrock exposure of the north-central Canadian part of the orogen called the North American Cordillera. They comprise late Proterozoic to early Mesozoic platform and outer shelf strata of ancient North America that were shortened during Mesozoic folding and thrust faulting. These strata are structurally overlain by thrust slices of metamorphosed Paleozoic sedimentary, volcanic and gneissic rocks that originated far from this continent. Both the in-place and overthrust rocks were intruded by plutons of late Mesozoic age. The map areas are bisected diagonally by the Tintina Trench which is underlain by a dextral strike-slip fault with 450 km of post-Cretaceous displacement.

This Open File report presents a general description of the regional stratigraphy of the two map areas and an interpretation of structure, metamorphism and igneous history as they were understood in the early 1980s. It builds upon field work dating back to Operation Pelly, a large and wide-ranging reconnaissance survey in central Yukon begun by the Geological Survey of Canada in 1958. The field observations upon which the stratigraphic units are based were principally made between 1973 and 1976. The report describes principal structures and develops a comprehensive structural interpretation with deductions of the relative timing of faults. The mineral deposits are placed in a regional tectonic framework. It is a classic report that extends from field observations to propose correlations of stratigraphy, structural models and imaginative geological hypotheses which can be tested by further geological work. These include the concept of discrete packages of platform stratigraphy overlapped on low-angle thrust faults, and the recognition of internally deformed sheets of allochthonous sedimentary, volcanic and plutonic rock overlying the platform strata. The latter idea resulted in a widely read and acclaimed paper (Tempelman-Kluit, 1979)

This report was written in the early 1980s in the traditional 'memoir' form to accompany 1:250 000 scale maps. At the time of writing the techniques for isotopic dating limited the interpretation of metamorphic rock units and for them this report is no longer relevant. Nevertheless some aspects of the regional tectonic model expounded here have served for 25 years, although many details have contemporary explanation. Following discovery of a copper-zinc sulphide deposits south of Finlayson Lake (1992), a boom in mineral exploration and revision mapping improved the stratigraphic framework northeast of Tintina fault (see Murphy et al., 2006 and references therein). Southwest of Tintina, the regional mapping and descriptions of stratigraphic units (primarily Pelly-Cassiar platform and slope facies) herein remain current. The latter are the prime motive for publishing this unfinished work – to make these descriptions available to subsequent workers. Most of Quiet Lake map area has not seen systematic bedrock mapping during the last 30 years.

Steve Irwin
Subdivision head
Geological Survey of Canada, Vancouver

GEOLOGY OF QUIET LAKE AND FINLAYSON LAKE (105F, 105G) MAP AREAS, SOUTH-CENTRAL YUKON

Abstract

The Pelly Mountains in central Yukon expose a cross-section through part of the collision orogen that is the Canadian Cordillera. Elements of this cross-section are the pre-tectonic autochthonous succession, klippen of sheared and metamorphosed allochthonous rocks, thrusts that repeat the autochthonous succession, late tectonic regional metamorphic rocks and associated granite batholiths, and the Tintina fault, a late tectonic dextral strike-slip fault.

Late Proterozoic through Triassic shelf strata, six to eight kilometres thick, comprise the autochthonous sequence. They are divided into four main divisions bounded by unconformities or a depositional hiatus. Oldest are Eocambrian and Lower Cambrian shale and siltstone, capped by an argillaceous limestone, which together form the Ketza group. Unconformably above it are Upper Cambrian and Ordovician slate, phyllite and alkaline basalt of the Kechika Group, and Silurian and Devonian siltstone, dolostone and quartz sandstone, of the Askin group. Profound and rapid facies variation characterizes this second division. Also part of this division is the shale-dominated Harvey group, which is time-equivalent to the Kechika and Askin groups. Unconformably above are Upper Devonian and Mississippian black slate and marine felsic volcanic rocks - the Seagull group. The fourth division is thin and restricted in distribution. It includes a Permian and an Upper Triassic siltstone separated by a disconformity.

The autochthonous succession extends to, and is like that in, the Cassiar Mountains; it has equivalent units in the northern Rocky Mountains. Broadly it is equivalent to the outer shelf of the North American miogeoclinal wedge. Three assemblages of ductile deformed rocks are thrust over the "in place" succession from the southwest. Although each assemblage is dismembered so that its internal stratigraphy is unknown, the three are not mixed, and exhibit a consistent structural order. Lowest is a mylonite schist, that represents Triassic (and older?) immature clastic rocks and intermediate volcanics. Next is an assemblage of Late Paleozoic amphibolite and serpentinite with flaser limestone. Highest are granodiorite gneiss and schist, the sheared and metamorphosed remnant of Paleozoic intrusive rocks. K-Ar ages show that all three assemblages were sheared and metamorphosed during the Late Triassic and Early Jurassic.

Four northeast directed thrusts repeat the "in place" beds beneath, and northeast of, the transported strata. Cumulative shortening is on the order of 100 kilometres. Each thrust carries a slice two or three kilometres thick and is itself deformed by newer thrusts, folds and tear faults. The thrusts, interpreted as splays of one basal detachment, formed in sequence from southwest to northeast during the Late Jurassic and Early Cretaceous.

On the southwest the structurally imbricated "in place" strata and the transported rocks above them, were regionally metamorphosed about the late Early Cretaceous to form the Big Salmon and Mink complexes. At the same time partial melting produced granites like the Quiet Lake, Nisutlin and Big Salmon batholiths. The metamorphic rocks are exposed in large structural culminations centred roughly on the batholiths; these fold early thrusts, giving a structural relief near eight kilometres. Some granites may be cut by thrusts below and the metamorphic culminations may be severed by the basal detachment.

Tintina fault is the locus of 450 kilometres of dextral slip as shown by offset of the autochthonous succession. Movement followed intrusion of most or all of the late tectonic granites. Movement may have been ceased by Early Eocene when fluvial conglomerate was laid down along the fault. Because its slip coincides with shortening in the Selwyn and Mackenzie mountains the Tintina is considered to be a listric branch of the same detachment surface on which shortening was accomplished. It is viewed as a giant tear fault in the detached slab.

A model in which the Pelly Mountains are the product of Middle Jurassic oblique collision of a volcanic arc with the then-western margin of ancient North America fits these data. Sheared metamorphic rocks, which are interpreted as the arc's subduction complex, were obducted in the Late Jurassic. Cretaceous imbrication and dextral translation of the autochthonous rocks are responses to obduction, while metamorphism and plutonism reflect heating of the overridden and imbricated "in place" slab.

FORWARD

This final report from fieldwork for the Quiet Lake-Finlayson reconnaissance mapping project has been changed from its original intended 'memoir' format to an 'open file'. The open-file series allows publication of this nearly finished manuscript with the author's original interpretations. A casualty of the change is that stratigraphic unit descriptions which were intended to formalize 12 new formations are here presented as informal formations (because an open file is not recognized as a formal publication by the North American Stratigraphic Code).

Despite a long delay in publication this final report contributes new information to the understanding of Yukon geology. The five maps at 1:50 000 scale and stratigraphic descriptions of units comprising the Pelly-Cassiar platform in Yukon were not previously available. The description of structures from field observations supported by annotated photographs of mountainside exposures result from classic reconnaissance mapping.

During resurrection of the manuscript, most original figures were located. Hand-drawn and mechanically lettered, these originals would have been reduced to photographic film, and degrees of shading introduced using interleaved screens of different densities. Equipment for this photo-mechanical transfer process is no longer available. Therefore the multi-sheeted originals were digitally scanned and color was substituted for black and white patterns. Photographs were digitally scanned from black-and-white prints that were annotated by the author. No corrections to photographs were possible and some abbreviations on them are therefore inconsistent with those on the up-dated maps.

The author published the two 1:250 000 scale maps in 1977 (GSC Open File 486). These and the five maps at 1:50 000 scale that the author prepared later have been re-drawn on a geo-referenced base and a common legend attempted. Some discrepancies between units shown at the different map-scales remain. As far as possible colors assigned to geological units are consistent between the various maps.

At the time of writing the author had a unifying vision for the geology of central Yukon. He considered strata of the Pelly-Cassiar platform to have time- and lithologic-equivalent units on the northeast side of Tintina fault. This time- and lithologic framework is no longer valid; these rock successions now have different time-boundaries and provenance. In the legend for maps accompanying this report, several allochthonous assemblages are divided and units of McEvoy platform (northeastern Finlayson Lake map area) are shown in a separate column. Regional correlation charts (Figs. 14, 31 and 36) are presented "as-is" and include obsolete interpretations.

The author's text has been edited to improve its clarity, consistency and grammar; footnotes from the critical reader and editor point the reader toward subsequent geological research or interpretations (citations in a second, post-1985 list of references).

The Geological Survey of Canada has a long history of reconnaissance mapping in the Yukon. It is a pleasure to present the final report of the Pelly Mountians project.

Charlie Roots
Geological Survey of Canada
c/o Yukon Geological Survey
Box 2703, K-102, Whitehorse, Yukon Y1A 2C6

Critical reviewer:
S.P. Gordey, 1985, 2007
Geological Survey of Canada, Vancouver, B.C. V6B 5J3

Original manuscript submitted: 1985-08
Final version approved for publication: 2012-12

CONTENTS

FIGURES *(linked to text)*

TABLES *(linked to text)*

LARGE FIGURES *(separate files; require plotter)*

INTRODUCTION

This description and interpretation of the geology of the Pelly Mountains in the Yukon Territory specifically concerns Quiet Lake and Finlayson Lake map areas between latitudes 61° and 62°N and longitudes 130° and 134°W, encompassing an area of 16,500 square kilometres (Fig. 1). The report is based on field work by Geological Survey of Canada field parties led by the author during the summers of 1973, 1975 and 1976 and part of the summer of 1974.

Access and Logistics

The South Canol Road and Campbell Highway (Fig. 2) provide two excellent routes through the project area. The Canol Road traverses the Pelly Mountains, passing from north to south through Quiet Lake map area. It was built by the U.S. Army in 1944 to service the oil pipeline from Norman Wells on the Mackenzie River to the Alaska Highway (Kindle, 1946, Fradkin, 1977). It is open only in summer and travel on it is slow because it is narrow, twisted and not paved. From the Alaska Highway near Johnson's Crossing the road traverses northward through two passes (the first across the Big Salmon Range at about 4000 feet elevation, the second at 3800 feet across the St Cyr Range) on the way to Ross River, a distance of about 220 kilometres (135 miles).

The Campbell Highway was built between 1964 and 1967 to provide access from Watson Lake to Ross River and from Carmacks to Faro. The two sections were joined and this road now crosses the Canol Road near Ross River. The Campbell Highway roughly follows the northern margin of the Pelly Mountains providing an east-west route through the project area. The road is appropriately named for Robert Campbell, the Hudson Bay Company explorer who canoed up the Frances and Finlayson rivers in 1852 and who first mapped the Pelly River, recognizing it as part of the Upper Yukon River drainage system. Campbell Highway is an excellent road, which, though not paved, allows fast year-round travel.

Ross River is the only settlement in the project area (Figs. 1, 2). It began as a trading post at the mouth of the Ross River established about 1903 (Money and East, 1975; Coutts, 1980) and attracted aboriginal people from the Nahanni watershed to the area. Later the post was later taken over by the Whitehorse firm of Taylor and Drury. When the Canol Road arrived in 1944 the village was moved to the south side of the Pelly River and its population grew. Today about 220 people live there and the town's main livelihood, outside government support, comes from catering to the mineral exploration industry which uses the town as a logistics base. Ross River had flourishing helicopter and small plane charter businesses during much of the 1970's as well as a store, hotel, restaurant and bar, post office, community hall, nursing station, police detachment, and a highway maintenance camp.

Faro, at the time of the mapping, was the second largest town in the Territory. It lies just north of the project area. This instant town was almost completed when the entire settlement was burned by a forest fire that swept the north side of Pelly River valley in July 1969. The town was immediately rebuilt and in 1979 housed about 1550 people, all concerned with operating the Cyprus-Anvil mine and concentrator or with services to that operation[1].

Summer helicopter operation in the Pelly Mountains poses no particular hazards; good landing sites are plentiful and weather conditions rarely prohibit flying. Many lakes are large enough to allow landing and takeoff of float-equipped aircraft.

Streams in the project area feed three watersheds: the Pelly River on the north, the Liard River on the southeast and the Teslin River on the southwest (Fig. 2). Important tributaries of the Pelly are the Hoole, Ketza and Lapie rivers and the main tributaries of the Liard are the Black and Ings rivers. The Teslin River is fed by a host of streams; they include the McConnell, McNeil, Rose, Nisutlin, Big Salmon and Magundy rivers. The larger streams are navigable for great distances. The details supplied by McConnell (1898), Dawson (1887) and Johnston (1936) may be useful to river travellers.

Climate, Flora and Fauna

Several comprehensive reports provide general information on the climate, flora and fauna of Yukon (Kendrew and Kerr, 1955; Hare and Thomas, 1974; Oswald and Senyk, 1977[2]; Hultén, 1968). Porsild (1951) and Rand (1945) gave exhaustive accounts of the flora and fauna in the Pelly Mountains specifically. The following is summarized from those reports.

The climate in the project area is continental with mean daily temperatures for January and July about -25°C and 12°C, respectively. The mean annual snowfall is about 200 centimetres, contributing to total precipitation of about 300 millimetres, of which most falls in summer. Valley bottoms are well forested to elevations of 4500 feet. Native trees include lodge-pole pine, black spruce, white spruce, balsam fir, birch, popular, and aspen. Porsild (1951) distinguishes the plant associations of the valleys and uplands, characterizing those of the valleys as comparatively

[1] The population of Faro declined after the mine closed in 1993. In May 2006 the town had 341 inhabitants (Yukon Government Bureau of Statistics)
[2] Updated by Smith et al. (2004)

modern and those of the uplands as older and more diverse. The fauna includes black bear, grizzly bear, marten, mink, weasel, wolverine, fox, coyote, wolf, lynx, beaver, muskrat, porcupine, rabbit, moose, caribou, Dall sheep, wood chuck, marmot, and many shrews, mice and voles. The fish include grayling and lake trout and the variety of birds is innumerable[3].

Physiography

The physiography of the northern Cordillera was described by H.S. Bostock (1948a). The names used here and the summary descriptions follow his usage. The Pelly Mountains form an east trending highland and are essentially confined to central parts of Quiet Lake and Finlayson Lake map areas (Fig. 3). They include the Glenlyon, Little and Big Salmon, Campbell and Simpson ranges, but their backbone is the St Cyr Range. This range, underlain largely by deformed Paleozoic sedimentary rocks, is a rugged northwest trending chain of which Fox Mountain (7886 ft; 2404 m) is the highest. A dozen or more other peaks attain elevations of 2000 metres or higher and with base level just above 1000 metres the relief approaches 1000 metres generally. The range is dominated by sharp northwest trending ridges that are separated by narrow V-shaped valleys, and this fabric is controlled by bedrock structure. The ridges generally have their gentler dip-slopes on the southern side. A few small glacierets remain on some of the higher mountains.

In Quiet Lake map area the Big Salmon Range is composed of gently dipping metamorphic rocks intruded by large quartz monzonite batholiths. Although it is generally not as rugged as St Cyr Range it has some steep glacier-carved cliffs, for example, Dropoff Mountain. The height and relief are about the same as St Cyr Range but the valleys are somewhat wider and more U-shaped.

The Big Salmon Range lacks a preferred grain and reflects the homogeneity of the rocks that underlie it. On the west the Big Salmon Range slopes evenly toward Teslin valley, falling 1000 metres in 15 kilometres.

In the southeastern Pelly Mountains are the Simpson and Campbell ranges. These are underlain by gently dipping metamorphic rocks and small plutons. The mountains are more rounded and open than those in Big Salmon Range, but equally homogeneous in their fabric. The ranges form an upland surmounted by isolated mountain groups and the valleys between are broad and U-shaped. The maximum elevation (2353 m), base level (1000 m), and average relief (about 1000 m) are the same as for other parts of the Pelly Mountains.

Tintina Trench is a deep gash marking the Tintina fault, an important Cretaceous[4] transcurrent break. Its southwest side, part of the boundary of the Pelly Mountains, is an abrupt, steep slope that rises directly from the valley floor at 800 metres to an elevation of about 1800 metres in a horizontal distance of 2 or 3 kilometres (Fig. 3). The trench narrows southeastward and its floor rises nearly 500 metres from Lapie River to Hoole River. Along part of its northeast boundary the trench merges imperceptibly with the MacMillan and Pelly mountains.

Nisutlin Plateau is a broad flat depression south of the Pelly Mountains with little relief at an elevation about 1200 metres. It is underlain by the same deformed Paleozoic sedimentary rocks found in the southern St Cyr Range. The two physiographic divisions merge and St Cyr Range rises gradually from the north side of Nisutlin Plateau. The Plateau is a drift-covered expanse of black spruce swamps and shallow lakes from which the Liard, Wolf and Nisutlin rivers issue. Most of the streams are only slightly entrenched into the Plateau and exposures are rare.

Pelly Plateau north of Tintina Trench and Pelly Mountains is a rolling upland with relief of 500 metres or more. It is drained toward the southwest by the Pelly and Ross Rivers, which are entrenched several hundred metres below the plateau surface. The plateau slopes gradually upward to the northeast merging with the Selwyn Mountains, some which reach elevations near 2000 metres. Pelly Plateau is underlain by Paleozoic sedimentary rocks that are equivalent to, but less intensely deformed, than those of Pelly Mountains.

The physiographic differences between the ranges of the Pelly Mountains reflect bedrock character and structure to a degree, but mainly they indicate fundamental differences in the Late Tertiary or Recent uplift history. For example, St Cyr Range and Nisutlin Plateau are on trend with and underlain by the same strata; they have survived the same Pleistocene glacial events. Yet they could hardly be more different. This probably indicates that the St Cyr Range was raised more recently than Nisutlin Plateau, and implies that differences may also exist in the uplift history of the other ranges. If the physiographic maturity of the ranges is a clue to their relative times of uplift the Nisutlin and Pelly plateaus are oldest, the Simpson and Campbell ranges younger and the Glenlyon, Big Salmon and St Cyr ranges youngest.

It is possible to draw a subjective map that delimits the uplift by drawing envelopes of the elevations of peaks. Such a map (Fig. 4) shows the St Cyr and Big Salmon uplifts are sharply truncated on the northeast by Tintina fault and slope gently west, and more gradually southeast into Nisutlin Plateau. This map also indicates

[3] Species lists and details of the Pelly Mountains and Yukon Plateau - North ecoregions are given in Smith et al., 2004.

[4] Tintina fault motion is mostly post-Cretaceous.

faults that separate the Glenlyon-Big Salmon-Little Salmon and St Cyr ranges. They are herein called the Drury and Glenlyon faults. The sketch further shows the Pelly, Simpson, Campbell uplifts as a gently warped surface truncated on the southwest by Tintina fault.

The St Cyr-Glenlyon-Big Salmon uplift can be viewed as a slab hinged on the west and southeast, but cut and uplifted as much as 1000 metres by a young fault that defines the south side of Tintina Trench. The displacement on this Tertiary reactivated Tintina fault diminishes along strike from a maximum near longitude 133°W. As part of the last uplift the Drury and Glenlyon faults allowed differential uplift between the St Cyr, Glenlyon, and the Big Salmon ranges. These faults can be traced southeastward into the upper valley of North Big Salmon River and probably control that prominent lineament. However, displacement diminishes southeastward and rock units are not displaced along the North Big Salmon River. This is evidence that the Drury and Glenlyon faults are normal faults without appreciable strike-slip movement. The Tertiary reactivation of the Tintina fault was probably also without strike-slip.

The physiographic development of the Pelly Mountains involved three stages. The earliest was planation to form the surface of which Pelly and Nisutlin plateaus are remnants. That broad surface once extended across Tintina fault. The second stage was general uplift of this surface, called the Yukon Plateau by Bostock (1948a). To the northwest lay the valley localized along Tintina fault by normal faults. Finally, renewed sharp uplift of a comparatively small part of the upland spared during the Cassiar events, formed the Glenlyon-Big Salmon-St Cyr ranges. In this stage uplift was localized by normal faults along and near the Tintina Valley.

Evidence for the age of the first two events lies outside the project area. Near Carmacks and Aishihik Lake (Tempelman-Kluit, 1974a) the mid-Cretaceous Mount Nansen and Carmacks groups are truncated by the Yukon Plateau. This physiographic feature therefore formed after the Cretaceous and was raised and dissected in the Tertiary. The Cassiar Mountains and Campbell-Simpson ranges may be Cretaceous remnants that stood above Yukon Plateau. Alternatively they may be parts of the Plateau rejuvenated before the St Cyr-Big Salmon-Glenlyon uplift. The St Cyr-Big Salmon-Glenlyon uplift predates the last glacial advance[5] and followed Tertiary sedimentary and volcanic rocks in Tintina Trench, which are truncated by the reactivated normal fault that raised the St Cyr and related ranges. Because those rocks are Eocene the latest uplift (i.e., St Cyr-Big Salmon-Glenlyon ranges) may be Eocene through Pliocene. The Cassiar-Campbell-Simpson ranges may be Early Tertiary.

[5] The glacial history is not addressed in this report. See Jackson et al. (1991) and Duk-Rodkin (1999) for Yukon-wide descriptions. Bond et al. (2002), Bond and Plouffe (2002) and Bond and Kennedy (2005) report studies in the Finlayson district and Pelly Mountains.

Previous geological investigations

Employees of the Hudson's Bay Company reported their exploration of interior Yukon in the 1850's, notably Robert Campbell (Wilson, 1970). Notes on the geology included in reports of these early journeys were summarized by Kindle (1946).

The geology of the Pelly Mountains has been studied in increasing detail since Dawson's first reconnaissance in 1887. The early investigations differ from the later in one important aspect: The first fifty years of work produced a set of progressively more accurate route maps along which the geology was well known, but between which little was known. The later work, beginning in this region with Operation Pelly by the Geological Survey of Canada, resulted in comprehensive coverage that eliminated the differences in level of understanding from place to place.

Dawson in 1887, McConnell in 1898, Johnston and Lees in 1935 and Kindle in 1944 and 1945 were part of the early route mapping phase. Dawson (1888) mapped the course of the Pelly River and reported on the rocks he saw. Among other things, he observed "Laramie" age sediments along the Pelly, which we now know were deposited in the fault-controlled Tintina Trench. On a canoe trip up the Big Salmon River to Quiet Lake, McConnell (1898) made the remarkably accurate observation that the rocks consist of three great divisions. These were: a basal series of quartz- and mica-rich schist and crystalline limestone, an intermediate division characterized by granular limestone, and an upper division of dark slate, green schist, tuff, limestone and serpentine. We now know that his lowest schist is of Proterozoic age and the marble division is Early Cambrian. His upper division included the Ordovician to Devonian Nasina shelf edge facies with slices of cataclastic and ophiolitic rock thrust over them.

Johnston (1936) mapped the rocks along Pelly River which follows Tintina Trench. Although he did not recognize that the valley marks a zone of dislocation his main stratigraphic divisions are accurate and with hindsight his map nicely defines Tintina fault.

Lees (1936) examined a large area that includes southwest Quiet Lake map area, recognizing that rocks on opposite sides of Teslin River reflect widely different geological histories; those on the northeast are now included in Omineca Belt, whereas those on the southwest are in the Intermontane Belt (Fig. 5).

Kindle (1946) mapped the geology of a strip along the Canol Road. This provided the first cross-section of structural divisions that later came to be called Selwyn basin, Cassiar platform and Omineca Belt. Kindle also mapped two of the important strands of the Tintina fault system in his transect and demonstrated the differences across the fault system, but the magnitude of movement remained unknown.

The second, comprehensive phase of the mapping, made possible mainly through expansion of the Survey's staff and the use of helicopters, provided more information in a single season than had been available before. In a decade the geological map of this region changed from strips to complete coverage through the work of Wheeler et al. (1960a, b), Poole et al. (1960), Wheeler (1961), Roddick and Green (1961a, 1961b), Mulligan (1963), Blusson (1966), Gabrielse (1967), Wheeler et al. (1960a, b) and Campbell (1967) (see Fig.1).

In 1956 when most of the resources of the Survey's Cordilleran Section were devoted to Operation Stikine in north central British Columbia, J.O. Wheeler maintained the GSC presence in the Yukon by finishing the southern part of Glenlyon map area (R.B. Campbell having completed the rest) and beginning the initial reconnaissance of Quiet Lake map area. The seven-man packhorse party included D. Gednetz, Kieller and Anderson. The cook was N. Scherf, and R. Mann and A. Porterfield were packers . In 1957 no work was done in the area because funds were lacking.

Operation Pelly began in 1958. Its aim was a rapid reconnaissance of a much larger region than the Quiet Lake map area; it included the surrounding Finlayson Lake, Tay River, Sheldon Lake, Nahanni, Frances Lake and the northern part of Wolf Lake map areas (Fig. 1). Under J.O. Wheeler (mainly responsible for Quiet Lake and Finlayson Lake) and J.A. Roddick (Tay River and Sheldon Lake) a central base camp was extablished at Jackfish Lake (on the Canol Road six kilometres south of Ross River) from which four two-man fly camps were set out by a float-equipped Piper Supercub flown by Ernest Harrison. Wheeler and Roddick were assisted by T.G. Allan, F. Cicierski, R.K. Germundson, J.C. Lamont, J.H. Montgomery, G.D. Pollock and M.G. Williams. The cook was A. Martin. It was a hot and dry, very smoky summer during which most of the forest around Jackfish Lake burned.

In 1959 L.H. Green replaced J.O. Wheeler (re-assigned to southeastern British Columbia), and the operation was supported by a Hiller 12E helicopter and a Beaver floatplane. Besides Roddick and Green the party included Bateman, S.L. Blusson, D.B. Craig, W.J.P. Crawford, G.L. Goruk, W.W. Nassichuk, A.C. Green (helicopter pilot) and Bob Harrison (Beaver pilot). A. Martin cooked for the party again until he died on Wolf Lake in August. The party in 1960 included J.A. Roddick and L.H. Green assisted by G.F. Finlayson, A.T. Jenik, A.P.D. Lorraine (cook), G.R. Turnquist, H.N. Wilkinson, R.W. Yole and A.C. Green (helicopter pilot). The results of Operation Pelly were reported in a set of preliminary maps (Wheeler et al., 1960a, b; Roddick and Green, 1961a, b; Poole et al., 1960), but no final map or report was published.

Geological fieldwork for this report

In 1967 the author began field work in the Anvil Range, part of the area of Operation Pelly (Tempelman-Kluit, 1972b). With this background an attempt was made to report Operation Pelly more fully, but experience in only a small part of the region proved too great a handicap and the product (Tempelman-Kluit, 1974b) was anything but authoritative. The writer began remapping in Quiet Lake map area in 1973 as the basis for a comprehensive account of the geology of the region. This report is the result. During 1973 and 1974 work was done largely from the roads and the logistics included minor helicopter support, but in 1975 and 1976 more extensive helicopter transportation was used.

Abbott (1977, 1981) documented the contrasting structural styles of Cambro-Ordovician slaty rocks and overlying Siluro-Devonian carbonate strata just outside the project area. The object was to resolve whether the different styles reflect early deformation and unconformity or contrasting response controlled by the lithologic difference. Read (1976, 1980) examined Lower Cambrian strata in east central Quiet Lake map area and documented the thick archaeocyathid build-ups, speculating on their controls and correlations. Gordey (1977, 1981a) mapped a geological transect southwest from Tintina fault to detail the structural variation of Early Paleozoic strata and to define the structural relations of metamorphic klippen that lie above unmetamorphosed strata.

The main findings of the present study were reported annually as the work progressed (Tempelman-Kluit et al., 1974, 1975, 1976 and Tempelman-Kluit, 1977b) and in preliminary geologic maps of Quiet Lake and Finlayson Lake map areas (Tempelman-Kluit, 1977a).

The regional coverage outlined the main stratigraphic divisions and defined its structural units. For example, Tintina Trench was recognized as the locus of a right lateral strike-slip fault with about 450 km (260 miles) of displacement (Roddick, 1967). The extent, near completeness and platformal nature of the Early Paleozoic stratigraphic record in the Pelly Mountains had been noted and contrasted with the Devono-Mississippian transgressive clastic sequence. Affinities of the Paleozoic sequence with that of the Cassiar Mountains had been recognized (Gabrielse and Wheeler, 1961). Some of the structural complexity of the Paleozoic succession, including its repetition by northeast directed thrusts, had been defined (Gabrielse and Wheeler, 1961) and some of the klippen of metamorphic rocks above unmetamorphosed strata were delimited (Wheeler et al. 1960a, b). An extensive metamorphic-plutonic episode of roughly mid-Cretaceous age had been mapped through much of the region and its apparent relation to stratigraphic level had been noted (Campbell, 1967). The disparity between the geologic history of the northern Intermontane and Omineca belts was well known (contrast Wheeler, 1961 with Campbell, 1967, for example). It remained to describe and define most of these features, for in the excitement of the initial reconnaissance phase, the follow-up reporting of the results lagged badly. In addition comparative details of stratigraphy and extent of structures remained to be studied, as was the relationship of the diverse geologic units to each other.

The principal contribution of this report is to document the regional stratigraphy and structure resulting from this and the earlier reconnaissances of the Pelly Mountains. Refinements in understanding the regional geology include:

1) the existence of two distinct volcanic assemblages, one Cambro-Ordovician (Cloutier formation), the other Devono- Mississippian (Felsic Volcanic formation). Formerly the volcanic rocks were all considered Devono-Mississippian, which led to problems in structural interpretation;

2) the recognition by conodont study that a thick Upper Triassic sequence (Hoole formation), formerly considered Mississippian, is locally preserved above Paleozoic strata. The presence of relatively young rocks indicates the youthfulness of regional deformation;

3) the documentation of profound facies variation in the Cambro-Ordovician, Siluro-Devonian and Devono-Mississippian platform and shelf sequences. In contrast, coeval strata in Mackenzie Mountains exhibit more consistent facies;

4) documentation of the deformation of the miogeo-clinal rocks and the recognition of about 100 kilometres of shortening on four main thrusts that imbricate the strata. These thrust sheets, named for their underlying faults, are depicted on page-size maps in this report;

5) recognition of the regional extent of allochthonous rocks; that they are distinctively sheared or cataclastic; that they can be separated into a sedimentary slice (Nisutlin allochthon), a dismembered ophiolite (Anvil allochthon) and a cataclastic granitic slice (Simpson allochthon). Furthermore these rocks occupy Teslin Suture Zone between the Omineca and Intermontane belts;

6) documentation of variations in the mid-Cretaceous plutonic-metamorphic culminations across Omineca Belt. On the southwest are large deep seated batholiths, rooted in the rocks from which they formed, but to the northeast are small, shallow, high-level discordant plutons; and

7) recognition that the autochthon-allochthon-granitic stories fit an arc-continent collision model that can explain the evolution of the Columbian orogen in the north and which may apply to the western Cordillera more generally.

Mineral exploration history

The Pelly Mountains have been prospected continuously since the Canol Road was opened to the public, but during this thirty year interval they have witnessed three flurries of mineral exploration activity - each focussed on different types of occurrences. The exploration history illustrates not only the way in which the industry works, but also the changing focus on Yukon mineral deposits and the feasibility of mining them.

The earliest exploration rush occurred in the late 1950s. It accompanied the initial phases of Operation Pelly when much new geological information suddenly became available. This was a time when only rich silver veins seemed likely to be economic in the Yukon and prospecting in the St Cyr Range concentrated on such occurrences to the exclusion of all else. The Tintina Silver, Ketzakey, Groundhog and Canol prospects, all silver-lead vein occurrences, date from this period. None of them proved economic despite much exploratory drilling and some underground work.

The second period of concentrated activity in the Pelly Mountains began about 1966. It followed discovery in 1965 of the Faro ore body, a metamorphosed strata-bound zinc-lead deposit in the Anvil Range. Exploration concentrated near Faro, but spilled over into the Campbell and Simpson ranges and onto Pelly Plateau. Rocks in these areas differ from those in Anvil Range and no new zinc-lead occurrences were found although several copper-zinc showings were turned up near Grass Lakes, North Lakes and Fyre Lake[6].

By the late 1960s and early 1970s interest had shifted to other parts of the Territory, notably the Dawson Range, but about 1973 the Selwyn Mountains became exciting through the discovery of a stratabound lead-zinc deposit of staggering size at Howards Pass. Also at this time, and for a combination of reasons, many exploration groups that had concentrated their efforts in British Columbia, were attracted to the Territory. Later recognition that strata of the same age and lithology as those at Howards Pass and Anvil also occur in the Pelly Mountains led to renewed interest and

[6] A volcanogenic massive sulphide deposit (Kudz Ze Kayah) was found near Grass Lakes in late 1992 (Schultz, 1994). The the area south of Finlayson Lake was then intensely explored (e.g. Burke, 2000; Hunt, 2002).

exploration in the St Cyr Range during 1975 and 1976. In 1977 the JA and Angie, both stratabound lead-zinc showings, were discovered there in rocks like those at Howard's Pass. The mid-seventies exploration rush also focussed attention on the Mississippian felsic volcanics in St Cyr Range which host the MM lead-zinc occurrence. These attracted interest as a volcanogenic massive sulphide target. They are younger than the Tom, an early Late Devonian stratabound lead-zinc-barite deposit near Macmillan Pass.

In the late seventies the Pelly Mountains were explored for uranium and for tin and tungsten. Several showings of these metals are known in the Big Salmon and St Cyr ranges where they are spatially associated with mid-Cretaceous quartz monzonite intrusions. With the rising price of silver (1980) the vein occurrences that first attracted attention to the Pelly Mountains are again under investigation.

Acknowledgements

The author was ably assisted in the field by university students, many working toward degrees in geological sciences. In 1973 the assistants were: J.G. Abbott, B.C. Read, J. Wrobel and R. Simson. In 1974 J.G. Abbott, S.P. Gordey, B.C. Read, D. Harper, G. O'Neill, D. Van Appelen and R. Turna were the assistants. In 1975 the assistants were B.C. Read, S.P. Gordey, D. Van Appelen and G. Cavey. M. Yandel and K. Kaninagan cooked for the party and W. Muise and L. Davidson were the helicopter pilots. In 1976 S.P. Gordey and K. Pivnick were the assistants, M. Yandel cooked and R. Taylor and G. Curtis were the helicopter pilots.

Abbott, Gordey and Read stayed with the project for several years and assisted with the mapping. They contributed much to the project by their perceptive and critical discussion of many of the ideas detailed here. As well they studied parts of the area for their graduate dissertations.

Compilation of field data was done during several winters. It is a pleasure to acknowledge the inspired and willing assistance of B. Struik, M. MacArthur and E. Fuller with this phase of the work. A first draft of the regional maps (1:250 000 scale) was released as Open File 284 in 1984. B. Vanlier typed the manuscript in an early word-processing program, and the first edition was submitted to R.B. Campbell, GSC Vancouver subdivision head, for critical review in the summer of 1985.

The critically reviewed report remained uncorrected as the author was involved in other field projects, managerial duties and secondments, until his departure from the GSC in 1994. The open file maps and unpublished information were used in the digital compilation of Yukon bedrock geology (Gordey and Makepeace, 2001, 2003).

In 2004 the Yukon Geological Survey requested access to the unpublished report, and subsequently contributed funding for digital cartography of both the 1:50 000 and 1:250 000 maps to accompany it. These were drafted by R. Cocking and R. Chan respectively. The legends for these maps were syncronized by C. Roots, who also undertook the changes requested by the critical reviewer. At Publication Services - GSC Ottawa, optical character recognition of the 1985 paper manuscript was arranged by A. Moore. The compiled report and maps were critically reviewed in 2007. Map revisions were made by R. Cocking and C. Wagner of GSC Vancouver. At Technical Quality Control - GSC Ottawa, E. Everett reviewed the changes. S. Irwin, the GSC-Vancouver subdivision-head, oversaw this project to its completion.

GENERAL GEOLOGY

Introduction

Quiet Lake and Finlayson Lake map areas provide a cross-section of the northern part of the Omineca belt, a subdivision of the Canadian Cordillera (Fig. 5). The Omineca and Foreland belts (Mackenzie and Rocky mountains) contain time equivalent Proterozoic and Paleozoic strata of different facies deposited close to the western margin of ancient North America. The Foreland belt is dominated by carbonate strata that constitute the miogeocline; Omineca belt includes its western, more shaly, eugeoclinal equivalents. The Intermontane belt, southwest of Omineca belt, contains Mesozoic volcanic and plutonic rocks and derived sedimentary strata, which bear no depositional relationship to those of Omineca belt. The Intermontane belt was tectonically juxtaposed next to the Omineca belt probably in the Jurassic.

Tintina Trench[7] slices diagonally across Quiet Lake and Finlayson Lake map areas (Fig. 3). It is a Late Miocene or Pliocene graben formed along a Late Cretaceous-Early Tertiary strike-slip fault with about 450 kilometres of dextral displacement. The effect of this displacement in southern Yukon has been to repeat the Omineca belt, but because the Tintina fault and the Belt are only slightly divergent a similar stratigraphic-structural sequence is exposed in the two map areas on opposite sides of the fault despite the tremendous displacement.

Quiet Lake and Finlayson Lake map areas are underlain by folded and faulted sedimentary rocks that range from Eocambrian[8] to Triassic. Cataclastic sedimentary rocks, ductile-deformed serpentinite, gabbro and basalt and sheared granitic rocks are thrust over the sedimentary strata and preserved above them as klippen[9] (Fig. 6a). The sheared and cataclastic rocks are mostly Late Paleozoic and Mesozoic and, being transported, or allochthonous, are depositionally unrelated to the sedimentary sequence (autochthonous) on which they rest. The autochthonous strata were folded and thrust faulted in Jura-Cretaceous time after the allochthonous cataclastic rocks were transported over them. Finally, the deformed sedimentary sequence and the cataclastic rocks emplaced on them, were metamorphosed and intruded by quartz monzonite batholiths and plutons that cooled during the Cretaceous.

AUTOCHTHONOUS STRATIGRAPHY

The autochthonous strata of the northern Omineca Belt in Quiet Lake and Finlayson Lake map areas form a fairly complete marine succession separated into four subdivisions bounded by unconformities or depositional breaks (Figs. 6a, 6b). The stratigraphy (Table 1) is summarized below. The oldest strata form the Ketza group, an Eocambrian to Lower Cambrian, fine grained clastic assemblage capped by carbonate and calcareous shale. The facies are comparatively constant. The second subdivision includes the Kechika, Askin and Harvey groups, which range from Upper Cambrian to Lower Devonian, and in which profound and abrupt facies variation is characteristic. From northeast to southwest in the report area this sequence changes from shale (Selwyn basin), through coeval volcanics capped by thick accumulations of carbonate (Pelly-Cassiar platform[10]), to finer grained siltstone and shale (Nasina outer shelf). The third assemblage, the Seagull group, comprises an Upper Devonian and Mississippian transgressive marine conglomerate. In part of the project area the unit is dominated by marine felsic volcanic rocks. The fourth subdivision, preserved locally, is a thin marine sequence of Permian shale (Starr formation) overlain by Upper Triassic marine calcareous siltstone (Hoole formation).

None of the Eocambrian through Triassic strata in the map areas, even the most southwesterly, are facies that likely accumulated on the continental slope. Probably they were deposited on the shelf or outer shelf of ancient North America, an unknown distance from the shelf-slope break.

Equivalent or related strata on opposite sides of Tintina fault are treated together even though they were deposited 450 kilometres from each other[11]. The descriptions in this report progress from the oldest to youngest rocks. Several new names are introduced to identify the main rock units, and some units are correlated with existing stratigraphic nomenclature. New unit names include the Ketza and Seagull groups

[7] "Tintina trench" refers to the Pliocene graben and "Tintina fault" to the older transcurrent fault

[8] "Eocambrian" in the original text refers to strata of unknown age stratigraphically below those containing Early Cambrian fossils. Modern useage would be Late Proterozoic or Ediacaran period without a lower age constraint.

[9] These klippen were later included in the 'Yukon cataclastic complex' (e.g. Tempelman-Kluit, 1979) and subsequently part of the Yukon-Tanana Terrane (Mortensen, 1992a).

[10] Cassiar platform was named by Gabrielse (1963, 1998] for Late Silurian and early Devonian carbonate in northern British Columbia, but expanded to Lower Cambrian through Middle Devonian shallow water strata (e.g. Tempelman-Kluit, 1979; Fritz at al., 1992). The lithologic and time-equivalence of these strata to those of the miogeocline of western cratonic North America is widely accepted. A gap in the semi-continuous oolitic shoal iIn southwest Washington and southern Idaho is a possible origin (Pope and Sears, 1997) translated northward along poorly constrained dextral strike-slip faults.

[11] The validity of this statement was questioned by the critical reader but the manuscript has not been altered. However the accompanying maps (1:250 000 and 1: 50 000 scale) depart from the report by showing separate legends for rocks northeast and southwest of the Tintina fault.

and the Pass Peak, McConnell, Gray, Groundhog, Cloutier, Ram, Nasina, Hogg, Barite Mountain, Porcupine, Starr and Hoole formations. Terms that are redefined include Big Salmon Complex and Askin and Kechika groups. Members of formations and units named after their distinctive lithology remain of informal status[12].

Three diagrams (Figs. 7a, 7b and 8) schematically portray the depositional relations and lateral variation of the rock units southwest, and northeast of the Tintina fault. These diagrams summarize the legends for the accompanying maps. The diagrams depict facies variations in a generalized northeast-southwest reconstructions of the now deformed sequence. Facies relations are projected onto these schematic cross-sections, which have no scale. Because variations in the third dimension cannot be portrayed some of the associations or relations are exaggerated. The stratigraphic and lateral relationships of these units are understood with various degrees of certainty ranging from hypothetical to well documented, as pointed out in the text.

Lower Cambrian and older (?) strata

The oldest strata in the project area are most extensively exposed southwest of Tintina fault in the Big Salmon and St Cyr ranges (Fig. 9) where they make up the Big Salmon Complex. Strata in the Big Salmon Range are regionally metamorphosed and migmatized and their stratigraphy is poorly known. By contrast the rocks in the southern St Cyr Range and Ings River area are unmetamorphosed, and make up the Ketza group.

Northeast of Tintina fault Eocambrian or older strata are known in a restricted area near McEvoy Lake in the northeast corner of Finlayson Lake map area. Highly metamorphosed rocks, which may include Eocambrian and/or older beds, occupy a large area northeast of Tintina fault in Finlayson Lake map area and are referred to as the Mink complex.

The Big Salmon and Mink complexes represent largely Eocambrian and Lower Cambrian strata, which with some younger rocks, were metamorphosed during the Cretaceous[13]. Neither complex is a stratigraphic entity, but represents metamorphosed rocks that are dominantly Eocambrian. These metamorphic complexes are described in a subsequent section; only the weakly metamorphosed or unmetamorphosed strata referred to as the Ketza group are described here.

[12] All new formation names in this report remain informal (Open File publication)

[13] Subsequent U-Pb zircon age determinations suggest the metamorphism is Mississippian and Permian (Stevens et al., 1993; Murphy and Piercey, 1999).

Ketza group

The name Ketza group is introduced for Lower Cambrian and older strata near the headwaters of Ketza River in Quiet Lake map area (Figs. 9 and 10). The rocks can be traced to White Creek and Lapie Lakes, and equivalent strata, for which the name is also used, occur near Ings River. The Ketza group comprises the oldest stratified rocks exposed in the project area. It is overlain unconformably by the Kechika Group and grades laterally (and downward) into metamorphosed rocks that are part of the Big Salmon Complex. Near Ketza River the Ketza group has been studied and described by Read (1976, 1980), and are summarized in Tempelman-Kluit et al. (1975).

The Ketza group includes a lower non-calcareous unit called the Pass Peak formation that is possibly as old as Late Proterozoic, and an upper unit of calcareous rocks named the McConnell formation that is Early Cambrian. Each is divided into several members (Fig. 10).

The Pass Peak formation (map unit lЄP)

his unit, named after Pass Peak in central Quiet Lake map area, includes greenish shale, siltstone and quartzite in various proportions. In the type area about seven kilometres east of Pass Peak as well as near Lapie Lakes and Ketza River, shale and siltstone generally predominate. By contrast, medium- to coarse-grained quartz sandstone, the Ings member, makes up the bulk of the formation near the mouth of Ings River. The Pass Peak formation also includes the Lapie member, a 200-metre-thick, mainly massive carbonate found on the ridges between Pony Creek and Lapie Lakes. The base of the Pass Peak formation is not seen, but at least 200 metres of these rocks are present in the type area and 400 to 700 metres are exposed west of Lapie Lakes. The Ings Sandstone is about 200 metres thick.

The McConnell formation (map unit lЄM)[14][15]

This unit is named after McConnell River which flows across it, includes about 600 metres of argillaceous lime-mudstone and calcareous argillite with discontinuous lenses of cleaner limestone (Fig. 11a, b, c). Southwestward the formation is dominated by cleaner carbonate and most of it is metamorphosed. The lowest member is a dark grey, thin bedded, silty lime-mudstone found near Ketza River and locally northwest

[14] A formal McConnell Formation exists in the Front Ranges of the Rocky Mountains. Strictly, the unit describedsin this report should be the McConnell River formation, but for internal consistency among the original figures is left intact..

[15] This unit is time-equivalent to Rosella Formation (Gabrielse, 1998) which was recognized in central Wolf Lake map area by Murphy (1988) and adopted by (Gordey and Makepeace, 2001, 2003).

of Lapie Lakes (Fig. 11d) discontinuous archaeocyathid carbonate mounds, locally 200 or 300 metres thick, occur commonly in the Ketza River area. They are called the White Creek member for the excellent exposures south of that stream and north of Grizzly Creek (Figs. 11e, f). In its southwestern outcrops near Ings River and west of Quiet Lake, the cleaner carbonate is named the Scurvy member after Scurvy Creek. In the Ketza area 100 metres of black pyritic slate overlie the White Creek member and form the uppermost unit of the McConnell formation (Fig. 11c). Near Ketza River and northwest of Lapie Lakes. Parts of the McConnell formation are replaced by irregular shaped patches of massive secondary orange dolomite up to hundreds of metres across (Fig. 11a, b) he dolomite is not separately named, but distinguished by a pattern on the map

Fossils indicate a hiatus between the Ketza and Kechika groups. The strata, however are lithologically similar (Fig. 11f) and equally incompetent, they are locally transposed and structurally mixed to such an extent that their depositional relations are obscure.

Age and Correlation

The age of the upper formation of the Ketza group is determined from numerous archaeocyathid and trilobite collections near Ketza River and Ings River and scattered other localities. Appendix 1 lists the collections; their age range is shown schematically in Figures 12 and 13. Most of the fossils are from the McConnell formation, particularly the White Creek member in the upper part of the Ketza group. Together these data show that the upper part of the Ketza group is Early Cambrian.

The Lower Cambrian is divided into three trilobite zones (Fritz, 1972); from oldest to youngest they are: the *Fallotaspis*, *Nevadella* and *Bonnia-Olenellus* zones. No fossils have been recovered from the Pass Peak formation, but because these rocks lie directly beneath the fossiliferous strata in an uninterrupted sequence they are considered Lower Cambrian or older. The lowest part of the McConnell formation, including the dark grey silty lime mudstone, is probably near the *Nevadella-Fallotaspis* Zone boundary. The Precambrian-Cambrian boundary may lie within the Pass Peak formation (Fig. 10).

Possible correlations of the Ketza group with Lower Cambrian strata elsewhere in the Cordillera are indicated schematically in Figure 14[16]. The Ketza group is lithologically like the Atan Group of the Cassiar Mountains (Gabrielse, 1963) and of the northern Rocky Mountains (Taylor and Stott, 1973) in which Fritz (1980) defined the upper carbonate as Rosella Formation and the lower siliciclastic unit as Boya Formation. The Ketza and Atan groups also have similar fossils dominated by archaeocyathids (Okulitch and Greggs, 1958). As in the Ketza group most fossils from the Atan Group were found in the upper calcareous unit, but are somewhat younger and are referred to the *Bonnia-Olenellus* zone (Gabrielse, 1963; Taylor and Stott, 1973). Some fossils in the upper part of the lower clastic part of the Atan Group indicate that it too is Lower Cambrian. If there are differences between the Atan and Ketza groups, they are slight. The lower part of the Atan contains more quartzite, in contrast to the Ketza which includes more siltstone except near Ings River where it too is quartzitic. The lime-mudstone in the upper Atan may be cleaner than that in the Ketza group. The Ketza group may have more archaeocyathid build-ups, and may be slightly older than the Atan Group.

More distant correlatives of the Ketza group are the Sekwi Formation in the Mackenzie Mountains (Fritz, 1976) and the Yanks Peak, Midas and Mural formations of the Cariboo Mountains (Campbell et al., 1973). The Sekwi differs from the Ketza group in that it includes a complete and thicker section of the Lower Cambrian, dominantly in carbonate facies. The Sekwi probably represents a more miogeoclinal equivalent of the Ketza group. On the other hand the Cariboo Mountains section is similar to that in the Pelly Mountains. The Yanks Peak and Midas formations are lithologically like the lower part of the Ketza group. They directly underlie the Mural formation, a carbonate with fossils ranging through the entire Lower Cambrian. The Yanks Peak and Midas formations are therefore probably equivalents of the Pass Peak formation whereas the Mural correlates best with the McConnell.

The black pyritic slate at the top of the upper carbonate formation of the Ketza group is comparable with the Dome Creek Formation of the Cariboo Mountains (Campbell et al., 1973). The units are lithologically alike, largely unfossiliferous, and overlie Lower Cambrian strata. Only about 100 metres of the black slate is preserved in the Pelly Mountains beneath the unconformably overlying Upper Cambrian Kechika Group whereas the Dome Creek is 1500 metres thick. The Eocambrian-Lower Cambrian strata of the Cariboo and Cassiar mountains compare closely to those in the Pelly Mountains and perhaps formed in similar positions with respect to North American shorelines of the Lower Cambrian.

Kechika Group (Cambro-Ordovician)

Slate and phyllite with thick volcanic accumulations unconformably overlie the Lower Cambrian strata in the Pelly Mountains. These Upper Cambrian and Ordovician rocks are confined to the region southwest

[16] The correlations discussed in this section are mostly of historical interest. Precision dating and facies analysis since this report was written show significant differences in age and paleogeography for most units mentioned.

of Tintina fault (Fig. 15). They resemble the Kechika Group[17] (Gabrielse, 1963) so closely that the same name is used here. Originally mapped as unit 2 by Wheeler et al. (1960a,b), these rocks were grouped with other slaty rocks of diverse ages.

In the Pelly Mountains the Kechika Group is divided into four laterally equivalent, interfingered facies capped by a thin discontinuous black slate. From northeast to southwest these facies are named the Ram, Cloutier, Groundhog and Gray Creek formations for different streams where they are well exposed. The black slate cap is called the Magundy formation (Fig. 16).

Subdivisions of the Kechika and correlative Harvey groups are mapped separately as formations and occur in narrow, northwest trending, discontinuous belts several kilometres wide (Fig. 15). Relations between the facies are inferred to be laterally gradational because they are commonly interbedded with each other. For example, the orange or buff platy limestone and phyllite that characterizes the Ram formation occurs within the darker, non-calcareous phyllite of the Groundhog formation and less commonly in the volcanic tuff of the Cloutier formation. Similarly lenses of flows and tuffs, characteristic of the Cloutier formation are common in the Groundhog and Gray formations and occur locally in the Ram and Magundy formations. Black graptolitic slate of the Magundy formation generally caps the Ram, Cloutier and Groundhog formations, but locally such slate occurs within the other units.

The boundary between the Kechika and Ketza groups is exposed on the western and northern sides of the Ketza group outcrop area near Ketza River (Fig. 9). It can also be studied on the south side of White Creek and on the west side of the Seagull Creek. In each place grey phyllite and slate (Groundhog formation), locally with intercalated volcanics (Cloutier formation) and elsewhere with buff calcareous phyllite (Ram formation) lies above calcareous argillite, limestone or the black siliceous slate of the Ketza group (McConnell formation). Because the rocks above and below are alike and because the relations are masked by low grade metamorphism and deformation the contact is not prominent and appears conformable. Although discontinuity of the units below the black slate may have resulted from stratigraphic truncation, it may equally be ascribed to facies variation. However fossils (Fig. 12) indicate that the contact is unconformable or disconformable because Middle Cambrian strata are missing between the two groups. Trilobites from the upper 50 metres of Ketza group are Lower Cambrian, probably Nevadella zone; those from the lowest 100 metres of overlying Kechika Group nearby are probably Upper Cambrian (Compare Collections 98A and 98B, Appendix 1).

Absence of Middle Cambrian strata in the Cassiar Mountains (Gabrielse, 1963) indicates a depositional break of the same age there. In the northern Rocky Mountains Taylor and Stott (1973) inferred that an angular unconformity separates the Lower Ordovician Kechika Group from the Lower Cambrian Atan Group.

Ram formation (map unit uЄOR)

The Ram formation occupies a narrow strip between the Tintina and St Cyr faults (Fig. 15). It is a recessive unit with most exposures covered by slabby or chippy scree. The talus weathers to a distinctive orange buff colour and has a prominent lustrous sheen that is evident from a distance. The rocks are homogenous, thin bedded and laminated, medium brownish grey calcareous slate. Locally, thin bedded platy limestone with argillaceous partings predominates (Fig. 17a). Such limestone is particularly common north of Tintina stock in Finlayson Lake map area (Fig. 17b). The slate locally contains up to 5 percent euhedral scattered pyrite a millimetre or less across. Where these are weathered the rock looks "spongy" and is pock-marked by limonite-lined pits. Some of the carbonate in the slate is ferrodolostone or ankerite which gives the slate its orange weathering colour.

Bedding, defined by compositional differences, consists of alternate layers with contrasting amounts of slate and limestone, a few centimetres thick (Figs. 17a, b). The limestone is a micritic lime-mud, without silt- or sand-sized material. A pervasive cleavage, spaced at a centimetre or less, generally cuts across bedding and produces a thin splitting habit which controls the size of scree. Small scale folds are common and spectacularly outlined by thin bedded rocks. Near Ram Creek the Ram formation contains interbedded black slate like that of the Magundy formation in beds about a metre thick, and the unit contains beds of grey slate like that of the Cloutier formation. The Ram formation is about 1000 metres thick on Ram Creek and a similar thickness is present north of Tintina stock. The base of the unit is not seen. The Ram formation is equivalent to Gordey's (1977, 1981a) 'Orange Slate Unit'.

Groundhog formation (map unit uЄOG)

This unit is named for Groundhog Creek (Fig. 15), the type locality, where the unit is extensively and well exposed. It includes at least 800 metres of medium grey slate or phyllite, tuffaceous slate or chloritic phyllite and greenstone or andesitic tuff (Figs. 17c, d). Most of the rocks weather recessively, but the greenstone tends to form resistant lenses. The phyllitic rocks have the same

[17] Subsequently redefined by Gabrielse (1998) as Kechika Formation. He re-assigned the upper graptolitic part of the former Kechika Group to the Road River Formation in northern British Columbia.

sheen as the Ram formation, but weather dark grey and lack the orange colour of that unit. Medium to light grey phyllite or slate with a greenish cast is the commonest rock. It contains silt-sized or finer quartz with chlorite and sericite (Fig. 17d). Unlike the slate of the Ram formation this rock is rarely calcareous. Interbedded with the slate and phyllite is greenish choritic phyllite, a weakly metamorphosed fine grained tuff (Fig. 17e). Its fragments are up to a centimetre across, generally flattened and difficult to recognize. It commonly contains prominent euhedral pyrite crystals that locally make up as much as 5 per cent of the rock. Tuff forms lenses several metres thick and several hundred metres long. The tuff, slate and phyllite are cut by a pervasive, closely spaced cleavage that has localized mica growth. This cleavage controls the size of the talus and lends slopes of these rocks their lustrous sheen.

Dark grey weathering greenstone, a finely crystalline massive rock made up of epidote, chlorite and rare relic augite phenocrysts, forms resistant lenses up to 30 metres thick within the phyllite. It has some flattened, calcite-filled vesicles and is presumed to have originated as flows. The greenstone is like the volcanics that make up most of the Cloutier formation.

A diabase sill, 30 metres thick, caps the ridges immediately east of Lapie Lakes (Quiet Lake map area). This resistant sill weathers dark grey and has dark blocky talus. The rock is medium green and, although partly altered, retains a "greenstone" texture. It is medium- to coarsely-crystalline in the centre with aphanitic to finely crystalline margins several metres thick. The rock contains saussuritized plagioclase with actinolite and chlorite pseudomorphous after augite. Ilmenite is a common opaque mineral. Diabase from 20 kilometres ESE of McNeil Lake (Finlayson Lake map area) (Gordey, 1977, 1981a) has the same relations and is considered equivalent. Gordey showed that the sills are chemically indistinguishable from the volcanic rocks of the Cloutier formation. The diabase sill and the massive greenstone lack the pervasive cleavage of the enclosing phyllitic rocks. Near Lapie Lakes the volcanic flows and tuff lenses are commonest in the upper part of the Groundhog formation and make up one quarter of the stratigraphic section. The diabase sill occurs at the top of the exposed part of the unit.

The Groundhog formation is equivalent to Gordey's (1977, 1981a) Tuffaceous Slate and Slate Diabase units.

The Cloutier formation (map unit uЄOc)

The Cloutier formation is named for Cloutier Creek, a tributary of Ketza River in Quiet Lake map area. The unit is well exposed there (Fig. 15) and farther southeast along the boundary between Quiet Lake and Finlayson Lake map areas.

Cloutier formation includes resistant dark grey weathering basaltic flow and volcaniclastic rocks and is more than 500 metres thick southeast of Peak 6762 (Quiet Lake map area). The rock is a massive to weakly cleaved greenstone (Fig. 17f) that consists of a mat of saussuritized plagioclase microlites, a fraction of a millimetre across, with intergranular opaque minerals and chlorite (Fig. 17g). Rare augite phenocrysts up to a few millimetres across, are largely altered to actinolite and epidote. The rocks are commonly amygdaloidal with calcite-filled, chlorite-rimmed, flattened ellipsoidal or disc-shaped amygdules. Finely disseminated hematite is a common opaque mineral that locally lends the rocks a distinctive reddish color. Massive and amygdaloidal flow breccia and tuff are most common, but volcaniclastic rocks are locally intercalated with them. The latter contain rounded cobbles of the volcanic rocks in a matrix of finer grained material. Samples of the volcanic rocks from the Indigo Lake area were analysed by Gordey (1977, 1981a) who compared their composition to alkali basalt.

Volcanic rocks constitute the bulk of the Cloutier formation. Phyllite intercalated with the volcanic rocks is common only in the lower part of the unit. It is darker grey than that of the Groundhog formation and contains considerable graphite and no carbonate.

Gray Creek formation (map unit uЄG)

Eight kilometres southwest of Big Salmon Lake in southwestern Quiet Lake map area (Fig. 15), sheared greenstone is interlayered with biotite quartz schist and black graphitic muscovite siltstone. Together these rocks are called the Gray formation. The three rock types are about equal in importance volumetrically and occur as alternate sheets several metres thick. They are considerably more metamorphosed than the Kechika Group elsewhere and their postulated equivalence to that group is based on lithology and stratigraphic position. They cannot be traced into each other. The black graphitic siltstone resembles that of the Nasina formation, but that unit contains no volcanics. The biotite schist may be the metamorphosed equivalent of the Pass Peak formation and the greenstone may be equivalent to the Cloutier formation. The rocks lie above schist that is considered Eocambrian and below the Nasina formation.

The thickness of the Gray Creek formation is about 400 metres.

Magundy formation (map unit SM)

The Magundy formation is a distinctive black graphitic slate that locally overlies the Ram, Cloutier and Groundhog formations. Outcrops in the valley of Magundy River in north central Quiet Lake map area form the type section. The unit occupies only a small area generally immediately southwest of the St Cyr fault, but the rocks are also exposed near McConnell Peak and Mount Hogg. In some areas the Magundy formation is not distinguished from other rocks of the Kechika Group as it is commonly covered by vegetation, weathers recessively and is inconspicuous except for rare sooty outcrops in stream gullies.

The Magundy formation is generally thin and discontinuous, but 300 metres were measured on Ram Creek and the base was not seen. At McConnell Peak 100 metres or more are present. Near Askin Lake nearly 200 metres were measured and near Hoole River a section of 500 metres is known.

The slate is black, finely fissile, locally silty and commonly graptolitic.; Compositional lamination is subtle and on a scale of millimetres.

Near Ragged Peak in Finlayson Lake map area the Magundy formation contains several resistant 5-metre beds of quartz sandstone (Gordey, 1977, 1981a), but these are not known elsewhere. In places the graptolitic slate includes greenstone lenses, up to 50 metres thick, which are flows and tuffs like those of the Cloutier formation (Fig. 21a, b, c). Locally the volcanics make up the greater part of the section, particularly near the boundary between Quiet Lake and Finlayson Lake map areas (see Hoole River section, Appendix 2).

Age and correlation of the Kechika Group

In Quiet Lake and Finlayson Lake map areas the age of the Kechika Group is inferred from trilobites in the Ram formation (Fig. 12; Appendix 1) and graptolites from the Magundy formation. Collections of diagnostic trilobites and graptolites, identified by W.H. Fritz and B.S. Norford, are listed in Appendix I. The locations where fossils were discovered in the Ram facies are shown in Figure 15 and the age range of the collections is shown schematically in Figures 12 and 13. The locations of fossil collections from the Magundy facies are also shown on Figure 15. The Ram facies includes scattered Late Cambrian (probably Dresbachian and Franconian) and earliest Early Ordovician trilobites whereas the Magundy facies includes a nearly complete succession of Ordovician graptolites.

As no fossils have been discovered in other facies of the Kechika Group the age of these other strata is inferred from relations with the fossiliferous units. Because the Groundhog and Cloutier facies are probably lateral equivalents of the Ram and Magundy facies, they are also Upper Cambrian and Ordovician.

The Kechika Group of the Pelly and Cassiar mountains are alike in age and facies and contain the same fossils. Although he did not name them, Gabrielse (1963) recognized each of the facies described above in the Cassiar Mountains and implied they are laterally equivalent[18]. For instance, strata he described under the heading "Rocks in the southwest and southern parts of the map area" (Gabrielse, 1963, p. 32) are like the Groundhog formation. Those headed "Rocks northeast of the McDame Synclinorium" are lithologically like the Ram formation. The unit Gabrielse (1963, p. 34) called the "upper part of the Kechika Group" is black slate with minor phyllite and quartzite, here included in the Magundy formation. Gabrielse further noted that volcanic rocks are common in the Kechika Group and, from his descriptions, they closely resemble the Cloutier facies. The "greenstone…. east of the north end of Looncry Lake" (Gabrielse, 1963, p. 34) may be equivalent to the gabbroic sills of the Kechika Group east of Lapie Lakes. Finally the facies in the Cassiar Mountains occur in the same northeast-southwest sequence as they do in the Pelly Mountains. Fossils from the Kechika Group in the Cassiar and Pelly mountains are similar (compare Appendix I with Gabrielse, 1963, p. 38-39). The Cambro-Ordovician strata of the Cassiar and Pelly mountains may have once been parts of a continuous belt but further study of the intervening Wolf Lake and Jennings River map areas, is required to demonstrate this continuity[19].

The phyllitic rocks in northeastern Finlayson Lake map area are like the Ram formation and can be traced southeastward discontinuously through Frances Lake (Blusson, 1966[20]) and into Watson Lake (Gabrielse, 1967, Abbott, 1977) map areas. Ordovician to Devonian chert and slate exposed still farther northeast in Sheldon Lake and Nahanni map areas constitute the Road River Formation (Jackson and Lenz, 1962). Although the two are partly time-equivalent and presumably interfinger their transition remains to be studied.

The name Kechika Group has also been applied to Cambro-Ordovician rocks in the northern Rocky Mountains implying identical stratigraphy to the type area in the Cassiar Mountains (Taylor and Stott, 1973). Strata in the northern Rockies include facies not seen in the Kechika Group of the Cassiar and Pelly mountains. Nevertheless the graptolitic shale of Taylor and Stott (1973) resembles the Magundy formation and their argillaceous limestone probably corresponds to the Ram formation. The Kechika Group of the Pelly and Cassiar mountains are much more alike than either is to the assemblage with the same name in the northern Rockies..

---

[18] This paragraph refers to stratigraphy of the Cassiar Mountains, defined by Gabrielse, 1998.

[19] Cassiar platform strata exposed in southeastern Wolf Lake map area were described by Lowey and Lowey (1986) and Murphy (1988). In northeastern Jennings River map area they are described by Nelson (1993).

[20] These rocks were subsequently regionally mapped by Murphy et al. (2001, 2002)

The Cloutier volcanics resemble basalt and related volcanics that are probably Ordovician in the Dawson map area (unit 4 of Green, 1972; unit 5 of Tempelman-Kluit, 1970a[21]). Near Dawson the volcanics are extensive and can be traced westward to Tintina fault. In the White Mountains north of Fairbanks on the southwest side of Tintina fault are rocks that are probably their offset equivalents. They are Ordovician to Early Silurian basalt, called the Fossil Creek Volcanics (Mertie, 1937; Chapman et al., 1971). The volcanics near Dawson and Fairbanks both occur with strata like other facies of the Kechika and are considered correlatives of the Cloutier facies (Fig. 14).

Grey phyllite, tuff and basalt, hosts to most of the zinc- lead deposits of the Anvil Range (unit 3 of Tempelman-Kluit, 1972b), are considered directly equivalent to the Groundhog and Cloutier facies of the Kechika Group and are therefore presumed to be roughly comparable in age[22].

Askin group (Siluro-Devonian)[23]

Introduction

The Askin group is a resistant carbonate and sandstone that forms most of the spectacular scenery in the Pelly Mountains. It is here named for Askin Lake in north central Quiet Lake map area southwest of Tintina fault where excellent exposures occur. Strata of the Askin group are distinctive, fossiliferous and well exposed. For these reasons and because they occur about the middle of the stratigraphic section, they are a geologist's bonanza and have helped the mapping immeasurably. The Askin group is Silurian and Devonian and correlates closely with the Sandpile Group of the Cassiar Mountains (Gabrielse, 1963). Strata of similar lithology and equivalent to the Askin group also occur northeast of Tintina fault within and beyond the project area (Fig. 18). Considering the magnitude of right-lateral displacement on the Tintina fault it is debatable whether the term Askin group should be used for strata on both sides, but this usage is adopted here[24].

Subdivision

The Askin group comprises four laterally equivalent, northwest-trending formations underlain by a basal siltstone that can be traced across most of the area (Fig. 19) and partly capped by grey limestone. The Askin group is also laterally equivalent to the Danger formation of the Harvey group. From northeast to southwest the subdivisions of the Askin group are:

- the **Porcupine formation**, a well-bedded shallow-water platform carbonate;
- the **Barite Mountain** formation, a more massive dolostone;
- the **Hogg formation**, an orthoquartzite and dolomitic sandstone, and
- the **Nasina formation**, a sequence of black graphitic siltstone and slate with interbedded quartzite and limestone.

The basal siltstone is referred to as the **Platy Siltstone formation**. Andesitic brecccia that occurs locally within the siltstone forms the **Orange Volcanics member** of the Platy Siltstone formation. The partly capping carbonate is called the Grey Limestone formation.

Internal relations

The Hogg formation occurs over a broad area. The Hogg and Nasina formations interfinger and are evidently roughly coeval. Localities where interfingering is seen are northwest of Mount St Cyr, on the east side of northern Quiet Lake and north of Indian Mountain, as well as north of McEvoy Lake (see Fig. 18). The Hogg, Barite Mountain and Porcupine formations each contain lithology characteristic of the others and are therefore considered to be laterally gradational. For example, the Hogg formation at Mount Hogg contains interbedded dolostone like that of the Porcupine and Barite formations and conversely, sections of the Porcupine formation in Porcupine syncline contain orthoquartzite like that of the Hogg formation. Lateral gradation of the Barite formation into the Danger formation can be seen just southwest of Halfmoon Lake and the Danger formation also contains minor interbedded siltstone, orthoquartzite and dolostone like parts of the Askin group.

Northeast and southwest of Cassiar platform, and locally on it, a dark grey marine limestone is distinguished within the Askin group. It is called the Grey Limestone formation and contains Middle and Upper Devonian fossils. The member is time equivalent to dolostone that generally constitutes the upper Askin group. The limestone and dolostone may interfinger, but this relationship was not seen. The McDame Group of Gabrielse (1963) contains equivalents of this limestone. The upper part of the Danger formation is also a thin-bedded to platy, dark grey, crinoidal calcarenite with the same Middle and Upper Devonian fossils. The presence of similar limestone in both the Harvey and Askin group provides a tie between them.

[21] Subsequently known as the Dempster volcanics (Roots, 1988) , these are part of an arcuate belt of Cambro-Ordovician tholeiitic to alkalic submarine volcanic centres (Goodfellow et al., 1995).
[22] This correlation is only correct in broad time-equivalent sense. The Anvil Range stratigraphy is described in Pigage (2001, 2004).
[23] The name Askin group was used previously by Campbell (1967) based on a report in preparation that was never published. The name therefore remains informal.
[24] The northeastern corner of Finlayson Lake map area has been referred to as McEvoy platform, a belt of Siluro-Devonian shelf rocks outboard of Mackenzie platform (Gordey, in press)

The relations of the Platy Siltstone formation to higher parts of the Askin group vary. Locally the Platy Siltstone formation is gradational with the overlying formations. For example, in the Twin Lakes section (Fig. 20) the siltstone grades upward into the overlying Porcupine formation (sandstone and dolostone) through an interval of several tens of metres without apparent break. On the other hand near McConnell Peak an angular discordance of several degrees exists between the Platy Siltstone and the overlying Hogg formation (Fig. 23c). At the McConnell River section (Fig. 20) the Barite Mountain formation lies discordantly (as much as l0°) above the Platy Siltstone (Fig. 22k), but elsewhere along its trend the two formations appear conformable. The orange weathering volcanics that are a member of the Askin group are conformable within the Platy Siltstone formation near Hoole River, but unconformable below the Porcupine formation (Figs. 21a, b, 22k). Although its thickness varies the Platy Siltstone is rarely missing and where it is, facies variation can explain its absence as convincingly as unconformable relations. The unconformity between the upper dominantly carbonate formations of the Askin group and the Platy Siltstone is therefore local and probably related to rotation of fault-blocks that were active in Early Devonian time. The local unconformity may correspond with an apparent lowermost Devonian faunal break (Fig. 13). Only two collections of fossils that are certainly Lower Devonian have been discovered in platformal strata of the Askin group (Collections 8A and 29; Appendix I).

There is no evidence that the unconformity between the Platy Siltstone and overlying unit extends into the basinal rocks northeast and southwest of Cassiar platform, although faunal control is lacking. The Danger formation northeast of the Cassiar platform is a thick, black, graphitic slate and siltstone without abrupt lithologic changes, part of Selwyn basin. It contains Lower Devonian fossils at four places (Collections 25A, 26, 42A and 8; Appendix I). Similarly the Nasina formation (part of the outer shelf) is an uninterrupted dark siltstone, quartzite and slate sequence. Both the Danger and Nasina formations are equivalents of the Platy Siltstone and higher formations of the Askin group.

Northeast of Tintina fault near McEvoy Lake similar relations are found. There the Hogg formation and underlying Platy Siltstone formation may be disconformable, as indicated by the abrupt contact. No such pronounced contact is seen in the adjacent Nasina facies, which is an unbroken succession.

A local unconformity within the Askin group on Cassiar platform, which dies out laterally into equivalent basinal strata, is compatible with the evidence of Gabrielse (1963) and Taylor and Stott (1973). In McDame map area, on the extension of Cassiar platform, Gabrielse (1963) suggested that the lower and upper parts of the Sandpile Group are separated by an unconformity. In the northern Rocky Mountains, Taylor and Stott (1973) found that the Lower Devonian Muncho-McConnell formation (the correlative of the Porcupine formation) and the Silurian Nonda formation (equivalent to the Platy Siltstone) are separated disconformably.

Platy Siltstone formation (map unit SP)

The Platy Siltstone formation is dominated by buff, brown or tan weathering, moderately resistant, thin bedded dolomitic siltstone. Fresh surfaces are light grey. Beds are a few centimetres thick and the rocks are commonly finely laminated (Fig. 21h). Quartz, sericite, feldspar and tourmaline, the detrital constituents, are cemented and partly replaced by carbonate. Quartz grains are coarse silt to very fine sand-sized and subrounded, but grain outlines are commonly masked by quartz grain interpenetration so that grain boundaries are sutured. Near Askin Lake the Magundy formation grades upward into the Platy Siltstone and in the transition black graptolitic slate becomes progressively more silty and dolomitic (Fig. 21c). The amount of graphite and fine clayey material decreases so that the rocks gradually lose their black colour.

Orange Volcanics member

In a narrow strip southwest of St Cyr fault, and near Ragged Peak, the Platy Siltstone includes prominent rusty/orange weathering volcanic rocks in its lower part called the Orange Volcanics member. These occur as discontinuous lenses up to 40 metres thick (Figs. 21a, b). They are resistant and on fresh surfaces are pale green and locally maroon. The rocks are mainly breccias, but include some flows (Figs. 21d, f, g). Breccia fragments contain phenocrysts of altered plagioclase with chlorite clots in an altered matrix. The fragments are enclosed in a carbonate matrix and cemented by calcite which partly replaces and fills veinlets in fragments. The matrix carbonate, a biosparite, commonly includes remains of crinoids, corals and brachiopods (Fig. 21e). The volcanics are probably andesitic in composition and resemble volcanics in the Magundy, Groundhog and Cloutier formations.

Porcupine and Barite Mountain formations (map units SDP and SDB)

The Porcupine and Barite Mountain formations (Fig. 19) are dominated by very resistant, thick bedded, sugary dolostone with varied amounts of detrital quartz, sand and silt. Where such detrital quartz is lacking the rocks are thick bedded and light grey but weather to pale grey and buff colours. With increasing sand and silt the rocks grade to dolomitic sandstone and siltstone and weathering colours deepen to yellow and orange buff shades. The two formations are alike, but the Porcupine formation is generally thicker. Rocks assigned to Porcupine formation are restricted to the Porcupine syncline above the Porcupine thrust, whereas the name Barite Mountain formation is used for similar and time equivalent rocks below the thrusts northeast and southwest of Porcupine syncline (Fig. 18). The Barite Mountain formation is therefore a widespread unit while the Porcupine formation represents a geographically and structurally confined part of it[25]. Both formations are named after features near where they are well exposed, Porcupine Creek and Barite Mountain.

The Porcupine formation includes six units or members of which the lowest three are widespread and mappable in most of the Porcupine syncline. The upper three are less common. The six units vary in thickness and they change laterally so that light grey dolomitic mudstone is replaced by sandy or silty equivalents and vice versa. Upwards from the base the six members of the Porcupine formation (Figs. 22e, j) are: a light grey dolomitic mudstone; an orange weathering silty dolostone; a second light grey dolomitic mudstone; a yellow sandy dolomitic mudstone; an orthoquartzite; and a yellow dolomitic mudstone (see Porcupine measured section, Appendix 2). In the Barite Mountain formation light grey dolomitic mudstone, like the lower and third members of the Porcupine formation, predominates, but sandy and silty beds, like higher members of the Porcupine formation, are present. Though these intervals are as much as 100 metres thick in places they are laterally discontinuous over distances on the order of a kilometre or two.

Most of the light grey dolostone of the Barite Mountain formation and of the two lower members of the Porcupine formation is fine- to medium-crystalline dolomitic mudstone or dolomicrite. Beds are about one metre thick and bed boundaries are partings without shale concentrations or marked compositional differences (Fig. 22d). Upper parts of many beds are more coarsely crystalline than their lower parts. Bedding is generally well defined, but more prominent from afar than up close (Figs. 22a, b, c, d). Most of the dolostone is unfossiliferous, but crinoidal and/or coral floatstone with crinoid columnals or corals in dolomitic mud are fairly common. Much of the dolomite is finely laminated with slightly wavy, parallel, laterally discontinuous layers, about a millimetre apart, which are considered to be cryptalgal (Figs. 22h, i). The laminae reflect subtle compositional differences and are most obvious on weathered surfaces. The dolomite is vuggy with randomly distributed, irregularly shaped cavities up to a centimetre across, partly filled with drusy or sparry dolomite; locally the vugs make up five per cent of the rock's volume (Fig. 22g). Birdseye and fenestral structures, which are bedding-parallel dissolution cavities thought to reflect algal origin, are also common. Irregular syneresis cracks are seen on some bedding surfaces in the dolomite (Fig. 22f). Horizontal burrows and vertical tubelike borings, generally marked by slight colour differences, are common in some beds.

The upper three members of the Porcupine formation consist of a medium-bedded, parallel-laminated and cross laminated yellow brown rock that ranges from a dominant sandy and silty dolostone to dolomitic sandstone or siltstone. In the sandy or silty dolostone 5 or 10 per cent of the volume is made up of rounded, fine sand to coarse silt-sized (0.2 to 0.5 mm) quartz grains in a matrix of medium or finely crystalline dolostone. In the dolomitic sandstone, detrital quartz constitutes more the half the volume and the dolostone is interstitial among the quartz grains. In many beds the quartz is concentrated in laminae or layers separated by relatively quartz-free dolostone. Such layers are several centi-metres thick. Silt and sand also commonly define thin cryptalgal laminae, being concentrated in layers a few grains thick. Small scale channelling and cross beds with foresets up to 20 centimetres high are common in sandy beds, and symmetrical or oscillation ripples are seen locally on bedding surfaces. In the quartz-rich beds horizontal burrows are common on the bedding surfaces.

Quartz sandstone or orthoquartzite in the upper Porcupine formation, like that of the Hogg formation, contains medium to fine sand-sized, well rounded, monocrystalline quartz cemented by quartz and dolomite. The rocks are thick bedded (Fig. 23e) commonly crossbedded, resistant and weather dark grey.

Yellow dolostone at the top of the Porcupine formation resembles carbonate lower in the unit, but weathers yellow brown and includes dolomitized crinoid and brachiopod floatstone and wackestone without the characteristic mudstone. The rocks are medium to thick bedded and like the other rocks they are mud-cracked and laminated.

Between Pass Peak and Groundhog Creek, and at the head of Groundhog Creek, the Barite Mountain formation is extensively replaced by secondary, coarse crystalline, sparry dolomite or ankeritic dolomite. Similar secondary red dolomite is seen in more restricted areas of the Barite Mountain formation along the upper reaches of McConnell River. Such replacements are lacking in the Hogg and Porcupine formations. These altered rocks are massive, resistant and weather to bright red colours. Many primary features including bedding, lamination and fossils are obliterated. The altered zones are irregular in shape and crudely controlled by bedding; they extend for several

[25] The critical reader recommended these units be considered as a single formation.

kilometres in plan without noted preferred orientation and generally replace the entire Barite Mountain formation from bottom to top. The alteration appears equally intense throughout and the transition from replaced to unaffected rocks is abrupt. Though it is a brighter red, the alteration resembles the secondary orange dolostone of the Ketza group, and the two probably originated the same way. The red secondary dolomite may be much younger than deposition of the Barite Mountain formation or it may be roughly contemporaneous. Thus it may be roughly coeval with the Jura-Cretaceous deformation of the rocks, or genetically related to the Mississippian extrusion of the Felsic Volcanic formation or may be as old as Devonian. Critical relations to distinguish between these possibilities were not recognized.

Internal Correlation. Because they include identical rocks of generally the same age the Barite Mountain, Porcupine and Hogg formations are considered lateral equivalents. However it is difficult to correlate from section to section in these beds (Fig. 20) and the thickness of the formations varies drastically from place to place. This may mean that the three are not precisely equivalent, and/or that parts of these formations were removed by erosion prior to deposition of the overlying Black Slate formation of the Seagull group. In particular the Barite Mountain formation may be equivalent to only the lower member (the light grey dolomitic mudstone) of the Porcupine formation instead of the entire formation. The same fossils, however, occur in all three formations. Although the fossil control is not detailed the faunal range of the Barite Mountain formation corresponds with that of the whole Porcupine formation, not only with that of the lower member (Figs. 12 and 13). Further, no stratigraphic break is indicated between the Askin group and Black Slate formation (overlying Seagull group). The Barite, Porcupine and Hogg formations are therefore considered lateral time-equivalent units, their lithologic variation represents facies variation from place to place and their thickness changes reflect depositional, and erosional, changes..

Depositional Environment. Dolomicrite of the Askin group formed on tidal flats, mainly under intertidal and supratidal conditions as indicated by the cryptalgal laminae, borings and dessication features. Fluctuation from intertidal to supratidal deposition probably occurred repeatedly and the Porcupine and Barite Mountain formations probably are a series of upward shoaling sequences (James, 1979). An expansive tidal flat is envisioned because the bulk of the rocks are mudstones, which indicate relatively low energy environments. Orthoquartzite beds in the upper Porcupine formation represent more active depositional conditions such as a beach. Quartz sand and silt in the dolomitic mudstone may be partly blown in, across the tidal flat from the shore-face zone, but most of the siliciclastic dolomitic mudstone probably reflects subtidal to intertidal deposition.

On the larger scale the Platy Siltstone formation and the three lowest members of the Porcupine formation represent two upward shoaling sequences in which the two dolomitic mudstones are predominantly supratidal and the siltstones mainly subtidal. The three upper members of the Porcupine formation reflect generally subtidal to intertidal conditions perhaps signalling the advent of normal deep marine conditions represented by the Black Slate formation (overlying Seagull group). The Grey Limestone Member also formed under generally marine conditions transitional to the basinal Black Slate formation.

In plan the Barite Mountain and Porcupine formations represent an extensive, carbonate mud-dominated, tidal flat with equivalent beach and slope clastics on the southwest (Hogg and Nasina formations) and basinal shaly rocks on the northeast (Harvey group). Thus the Hogg formation is probably the deposit of a relatively high energy zone on the outer side of the tidal flat, whereas the Nasina formation, farther southwest, may be the open marine slope sequence dominated by siliciclastic detritus and calcarenite. On the northeast the Danger formation also is a marine to slope succession with depositional conditions like the Nasina formation, but restricted from open water by the extensive tidal flats of the Porcupine and Barite formations to the southwest.

Hogg formation (map unit SDH)

The Hogg formation is dominated by resistant quartz sandstone and dolostone. The sandstone weathers a distinctive silvery white to light grey and is white on fresh surfaces. It is characterized by beds several metres thick. Parallel lamination, spaced at a centimetre or two, that results from differences in the amount of carbonate cement is common. Large scale, planar crossbedding with foresets up to a metre thick is seen throughout (Fig. 23h). Mud cracks, vertical burrows, and symmetrical ripple marks are also common (Figs. 23e, g).

The Hogg formation in most places is divisible into four superposed units (Figs. 23a, b; see Mount Hogg section, Appendix 2). Lowest is a thick bedded, light grey to buff dolostone 20 to 30 metres thick. It is overlain by 50 to 60 metres of dolomitic sandstone, followed by 30 metres of orthoquartzite which, in turn, is covered by 30 metres or more of sandy dolostone and dolomitic sandstone. Variation in the proportion of carbonate and quartz in the three upper sandy units result in lateral gradations, so that the units are locally separable. The lowest dolostone is not gradational with the sandy units and in most places constitutes a distinct member. It may be equivalent to the lower dolomitic mudstone of the Porcupine formation.

Locally the lowest dolomitic member is truncated, apparently by erosion, beneath the overlying sandy units (Fig. 23c). This implies that in places the Hogg formation includes a depositional break. Probably this break represents an interval of erosion following tilting of local fault blocks.

Cyclic units, each made up of orthoquartzite-dolomitic sandstone couplets dominate the sandy part of the Hogg formation. Each couplet has a lower orthoquartzite of massive to crossbedded, bioturbated and horizontally burrowed, clean sandstone about a metre thick. The upper unit, about two metres thick, is thinly laminated, dolomitic sandstone or sandy dolostone with abundant bioturbation, soft sediment deformation and mud cracks. The couplets may represent upward shoaling cycles in which the clean sandstones were the active intertidal or beach deposits, with the dolomitic sandstone the product of less agitated supratidal conditions.

Quartz sandstone is made up of extremely rounded and well-sorted medium- to fine-sand-sized, monocrystalline and unstrained quartz grains cemented by optically continuous quartz overgrowths (Fig. 23f). The sandstone also contains rare rounded tourmaline and zircon grains. Locally the quartz grains and cement are replaced by rhombs of dolomite.

Dolostone near the base of the Hogg formation, like that of Porcupine and Barite Mountain formations, weathers buff and forms regular well-developed beds several metres thick. The beds are commonly massive, but locally parallel lamination on a millimetre scale is well developed. The dolostone is a fine crystalline mosaic of dolomite rhombs less than 0.1 millimetre across, which probably replaced original carbonate mud. Dolostone interbedded with sandstone commonly contains well rounded detrital silt- or sand-sized quartz grains as well as albite and muscovite, the same constituents as the quartz sandstone. The detrital grains are matrix supported and mostly concentrated in laminae. They generally make up less than half of the ck's volume. The laminae in the dolostone, on which the detrital grains are concentrated, may be of algal origin.

Foreset beds were measured at 117 places in a stratigraphic section on Mount Hogg (Fig. 24). No systematic variation in orientation with stratigraphic position was noted and the foresets are used as a single data set. They have bimodal orientation and are interpreted as tidal foresets. The majority dip west-northwest (outgoing tides?); fewer dip southeast (incoming tides?). The foreset orientation implies a west-dipping paleoslope during deposition of the Hogg formation but gives no clue about the regional transport direction of the sand.

The source of detrital quartz in the Hogg formation is problematic. The sandstone is widespread along the west margin of Cassiar platform, but none occurs to the east in Selwyn basin or on Mackenzie Platform. A source directly to the east is therefore improbable, if not impossible. Paleogeographic reconstruction of the northern Cordillera by restoring the Cretaceous or early Tertiary strike-slip on Tintina fault shows that the quartz sand was probably transported along the west side of Cassiar platform. If the quartz was derived from the early Paleozoic North American craton, it may have been transported west across the carbonate platform about the latitude of northern British Columbia and thence north, perhaps by long-shore currents, along Cassiar platform. This interpretation is shown in Figure 25. Quartz sandstone is prominent in the Lower Devonian Wokkpash and Muncho-McConnell formation (Taylor and Stott, 1973) of northeastern British Columbia, where the sand was carried west across the carbonate platform.

Nasina formation (map unit ODN)

The Nasina formation is a dark grey to black graphitic sandstone exposed in southwestern Quiet Lake map area. It occurs mainly near the north end of Quiet Lake, near Sandy Lake and to the northwest near the mouths of Grey and Scurvy creeks. Good exposures are in stream cuts and on ridges between Brown and Grey creeks. Scattered outcrops are known in northern Finlayson Lake map area about 20 kilometres east of Weasel Lake and particularly north and south of McEvoy Lake.

The dominant and diagnostic rocks of the Nasina formation are recessive weathering, dark grey to black, platy, thin bedded siltstone or fine grained quartzite (Fig. 26c). They are generally calcareous or dolomitic and contain much graphite which gives the unit a distinctly sooty appearance in specimens and from a distance. The Nasina formation includes discontinuous lenses of interbedded, fine grained impure quartzite and silty slate as well as dark grey bioclastic limestone and minor grey dolostone (Figs. 26a, b, c). Most such lenses are generally of mappable size, several kilometres long and one or two hundred metres thick. The rocks weather dark grey and black. Siltstone and fine grained quartzite of the Nasina formation resemble that of the Platy Siltstone formation, but are dark grey or black instead of brown. The detrital constituents, mainly quartz and less feldspar are the same as those in the Platy Siltstone formation. Quartz grains are intergrown and detrital boundaries are modified and sutured; the rocks are partly replaced by calcite or dolomite in irregularly shaped patches. Calcite- filled fractures and veinlets are common. Locally brownish siltstone like that in the Platy Siltstone formation is interbedded with the dark grey siltstone and quartzite that is more characteristic of the Nasina formation (e.g., on the ridge twelve kilometres north of the north end of Big Salmon Lake). The quartzite and dolostone in lenses within the Nasina formation, like the large lens five kilometres southwest of Big Salmon Lake, is like that in the Hogg formation and is interpreted as tongues of that unit in the Nasina. Dark crinoidal limestone, which forms tongues up to 50 metres thick, is considered equivalent to the Grey Limestone formation.

Siltstone of the Nasina formation generally forms thin, parallel beds with laminations about one centimetre or less, but small scale crossbedding is common. In many places a closely spaced slaty cleavage is developed at various angles across the compositional layering.

Grey Limestone formation (map unit muDG)

The Grey Limestone formation, named for its characteristic rock type, lies at the top of the Porcupine and laterally equivalent Barite Mountain formations in the Askin group (Figs. 7b, 19). Like the Platy Siltstone formation it was probably deposited across the narrow facies belts defined by other formations of the Askin group, but unlike the Platy Siltstone the Grey Limestone is limited in extent. Outcrops are known above the Porcupine formation north of Fox Mountain and above the Barite Mountain formation at the head of McConnell River; still other restricted occurrences are above the Hogg formation near Mount McConnell (Fig. 27). The spotty map distribution may reflect the discontinuous distribution of the member over a wide area, or possibly that the unit was not recognized in some places. The Grey Limestone formation also occurs southwest of the St Cyr fault near Halfmoon Lake in Finlayson Lake map area.

Characteristically the Grey Limestone formation is a bluish grey, resistant weathering, grey, medium- to thin-bedded, fetid, bioclastic and sparry limestone. It also includes resistant black and orange-yellow weathering, massive thick bedded, finely sugary dolomite. Where the characteristic grey limestone was not mapped the unit is included in the Barite Mountain and Porcupine formations.

The Grey Limestone formation was measured only in the McConnell River section where it is 150 metres thick. Similar thicknesses are present in other places where the formation was recognized. The unit probably represents local, accumulations of bioclastic limestone within laterally equivalent dolostone.

Thickness

The thickness of the Askin group varies as dramatically from place to place as do its facies (Figs. 18, 20). The Porcupine formation near Porcupine syncline consists of 1000 metres of dolostone with minor siltstone, orthoquartzite and volcanic rocks. Along trend the measured thickness of this formation is 400 metres at Hoole River, 600 metres at Twin Lakes, 200 metres at Ram Creek and 400 metres at Askin Lake. The thickness of the Hogg formation at Mount Hogg is 500 metres, while at McConnell Peak, 10 kilometres southwest, it consists of 250 metres of sandstone and dolostone. The volcanic rocks in the Askin group are generally 100 metres or less thick, but are substantial in section Hoole River #1 (Fig. 20).

The thickness of the Platy Siltstone is more consistent than that of other units of the Askin group. At Askin Lake 500 metres were measured, at Ram Creek 200 metres, at Twin Lake 70 metres, at Mount Misery 150 metres, at Mount Hogg 100 metres, and at McConnell Peak 250 metres. At the western edge of Quiet Lake map area the thickness of the Nasina formation is estimated to range from 500 metres near Dunite Mountain to 1500 metres south of Big Salmon Lake. On the northeast side of Tintina fault the Hogg formation is estimated to be 500 metres thick; the Platy Siltstone about 200 metres and the Nasina formation 500 to 1000 metres. Equivalent rocks of the Harvey group, the Danger formation, may be 1000 to 1500 metres thick.

Relations between the Askin and Kechika groups

Close to and southwest of Tintina fault, the platy siltstone strata at the base of the Askin group lie on black slate of the Magundy formation (Figs. 21a, c, 22k). In many places (e.g. east and west of Askin Lake, near Ram Creek and at the Hoole River section; Appendix 2) the two units grade into one another so that black slate is replaced up section by platy siltstone over an interval of several tens of metres. Furthermore, graptolite collections from several measured sections (Ram Creek, Askin Lake, Hoole River) in this zone are complete enough to show that no large depositional break exists between the groups. (See fossil collections 17A-J, 52A-J, 123A-J, Appendix I).

To the southwest relations between the two groups differ. Near Mount Hogg, for example, the Platy Siltstone formation lies abruptly above phyllite and slate without gradation. No fossils have been recovered from the slate, but the siltstone is Lower Silurian to Wenlockian. Possibly the relationship there is disconformable or unconformable as shown in Figure 19. Such a relationship is also required to explain the absence of Cambro-Ordovician strata between the Nasina formation and Lower Cambrian rocks in the Big Salmon Range. The data therefore suggest that the Askin and Kechika groups may be unconformable on the southwest, but that they are conformable on the northeast. There is no data to determine if this unconformity bevels the section progressively, indicating broad tilting, or if it ends abruptly. The latter would imply that Late Ordovician faults controlled deposition there. The possible fault or hinge line separating bevelled strata from those with conformable relationships is shown in Figure 15.

Relations between the Askin and Seagull groups

The Askin group is overlain in different places by two distinct units of the Seagull group, the Black Slate and Felsic Volcanic formations. The Danger formation, the northeastern equivalent of the Askin group, is also overlain by the Black Slate formation. In the Porcupine syncline, above and below the Porcupine thrust, the Porcupine formation is overlain (conformably?) by black slate with minor chert-granule grit and siliceous tuff. The relationship is well exposed. The same slate and grit lies above the Barite Mountain and Hogg formations at other localities. Although evidence of discordance is lacking on the outcrop scale in those places and although the Seagull group generally rests on the Askin group, the two groups are unconformable on the divide between McConnell River and Porcupine Creek. There the Black Slate formation truncates several members of the Barite Mountain formation (Fig. 28). Some of the thickness variation of the Barite formation is the product of unconformable truncation, but some probably resulted from changes in depositional thickness. The general thinness of the Barite Mountain formation in the Cloutier thrust sheet may reflect stratigraphic truncation. If so, general northeastward bevelling beneath the Seagull group is indicated. More probably the unconformable truncation is related to locally tilted fault blocks. In this way profound variations in thickness of the Askin group can be reconciled with its general presence below the Seagull group. Relations between the Danger and Ankeritic Slate formations, the presumed equivalents of the Barite Mountain and Black Slate formations in the Harvey group, are not clear. These two formations appear conformable, but critical data are lacking.

Near the headwaters of Seagull Creek felsic volcanic rocks predominate in the Seagull group, and there the Barite Mountain formation is overlain directly by the volcanic rocks. Because the volcanics are massive the relationships are not clear, but there is no obvious stratigraphic truncation of the Askin group. The volcanic rocks were apparently extruded on the carbonate rocks as submarine flows with related flow breccias. A 20-metre-thick gossan commonly marks the contact between the volcanic and carbonate rocks near Seagull Creek. It probably represents alteration related to the present weathering cycle. Whether it also marks a relic soil developed on the carbonate during an Upper Devonian subaerial stage, before volcanism, is not known.

Age of the Askin group

.Most facies of the Askin group are fossiliferous and contain brachiopods and corals. The Platy Siltstone is dominantly graptolitic, but also has shelly fossils. Collections of diagnostic fossils from the Askin group have been identified by T.E. Bolton, B.E.B. Cameron, D.J. McLaren, B.S. Norford, A.W. Norris, R. Thorsteinsson and T.T. Uyeno, paleontologists of the Geological Survey. Appendix I includes identifications and correlations of the fossils from the written reports of those experts. Figure 12 shows the range in age assigned to the fossils and the map-unit from which they were collected. The diagram is intended to show differences in the degree of fossil- control of various units as well as their age range.

The Askin group ranges from Lowest Silurian to Upper Devonian[26] and contains a preponderance of Silurian and Middle Devonian fauna with few of Lower Devonian age (Figs. 12 and 13). Its oldest fossils, from the Platy Siltstone, are Landoverian (e.g., Collections 44B and 43); its youngest, from the Porcupine and Barite Mountain formations, are Upper Devonian. The Platy Siltstone ranges through the Silurian, from Landoverian through all or part of the Ludlovian. The thick carbonate units of the Askin group, the Porcupine and Barite formations, are dominantly Lower and Middle Devonian. However, part of the Porcupine and Barite formations is Silurian, ranging from Landoverian or Wenlockian - through Ludlovian (e.g., collections 89, 114, 115, 151, 90, 99, 87, 118, 1, 50A, 9, 152, 28, 8b, 43, see Appendix 1 and Fig. 12).

The Hogg formation is probably Lower Devonian (Collection 108, Appendix I), but parts are Upper Silurian (Wenlockian, Collection 103, Appendix I). Finally the orange weathering volcanics are also Silurian (Upper Landoverian to Wenlockian; Collection 88a, Appendix I). Therefore, although the Platy Siltstone is broadly Silurian and the carbonate and sandstone formations generally Devonian, all facies were evidently deposited synchronously for a time in the Late Silurian (about the Wenlockian and Ludlovian) (Fig. 29).

The Grey Limestone is Middle and Upper Devonian, same as the upper Barite Mountain and Porcupine formations, based upon its numerous coralline and shelly fossils. One fossil collection (Collection 111, Fig. 12 and Appendix 1) bears directly on the age of the Nasina formation, but the postulated lateral relations with the remainder of the Askin group, mainly the Hogg formation, imply that the Nasina formation ranges through the Devonian and probably extends also through the Silurian into the Ordovician. A poorly preserved biserial graptolite, not reported in Appendix I, was found in black graphitic slate assigned to the

[26] That any part of the Askin group ranges into the Upper Devonian can be questioned. Only one locality of this age is known from either of the Barite Mountain and Porcupine formations (nos. 36 and 150, Appendix 1), and both of these are based on the identification of a single genus Cyrtiopsis. A third locality (4, Appendix 1), from Tay River map area, is from beds different from and likely erroneously assigned to the Barite Mountain formation (Gordey, in press). Likewise, the only Upper Devonian locality reported for the Grey Limestone formation is from northwest Tay River area where strata there are lithologically unlike Askin group and have been more recently mapped as the Earn Group (Gordey, in press).

Nasina formation at Slate Rapids on the Pelly River. It indicates a Middle Ordovician to Early Silurian age for part of the Nasina formation.

The fossiliferous limestone lenses in the Nasina formation can probably be correlated with the Grey Limestone formation and with certain of the crinoidal limestones in the Danger and Ankeritic Slate formations (Harvey group).

Correlation of the Askin group

The Askin group includes a variety of interbedded and laterally equivalent carbonate and clastic strata that demonstrate on a restricted scale how some of the varied mid-Paleozoic strata of the northern Cordillera might be interrelated.

Correlations of the Askin group with other strata in the northern Cordillera are shown diagramatically in Figure 14. Although direct continuity is not necessarily implied, approximate time equivalence and close lithologic similarity is suggested in these correlations.

Many of the subdivisions of the Askin group can be matched with strata in the Sandpile Group of the Cassiar Mountains. Fossils in the two groups are also alike. Norford (1962), who made a detailed study of the fauna in Sandpile Group, also described many of the fossils that also occur in the Askin group. Although the two groups are correlatives they have important differences. The Sandpile Group is dominated by cherty dolomite (Gabrielse, 1963[27]), a carbonate with lenses and pods of chert that may be depositional or diagenetic. Such cherty dolomite is lacking in the Askin group except in outcrops five to ten kilometres south of Halfmoon Lake. Other strata of the Sandpile Group are like the Barite Mountain and Porcupine formations. Lithologies in the Platy Siltstone formation of the lower Askin group also occur in the lower Sandpile Group (Gabrielse, 1963, p. 12).

The main difference between the Askin and Sandpile groups is in their age range. In addition to numerous Silurian fossils like the Askin group the Sandpile contains Upper Ordovician forms (Gabrielse, 1963, p. 46-48) not known in the Askin group (Fig. 30). Gabrielse reported no Devonian fossils from the Sandpile Group, but considered it may range into the Devonian System. In contrast the Askin group contains abundant Middle Devonian and some Upper Devonian fossils. The upper part of the Askin group contains the same fossils as the McDame Group (Gabrielse, 1963) and is its time equivalent, but it is mostly dolostone like lower parts of the Askin group. The McDame Group on the other hand is lithologically like the Middle and Upper Devonian dark grey limestone, found only locally in the Askin group.

Stratigraphic relationships between the Sandpile and Kechika groups in the Cassiar Mountains are like those between the Askin and Kechika groups. Gabrielse (1963) noted conformable relations on the northeast, but an unconformity on the southwest in McDame Synclinorium.

Unit 8 (Green, 1972) in the southern Ogilvie Mountains[28] is a varied, thick dolostone similar to the Porcupine and Barite Mountain formations, but lacks sandstone like the Hogg formation. Unit 8 contains a varied shelly and coral line fauna and ranges from Lower Ordovician, through the Silurian, and possibly into the Lower Devonian (Fig. 30). The unit also contains intercalated Upper Ordovician or Lower Silurian tuff resembling the orange volcanics in the Askin group. In Ogilvie Mountains strata that are time equivalent to the upper part of the Askin group are divided into three dark grey limestones, units 10, 11 and 12 that resemble the restricted grey limestone facies of the upper Askin group and the McDame Group. Units 8, 10 and 11 are conformable and unit 12 is the lateral equivalent of all three. Unit 11 contains Middle Devonian fossils resembling those in the Askin group, but unit 12, has many unique fossils that range from Upper Silurian to Middle Devonian.

The Askin group is time- equivalent to units 8, 10, 11 and 12 in the Ogilvie Mountains and to the Sandpile and McDame groups of the Cassiar Mountains (Fig. 30). In both those areas Middle and Upper Devonian strata are represented mostly by dark grey limestone that is distinct from the underlying dolostone. In the Pelly Mountains such limestone occurs as a restricted facies of the Askin group and most Middle and Upper Devonian rocks are dolomitic.

In the Northern Rocky Mountains, the Nonda (Silurian), Muncho-McConnell (Lower Devonian), Wokkpash (Lower Devonian), Stone and Dunedin (Middle Devonian) formations (Taylor and Stott, 1973) are coeval with the Askin group (Fig. 31). Specifically the Nonda is time-equivalent to the Platy Siltstone, but the two are lithologically unlike, as the Nonda is a platform dolostone and sandstone. The Muncho-McConnell formation resembles the Porcupine and Barite formations and the Wokkpash is a sandstone like the Hogg formation. The Dunedin is directly correlative with the grey limestone in the upper Askin group. Although one to one correlation of all the northern Rocky Mountains subdivisions with those in the Askin group is not possible, the similarities are more compelling than the differences.

[27] This report is superceded by GSC Bulletin 504 (Gabrielse, 1998). The Sandpile Group is re-defined as Sandpile Formation (lower) and Ramhorn Formation (upper).

[28] Now defined as the Bouvette Formation (Morrow, 1999); this is a widespread unit of the Mackenzie Platform.

Distant southern correlatives of the Askin group, and more specifically of its Nasina formation, may include the Black Stuart formation in the Cariboo Mountains (Campbell et al., 1973) and the Triune, Ajax and Sharon Creek formations in the Kootenay Mountains (Fyles, 1964). Both include dark grey argillaceous rocks and quartzite and are roughly mid-Paleozoic.

The Askin group is partly coeval with the Tolovana Limestone (Mertie, 1937, Chapman et al., 1971), a massive thick bedded carbonate of Silurian age in the White Mountains north of Fairbanks, Alaska (Fig. 14). Like the Askin group the Tolovana lies on Cambro-Ordovician slate, chert and volcanic rocks including the Livengood Chert and Fossil Creek Volcanics. The Tolovana is therefore equivalent in age and lithology to the Silurian part of the Barite and Porcupine formations and is correlative with the Nonda Formation. Devonian strata may be absent from the Tolovana limestone (Mertie, 1937). If so, strata equivalent to the upper part of the Askin group above the local unconformity are absent in the White Mountains.

Harvey group

The Harvey group, exposed in a five to ten kilometre wide strip between the Tintina and St Cyr faults (Fig. 32d), includes four dominantly slaty formations (Figs. 7b, 16, 19, 33). The two lower units have correlatives in the Cassiar platform: the Danger formation is probably equivalent to the Askin group, and the Canyon formation probably continues into the Kechika Group. The upper two, the Siliceous Slate and the Ankeritic Slate formations, are probably equivalent to the Seagull group. Detailed correlations are discussed within each formation description.

Canyon formation (map unit ЄOc)

Excellent and easily accessible exposures of the Canyon formation are seen in the canyon of Lapie River upstream of the bridge-crossing along the Canol Road. The rocks are exposed in a narrow strip to the northwest and southeast of the Lapie River where it emerges from the Pelly Mountains into Tintina Trench (Fig. 15).

The rocks are resistant and weather to a distinctive reddish brown colour. They are thin bedded to laminated on a scale of centimetres. Few beds are thicker than a half metre. Medium grey finely recrystallized limestone with brownish grey phyllite and lighter grey micaceous quartz phyllite partings make up the unit. Phyllite predominates over limestone and makes up somewhat more than half the unit. Although inhomogeneous on outcrop scale the Canyon formation is remarkably consistent throughout the region. The phyllite is fine grained, with quartz, muscovite and biotite seen in thin section. Locally the phyllite is dark grey and probably graphitic. The micas are preferentially oriented giving the rock a good foliation. Bedding and the micaceous foliation are coincident and contain nearly isocinal folds at microscopic to outcrop scales.

The thickness of the Canyon formation was not measured because of its numerous small-scale folds, but is estimated to be between 500 and 1500 metres.

The rocks of the Canyon formation are metamorphic, in contrast with its unmetamorphosed probable equivalents in the Kechika Group. Before metamorphism, the Canyon formation may have resembled the Ram formation.

Danger formation (map units OD-S, SD-R)

The Danger formation consists of two lithologically distinct units, a lower black, calcareous Sooty Slate member, and an upper buff calcareous siltstone and orthoquartzize, the Mount Ross member. The formation, named after Danger Creek where it is well exposed, is confined largely to the block between the Tintina and Kumquatly faults, but also occurs immediately southwest of that block. Although the width of the exposed Danger formation is only about 5 kilometres its strike length is 120 kilometres, from the Glenlyon Range, northwest of the project area, to the eastern edge of Quiet Lake map area. Exposures of the Danger formation lie a few kilometres southwest of the Lapie River bridge on the South Canol Road.

As with other formations of the Harvey group, the connection between Danger formation and other Pelly Mountains strata is speculative. The Danger formation is roughly equivalent to the Askin group (Fig. 19). The transition from the Askin group to the Danger formation is abrupt across the St Cyr fault with little or no lithologic interfingering, but near Halfmoon Lake (Fig. 32e) beds like those of the Danger formation are interbedded with strata south of St Cyr fault more characteristic of the Barite Mountain formation.

The Sooty Slate member (SD-R) resembles parts of the Nasina and Magundy formations. It is dominated by thin bedded, graphitic, rich black, slightly calcareous slate. Minor interbedded platy, medium to dark grey crinoidal and micritic limestone is seen in discontinuous beds, a metre or two thick (Fig. 32c). A few massive beds of grey orthoquartzite or quartz sandstone, each about a metre thick, occur more commonly. The slate is recessive and weathers black; it lacks the rusty streaks of the equally black-weathering but younger Siliceous Slate formation. The sooty slate is intricately folded and transposed on a generally southwest-dipping cleavage, like rocks of the Canyon formation, but unlike that unit the sooty slate is not appreciably metamorphosed.

The Mount Ross member consists of moderately resistant, orange-buff weathering medium- to thin-bedded calcareous quartz siltstone and fine grained sandstone. The rocks are commonly crosslaminated and bedding partings are marked by thin brown shale. The member includes massive beds of orthoquartzite. Broadly it resembles the Platy Siltstone formation and parts of the Hogg formation as well as the Hoole formation. The rocks are deformed, but folds are larger and more open than those in the sooty slate, presumably reflecting the difference in competency.

Internal and external relations of the Danger formation are speculative; relations shown in Figure 7b are considered the most probable. The Sooty Slate member is thought to be the lowest part of the Danger formation and the Mount Ross member presumably overlies it. On the north side of Mount Ross the two units are in contact, but may be a fault relationship. On the ridges between Danger Creek and Lapie River immediately southwest of Tintina fault the Sooty Slate member overlies the Canyon formation. The contact is considered stratigraphic, but may be separated by a fault. On the south flank of Mount Ross the Mount Ross member is overlain (stratigraphically?) by the Ankeritic Slate formation, but northwest of Lapie River between the Kumquatly and Tintina faults the opposite relationship exists. Northwest of Lapie River the contact is thought to be a thrust, folded over the Danger anticline. The thrust is assumed to lie below the Mount Ross member and above the Ankeritic Slate formation. Most probably the Mount Ross unit is the upper member of the Danger formation and a northeastern equivalent of the Platy Siltstone and/or Hogg formations.

The Sooty Slate member is estimated to be roughly 500 metres thick and the thickness of the Mount Ross member is similar.

The Sooty Slate includes a limestone with distinctive "two-holer" crinoids which are possibly Emsian (Lower Devonian, Collection 26). If the relations suggested in Figure 7b are valid most of the Danger formation is older than these fossils and it probably ranges through the Ordovician, Silurian and Lower Devonian.

Ankeritic Slate formation (map unit DMAS)

The Ankeritic Slate formation includes about 700 metres of distinctive, rusty orange weathering, phyllitic slate with crinoidal limestone beds and lenses of vesicular volcanics. The unit weathers recessively, but its limestone and volcanic rocks are resistant compared with the slate. The formation is named for its distinctive and dominant rocks, which are moderately recessive phyllitic slate with rusty weathering pits which are remnants of ankerite or ferrodolomite filled "vesicles" one or two millimetres across (Fig. 32a, b). Some of these rusty spots have remnant pyrite centres. The rusty spots are randomly scattered through the rock and locally make up 5 or 10 per cent of their volume. The slate weathers to orange-ochre colours and is grey brown on fresh surfaces. Its regular lamination, enhanced by weathering, results from slight compositional differences between laminae. The slate also contains disseminated rhombs of dolomite (or another carbonate).

The Ankeritic Slate formation is well exposed in a moderately southwest dipping section on the ridge leading southwest from the peak of Mount Ross. Lowest in the sequence there is about 300 metres of thinly laminated limy slate that weathers ochre and dark grey. The lower beds apparently overlie the Mount Ross member of the Danger formation conformably, but this relationship is not certain. The limy slate is overlain by about 200 metres of ochre-brown weathering, grey, regularly and thinly laminated silty argillite with phyllitic partings. Another 200 metres of ochre to rusty weathering, dark grey slate, with rusty ankeritic pits after pyrite, are next, and followed by about 100 metres of thin bedded, platy, calcareous shale, or grey shaly limestone. The sequence is capped by 20 metres of fetid, grey, medium bedded, crinoidal biosparite which includes crinoids with star-shaped and pentagonal axial canals. On trend and about 5 kilometres southeast of Mount Ross, rusty orange weathering volcanics form a 100-metre-thick lens near the top of the formation, perhaps as a tongue within the upper slate. The upper crinoidal limestone of the Ankeritic Slate formation is conformably overlain by the Cherty Tuff formation.

About four kilometres northwest of Mount Cook the upper crinoidal limestone is thicker than on Mount Ross and fully 50 metres are present. There two volcanic lenses each 20 metres thick, separated by about 10 metres of crinoidal biosparite, replace the single volcanic lens and rest above the limestone. In that place the succession lies conformably beneath the Black Slate formation.

The limestone and volcanics in the upper part of the Ankeritic Slate formation are laterally discontinuous, but the distinctive thinly laminated ankeritic slate is continuous and diagnostic. Volcanic rocks in the Ankeritic Slate formation range from massive amygdaloidal flows to volcanic breccia. They are most probably intermediate to basic submarine flows and flow breccias.

The age of the Ankeritic Slate formation is Devonian and Lower Mississippian, based on numerous conodonts and other fossils recovered from crinoidal limestone in the unit (Figs. 12 and 13). The fossils range from Early Devonian (Emsian, Collections 47A and 25A) through Middle and Upper Devonian (Collections 60, 22c, 59 and 42B) into the Carboniferous (Collection 62). The slate is conformable beneath the Cherty Tuff and Black Slate formations and its upper part may be roughly time equivalent to the Black Slate formation. The range of fossils overlaps with those in the upper Askin and Lower Seagull groups; the Ankeritic Slate is thought their equivalent. Crinoidal limestone and volcanic lenses in the upper part of the Ankeritic Slate are probably equivalents of the limestone lenses within the Felsic Volcanic formation from which Lower Mississippian fossils were recovered. Although the upper beds resemble strata of the Seagull group the lower slate is unlike anything in the Askin group and their lateral equivalence is based on the fossil evidence. Regionally the Ankeritic Slate formation has no known lithologic equivalent.

A bright orange weathering slate occupies a 500-metre-wide strip that was traced for 65 kilometres along the north side of St Cyr fault through most of Finlayson Lake map area (Fig. 32d). Although it is lithologically like the Ankeritic Slate formation it is mapped separately because its weathering colour is several shades brighter orange than the rest of the formation. The position of this bright orange weathering member within the Ankeritic Slate is unknown, but fossils from a thin crinoidal platy limestone (Collection 84) in it are Lower Devonian (Emsian) just older than the Grey Limestone formation in the upper Askin group.

Siliceous Slate formation (map unit DMss)

The Siliceous Slate formation is confined to a narrow panel between the Tintina and Kumquatly faults. It is dominated by moderately resistant uniformly black, graphitic, non-calcareous slate that weathers black with rusty streaks. The slate is thin bedded and laminated, but because the contrast between beds is generally slight outcrops commonly appear massive. Closely spaced cleavage across bedding is common. The rocks carry up to 5% disseminated, finely crystalline pyrite locally concentrated along laminae. Because the general orientation of bedding is unknown the thickness is estimated roughly as several hundred metres. The formation is named to emphasize its non-calcareous nature which contrasts with the underlying Danger formation.

Generally the Siliceous Slate formation resembles the Black Slate formation, but it lacks interlayered volcanics and it is even more graphitic. Occasionally some interbedded slate like that of the Ankeritic Slate formation is seen and one or two dark grey crinoidal limestone lenses like those in the Ankeritic Slate were noted. Relations to other units are uncertain because the formation is generally faulted. Northeast of Mount Ross in the zone between the Tintina and Kumquatly faults the Siliceous Slate apparently overlies the Danger formation conformably, but this relationship is open to interpretation.

Collections of conodonts from the Siliceous Slate formation, one from a crinoidal limestone, the other from a silty limestone are Upper Devonian (Famennian?; Collections 195, 202). On this basis and on the tenuous stratigraphic evidence and lithologic similarity to the Black Slate formation (Seagull group) the Siliceous Slate is considered Mississippian. It is therefore roughly time equivalent to the Ankeritic Slate formation and even though no transitional relations between the two units were seen, they are assumed to be correlative (See Figures 7b and 33).

Depositional Conditions of the Harvey group

The lithology generally reflects conditions of deeper water deposition than for probable time equivalents to the southwest. While strata of the Ankeritic Slate, Siliceous Slate and Danger formations were laid down water depth was probably more than for correlative beds of the lower Seagull and Askin groups. The lithology of most of the Harvey group also reflects limited coarse clastic input. Little of the quartz sand in the Mount Hogg formation is found in the Mount Ross Member, its equivalent in the Harvey group. The chert conglomerate found locally in the Seagull group, is also absent from the Harvey group. Volcanic rock, common in the Kechika and Seagull groups and in the lower part of the Askin group are uncommon in the Harvey group. Crinoidal limestone in parts of the Harvey group may be the deposits of crinoidal debris flows derived from a shallower water source. The shaly Harvey group is considered the basinal equivalent northeast of more platformal beds. This basinal element is the southwest part of Selwyn basin, displaced by Tintina fault, while the shallower area is Cassiar platform. Selwyn basin is a 500 kilometres long, 200 kilometres wide northwest trending area of shaly Lower and Middle Paleozoic rocks that lies between time equivalent carbonate beds of the Mackenzie and Cassiar platforms. It occupies most of the Selwyn Mountains and is largely northeast of Tintina fault. Most of its shaly rocks are included in the Road River Group,

Seagull group

Introduction

The name Seagull group is introduced here for a Mississippian assemblage of slate, greywacke and felsic volcanic rocks that overlie the Askin group mainly conformably but locally discordantly. The volcanic part of the group is well exposed east of the upper reaches of Seagull Creek, from which the name is taken. Slate, volcanic rocks, trachyte and siliceous tuff are mapped separately as informal formations in the group. Wheeler et al., (1960a, 1960b) first noted that the volcanic and slaty rocks are equivalent and suggested a genetic link to several nearby syenite bodies.

Distribution

Southwest of Tintina fault strata assigned to the Seagull group are exposed in a 60-kilometre-wide strip parallel to the fault (Fig. 34). Volcanic and trachytoid rocks are restricted to an area of 40 x 100 kilometres within this belt. Elsewhere slate and siliceous tuff predominates. Northeast of Tintina fault the Seagull group includes slate and volcanic rocks northeast of Finlayson Lake (Fig. 34).

Volcanic rocks of the Seagull group attain a thickness of about 700 metres between Seagull and Cloutier creeks and near the headwaters of McConnell River. Away from there they are thinner and are replaced by slate, minor greywacke and cherty tuff that form a blanket about 300 to 500 metres thick. In Finlayson Lake map area northwest of Tintina fault the slate is also about 300 metres thick.

Subdivisions and internal relations

The Seagull group is subdivided according to the scheme of Figure 33. The Black Slate formation at the base is the lowest incompetent black unit above light coloured carbonate of the Askin group. It forms a nearly continuous cover, up to 500 metres thick, over the Askin group and is itself overlain by volcanic rocks. The Black Slate formation is well exposed in stream cuts of Ram Creek tributaries. This is taken as the type locality. The Felsic Volcanic formation contains intercalated black slate, for example, on Peak 7001', and the Black Slate includes volcanic lenses, for example, on Mount Green. Though generally superposed with conformable relations, the two formations interfinger and are considered time equivalent. The Felsic Volcanic formation, about 500 metres thick in places, is well exposed in the high area between Cloutier Creek and McConnell River and on Peak 7001' (its type area).

The Cherty Tuff formation of the Seagull group is laterally equivalent to the Felsic Volcanics. It is generally about 100 metres thick and overlies the Black Slate gradationally and conformably. In most places it constitutes the upper part of the Seagull group. Good exposures occur four kilometres north of Mount Green, the type area for the formation.

Trachyte is a lithologically distinct part of the Seagull group found in a restricted area (Fig. 34). One twelve-kilometre-long body straddles the upper McConnell River and two smaller plugs, a kilometre or two across, are shown on the maps but smaller bodies are not. The trachyte rarely exhibits clearly intrusive relations with the surrounding rocks. Commonly it interfingers with the Felsic Volcanics and is texturally gradational with them. It is considered the subvolcanic part of the volcanic assemblage and is mapped as a member of the Felsic Volcanic formation. Bioclastic crinoidal limestone lenses, a few metres thick, are found at various places in the Felsic Volcanic formation. They are too small to be mapped at the present scale of investigation.

Northeast of Tintina fault the Seagull group is dominated by the Black Slate and Cherty Tuff formations. Volcanic rocks of intermediate to felsic composition, considered equivalent to the Felsic Volcanic formation, are restricted to a small area north of Wolverine Lake. There the Cherty Tuff overlies the Black Slate conformably as it does southwest of Tintina fault. Relations between the Cherty Tuff and Felsic Volcanics, are assumed to be gradational north of Wolverine Lake.

Black Slate formation (map unit DMBS)

The Black Slate formation is dominated by thin bedded, black, fissile slate with a pervasive, closely spaced cleavage, generally at angles to bedding. Lamination, defined by subtle colour contrasts and slight grain size variation, is common. The rocks locally range to slaty siltstone and contain rare beds of dark greywacke up to a metre thick. The rocks weather recessively to a rich blue-black colour and in most places are covered by vegetation. The best outcrops are recently cut stream banks. The slate is generally siliceous and locally grades to slaty chert which is paler grey than the slate. Pyrite, a common constituent, occurs as disseminated cubes and locally makes up five per cent of the rocks.

Greywacke, interbedded with the slate, includes a high proportion of subrounded chert grains, much quartz, minor feldspar and mica and rare altered volcanic grains. It is a fairly well size-sorted, medium- to coarse-sand-sized wacke, which locally contains coarser chert varying from granules to pebbles. The granule wacke grades laterally into chert conglomerate. The greywacke-conglomerate forms resistant beds in the slate, about ten metres thick and laterally continuous for hundreds of metres.

Resistant lenses of intermediate volcanic rocks, up to 50 metres thick, are common within the Black Slate formation between Porcupine syncline and St Cyr fault. These flow rocks are equivalent to, and lithologically like, the Felsic Volcanics.

The Black Slate resembles the Magundy formation but is more siliceous, blue-black weathering and not "sooty" like the Magundy formation. The associated rocks also help distinguish the two units; greywacke and chert conglomerate are not known in the Magundy formation.

In Porcupine syncline above Porcupine thrust the Black Slate formation reaches a thickness of 250 metres, but below the thrust northeast of the syncline this unit is 500 metres thick. In the Cloutier thrust sheet around McConnell River the Black Slate formation is only about 100 metres thick. Above the Seagull thrust near Groundhog and Upper Sheep creeks the formation is about 300 metres thick. In the McConnell thrust sheet east of McNeil Lake about 500 metres are known. Southwest of the St Cyr fault, in eastern Quiet Lake map area, only about 150 metres of the Black Slate formation are seen.

Barite member. Thin-bedded barite occurs in the upper part of the Black Slate formation at several places. This member is lenticular and individual occurrences can be traced for one or two kilometres. Near Mount Cook a lens about 1000 metres long and 100 metres thick lies some 200 metres above the base of the Black Slate formation and 50 metres below the Cherty Tuff. The upper and lower contacts are gradational with the black slate and similar relations are postulated where the lens terminates laterally in the Black Slate. The barite consists of alternate laminae, a millimetre to a centimetre thick, of very finely crystalline, light grey barite and dark grey siliceous slate or chert. Laminations are commonly wispy and discontinuous, and lenticular chert nodules locally assume the place of the siliceous layers. Aside from the Mount Cook area, the Barite member is known on the northeast side of Porcupine syncline, near McNeil Lake and near the mouth of Cloutier Creek, but as the member is a stratigraphic horizon, it probably occurs more widely. At several places small nodules or thin wispy barite was noted in the upper part of the Black Slate formation.

A dark grey, discontinuous limestone bed up to a metre thick and made up of coarsely crystalline black calcite appears near the base of the barite at several localities.

Cherty Tuff formation (map unit MC)

The Cherty Tuff formation includes dark grey, greenish grey and pale, apple green, tuffaceous chert and siliceous tuff that forms a resistant, rusty orange weathering cap above the Black Slate formation. Beds are about five centimetres thick and separated by argillaceous partings a few millimetres to tens of centimetres thick. The rocks are a cherty textured, extremely fine grained mixture of chlorite and siliceous material in a wide range of proportions. Where the chloritic material predominates the rocks are slate or tuff and where the siliceous component is dominant they are chert. Pyrite, in small cubes, is a common constituent.

The rocks locally have a widely spaced fracture cleavage, but lack the slaty cleavage of the incompetent Black Slate formation. Generally the Cherty Tuff formation close to the Tintina fault is about 100 metres thick, but in southern Finlayson Lake map area 200 to 300 metres are present.

Felsic Volcanic formation (map unit MFV)

The Felsic Volcanic formation is a heterogeneous unit, about 500 metres thick, which includes a variety of light coloured, moderately resistant tuff, breccia and flow rocks, dykes, sills and plugs, whose interrelations were not mapped in detail. The breccias, tuffs and flows form interlayered units which are several metres or tens of metres thick and not generally prominent. Units are distinguishable by grain-size differences and compositional variation. Individual units are laterally discontinuous and sections more than a kilometre or two apart generally cannot be correlated. A section measured at the head of Cloutier Creek is given as an example of the variation and heterogeneity of the Felsic Volcanic formation (Fig. 35a).

On fresh surfaces the rocks are pale grey or near white to light, medium and dark green, medium grey and locally brown; they weather to various light colours. Brilliant orange gossans are locally prominent. In places black siliceous slate, several metres thick, is interbedded with the Felsic Volcanics.

The tuffs range from extremely fine grained siliceous varieties like the Cherty Tuff to lapilli and bomb breccias. Individual tuffs include angular volcanic fragments of a range of sizes, but generally of fairly restricted mineralogy and texture. By contrast fragment variation between units is extreme.

Flow rocks and fragments in the tuffs are made up of abundant subhedral perthitic feldspar and albite with minor quartz, hornblende, biotite and opaques with grain size generally less than one millimetre (Fig 35b). The texture varies from porphyritic (K-feldspar), to trachytoid, to irregularly intergrown or felted.

The feldspars are commonly saussuritized and the mafic minerals are replaced by chlorite. Just as the fine grained tuffs grade into the Cherty Tuff formation the trachytoid felsic volcanics grade into the Trachyte Member.

Fine grained tuffs locally have an irregular spaced cleavage along which chlorite is recrystallized.

Trachyte member (map unit MFV_T). The trachyte is a resistant, medium grained, equigranular rock that weathers into irregular blocks up to a metre across. Euhedral light grey tablets of perthitic K-feldspar make up more than half the rock volume. Another quarter consists of albite or andesine laths. Dark green biotite (10%) occurs as clusters of tiny flakes, pseudomorphous after an original interstitial amphibole (Fig. 35c). The rocks commonly contain as much as 5% intergranular mauve fluorite. Apatite and opaque minerals, locally galena, are accessory minerals in the rocks. The trachyte is locally partly replaced by dolomite, which occurs as scattered tiny rhombs. The feldspar is variably sericitized; some grains are entirely and evenly altered others are affected only in zones. The trachyte is "granite" textured with randomly oriented feldspar tablets. Locally the tablets are crudely aligned and give a trachytoid texture.

External Relations

In the northeast-most outcrop area of the Seagull group (parallel to Tintina fault) the contact between the Seagull and Askin groups is sharp, and possibly conformable. In this area the Askin group is always present beneath Seagull group strata and variation in the rocks below may reflect depositional differences rather than stratigraphic truncation. Farther south near Mount Hogg and McNeil Lake similar relations hold generally, but near Indigo Lake Gordey (1977) showed that the Black Slate formation lies directly on the Platy Siltstone and at one place on Kechika Group strata, which implies unconformable relations there. Block rotation on faults active before or during deposition of the Seagull group (i.e., during Late Devonian time) best explains the relations between the Askin and Seagull groups. The rapid facies variation within the Askin group may also have been fault controlled during deposition of that unit.

The locus and nature of such older faults is uncertain because most may have also been active during the Jura-Cretaceous foreshortening of the Pelly Mountains. The faults are assumed to have had normal movement. The Seagull fault (Fig. 34) juxtaposes the Felsic Volcanic and Black Slate formations and may be such an old structure, active during the Late Devonian or Early Mississippian, and reactivated later. The White Creek and McNeil faults may also have moved as normal faults about this time. Along the northeast boundary of the Felsic Volcanics they interfinger with the Cherty Tuff and Black Slate; this boundary is assumed to be unfaulted. The Felsic Volcanic formation evidently accumulated as submarine and locally subaerial flows and tuff in a half graben or unequal graben with a southwest wall higher than the northeast wall. The southwest wall may have been the Seagull-White Creek-McNeil fault system.

In the area 20 kilometres west-southwest of the community of Ross River, The Black Slate formation conformably overlies limestone of the upper Danger formation. The relations are exposed about five kilometres northwest and two kilometres southeast of Mount Cook (Quiet Lake map, Sheet 1, in pocket). Farther northeast, nearer Tintina fault, the Black Slate and Cherty Tuff formations are replaced by two distinctive slate units that contain fossils of roughly the same age as the Seagull group and that have the same conformable relations to the Danger formation. They are the Ankeritic Slate and Siliceous Slate formations of the Harvey group, facies equivalents of each other and of the Seagull group. The extent of these formations across strike is small compared to that along strike.

The boundary between the time equivalent Seagull and Harvey groups coincides with that between the Askin group and Danger formation and that separating the Kechika Group and Canyon formation. This boundary, close to or at, the St Cyr fault, is an abrupt facies front in Upper Cambrian to Mississippian rocks. It may be localized on a long-lived fault system that marks the northeast edge of Cassiar platform in the Pelly Mountains. The fault, indicated on Figure 34, probably had Paleozoic normal movement which maintained the southwest side above the northeast. The profound change in facies in all the Paleozoic rocks across the fault may be evidence also for Mesozoic transcurrent movement on this fault. The St Cyr fault may be localized along such an old, long-lived fault. This hypothetically old fault marking the northeast side of Cassiar platform extends northeast of Tintina fault (Tempelman-Kluit, 1980a) through Glenlyon, Tay River, Sheldon, and Finlayson Lake map areas and where it juxtaposes different strata of the same age.[29]

Chemistry.

Analyses of ten samples of the Felsic Volcanics from Finlayson Lake map areas were reported by Gordey (1981a,b). These analyses show that the rocks are subalkaline to alkaline and anomalously rich in potash; the volcanics are calc-alkalic trachyte or rhyolite. Morin (1977) gave an AFM plot and alkali-silica diagram for 36 samples of the Felsic Volcanics from all parts of the project area, which corroborates these conclusions. Morin further reported high barium (1813 ppm), thorium (31 ppm), and fluorine (1643 ppm) and low uranium (0.6 ppm) concentrations in these rocks.

Mortensen (1979) included a third set of analyses of these volcanic rocks along with an exhaustive discussion. His data corroborated Gordey's and Morin's, but he emphasized the metaluminous nature and extensive alteration of the volcanic rocks and presented evidence of important alkali exchange with sea water during extrusion. He concluded that the major and minor element chemistry of the Felsic Volcanics compares closely to that of the Kenyan Rift zone.

Origin

The Trachyte, Felsic Volcanics and Cherty Tuff are the products of one alkalic volcanic event. The trachyte bodies are interpreted as subvolcanic plug domes intruded into the platformal Askin group during the Mississippian. Other trachytes were extruded as submarine flows while the Black Slate formation was deposited. Plugs and dykes were injected along faults

[29] This fault was discounted by the critical reader and has not been recognized in subsequent geological mapping.

and fractures that cut Devonian and older strata. These faults were active in the Mississippian, but do not affect facies of the older strata and are not considered to be still older, reactivated faults. The tuffs and breccias are submarine volcanic flow deposits, perhaps derived as debris flows and sheet washes eroded from those plug domes that had built to the water surface. Columnar-jointing suggests the flows were extruded subaerially when the volcanic vent had built above water. The coarser breccias may have been deposited close to the volcanic centre and the finer farther away. The extensive Cherty Tuff is thought to be the most distal equivalent and may be partly a chemical deposit. The Barite member in the Black Slate is similarly thought to be a widely dispersed chemical deposit from one or more sources related to the Felsic Volcanics.

The maximum thickness of the preserved volcanic sequence, about 500-700 metres, is a minimum measure of the depth of water into which the volcanics were extruded if the volcanics built at least to the water surface, and if there was little tectonic uplift or subsidence during volcanism. The second assumption is problematic.

Age

The Seagull group is Upper Devonian (fossil collections 191, 24A, 24B, 62 in Appendix 1) to Lower Mississippian (Tournaisian and Visean). It may range into the Upper Mississippian (Namurian),.

The Seagull group contains Mississippian conodonts at many localities southwest of Tintina fault (Figs. 12, 13, 33; Appendix I). Two are from thin limestones within the Felsic Volcanics (Collections 70 and 80) and another (Collection 63) is from a limy part of the volcanics themselves. Two are from thin limestones within the Black Slate formation where it is overlain by the Felsic Volcanics (Collections 81 and 82) and several others are from thin limestones in the Black Slate about 100 metres below the Cherty Tuff (Collections 47A, B, 22A and 194). Collection 155 is from within the Cherty Tuff near Indigo Lake. The fossils were all assigned to the Early Mississippian by B.E.B. Cameron, who considered most to be Tournaisian to Early Visean. Because the upper limestone of the Danger Creek formation is Upper Devonian (Frasnian and Famennian) and Lower Mississippian, the overlying Black Slate probably does not include Upper Devonian rocks.

The Barite member is also considered Tournaisian or Visean because at Mount Cook and Porcupine syncline limestone within this unit contains Mississippian conodonts (probably Tournaisian or Visean, Collections 22A 47A, 47B; Appendix I).

Northeast of Tintina fault only one collection of conodonts (Collection 185, Appendix I) bears on the age of the Seagull group. This collection, from a thin bioclastic limestone in the lower part of the Black Slate, is tentatively assigned a Mississippian age. The fossil evidence confirms that the Black Slate, Felsic Volcanics and Cherty Tuff are time equivalent. The gradational relations of the Trachyte and the Felsic Volcanics implies that it is of like age.

A K/Ar date of 185 Ma determined for biotite from the largest trachyte body probably reflects partial argon loss from the rock during a later thermal event. Rb/Sr isochron ages have been determined for the Felsic Volcanics and Trachyte by Mortensen (1979) and by Chronic (1979). The best determination (Chronic, 1979), a mineral isochron, is 333 Ma, with an initial ratio of 0.7105, which corresponds with a K/Ar date of 319 Ma on mica from the same rock. On presently accepted time scales these dates corroborate the fossil evidence.

Correlation

From Finlayson Lake the Seagull group can be traced intermittently northwest along strike to Glenlyon map area and southeast through Frances Lake and Watson Lake map areas. Felsic volcanic rocks have not been noted in these areas, but slate, locally with cherty tuff, and commonly interbedded with, or overlain by, voluminous chert conglomerate predominate, particularly in Glenlyon map area. Strata of the Seagull group also trend southeast along strike through Wolf Lake and Jennings River map areas to the McDame area, but volcanic rocks are absent, leaving only the slate.

Some of the volcanic rocks of the Kechika Group were confused with those of the Seagull group and mapped together (e.g. Wheeler et al, 1960a,b). The Seagull group is equivalent to, and lithologically like, parts of the Sylvester Group (Gabrielse, 1963), the Earn Group (Campbell, 1967) and the "Black Clastic", an informal name for Devono-Mississippian slate and chert in the Selwyn Mountains[30].

The Black Slate formation is similar to slate in the lower part of the Sylvester Group of the Cassiar Mountains (Gabrielse, 1963). The Felsic Volcanics and Cherty Tuff associated with it in the Pelly Mountains have not been recognized in the Cassiar Mountains.

The Black Slate formation is lithologically similar to, and a time equivalent of, the Besa River formation of the Northern Rocky Mountains (Bamber et al., 1968; Taylor and Stott, 1973), but that unit includes Upper Devonian as well as Lower Mississippian strata (Fig. 36). No volcanics are known in the Besa River formation, but the western facies of the Prophet formation, which conformably overlies, and laterally grades into, the Besa River, is a chert with Meramecian

[30] On the Mackenzie Platform (autochthonous stratigraphy) this unit is now referred to as Earn Group (Gordey and Anderson, 1993; Gordey, in press).

fossils (Bamber et al., 1968), the same age as the Cherty Tuff (Visean). The Besa River and Prophet formations together are correlated with the Seagull group.

The Black Slate formation is further correlated with unit 13 of Green (1972) in the Ogilvie Mountains but the latter does not contain rocks like the Cherty Tuff formation.

In Tay River map area a sequence of black slate overlain by cherty tuff like the Seagull group occurs 25 kilometres north of Mount Mye; both are included in unit 5 by Roddick and Green (1961a). Near TwoPete Mountain about 30 km northwest these strata have interbedded limestone with Mississippian fossils (i.e., Collection 171, Appendix I). They can be followed northwest into the northeast part of Glenlyon map area, where they are included in the Earn Group. As mapped, the Earn Group probably includes strata of Ordovician to Mississippian age, partly in a facies like the Seagull group, and the two cannot be correlated directly. To the southeast the Seagull group is correlated with black slate and chert wacke in Watson Lake map area (mapped as unit 7 by Gabrielse, 1967, and as unit 5 by Abbott, 1977).

Starr formation (map unit Ps)

Thin shale and siltstone found stratigraphically between the Seagull group and the Hoole formation occupies a narrow zone close to, and southwest of, Tintina fault and occurs also south of McNeil Lake (Fig. 37), The unit is best exposed in the outcrop belt between the St Cyr and Cloutier faults about longitude 132, near the headwaters of Starr Creek from which the formation is named. The unit was separated from the Hoole formation only in later stages of the work and its distribution on the accompanying maps is probably incomplete. Locally it is about 100 metres thick.

The Starr formation rests apparently conformably on the Cherty Tuff formation. It is disconformably(?) overlain by the Hoole formation (Upper Triassic; Fig. 38a). Rocks of the Starr formation are thin bedded and thinly laminated, bioturbated shale and siltstone (Fig. 38b, c). The lamination is irregular and consists of alternate layers of brown shale, one or two millimetres thick, and somewhat thicker (2-5 mm) layers of buff to yellow fine silt. The lamination is discontinuous because it is disrupted by horizontal borings and in places is altogether absent. The unit is moderately resistant and weathers dark brown, distinctly darker than the overlying Hoole formation. The rocks are not calcareous in contrast with the siltstone of the Hoole formation. The chief detrital constituent of the silty laminae are angular quartz grains with sutured, interpenetrating boundaries. The shale layers are locally pyritic lending the rocks a rusty weathering colour.

One sample of the Starr formation contains Permian (Leonard) conodonts (Collection 85, Appendix I). The Starr formation is homotaxial with, and comparable in age, lithology and thickness to, the lower and middle parts of the Kindle formation in Racing River Synclinorium in the Northern Rockies (Bamber et al., 1968) (Fig. 36). In the southern Mackenzie Mountains shale, sandstone, limestone, and siltstone of the Mattson Formation and the Stoddart Formation further south in the Northern Rocky Mountains, is interposed between the Prophet and Kindle formations (Bamber et al., 1968). Its equivalent in the Pelly Mountains may be included in the Starr formation. If so, the Starr formation is like the Mattson, and includes Stoddart and Kindle time-equivalents.

The locality in the Pelly Mountains contains the only known Permian strata southwest of the Tintina and Northern Rocky Mountain faults that resemble those of the Northern Rockies. Their presence in a sequence that otherwise is also similar to Northern Rockies strata supports the concept that the Pelly Mountains are the offset continuation of the Northern Rockies and implies right lateral movement on the Tintina-Northern Rocky Mountain Trench fault.

In the Ogilvie Mountains, Green (1972) included strata that are time-equivalent to the Starr formation in his unit 14. That limestone and shale formation is much thicker, and although coeval, is lithologically dissimilar.

Hoole formation (map unit uTH[31])

The youngest stratified rocks in the Pelly Mountains are calcareous siltstone with minor limestone that commonly contains Upper Triassic conodonts. They occur in narrow, structurally low fault blocks which trend northwest along both sides of the St Cyr fault between longitudes 131° and 133° (Fig. 37). The type area is 25 kilometres southwest of the mouth of Hoole River. These strata are also found south of McNeil Lake and northeast of Tintina fault, 20 kilometres north and 20 kilometres west of Finlayson Lake. Most information about the Hoole formation came from its exposures southwest of Tintina fault.

In the earlier work of Wheeler, Green and Roddick (1960a, 1960b), the siltstone was considered Mississippian, because its fossils were misidentified. The siltstone unit was not named in the earlier reconnaissance and the name Hoole formation is introduced here

Thickness and stratigraphic relations

The thickness of the Hoole formation beneath structurally superposed rocks is at least 200 metres (Appendix II). The estimated preserved thickness in southwestern Finlayson Lake map area is close to 500 metres. The relatively uniform thickness and facies show that the Hoole formation was deposited as a blanket over a larger area than that wherein it is preserved.

Unlike many other rock units in the Pelly Mountains the Hoole formation is comparatively facies constant and its limited exposures do not show the rapid and profound lateral variations of some of the older rocks. Its two rock types, predominantly calcareous to dolomitic siltstone, and less important phosphatic limestone are laterally intergradational and contain identical fossils. Limestone also forms a discontinuous member up to 100 metres thick, that is confined to middle or upper parts of the formation and is only exposed close to Tintina fault. Most of the exposures of the Hoole formation near Tintina Trench are dominated by siltstone with comparatively little interbedded shale, whereas the more southwesterly exposures near McNeil Lake are shalier and thinner bedded.

In the type area and along strike the Hoole formation overlies the Starr formation apparently conformably (Figs. 38a, 39a, b). The long faunal hiatus between the Starr (Permian: Leonardian) and Hoole formations (Upper Triassic: Carnian and Norian) suggests a disconformable contact.

In the two exposures of the Hoole formation in northeastern Finlayson Lake map area underlying strata are not exposed. The map pattern there suggests that the Hoole may lie directly on the Nasina formation[32] omitting the Seagull group, but relations along strike farther northwest suggest differently. In Tay River map area, 30 kilometres north of Mount Mye, the Hoole formation[33] overlies Permian chert[34]..

In the zone adjacent to Tintina fault the Hoole formation is capped structurally by older rocks of the Kechika, Askin or Seagull groups. In the most southerly exposures, and in the two outcrops near Finlayson Lake, cataclastic rocks are thrust over the formation (e.g. the McNeil klippen).

Lithology

The Hoole formation is thin bedded (ten centimetres) and includes tan or brownish weathering limy siltstone to silty limestone with thin beds of bioclastic limestone. Generally the rocks break into platy or thin slabby talus. The siltstone is laminated on a millimetre scale (Fig. 39c) and is commonly cross laminated (Fig. 39d). Bioturbation is a general feature of the rocks with horizontal borings more common than vertical. In places the rocks have been so thoroughly homogenized that the lamination is extensively destroyed. The lamination, cross lamination and burrows are prominent and are defined by contrast in weathering colours resulting from differences in the proportion of silicate and carbonate minerals (Fig. 39c). The siltstone contains angular to moderately rounded, moderately size-sorted, silt to fine sand sized monocrystalline quartz as the main detrital constituent, with less feldspar. Muscovite, biotite and opaque minerals are less common. Tournaline and zircon are rare. Finely crystalline calcite forms matrix and typically replaces silicate grains along grain boundaries, so that original detrital outlines are obscure. Detrital grains constitute somewhat more than half the rock's volume and are generally matrix supported so that the rocks are silty limestone. Where the detrital constituents are grain supported the quartz has sutured, interlocking boundaries and calcite occurs interstitially or as replacements. In places the calcite matrix is partly dolomitized.

Grey limestone interbedded with the siltstone contains numerous phosphate nodules, a millimetre or less across, as well as fish teeth and other phosphatic debris (Fig. 39e). Together the phosphatic material may make up five per cent of the rock. In addition the limestone contains a small proportion of the same detrital grains as the siltstone. The limestone is recrystallized to finely crystalline calcite, but probably formed as micrite.

The discontinuous limestone member in the Hoole formation, found particularly near Mount Cook, is massive light grey limestone with ten per cent sand-sized quartz grains and phosphatic and bioclastic debris. It differs from the interbedded limestone mainly in thickness and debris size; the limestone is crystalline and its origin may be largely detrital rather than lime mud like the thinner beds.

Siltstone of the Hoole formation resembles the Platy Siltstone formation of the Askin group. In the former the matrix and cement are generally calcite whereas dolomite is more common in the latter. The stratigraphic association and interbedded lithologies further distinguish the units.

Siltstone and limestone of the Hoole formation represents the product of normal marine deposition on a shallow shelf at considerable water depth, but above the carbonate compensation level. None of the rocks formed in shallow water of the intertidal zone or in agitated beach environments. Although the carbonates are detrital biogenic accumulations, none are thought to have formed as reefs.[35]

Age

The Hoole formation contains few pelecypods or ammonites in Quiet Lake and Finlayson Lake map areas

[32] Northeast of Tintina Trench), the Nasina formation is now considered part of the Earn Group.

[33] Northeast of Tintina Trench, this is the Jones Lake Formation (Gordey and Anderson, 1993).

[34] The Mount Christie Formation (Gordeyand Anderson, 1993)

[35] An important map unit (uTHv) of massive, dark green volcaniclastic sandstone and minor tuff, is not described in the 1985 manuscript.

and its assigned age is based primarily on abundant conodonts recovered from the unit. The oldest conodonts are Upper Ladinian or Lower Carnian (Collections 67C, 57B, 58, Appendix I, Figs. 12, 13)[36] and the youngest are Middle to Upper Norian (Collections 83, 86A, B, C, Appendix I), Most microfossils are Carnian or Norian. In measured sections no faunal hiatus is evident (Collections 57A to F from the section north of Mount Green and Collections 68A-E from the section near Cloutier Creek, Appendix I). The limestone member contains Carnian conodonts (Collections 27A, B and 67A, B, Appendix I, Fig. 12). The few macrofossils collected (Collections 66, 133A, B, Appendix I) are Norian, but no Carnian pelecypods or ammonites were found. No conodonts were found in samples with pelecypods.

Correlation

Outside the project area the Hoole formation has no known equivalents southwest of the Tintina - Northern Rocky Mountain fault system. Its best match is the Ludington formation of the Northern Rocky Mountains northeast of this fault. It is the same age as both the Ludington and Charlie Lake formations that are laterally equivalent units (Gibson, 1975; Taylor and Stott, 1973) (Fig. 40). In addition, the discontinuous limestone member of the Hoole formation may be equivalent to the Baldonnel Formation in the Northern Rockies. This Carnian to Lower Norian limestone is equivalent to part of the Ludington Formation and rests above the Charlie Lake Formation, showing relations like those between the discontinuous limestone and the remainder of the Hoole. The Hoole formation is therefore correlated with the Ludington, Charlie Lake and Baldonnel formations.

Unit 16 of Green (1972) in the Ogilvie Mountains is Upper Triassic and roughly equivalent to the Hoole formation. In detail this correlation breaks down (Fig. 40). Most of unit 16 is lithologically like the Pardonet and contains Norian pelecypods like that formation. It is, therefore, younger than most of the Hoole formation. The Triassic in the Tombstone area (Tempelman-Kluit, 1970a), a part of the Ogilvie Mountains, apparently includes correlatives of the Toad and Pardonet formations and lacks equivalents of the intervening Liard and Ludington. Those rocks therefore are only roughly comparable in age to the Hoole formation.[37]

TRANSPORTED (ALLOCHTHONOUS) ROCKS

In the Pelly Mountains cataclastic rocks occur as klippen above autochthonous strata on both sides of Tintina fault (Fig. 6b). Rocks in the klippen differ from those in the autochthonous succession in lithology, fabric and metamorphism, and do not represent transported lateral equivalents of the autochthonous succession. Instead they are considered sequences that originated far away and subjected to a metamorphic and cataclastic history not experienced by the autochthonous strata. They were thrust above the autochthonous strata only after they had acquired their metamorphic state and penetrative structure (see Fig. 41).

The transported rocks are divided into three allochthonous assemblages which maintain a consistent structural order in the klippen. Broadly each assemblage probably represents a set of depositionally related strata that were mixed tectonically. Although now superposed the three assemblages may be as unrelated to each other in origin as they each are to the autochthonous rocks[38].

Because the lithologic units are tectonic slices bounded by faults, and not depositional bodies with stratigraphic boundaries, normal stratigraphic terminology is inappropriate. The lithologically distinct units are called 'slices'. Groups of related slices are referred to as allochthonous assemblages or allochthons. The stratigraphic relations and even the age of the various map-units are unknown; only their lithology and structural sequence is clear.

The three allochthonous assemblages are here called Nisutlin, Anvil and Simpson (as in Tempelman-Kluit, 1979), from lowest to highest in the structural stack. Although consistent in sequence, all three assemblages are not everywhere present. The allochthons pinch and swell as do individual slices within them so that any of them may be absent from the structural sequence.

Structurally lowest are metasedimentary and metavolcanic rocks grouped as the Nisutlin Allochthon. It includes muscovite quartz schist and chlorite schist. The muscovite quartz schist is laterally and vertically gradational with gritty feldspathic sandstone, which is considered its precursor. In the same way the chlorite schist can be traced into tuffaceous volcanics of intermediate composition that are presumed its protolith. Nisutlin assemblage locally includes dark grey phyllonite and slate as well as large bodies of marble and flaser limestone. The assemblage is generally sheared and recrystallized (Fig. 42), but the degree of shearing and mineral regrowth varies so that the rocks

[36] M.J. Orchard recently re-examined some of the conodont collections. He considers that several collections including 68a, 139, 143 and 144 are Middle Triassic. Of these he thinks 139, the oldest, is Anisian (Appendix 1).

[37] Not described in the original report is a Tertiary conglomerate unit, northeast of Tintina fault and southwest of the community of Ross River.

[38] The interpreteation of three coherent sheets is an oversimplification. See Monger et al. (1991), Mortensen (1992) and Colpron et al. (2006) for subsequent interpretations.

range from beds that are transposed on a well developed crenulation cleavage to mylonite, blastomylonite and mylonite schist. Gordey (1981a,b) described the rocks of this allochthon in the McNeil klippen in detail.

Structurally above the sheared siliceous rocks are amphibolite and greenstone, cherty tuff, serpentinite, gabbro and peridotite. This, the second of the three transported groups, is called the Anvil Allochthon[39]. The mafic rocks are sheared like the siliceous strata of the Nisutlin assemblage, but the strain is concentrated in zones a few metres thick and these are separated by less penetratively sheared rocks, tens of metres thick.

The third and highest group of transported rocks is dominated by chlorite granodiorite schist and gneiss, and locally its less sheared equivalent, hornblende granodiorite. It also includes hornblende quartz monzonite and other granitic and intermediate volcanic rocks. This is the Simpson Allochthon. It is sheared and metamorphosed like the other transported rocks, but the degree of strain is more homogeneous than in the Anvil assemblage.

Klippen of the allochthonous rocks are preserved in structural depressions between younger intrusions. Some outliers are only a few kilometres across, but others are so large that they extend beyond the project area (Fig. 6b). Southwest of Tintina fault most of the klippen are confined to the large structural depression between the Quiet Lake and Nisutlin batholith, but northeast of the fault an area about 50 kilometres wide next to the fault makes up part of a giant sheet truncated by the Tintina fault (Fig. 6b). The klippen are generally sub-horizontal, but where invaded by younger plutons they are arched and dip steeply.

Basal contacts of the klippen are abrupt and sharp, with no structural mixing of the autochthonous underlying strata and overlying allochthons. The transported strata are generally sheared and their flaser fabric is broadly sub-horizontal, like the klippen themselves.

The klippen are considered outliers of a once extensive sheet or set of sheets thrust northeastward above the autochthonous succession during the Late Jurassic and Early Cretaceous[40].

Age of transported rocks

Five radiometric age determinations (Table 2; see also Fig. 49) on samples from the project area bear on the age and affiliation of the plutonic rocks. The two youngest, 183 ± 7 and 201 ± 7 Ma, on muscovite from strongly foliated chlorite muscovite quartz monzonite, from deformed and metamorphosed parts of the unit, indicate Upper Triassic to Lower Jurassic cooling, following strain and blastesis but give no clue as to the time of intrusion. Determinations on fresh hornblende from two widely separated samples of foliated hornblende quartz monzonite gave 316 ± 18 and 344 ± 19 Ma; these indicate that the rocks probably crystallized in the Early Mississippian. A whole rock K-Ar determination of 251 ± 10 Ma (Permian) was also made on fairly fresh intermediate volcanic rocks of Money klippe included in the Simpson Allochthon.

Outside the project area, south of Dawson, a nearly concordant-minimum U/Pb age of 276 Ma (Lower Permian) was determined for zircon in sheared hornblende granodiorite with the same field relations as the Simpson assemblage (Tempelman-Kluit and Wanless, 1980). Together the data indicates that the granitic rocks were intruded about Carboniferous time.

Depositional relations of the strata in the Nisutlin Allocthon are obscure, because the rocks are a metamorphosed and penetratively deformed structural stack, in which stratigraphic integrity is retained only locally. Relations to other groups of rocks are uncertain because external contacts are faults. The assemblage consists of texturally and compositionally immature clastic rocks and intercalated volcaniclastics of intermediate composition. None shows evidence of shallow water deposition. The "synorogenic sandstone" (Tempelman-Kluit, 1979), a rock with clasts of mylonite (Fig. 42) that is itself the protolith for mylonite, is interpreted to show that ductile deformation and deposition of at least part of the unit were synchronous. The sandstone is interbedded with impure limestone from which Late Triassic conodonts were recovered (Appendix 1, collection 182A, B); thus the sandstone and the shearing are thought to be Upper Triassic. The K-Ar dates from metamorphic micas of Nisutlin assemblage range from 170 Ma to 230 Ma and mark the time of cooling that followed shearing. This reasoning suggests that most or all of the Nisutlin Allochthon is Upper Triassic and Lower Jurassic.

Flaser limestone, which forms a prominent discontinuous tectonic unit between the Nisutlin and Anvil assemblages near Finlayson Lake, is here considered part of the Anvil Allochthon[41]. It contains Pennsylvanian conodonts (Appendix I, collection 187A, B) and is therefore older than the Nisutlin assemblage, but it is a separate tectonic slice and its depositional relations are unknown. The limestone is likely equivalent to Gabrielse's (1967) unit 9a of Watson Lake map area and Abbott's (1981b) unit CPC in Wolf Lake

[39] The Anvil Allochthon as described here is now known to contain mafic volcanic rocks of various ages (Murphy et al, 2006; Piercey, 1999, 2000) . To emphasize a first-order distinction on the accompanying maps, these rocks northeast of Tintina Trench are shown as Anvil Range group on the maps.

[40] Further description of individual allochthons is given in the structural geology chapter.

[41] Murphy et al. (2001) interpreted this carbonate within a structurally imbricated package of Yukon-Tanana Terrane.

map area[42], which have similar relations, but which are thicker and not as penetratively deformed.

Metamorphism of the allochthonous rocks

Erdmer (1982)[43] has studied in detail the metamorphic mineral assemblages of the allochthonous rocks in several parts of the project area and has determined their conditions of metamorphism. This summary includes his findings as well as those of Gordey (1981a,b).

Metamorphic minerals of the greenschist, and locally amphibolite facies of regional metamorphism characterize the transported rocks, and relic eclogite is preserved locally. Nisutlin assemblage is commonly represented by quartz, albite and muscovite, with or without chlorite, biotite and epidote. The Anvil assemblage contains actinolite and albite, with all or some of epidote, chlorite and biotite, but in places these rocks are formed by the higher grade assemblage of hornblende, plagioclase, almandine, with or without biotite or epidote. Quartz, albite, chlorite, biotite, epidote and muscovite are the common minerals in the Simpson assemblage.

Higher grade metamorphic minerals are preserved locally in the transported rocks near, but outside the project area. Notable is the eclogite near Faro, a rock with green omphacite and red garnet, formed as foliation-controlled lenses, a few metres thick, enclosed in quartz muscovite schist of the Nisutlin assemblage (Tempelman-Kluit, 1970a). The eclogite is commonly partly replaced by hornblende and chlorite. Similar eclogite has been recognized elsewhere as relic assemblages in the allochthonous rocks (generally the Anvil assemblage) northeast of Tintina fault by Greg Jilson (pers. comm., 1980). Retrograde eclogite was also recognized by Erdmer (1981) in the Teslin Suture Zone just west of the project area[44].

The relic eclogite lenses indicate that extreme conditions of prograde metamorphism were achieved at least locally in the transported rocks. The greenschist metamorphism of the bulk of the rocks may reflect the conditions of retrograde metamorphism, following such extreme metamorphism. Alternatively, the eclogite - greenschist association may result from differences in water pressure, so that the eclogite formed under relatively dry conditions and the enclosing greenschist under higher water pressure. By itself the greenschist facies does not pinpoint the pressure during metamorphism, as the facies is stable over a wide pressure at temperatures between 450° and 500°C, but the eclogite required pressure near 10 Kb, which indicates depths near 40 kilometres. The relations imply that a comparatively high pressure part of the greenschist stability field is represented by the transported rocks (Fig. 43). Temperatures near 500'C with depths between 20 and 40 kilometres may have prevailed..

Metamorphic and Deformed Plutonic Rocks

Cretaceous metamorphic and plutonic rocks underlie much of southwestern Quiet Lake map area, and central Finlayson Lake map area northeast of Tintina fault (Fig. 9). In both places the metamorphosed rocks are the Late Precambrian to Paleozoic "in place" strata within the structurally superposed "transported" cataclastic rocks. The metamorphic rocks in Quiet Lake map area are called the Big Salmon Complex (Mulligan, 1963), those in Finlayson Lake map area, the Mink complex (new name). In the main the Big Salmon Complex includes the metamorphosed autochthonous strata where as the Mink complex, at its level of exposure, also contains metamorphosed allochthonous rocks. The Mink complex therefore includes strata that are polymetamorphosed, with an early, comparatively high pressure (Early Jurassic) event overprinted by a moderate temperature episode (Early Cretaceous). The Big Salmon Complex is underlain mainly by strata subjected only to the later event (although the earlier episode is also recognizable in the transported rocks of this complex). The boundaries of the metamorphic complexes (Fig. 9) are drawn roughly at the biotite isograd, but this boundary was not mapped rigorously in the field. In places the stratigraphic units are sufficiently distinctive to be traced far into the metamorphosed complex and elsewhere, such distinction was not possible.

Plutonic rocks in the project area are biotite quartz monzonite, granite and granodiorite which are considered broadly cogenetic. They range from large, semi-concordant plutons, with metamorphic and migmatized mantles several kilometres thick, to small, cross cutting intrusions that invade metamorphosed and unmetamorphosed strata. These have their own narrow thermal metamorphic halos. Phases of the intrusions and the nature, correlation and distribution of the metamorphic screens have not been carefully studied or mapped. Separate intrusions of the composite batholiths display a variety of relations. For example, the central part of Nisutlin batholith is concordant with its enclosing rocks while its northwest part is distinctly intrusive. Smaller plutons are generally homogenous, single intrusions with few screens and xenoliths.

Although the intrusions have been dated, the available radiometric ages are between 70 and 99 Ma, whereas dates for the Black River batholith are 47 and 66 Ma. Because of their similar lithology all plutons are presumed to be Upper Cretaceous except those near Black River. Enclosing metamorphic rocks in central Quiet Lake map area give similar metamorphic ages indicating regional metamorphism in Cretaceous time.

[42] Screw Creek Limestone of mid-Mississippian and Pennsylvanian age (Abbott, 1979; Roots et al., 2002).
[43] See also Erdmer (1985; 1987) and Erdmer et al. (1998).
[44] See Creaser et al. (1999), Devine et al. (2004, 2006).

Big Salmon Complex[45] (map unit PBS)

The Big Salmon Complex was defined by Mulligan (1963) to include all the metamorphic rocks in the Big Salmon Range. Mulligan did not distinguish the allochthonous rocks, metamorphosed during cataclasis in the Late Triassic to Lower Jurassic, from the autochthonous Eocambrian and early Paleozoic strata that were regionally metamorphosed under quite different conditions during the Cretaceous. The name Big Salmon Complex is therefore restricted here to those rocks in the Big Salmon Range that were metamorphosed during the Cretaceous event regardless of their earlier history. Rocks that escaped important overprinting by the Cretaceous episode, namely the high grade metamorphic allochthonous rocks, are not properly part of the complex.

The Big Salmon Complex is exposed in two large northwest trending culminations (Figs. 9, 46). These arches, named the Big Salmon and Quiet Lake culminations are separated by a structural trough, the St Cyr depression. The cores of the culminations expose quartz monzonite of the Cretaceous Big Salmon, Nisutlin and Quiet Lake batholiths. In most places the foliation of metamorphic rocks that flank the intrusions lies parallel to the contact with the plutonic rocks, or they display a, gradational, interleaved relationship. Toward the culminations, in stratigraphically low rocks, the metamorphic grade increases and toward the structural lows; where higher strata occur, the metamorphic grade falls (Fig. 44a). Essentially unmetamorphosed strata outside the culminations grade successively through chlorite schist, biotite schist, biotite granodiorite gneiss (Fig. 44b, c), to gneiss with progressively more and larger sills and dykes of aplite and felsite into granodiorite with progressively fewer metamorphic screens and schlieren. The gradation from essentially unmetamorphosed strata, to plutonic rocks with few screens, occurs across widths of five to ten kilometres on the southwest and southeast margins of Big Salmon batholith and across five kilometres on the northeast side of the Quiet Lake batholith. As metamorphic grade increases the rocks are more penetratively deformed by small scale folds.

The Big Salmon batholith is concordant on all sides, but Quiet Lake batholith is partly transgressive and partly concordant. For example, the northeast side is concordant for most of its length, but on the eastern end is a lobe that cuts across the surrounding rocks. Nisutlin batholith similarly shows largely concordant relations along its northeast flank and around the margin of its northeastern off shoot, but the southwest margin truncates strata and the lobe on its northwest end is similarly discordant (Fig. 48).

Correlation. The Big Salmon Complex is like the metamorphic culmination in the Horseranch Range (Gabrielse, 1963) and the large metamorphic complex in central Frances Lake map area (Blusson, 1966), that expose Precambrian rocks around Cretaceous batholiths. The Horseranch Group includes schist and gneiss formed from Cambrian and Eocambrian strata, probably during the Cretaceous and metamorphism of the Frances Lake complex culminated probably at the same time. The Anvil Arch (Tempelman-Kluit, 1972b) also represents a metamorphic culmination around a Cretaceous pluton, but it exposes metamorphosed Cambro-Ordovician and younger strata.

Mink complex[46]

Augen gneiss, quartz mica schist and minor marble in central Finlayson Lake map area northeast of Tintina fault constitutes the Mink complex, named after Mink Creek, a stream where the rocks are well exposed (Figs. 9, 46). The complex was regionally metamorphosed during the Cretaceous, like the Big Salmon Complex, but the age of its protolith is not clear, because the metamorphosed rocks cannot be followed into unmetamorphosed strata, unlike the Big Salmon Complex. In places gneiss and schist of the Mink complex grades upward into the cataclastic and sheared strata of the Nisutlin Allochthon. Elsewhere lower parts of the Mink complex resemble the Big Salmon Complex and the protolith may be the same Cambrian and Eocambrian strata from which the Big Salmon Complex is derived.The Mink complex may also include older rocks that are basement to the Eocambrian strata.

The Mink complex includes three rock units mapped separately. Lowest is a coarse augen gneiss (Fig. 45) overlain by garnet mica schist with minor marble. These are in turn succeeded by quartz mica schist and amphibolite.

Metamorphism of the autochthonous rocks in the Big Salmon and Mink complexes

Autochthonous rocks of the Big Salmon Complex contain mineral assemblages of the biotite and garnet zones of the greenschist facies and of the staurolite and sillimanite zones of the amphibolite facies (Fig. 44d). The shallow, medium and deep levels of intrusion as indicated by metamorphic minerals have not been

[45] This unit extends southward to Teslin and Jennings River map area (Stevens, 1992, 1994; Mihalynuk et al., 1998) and is considered part of Yukon-Tanana terrane (Colpron et al., 2006).

[46] This section is of historical interest only. Subsequently called the Finlayson Lake district, this area contains a sequence of Devono-Mississippian sedimentary and volcanic units unconformably by Permian metavolcanic rocks. See Murphy and Piercey (1999, 2000); Murphy et al. (2006) and Piercey et al. (2001, 2004).

determined by mapping, but the metamorphic grade generally increases toward the plutonic cores of the culminations. Because the plutonic rocks likely formed by partial melting they provide a largely pressure-independent limit to the temperature the rocks attained during metamorphism (Fig. 43). The pressure conditions of the metamorphism are not closely restrained, however. If the temperature gradient was relatively normal between 20 and 30°C/km the 600°C required for partial melting would have been achieved at depths between 20 and 30 kilometres. Intrusions within Big Salmon Complex may be essentially enclosed by the rocks in which they formed, but other plutons are intrusions which probably cooled at depths near five kilometres[47].

Plutonic Rocks (map units Kqm, Kqmp)

Granitic rocks form bodies of widely different sizes and underlie much of the project area on both sides of Tintina fault. The total area of granitic rocks, 3327 square kilometres, represents nearly a quarter of the project area. Three intrusions, the Quiet Lake, Nisutlin and Big Salmon batholiths, account for about two-thirds of this. The distribution of intrusions in relation to the general geology is shown in Figure 46. This sketch map names the plutons, shows their radiometric ages, and distinguishes the semi-concordant, deeper plutons from medium and higher level intrusions. On both sides of the Tintina the large semi-concordant intrusions are generally southwest of the smaller, discordant plutons (Fig. 46). Names are taken from nearby topographic features[48]. Table 2 gives the areas of plutons and average specific gravity as determined from hand specimens.

Granitic rocks show much textural and compositional variation within and between individual intrusions. These variations were not carefully studied or mapped and the following is only a general description of their petrology. The granitic rocks are commonly resistant to weathering and form talus of large angular blocks. Slopes underlain by these rock s have a distinctive medium grey colour. Although the granitic rocks are resistant to weathering, large areas of talus restrict the amount of outcrop; it is not greater than that for most other rocks. The granitic talus is more commonly and extensively covered by black lichen than that of other rocks. The lichen darkens the general aspect of these rocks from a distance and hampers observation of texture and structure. Black River batholith is an example where the rocks are resistant and form the highest ground with many sharp ridges and peaks and steep-sided valleys (Fig. 47b). Granites of the three large batholiths are also generally resistant. Nisutlin batholith in particular has some spectacular mountains east of the Canol Road around the head of Big Creek. Quiet Lake batholith is somewhat more recessive and the southwest part of Big Salmon batholith weathers readily to gruss, making that part of the intrusion recessive. Most of the plutons northeast of the Tintina fault are recessive and occupy topographic lows, but the Fortin and McEvoy stocks are resistant. Hornfelses in the thermally metamorphosed halos around some small stocks are more resistant than the intrusions themselves (Fig. 47a, c). Such resistant hornfels halos surround bodies like the Brown, White Creek and Tintina stocks with rusty weathering peaks as high as those underlain by the granite itself.

The granitic rocks are broadly alike although there is much variation even in small intrusions. Compositionally they range from granite to quartz monzonite with less granodiorite; they are generally unfoliated, though part of the Big Salmon, Nisutlin and Quiet Lake batholiths (particularly their margins) are foliated and even gneissic (Fig. 44b, c). Foliated rocks are also seen in the North, Fyre and Money intrusions. Foliated types are generally closer to granodiorite, while those that lack fabric are compositionally closer to granite. Most of the smaller discordant plugs are homogeneous, hypidiomorphic and equigranular textured like central parts of the larger batholiths (Fig.47e, g). Parts of the Nisutlin and Big Salmon batholiths and the Red Mountain stock are megacrystic with thick tabular pinkish potassium feldspars several centimetres across making up as much as a third of the volume (Fig. 47d, f). Plagioclase ranges from albite to labradorite and is generally subhedral as twinned tabular white crystals up to five millimetres long; it is commonly zoned and has inclusions of K-feldspar and commonly muscovite. K-feldspar is generally microperthitic orthoclase (Fig. 47h), which forms anhedral or subhedral grains that enclose small crystals of plagioclase, quartz and biotite. Where porphyritic, the phenocrysts are perthitic orthoclase with Carlsbad twinning. Quartz occurs as greyish, anhedral, equant grains a few millimetres

across, interstitial to the feldspars. Quartz and K-feldspar are myrmekitically intergrown along crystal boundaries (Fig. 47h, i). Biotite, generally pleochroic in browns, and commonly the sole mafic mineral, forms ragged, corroded flakes commonly chloritized at the edges and along cleavage. A few grains of hornblende are present locally. Muscovite is common in small proportions. Zircon and sphene are general, but not plentiful accessory minerals.

[47] These statements reflect interpretations typical of the late 1970s. Modern deductions are based upon single grain zircon analyses, fluid inclusions and stable isotopes. Studies for southeastern Finlayson Lake map area include Grant (1997, Grant et al. 1996) and Devine et al. (2006).

[48] Quiet Lake batholith, Nisutlin batholith, Big Salmon batholith, Fox batholith, Dycer stock and Black River batholith are considered formal names as they are in current usage

Age[49]

The granitic rocks in the project area have given K-Ar ages between 46 and 99 Ma including three dates of 47, 53 and 66 Ma on samples from the Black River batholith and ten others in the span 70-99 Ma (Fig. 46, Table 2). The second set of dates match those obtained elsewhere in Yukon on kindred rocks and are thought to represent the time when most of the intrusions cooled, namely the Late Cretaceous. The radiometrically determined cooling dates coincide with the rise of the Quiet Lake and Nisutlin-Big Salmon arches.

A problem with the dates is their long range, a spread of roughly 30 Ma, and the variety of dates even from one intrusion. The limited sampling precludes determining whether the variations are meaningful, but the dates reflect the last time the rocks began to retain radiogenic argon. The condition for retention is generally thought to be a temperature barrier. An entire pluton may not pass through the blocking temperature at the same time, especially since the barrier is a temperature zone, different for different minerals, rather than a sharp boundary; variation in dates obtained from different samples is therefore reasonable. A pluton may also oscillate through the retention barrier as it is tectonically raised or lowered. Later thermal effects related perhaps to circulating warm water may significantly alter the age, thus the 15 Ma difference in the two ages for samples from the Quiet Lake batholith implies several possibilities. One is that the batholith is composite with older and younger parts that cooled at different times. Another is that the younger date is the thermally reset age of a rock that first cooled 85 Ma ago. Yet another possibility is that, both ages are "correct" and reflect different degrees of argon retention of two samples that passed through the argon retention barrier at different times and rates and/or have been differently reset by one or more later "thermal" or hydrothermal events. The dates are therefore not rigorously interpreted; but imply general Late Cretaceous cooling of the granitic rocks. Differences in the ages may not be related to variations in the intrusive age of the plutons.

The young dates for the Black River batholith may indicate a real difference in its emplacement age because the dates lie outside the range determined for the quartz monzonite suite regionally. But because the rock is lithologically indistinguishable from the other plutons, it can also be argued that the batholith only cooled more recently, passing through the argon retention isotherm about 20 Ma after other intrusions. Perhaps the Tintina fault controlled uplift of the Black River batholith and its K-Ar age dates its rise in the Early Eocene.

Stratigraphic evidence that bears on the age of the plutonic rocks corroborates, but does not refine the Upper Cretaceous age given by the radiometric data. The intrusions postdate thrusting and folding, which involves strata as young as the Upper Triassic Hoole formation. The oldest strata that contain clasts and feldspars derived from the granitic rocks northeast of Tintina Trench are the Early Eocene sedimentary rocks exposed near Ross River. Regionally the granitic rocks are best bracketed in the Tombstone area near Dawson where the Tombstone batholith intrudes folded and thrust-faulted Lower Cretaceous strata and yield a K-Ar age of 134 Ma (Tempelman-Kluit, 1970a). This indicates an Early Cretaceous emplacement.

Intrusive relations and origin

Relations of the Nisutlin batholith and the mantling metamorphic rocks suggest that the plutonic rocks formed in situ by partial melting and although parts have moved discordantly into the cover rocks other parts are essentially where formed (Fig. 48). Between the arms of the batholith are quartz mica schist with interfoliated lenses of marble. This sequence extends eastward where it is less metamorphosed, and at the southeast opening between the batholiths two prongs the strata are recognizable as the Eocambrian siltstone and Lower Cambrian limestone of the Ketza group. The swarm of marble lenses is traceable westward into the southern arm of Nisutlin batholith where it forms steep-dipping sheets of coarsely crystalline white marble engulfed in quartz monzonite. The marble sheets are oriented in the same way as those surrounding the batholith and have not been rotated. Lower Cambrian limestone and underlying schist reappear on the southwest edge of the batholith as a recognizable stratigraphic sequence. Apparently the siliceous country rocks were melted in place to produce the granitic rocks, but the limestone was left as evidence of this transformation. Part of the melt produced this way was apparently mobilized into its cover[50] so that contacts with the regionally metamorphosed rocks on the southwest and west sides of the batholith are abrupt and the country rocks are hornfelsed or baked. Other contacts with host rocks, notably between the arms of the batholith, are gradational with sheets of granitic rocks in the schist and screens of mafic metamorphic rocks in the granite at many places. Considering that part of the batholith is still rooted in its birth place the mobilized parts have probably not moved far, perhaps on the order of a kilometre or two. This suggests that the other plutons with mobilized characteristics, which are surrounded by the Eocambrian or Proterozoic rocks, but that lack the clear evidence of the marble lenses, may

[49] This section is mostly of historical interest. U-Pb, Nd and Ar isotopic dating has resolved late Devonian, early Mississippian and Cretaceous emplacement and metamorphic episodes. A compilation of isotopic dates to 2002 is available as Breitspecher and Mortensen (2004); and was updated for most of Finlayson map area by Murphy et al. (2006).

[50] An alternative interpretation is that these are screens of host rock engulfed in the intrusion.

similarly be close to their place of inception. The Big Salmon batholith is an example of an intrusion that is autochthonous and the northwest part of Quiet Lake batholith may be the same although the southern margin looks intrusive.

Despite their different relations, the various plutons are roughly the same age and apparently coeval with the metamorphism that affected the surrounding rocks. This implies that the large concordant plutons are in situ, ultrametamorphic melted parts of the metamorphosed stratigraphic sequence and that the smaller,crosscutting bodies represent the mobilized parts of these melted sedimentary rocks

Lamprophyre Dykes

A few dykes of dark brown weathering biotite lamprophyre (minette), generally several metres wide and traced a kilometre or two occur at widely separated localities in Quiet Lake map area. Prominent examples are five kilometres east of the Canol Road opposite Barite Mountain and on the ridge north of Groundhog Creek, eight kilometres above its confluence with the Lapie River. At both places several steeply dipping dykes, that trend generally northward but with wide variations in strike, cut the Askin group. At a third locality, on the ridge between Seagull Creek and McConnell River several kilometres north of the Mississippian syenite, two or three of these dykes intrude folded and thrust faulted rocks of the Seagull group, but are themselves undeformed. They have chilled margins and are generally planar.

Fresh biotite from a sample of the third locality was dated and gave a K-Ar age of 112 ± 4 Ma. Because the dyke is thin it probably cooled quickly and as it is unaltered its age is taken as the time of intrusion of the dyke. It is surprising that the dyke is older than the quartz monzonite batholiths and stocks some of which also postdate the deformation.

Sedimentary and Volcanic Rocks in Tintina Trench

Introduction

Three groups of rocks that are younger than the deposition, thrusting, deformation and metamorphism of the autochthonous and allochthonous rocks, and which also postdate metamorphism and intrusion of those strata by the quartz monzonite are confined to Tintina Trench. They include an Eocene immature clastic unit, a younger set of quartz feldspar porphyries, and basalt. The first are terrestrial clastics that may have been deposited during transcurrent movement in structural depressions. The second group is perhaps related to the closing stages of transcurrent movement, whereas the basalt may be connected with the latest Tertiary normal movement that formed the trench.

Ross River Conglomerate (map unit Ts)

Outcrops of slightly indurated Tertiary clastic rocks with interbedded coal lie about 2 kilometres west-southwest of Ross River along a 1-1/2 kilometre stretch of road, and on the west bank of Lapie River 5 kilometres above its mouth. The exposures occur in fault blocks whose relative offset is unknown. Although they are equivalent they can not be correlated in detail.

Outcrops along Lapie River are of massive, thick bedded, buff weathering, pale greenish conglomerate, sandstone and mudstone. The section dips southwest at moderate angles and about 500 metres of beds are present. Bedding is planar, but some two or three metres thick conglomerate beds wedge laterally and probably represent channel fills. Some conglomerate beds are 15 metres thick. Crossbedding is prominent locally. The conglomerate is immature with angular to subrounded pebbles, up to several centimetres across, in a coarse grit matrix. Pebbles of quartz, quartzite and schist predominate and their lithologies indicate derivation from the northeast side of the Tintina fault. The matrix contains abundant feldspar and quartz with prominent muscovite. Greenish brown mudstones form the partings between coarse beds; they are thinly laminated with much muscovite on bedding planes, common plant impressions and local coaly partings.

Along the road into Ross River are about 500 metres of equally immature clastic rocks in a moderately southeast-dipping sequence. It has coaly beds in mudstone near the base and coarsens upward to sandstone and conglomerate. The mudstone is dark brownish grey and thin bedded, parallel laminated, with abundant detrital muscovite and many carbonaceous plant impressions. Conglomerate and sandstone beds occur higher in the sequence as beds up to ten metres thick. Whereas the conglomerate tends to be internally massive the sandier beds are parallel bedded or crossbedded. The conglomerate is moderately indurated, but breaks around clasts. The pebbles are angular to subrounded with low sphericity, generally less than five centimetres across, of white quartz, chert of many colours, grey to black sheared quartzite, minor greenish cherty tuff and rare jasperoid pebbles. The clasts derive from northeast of Tintina Trench. The conglomerate's sandy matrix and the sandstone are predominantly feldspar and quartz.

Hughes and Long (1980) studied and described the occurrences more comprehensively than is done here and found that the coal is of anomalously high rank[51]. They attribute this to an exceptionally high geothermal gradient during burial of the coals and suggest that the coals may rest above shallowly buried intrusions

[51] The coal was mined and trucked to Faro to dry the lead-zinc concentrate from the Anvil Range until 1977.

emplaced during, or following deposition of the clastic rocks. Felsic and basaltic volcanic rocks in the region are roughly the same age as the clastic rocks.

Plant microfossils from the Lapie and Ross River road localities were studied by S. Hopkins of the Geological Survey of Canada and interpreted as Eocene and more probably Early Eocene (Hughes and Long, 1980; p. 8). [52] Previously plant remains collected by Kindle (1946) from the Lapie River beds were considered Paleocene. The conglomerate is the same as unit 15 of Tempelman-Kluit (1972b, p. 24) in Tay River map area[53].

Lapie River Quartz Feldspar Porphyry (map unit KTqfp)

About eight kilometres above its confluence with the Pelly River, the Lapie River cuts through crumbly, homogeneous buff to pale orange weathering, felsic porphyry. This exposure, and another about five kilometres to the northwest, represent a block between the Tintina and Grew Creek faults. The rocks have a white-altered, aphanitic, quartzofeldspathic groundmass with recognizable glassy quartz and white feldspar phenocrysts one or two millimetres across. Most likely they represent shallow subvolcanic intrusions, although an extrusive origin cannot be discounted.

The age and stratigraphic relations of the porphyry are unclear. Because it is confined to fault blocks in Tintina Trench, like the clastic rocks that have Eocene microfossils, the porphyry is considered to be the same age. The porphyry is the same as unit 14a in Tay River map area (Tempelman-Kluit, 1972b, p. 22-23) with which it is continuous. If the porphyry contributed to the anomalously high rank of the coal in the Eocene clastic rocks (Hughes and Long, 1980, p. 8), as it probably did, it postdates those strata. The porphyry is absent as clasts in the Eocene strata, but is cut by faults in Tintina Trench, the youngest of which are considered Pliocene on evidence of scarp retreat rates (Tempelman-Kluit, 1980a). Its age, between Eocene and Pliocene, is perhaps closer to the former.

Ketza River basalt, Hoole River basalt (map unit QTvbo)

Poor exposures of chocolate-brown weathering basalt are found at several localities near Ross River in Tintina Trench. One accumulation that covers about 50 square kilometres is just north of where the Ketza River enters Tintina Trench. Another lies astride Starr Creek about ten kilometres above its mouth. Both localities are breccias with fragments of dark green basalt in a basaltic sand matrix. Volcanic blocks contain sparse olivine crystals in a microcrystalline groundmass of feldspar laths.

The basalt breccias at the two localities and at a third immediately northwest of Grew Creek in Tay River map area (map unit 15b of Tempelman-Kluit, 1972b), are considered to be debris flows or lahars above beds of sandstone and conglomerate within Tintina Trench. They are confined to the block between the Grew Creek and Danger Creek faults and the fissures that fed them are in this block. Because basalt dykes cut the Eocene conglomerate and the quartz feldspar porphyries they are younger than both.

Basalt at two other localities occurs as gently dipping flows unconformably on older rocks. One locality is a kilometre above the mouth of Hoole River and the other is 15 kilometres northwest of Finlayson Lake. At both places olivine basalt flows a few metres thick may be intercalated with glacial gravels at their base and these flows are probably Late Tertiary or Pleistocene. [54]

The basalt flows are considered to be coeval with the breccias of the volcanic build ups. Both are regionally correlated with similar volcanics in Watson Lake map area (Gabrielse, 1967). In Carmacks map area equivalent strata are known as the Selkirk Lavas (Bostock, 1936) and in Whitehorse map area as the Miles Canyon basalt (Wheeler, 1961[55]).

STRUCTURAL GEOLOGY

Introduction

The project area displays both penetratively sheared allochthonous rocks and internally undeformed sedimentary strata imbricated on widely spaced brittle thrusts (Figs. 49 and 50). Two types of Mesozoic and Tertiary faults dominate the structure.

Most prominent on the map is the Tintina fault system which bisects the area from northwest to southeast. It is a transcurrent fault with about 450 kilometres of Late Cretaceous-Early Tertiary dextral displacement (Roddick, 1967; Tempelman-Kluit, 1977b; Gabrielse, 1985[56]). The trench along it results from probable Pliocene normal faults, which reactivated the feature.

The thrust faults are less obvious in the landscape, but are as significant in the geological map

[52] Tracks of Early Cretaceous hadrosaurs were discovered in the roadside outcrop in 1998 (Long et al., 2001; Gangloff et al. 2004).

[53] Unit Tcg by Gordey (in press).

[54] Most, if not all the localities of young volcanics (both felsic and mafic) indicated herein have been further described by Jackson et al (1986) who have also determined K-Ar ages ranging from 48 to 55 Ma (Eocene) from eight different localities.

[55] Most are Pliocene (6-17 Ma) by $^{39}Ar/^{40}Ar$ dating (Hart and Villeneuve, 1999)

[56] See also Gabrielse et al. (2006).

pattern. Some thrusts placed far-travelled rock slices (allochthons) above the native (autochthonous) Paleozoic strata and others imbricated these autochthonous beds. The thrusts are post-Upper Triassic and pre-late Early Cretaceous, dip generally southwest, and most verge northeastward.

Southwest of Tintina fault, ductile-deformed and metamorphosed allochthonous rocks are thrust over autochthonous beds from the southwest with a preserved overlap of 115 kilometres,. At their northeast edge the allochthonous rocks rest on the imbricated autochthonous succession (Fig. 51). Further to the southwest they lie above the thinned, strongly metamorphosed, mobilized and intruded autochthonous rocks. Similarly the surface that superposes the allochthonous rocks near the leading edge is abrupt and sharp over a metre or two, like thrusts in the autochthonous succession. Here the metamorphic inversion across the contact is profound with metamorphosed ductile deformed rocks above and unmetamorphosed internally undeformed beds below. In the southwest part of the map area the contact shows no metamorphic contrast although the lithologic change is sharp. The rocks and the contact have been "healed" by late tectonic metamorphism. The transition between these contrasting relations roughly follows the northeastern sides of the large batholiths. In Figure 49 this boundary is shown as that between involved and passive basement.

Southwest of Tintina fault autochthonous strata are imbricated by four important thrust faults (Figs. 49, 50 and 51). From the top down they are named the Pass Peak-McConnell-Liard, the Porcupine-Seagull, the Cloutier and the St Cyr thrusts. Each carries a slice two or three kilometres thick and is thought to merge down dip with the next lower thrust. Dip-slip on each thrust may be ten or more kilometres as demonstrated by offset facies. The cumulative shortening may amount to between 80 and 135 kilometres. The thrusts dip gently southwest and some are folded so that they are exposed across strike for up to 25 kilometres. Most thrusts have splays. Some of the splays are restricted to the rocks below the next higher thrust, but others cut and displace higher thrusts. Tear faults are developed on some thrusts that have allowed folds to attain different shapes on either side. Some thrusts are folded on a grand scale above a lower one that acted as the base of folding. Although most thrusts are northeast directed, a few southwest verging reverse faults are present. These are considered to be splays of the main northeast directed faults. Structures in the autochthonous rocks probably developed over a substantial period, perhaps the whole Early Cretaceous, and an average shortening rate of two millimetres per year is calculated. Southwestern thrusts formed early and lower, more northeasterly faults later. thrusting pre-dated strike-slip movement of the Tintina fault.

thrust faulted and metamorphosed strata southwest of Tintina fault are bowed over two broad northwest trending structural culminations about 30 kilometres across (Figs. 49 and 50). These are the Quiet Lake and Nisutlin arches. Their long axes coincide with the maximum dimension of the Quiet Lake and Nisutlin batholith respectively. St Cyr syncline, a 20 kilometre wide structural trough, separates these features. Rocks in the Quiet Lake, Big Salmon and Nisutlin arches are five to ten kilometres structurally higher than equivalent strata in St Cyr syncline. These structural highs and lows formed later than most of the thrusts. They coincide with intrusion of the Quiet Lake and Nisutlin batholiths, and occurred late in the metamorphism associated with these batholiths. The arches formed about the Early Cretaceous when the cores of the batholiths rose.

Northeast of the Tintina fault the ductile deformed allochthonous rocks predominate over the in-place strata. A slice of these rocks, 2 or 3 kilometres thick with only about 2 kilometres of structural relief, can be traced up to 65 kilometres across strike. These strata and their flaser fabrics are gently warped over structural culminations and displaced by steeply dipping faults. Gneiss beneath the allochthonous rocks may be crystalline basement mobilized into its cover. The autochthonous strata northeast of Tintina fault may be imbricated on thrusts, but no such faults reach surface there.

Ductile strain of the allochthonous rocks is isotopically dated as Early (and Middle) Jurassic. Emplacement of the transported beds by thrusting probably occurred in the Late Jurassic. Iimbrication of the subjacent autochthonous rocks began in the Late Jurassic and continued through the Early Cretaceous. Granitic intrusion probably overlapped with some of the thrusts in the early Late Cretaceous. Transcurrent movement on the Tintina fault was initiated in the Late Cretaceous and complete by the Eocene. The ductile strain, emplacement of the strained rocks on the autochthon, imbrication and folding of the in-place beds, and intrusion and transcurrent movement are apparently parts of a continuum, without evidence for long breaks between them.

Structures southwest and northeast of the Tintina fault are described separately and in that order. The structure of the autochthonous rocks southwest of the Tintina is explained as a series of four thrust sheets named after the four sole thrusts; from the highest of these sheets (including its internal structures) to successively lower sheets. Next the allochthonous klippen southwest of the Tintina fault are described. The Tintina fault zone and the steep faults immediately southwest are then discussed.

Structures northeast of the Tintina fault include the subhorizontal breaks within the sheared allochthonous rocks and younger steep faults and open folds that deform them. These structures are treated from structurally highest to lowest and a discussion of the detachment and the Mink complex follows.

Structures southwest of Tintina fault

McConnell thrust sheet

The McConnell thrust and thrust sheet are close to, and named for, McConnell River in east central Quiet Lake map area. Unlike the famous McConnell thrust of the southern Rocky Mountains this fault was not recognized by R.G. McConnell, but the river was named as a reminder of his reconnaissance in this part of Yukon and it seems particularly appropriate to emphasize that McConnell's geological work extended beyond the southern parts of the country by applying his name to this thrust fault.

The McConnell thrust sheet (Fig. 52) is the structurally highest involving the autochthonous rocks in the southern Pelly Mountains. It includes strata ranging from late Proterozoic to the Carboniferous. The sheet is between two and four kilometres thick and it demonstrates a profound change in structural style from the base upward. At the base, the Lower Cambrian McConnell limestone is imbricated and repeated by as many as six thrusts above the McConnell thrust, At the top, the competent quartz sandstone and dolostone of the Hogg formation is folded into open, broad folds, that are broken by steeply dipping reverse(?) faults, The intervening phyllitic rocks of the Kechika Group are internally deformed by small scale folds and transposition structures. The McConnell thrust sheet dips generally southwest towards the Nisutlin batholith which truncates it (Fig. 50).

The McConnell thrust sheet is cut by the Mount Hogg faults, many of which are interpreted as steeply dipping reverse faults that merge with the McConnell thrust at depth. Structures like the Twin Lakes and Lonely Creek faults cut the thrust sheet and its sole; they are interpreted as tears, rooted in lower thrusts, which have cut the thrust sheet following its emplacement.

Generally the McConnell thrust sheet carries a thick section of Lower Cambrian strata of the Ketza group (White Creek and McConnell facies) overlain by the Groundhog formation of the Kechika Group, the Platy Siltstone and Hogg formations of the Askin group and Black Slate of the Seagull group.

At the base of the McConnell thrust sheet are the McConnell, Liard, Pass Peak and MM thrusts, which are interpreted to have formed as a single surface that is broken by later faults. The separate names are used to emphasize that equivalence is not proven and to simplify description. The sole thrusts generally carry the McConnell or White Creek formations in the immediate hanging wall and their footwall is in the Kechika Group.

McConnell thrust

Segments of what are likely the McConnell thrust (Figs. 49, 52) i.e. the Pass Peak and MM thrust, are separately named because the connection between them is uncertain. The McConnell thust is clearly indicated south of White Creek where it places the lower part of the McConnell limestone above the middle or upper Groundhog formation (Lower Cambrian above Cambro-Ordovician) with a stratigraphic throw of 1200 metres. Along strike the fault remains in the same strata and can be traced from McConnell River southeastward about 20 kilometres, where it is cut by the Twin Lakes fault. East of that fault the offset continuation does not carry Lower Cambrian strata, but doubles the Groundhog formation. This segment can be traced southeastward for another forty kilometres to the valley of Lonely Creek with the same relationship across it and a stratigraphic separation on the order of 1000 metres. Gordey (1981a,b) recognized the repetition within the Cambro-Ordovician strata south of Peak 6656 and mapped a thrust, which he did not name.

For most of its length south of White Creek the McConnell thrust is the lowest of as many as six imbricate thrusts (e.g. Fig. 11e, f) that converge and diverge to give an anastomozing map-pattern. This pattern of interwoven hanging-wall imbricates is characteristic where Lower Cambrian strata are thrust over younger rocks. It is also seen in the Big Salmon thrusts and the Liard thrust. This "woven" pattern is not seen with thrusts involving other strata and may be a function of the depth of deformation and the competence of the rock unit. Recognition of the hanging-wall imbricates depends on the rock type. The thrusts are easily recognized where changes in the thick limestones signal a fault separation, but may be less obvious in homogeneous strata or fine grained clastic rocks.

The Liard thrust is most likely the faulted continuation of the McConnell thrust east of the Lonely Creek fault as it superposes the McConnell formation upon Cambro-Ordovician strata of the Ram formation, the same relations as formed by the McConnell thrust. If the Liard-McConnell and Pass Peak thrusts are the same, the strike-length on this fault system is about 140 kilometres, and for that entire distance the thrust follows roughly the same glide zone near the base of the McConnell limestone in the hanging wall and in the upper part of its Kechika Group in the footwall. In the segment between Lonely Creek fault and Twin Lakes fault, however, the hanging-wall glide zone is in the lower part of the Kechika Group. Because the thrust follows the same glide zones for long distances its trend is probably perpendicular to the transport direction. The fault does not give independent evidence for the sense of this movement, which is assumed to be to the northeast. The McConnell-Pass Peak-Liard thrust is the highest thrust in the autochthonous strata northeast of the Nisutlin-Big Salmon Arch. It is folded over the arch or truncated by the plutonic rocks, but may reappear in the St Cyr syncline as the Moose Creek thrust. If so the width of the thrust sheet across strike is 45 kilometres (distance from the Pass Peak to the Moose Creek thrust). The stratigraphic throw of nearly 1.5 kilometres is small in proportion to its possible transport.

If the Moose-McConnell-Pass Peak are one thrust, the fault cuts up-section to the northeast in the hanging wall, but down-section in the same direction in the footwall. The first implies northeast transport, the second the reverse. A possible explanation of this anomaly, assuming the faults are the same, is that folding preceded thrusting or that old faults diverted the thrust in the footwall. Such old faults may control the profound facies changes in the Askin group in the McConnell thrust sheet.

Pass Peak thrust

About five kilometres south of Pass Peak in central Quiet Lake map area is the Pass Peak thrust, a well exposed, clearly defined fault that dips moderately southwest (Fig. 52). It superposes Lower Cambrian limy schist generally onto beds of the Askin group, but locally onto the Kechika and Seagull groups. Stratigraphic throw is between one and two kilometres. The fault can be followed from its intersection with Groundhog fault-2 (Fig. 53; refer to 1:50,000 scale map sheet 105F/10) southeastward for about 20 kilometres. In its hanging wall it consistently carries the Pass Peak formation, but the footwall glide zone is at the top of the Askin group and jumps to the upper Kechika Group across Groundhog fault-7, and gradually drops to the Ketza group southeastward. The thrust, therefore, loses stratigraphic separation south-eastward and may die in that direction.

What is interpreted as a small klippe of the Pass Peak thrust sheet lies between Seagull Creek and Groundhog fault-3 (Fig. 53). On a rounded low peak is a sheet of phyllite of the Kechika Group above slate of the Seagull group (also discussed under "Groundhog faults"). As the thrust separating the two is structurally above the Seagull thrust this is considered an outlier of the Pass Peak thrust sheet. The hanging wall and footwall glide zones at this place are both stratigraphically above those of the Pass Peak thrust to the south indicating northeastward transport on the fault, assuming the outlier interpretation to be correct. The distance between the klippe and the Pass Peak thrust is eight kilometres, the minimum overlap of the thrust. The amount of dip slip is unknown.

To the southeast the Pass Peak thrust ends against an unnamed, east-northeast trending fault. Across that fault are the twin MM thrusts, the likely extension of the Pass Peak and themselves probably continuous with the McConnell thrust. Although the connection between the segments is tenuous, the Pass Peak thrust is apparently part of the McConnell-Liard thrust which generally brings the Ketza group over the Kechika Group or younger beds.

MM thrusts

The MM thrusts are a pair of subhorizontal faults, beautifully exposed, one above the other, in a small valley near the confluence of McConnell River and Seagull Creek (Fig. 54). Calcareous brown phyllite of the Pass Peak formation is thrust over dolostone and sandy dolostone of the Askin group, transitional between the Hogg and Barite Mountain formations. This is in turn superposed on black slate with intercalated volcanics of the Seagull group. The middle slice is discontinuous; at its thickest it attains 200 metres, but it thins and disappears so that the upper and lower thrusts merge (Fig. 52). The rocks in all three slices are metamorphosed generally to upper greenschist facies, and the thrusts are folded over an open northward trending antiform with an arcuate trace which is about three kilometres across. The metamorphism results from warming next to Nisutlin batholith and the folds formed during arching while the batholith rose. Both the metamorphism and folding postdate thrusting. A fairly tight Z-shaped fold (looking down the plunge to the northeast) with an amplitude of 250 metres also folds the thrust between the Askin and Seagull groups. This fold also formed during metamorphism and intrusion. It relates not to the thrust, but to upwarping of Nisutlin Arch.

Because the superposition relations of the MM thrusts are like those of the McConnell and Pass Peak thrusts they are considered one, but junctions between the three fault segments are not exposed and their locus is not closely defined. On the south the MM thrusts dip toward Nisutlin batholith and the intrusion presumably cuts the faults beneath the valley of Sheep Creek.

Liard thrust

The name Liard thrust is introduced here for the fault that superposes Lower Cambrian on Cambro-Ordovician rocks in south-central Finlayson Lake map area. It is best exposed on two ridges nine kilometres north of Mud Lake. The fault can be traced southeast from the Lonely Creek fault to the southern edge of Finlayson Lake map area. In Wolf Lake map area (south of Finlayson Lake map area) the fault mapped along Liard River is undoubtedly its extension (Poole et al., 1960). The Liard thrust is probably the continuation of the McConnell thrust in Quiet Lake map area. The McConnell thrust together with its probable extensions have a strike length of 200 kilometres from Pass Peak to the western edge of Watson Lake map area.

The Liard thrust has a stratigraphic throw on the order of 2000 metres as it brings quartzite from the lower part of the Ketza group (Pass Peak formation) to above high parts of the Kechika Group's Ram formation. The fault shows the same repetition of the Ketza group above it, seen also in the McConnell thrust, so that structurally higher faults that duplicate the succession merge with each other and with the basal thrust along strike. The thickness to length ratio of the imbricated sheets immediately above the base is on the order of 1:30 so that the sheets are thicker in relation to their length than those involved in the Big Salmon thrusts, but thinner, or about the same as, those above the McConnell thrust.

Tintina plug, an eight-kilometre-long quartz monzonite intrusion, cuts squarely across the Liard thrust. The intrusion is not dated, but is lithologically like the nearby Black River batholith which gave KA ages of 46.9, 53.9 and 66 Ma. The rock also resembles other batholiths dated in the range 70 to 95 Ma. Liard thrust is therefore Latest Cretaceous or older and is speculated to have been active during the Early Cretaceous.

Big Salmon thrusts

On both sides of the Big Salmon River in west central Quiet Lake map area are as many as ten superposed, relatively thin, extensive sheets of the Scurvy limestone, separated by metamorphosed clastic rocks (Fig. 55). These limestones dip into the St Cyr syncline conforming with the fabric of the metamorphic rocks. North of Big Salmon River they dip southwest and south of the river they dip northeastward. Similar fossil indicate one Lower Cambrian limestone exists, and away from the small area of superposed sheets only a single thin limestone bed is seen. These limestones, therefore, are considered to be a structurally multiplied stratigraphic unit, with thrusts at the upper or lower contacts of each of the limestone lenses. The thrusts are not individually named or labelled. Single thrust sheets are up to 500 metres thick with as much as 300 metres of limestone in some sheets. Most thrust sheets are about 150 metres thick. The largest sheets can be traced along strike for 15 kilometres, giving a thickness-to-length ratio of between 1:50 and 1:l00.

The southern and northern Big Salmon thrust packages each strike roughly northwest, parallel to the regional trend of the St Cyr syncline. This direction is related to the latest thrusts and folding of the rocks but does not necessarily reflect the orientation of the thrusts when they were initiated; nor does it intrinsically give the direction of transport during thrusting. The northern and southern Big Salmon thrust packages together trend about 15° west of north across St Cyr syncline. If the sheets were structurally stacked at right angles to this north 15° west trend a transport direction for the thrusts of about north 75° east is implied. It is not known whether this direction of movement (if valid) is general for other thrusts in the Pelly Mountains or whether it applies only to the Big Salmon thrusts. The trend of the zone of stacking may be related to a step in the subjacent thrust sheet and/or the basement on which these thrust sheets were piled. The indicated transport direction then reflects the orientation of this step and not the general movement of the thrusts.

The northern Big Salmon thrusts lie above the Moose thrust and the southern Big Salmon thrusts are north of (above) the Scurvy fault. The relationship of the Big Salmon thrusts to what may be a single master surface, the Moose-Scurvy thrust, implies that the displacement was not concentrated on one surface but on a series of surfaces in a relatively restricted zone above the main surface. It further suggests that the Big Salmon thrusts are upward splays of the Moose thrust forming an anastomozing pattern of convergence and divergence similar to that displayed nicely by the southern Big Salmon thrusts. The Big Salmon thrusts are therefore thought of as splays of the Moose thrust, like the branches immediately above the McConnell and Liard thrusts.

Mount Hogg faults and folds

faults near Mount Hogg and McConnell Peak are not individually named, but are numbered for description and, with the intervening folds, are collectively referred to as the Mount Hogg folds and faults (Fig. 52). They are in the upper part of the McConnell thrust sheet and contrast sharply with the imbricate thrusts in the less competent strata at the base of the thrust sheet. Their occurrence in the same sheet implies a possible connection, and the imbricate glide zones at the base of the thrust sheet may pass upward to become the steep northwest-trending faults that separate the open folds. This interpretation is shown in the cross-sections.

The Mount Hogg faults include two sets. Most prominent are the northwest-trending structures with straight traces, roughly parallel to the regional grain. These include faults 1 to 8 (Fig. 52), which can be followed for up to 25 kilometres each. Their northeast side fell relative to the southwest by amounts between 100 and 1000 metres with the largest displacement on faults 6 and 8. Most of these faults juxtapose different parts of the Askin group, but younger slate of the Seagull group and phyllite of the Kechika Group are also involved. The faults dip steeply, and it is speculated that movement was reverse so that the dip is probably southwest. Reverse movement on the faults is compatible with the compressive environment that existed during folding. The faults are easily followed in the competent distinctive beds of the Askin group (Figs. 23a, b), but only one (fault-8) was traced through the underlying Kechika Group towards the sole of the thrust sheet. Northwest of McConnell River fault-8 splits and juxtaposes the Ketza and lower Kechika groups, omitting about two kilometres of strata. This fault cuts a thrust near the sole of the McConnell thrust sheet and is itself truncated by Nisutlin batholith. fault-8 may have movement unrelated to that on the other Mount Hogg faults. It is a few kilometres northeast of, and parallel to, the margin of Nisutlin batholith. Mid-Cretaceous normal movement related to emplacement of the batholith while Nisutlin-Big Salmon Arch formed is probable on this structure. fault-8 may have originated, like the other Mount Hogg faults, during Jura-Cretaceous thrusting and folding extending into the mid-Cretaceous during the intrusion of Nisutlin batholith.

Other steep faults in the Askin group cannot be traced downward and may flatten sharply in the Kechika Group. Phyllite of the Kechika Group is intricately deformed by small scale subisoclinal to recumbent folds, a few centimetres or tens of centimetres across. The folds are commonly rootless and transposed on a closely spaced crenulation foliation that generally dips moderately southwest. If the steep faults do flatten down-dip they probably follow the crenulation foliation as they pass into the Kechika Group. Movement may remain concentrated on a single surface or may be spread over a zone of transposition surfaces. After following the foliation down-dip and behaving as foliation glides, the faults may merge with thrusts at the base of the thrust sheet in the Lower Cambrian strata. The transition from steep to moderate dip on the faults seems most likely to be abrupt and to coincide with the competency boundary between the Askin and Kechika groups. The second set of Mount Hogg faults strikes northeast. They are shorter, also dip steeply and are cross-structures to the first set. They are assumed to be tear faults in the McConnell thrust sheet that allowed differential deformation of the sheet. Southeast sides moved down on faults 9 and 10 and upon fault-11, with stratigraphic throw of one or two hundred metres. These faults were not traced downward into the Kechika Group.

Folds between the Mount Hogg faults are concentric and open. synclines are about five kilometres across and anticlines are tighter, one or two kilometres across, as measured at the present level of exposure in the Askin group. The folds have subhorizontal axial traces that are sinuous and trend northwest. Broadly the fold-trend corresponds to the regional grain, but some folds strike more westerly than the main Mount Hogg fault-set. The geometry of these oblique folds suggests that, at the level of exposure near Mount Hogg, the folds are closer to the base rather than the top of the concentrically folded strata. If true, the level of maximum flexure may be near the top of the Askin group. This in turn implies that between one and two kilometres of beds (now eroded) were also folded concentrically above the present level of exposure.

Moose thrust

Northwest of Big Salmon River between Northern Lake and Moose Creek the Moose thrust separates Proterozoic and Paleozoic metamorphosed rocks from similar strata (Fig. 52). The fault is well defined at an abrupt and sharp contact where Proterozoic schist rests on the metamorphosed Nasina formation (Fig. 56). No minor structures indicative of significant movement are seen at the contact and only the stratigraphically-incorrect superposition of strata indicates the fault. Northwestward from Moose Creek, where the Nasina formation is cut out, the trace of the fault is difficult to discern, Despite its relatively small stratigraphic separation, perhaps between 500 and 1000 metres, the fault probably continues far northwest from where it is clearly indicated. The fault strikes northwest and dips gently southwest, concordant with the foliation of the enclosing strata.

The Moose thrust may be the master fault in a set that includes the northern Big Salmon thrusts. The relative thinness of the Big Salmon thrust sheets compared with their lateral extent implies fairly plastic behaviour during transport. These thrusts likely formed at considerable depth, probably when metamorphism had already begun. However, metamorphism continued after thrusting, as indicated by the metamorphic fabrics, where micas are grown across foliation surfaces. The Moose thrust is therefore most probably Late Jurassic or Early Cretaceous. Movement on it had ceased when it was arched over the Big Salmon-Nisutlin culmination about the mid-Cretaceous.

Along strike to the northwest the Moose thrust cuts down-section in the footwall and remains at roughly the same stratigraphic level or cuts down in the hanging wall. This implies an eastward component of movement on the thrust if it formed before significant folding. Considering the plasticity of the rocks and probable depth at which it formed the assumption of no significant folding may be inappropriate. No conclusion is therefore possible about the direction of movement.

The Moose thrust may be the continuation of the Scurvy fault on the northern side of St Cyr syncline. A second possibility, which does not exclude the first, is that it is also the southern continuation of a thrust northeast of the Big Salmon-Nisutlin Arch. Being the highest fault in autochthonous strata in St Cyr syncline the Moose thrust might be the extension of the McConnell thrust, the highest two faults northeast of Nisutlin batholith. Connection across the 20-kilometre-wide structural arch implies about 35 or 40 kilometres of across-strike extent on the McConnell thrust system, suggesting larger amounts of transport on that fault than might be expected from the kilometre or two of stratigraphic throw. If the Moose and McConnell thrusts are one, the fault remains in roughly the same bedding glide zone in both walls across Big Salmon-Nisutlin

Arch although the facies in the footwall strata of the Askin group changes drastically.

Scurvy fault

Relations between the Proterozoic and Paleozoic strata along Scurvy Creek indicate a fault along that stream. This inferred fault places the top of the Proterozoic mica schist next to the upper part of the Nasina formation, so that the northeast side is raised with respect to the southwestern. Stratigraphic throw is difficult to estimate, because no sections of the Nasina formation were measured in this area, but a kilometre or more is probable. The fault trends west-northwest, but its dip is unknown. On the west the Scurvy fault is cut by the Dycer Creek stock, a quartz monzonite dated as Late Cretaceous. On the east the termination of the fault is speculative; it may cut the sole of a klippe of the Anvil assemblage or it may follow the valley of Quiet Lake.

Relations are inconclusive, but if Scurvy fault is a thrust it may dip northeast at about 30 degrees concordant with the foliation of the rocks on both sides. In this instance it may be the continuation of the Moose thrust, exposed on the opposite (northern) side of the St Cyr syncline. The Moose thrust juxtaposes the same rocks as the Scurvy fault and the two could simply be exposures of one synclinally folded thrust. If this interpretation is correct the width of the thrust sheet across the syncline is roughly 20 kilometres. Equivalence of the Moose thrust and Scurvy fault means that the thrust remains at roughly the same stratigraphic level in the upper plate across the syncline, but that it cuts down section toward the northeast in the footwall. In this interpretation the southeastern end of the fault probably lies in the valley of Quiet Lake..

Alternatively the Scurvy fault may dip more steeply northeast and cut the Moose thrust. In this case the dip-slip movement on the Scurvy thrust is on the order of two kilometres as determined from the offset of the allochthonous rocks across it.

If the Scurvy fault is a normal fault it probably dips steeply southwest. It may then be genetically related to the Quiet Lake Arch, just as the Pony Creek fault is related to the Nisutlin Arch. However, normal displacement seems unlikely, given the general compressional deformation. Consequently the steep reverse-fault interpretation is preferred and shown on cross-sections (e.g., Cross-section sheets, section E)

Caribou fault

A ten-kilometre-long north-northeast trending fault is inferred between the lower reaches of Moose and Caribou creeks (Fig. 52). Unfortunately exposures are not good enough to pinpoint the fault or even to confirm its existence. It cuts Paleozoic rocks in the core of the St Cyr syncline, juxtaposing the Nasina formation with roughly time-equivalent strata of the Hogg formation and with slate of the Seagull group. This fault had relatively minor displacement with the eastern side down perhaps a few hundred metres relative to the western side. Movement probably occurred when the St Cyr syncline formed in mid- or Late Cretaceous time. Furthermore the fault allowed the eastern part of the syncline to extend to greater depth than the western part. It may be an older structure that controlled depositional facies, which was reactivated in the Cretaceous.

Gray Creek fault

The northeast-trending fault between upper Gray and Scurvy creeks in southwest Quiet Lake map area is called the Gray Creek fault (Fig. 52). It is well exposed on two ridges and the existence and displacement are clearly demonstrated: the fault can be traced for about 15 kilometres. Gray Creek fault disrupts the northern margin of Quiet Lake batholith and the regionally metamorphosed Proterozoic and Paleozoic strata flanking it. Its apparent displacement is 5 kilometres left-lateral or northwest side down 3.5 kilometres relative to the southeast side. The short fault trace implies that this is not a strike-slip, but a tear fault so that the second option for displacement is more probable.

Gray Creek fault probably moved while the Quiet Lake batholith was emplaced or perhaps when part of it had cooled. K-Ar ages (83 and 68 Ma for two Quiet Lake batholith samples) indicate this was during the Late Cretaceous. The fault most probably behaved as a tear which enabled the eastern part of the batholith and its northeastern flank to rise about 3.5 kilometres farther than its northwestern part, while the plutonic rocks and their metamorphic halo rose. The Quiet Lake and Nisutlin-Big Salmon Arches are between 8 and 10 kilometres structurally above the intervening St Cyr syncline and 3.5 kilometres of relative vertical displacement on Gray Creek fault is reasonable in this context.

Porcupine-Seagull thrust sheet

Introduction

The Porcupine-Seagull thrust sheet occupies much of northeastern Quiet Lake and southwestern Finlayson Lake map areas and includes the strata between the Porcupine and McConnell thrust systems, i.e. between the Porcupine-Seagull-Mount Vermilion-Ragged Peak thrusts and the Liard-McConnell-MM-Pass Peak thrusts (Fig. 53). The thrust sheet is the second highest in the autochthonous sequence and lies above the Cloutier thrust sheet. The Porcupine-Seagull sheet varies in thickness from a few hundred metres near the lower reaches of Seagull Creek to 3 kilometres north of Pass Peak and may average 1-1/2 to 2 kilometres. The separately named faults at the base of the thrust sheet are thought to be parts of an initially continuous surface that was disrupted, but because this interpretation cannot be proven separate names are used for physically disconnected segments.

Carbonate rocks of the Askin group dominate the northeastern part of thrust sheet and form its most spectacular exposures. The more recessive Kechika Group generally predominates in southwestern parts. The thrust sheet also carries strata of the Ketza and Seagull groups. Generally the Askin group is represented by the Porcupine and Barite formations, the Kechika Group by the Groundhog and Cloutier formations and the Seagull group by the Black Slate formation. The hanging wall of the Porcupine-Seagull thrust is commonly in strata of the Kechika Group and the footwall in those of the Seagull group.

From northwest to southeast the thrust sheet was traced for about 130 kilometres along strike and up to 30 kilometres across strike. Large open folds, including the Porcupine anticline and syncline, deform the Porcupine-Seagull thrust sheet into broad, symmetrical folds traced for tens of kilometres, with wavelengths of 10 or 15 kilometres and amplitudes of several kilometres. Steeply dipping faults cut parts of the Porcupine-Seagull thrust sheet. Some, like the Groundhog faults, acted as tears and merge with the basal thrust. Others, like the Lapie, Twin Lakes, Starr Creek and Lonely Creek faults, are tears that cut the thrust sheet, but which are probably rooted on lower thrusts. Still others, like the McNeil faults, are interpreted to be steep reverse faults that merge with lower thrusts. The folds and those faults that cut the sole probably formed when the Porcupine-Seagull thrusts had ceased moving and are a response to motion on lower faults such as the St Cyr thrust. The size of the folds indicates this lower discontinuity may be at considerable depth.

Most of the competent strata of the Askin group in the thrust sheet are deformed by near-surface brittle faults, but shaly beds of the Kechika and Ketza groups are internally deformed by a penetrative cleavage and by innumerable small-scale folds.

Porcupine thrusts

The Porcupine thrust[57] (Fig. 53) is spectacularly exposed in some places and is inconspicuous in others. On the northeast side of Porcupine syncline it makes the most grandiose scenery in the Pelly Mountains superposing a 300 or 400 metres high cliff of buff-coloured carbonate rocks of the Askin group above recessive weathering, vegetated slate of the Seagull group for a strike length of 35 kilometres (Fig. 57). This imposing wall faces northeast and is breached only by Ram and Cloutier creeks. The fault was first mapped by Wheeler and others (1960) in this area. They did not trace the fault under the syncline. East of Ketza River the Porcupine thrust places basalt and slate of the Kechika Group above slate and felsic volcanics of the Seagull group. There the thrust underlies a high sombre mountain (Peak 6762') with a more subdued vista (Fig. 58). The Porcupine thrust is exceptional because it can be followed from north central Quiet Lake map area southeastward along strike for 125 kilometres and across strike for long distances. The thrust dips gently and is folded, so that it also surfaces across the strike. At its widest the thrust sheet measures 25 kilometres.

The Porcupine thrust can be traced around the Porcupine syncline to Cloutier Creek and thence followed to the small, faulted syncline between Cloutier Creek and Ketza River, the Cloutier syncline. East of Ketza River the fault is seen in the Twin Lakes syncline and anticline (Fig. 56). Farther southeast the Porcupine thrust continues as the Ragged Peak thrust of Gordey (1981a,b). South of the McNeil fault-1 (Fig. 53) the Porcupine thrust is repeated in the Ketza River area on several steeply dipping faults which step the thrust up toward the south. Across the southernmost of these steps[58] is the Seagull thrust, which is considered the equivalent of the Porcupine thrust although it is separately named. The Porcupine thrust is similarly repeated by the McNeil fault-2. South of that fault the Mount Vermilion thrust with a lower splay is its continuation.

Generally the Porcupine thrust places the Kechika and Askin groups on the Seagull group. In northwestern exposures of the hanging wall the thrust carries the upper part of the Kechika group or the lower part of the Askin group. Near its southeastern end the hanging wall is composed of lower Kechika group rocks, i.e. the fault cuts up-section in the hanging wall toward the northwest. Footwall strata are slate and volcanics of the Seagull group, but locally the Askin group underlies the thrust directly and elsewhere the Hoole formation is overridden. In general, the fault cuts up-section in the footwall toward the southeast. No significance is seen in this, nor for the northwestly direction of rise in the footwall . Across strike no systematic change is seen from the Seagull thrust to the Porcupine thrust. Both faults are in the same bedding glide zone in the footwall and hanging wall. However, around the Porcupine syncline the Porcupine thrust cuts up-section in the hanging wall and footwall toward the northeast.

On the map scale the Porcupine thrust is a single surface in most places, but in detail the hanging wall is imbricated and small subsidiaries are common above the main thrust. Where the Askin group is carried in the hanging wall these splays repeat lenses of the carbonate

[57] Here refers to that labelled Upper Porcupine thrust on the 1:50 000 scale maps. The Lower Porcupine thrust is described four paragraphs later.

[58] This is Ketza fault-1 on the Cloutier thrust sheet, shown on Fig. 62.

ten metres or more thick and traceable for hundreds of metres. Where hanging-wall strata are the Kechika Group, imbricate slices are less obvious, presumably because they follow the cleavage or foliation of the rocks. The imbricated zone of the Porcupine thrust is perhaps 100 or 200 metres thick in places, but is by no means uniform or present everywhere. The basal surface itself is abrupt and impressive - in some exposures on the north side of Porcupine syncline it is a two metre wide zone of intensely cleaved phyllite.

The Porcupine thrust has an important splay under the western part of the Porcupine syncline called the Lower Porcupine thrust (Fig. 53). This fault is roughly 1500 metres below the main thrust at the Lapie fault, but migrates gradually upward to join the main strand near the east end of Porcupine syncline. The Lower Porcupine thrust carries the Askin group over the Seagull group like the Porcupine thrust itself. Just as the Porcupine thrust is considered the continuation of the Seagull thrust the Lower Porcupine was probably connected with the Lower Seagull thrust over the Porcupine anticline (Cross-section D, cross-section sheets).

Immediately northwest of the Lapie fault (Cross-sections A, B, cross-section sheets) the lower splay of the Porcupine thrust southwest of the Porcupine syncline superposes the Silurian Platy Siltstone formation above Proterozoic or Lower Cambrian phyllite omitting the Kechika and Ketza groups (Fig. 59). This abnormal superposition of younger on older rocks implies that the lower splay moved after the Porcupine thrust was already partly folded or displaced by older faults, that it is a minor fault by comparison to the Porcupine thrust, and that it is probably allied genetically and in time with the Cloutier and not the Porcupine thrust. Northwestward the lower splay is truncated by the Bacon Creek stock, and apparently loses displacement rapidly[59].

The lower splay of the Porcupine thrust northeast of the Porcupine syncline and southeast of the Lapie fault doubles the Seagull group in most places, but locally includes part of the Askin group in the hanging wall. Across the Lapie fault the continuation of the lower splay is named the Magundy thrust (Fig. 56). It doubles the Askin group, but apparently dies in the bedding of these strata toward the northwest.

Across Porcupine syncline the vergence is to the northeast (Cross-section D). The minimum overlap, disregarding correlation with the Seagull thrust, is about six kilometres (the distance across the Porcupine syncline). If the Porcupine and Seagull thrusts are equivalent, the minimum northeastward overlap is 25 kilometres. The contrasting facies between the Seagull group in the hanging wall at the leading edge of the thrust (black Slate without volcanics) and that in the footwall on the trailing edge (felsic volcanics with little slate) suggests that this is a minimum. Similar large displacement is also possible on the McConnell and Cloutier thrusts.

The Porcupine thrust was folded when all or most movement on it was completed and this folding occurred during displacement on the Lapie, Twin Lakes and Starr faults, probably while the St Cyr or Cloutier thrusts moved. The youngest strata involved in the Porcupine thrust are Upper Triassic beds of the Hoole formation, which occur beneath it near Starr Creek. The oldest rock that postdates movement on the Seagull thrust, and by extension the Porcupine thrust, is quartz monzonite of the White Creek stock, which intrudes this thrust (Fig. 47c). It gave a K-Ar age of 99.2 Ma. The Porcupine thrust, therefore, moved during the Jurassic or Early Cretaceous.

The Porcupine thrust northwest of Lapie fault is interpreted to end at depth on the Fox Creek thrusts (Fig. 53). Its offset continuation south of the Nisutlin-Big Salmon Arch should be between the Moose thrust and Big Salmon batholith. No thrusts were recognized there as no stratigraphic repetition was noted. However, eight to ten kilometres of Proterozoic strata may occur southwest of Big Salmon batholith in the footwall of the Moose thrust. The anomalous thickness of Proterozoic rocks may indicate repetition of these beds. A thrust is possible parallel to, and about two kilometres northeast of, the Moose thrust where Proterozoic schist of the upper greenschist facies (map unit Pns) lies above equivalent gneiss (map unit Pn+). This possible thrust would be conformable with the foliation so that it dips southwest and trends northwest for possibly 40 kilometres. To the southeast it may be truncated by the Nisutlin batholith and to the northwest it runs perhaps into adjacent Laberge map area. The contact is an abrupt boundary that coincides with topographic lows.

If the possible thrust is the folded continuation of the Porcupine thrust the combined structure has a width across strike of 40 kilometres. The combined structures cut up-section in the hanging wall and footwall toward

[59] Reviewer's comment:
There is a strong case to be made that the lower thrust does not exist northwest of the Lapie fault. Its existence northwest of the Lapie fault is due to over-tightening of the Porcupine syncline, indicated by the high dips there. The lower thrust(s) are simply out-of-syncline faults of minor displacement which accommodated the minimal extra shortening compared to southeast of the Lapie fault. Also notice the trend of porcupine syncline is slightly offset across the Lapie fault; which was a tear fault so that the lower thrust need not continue northwest of it. This interpretation eliminates unusual "younger over older" relationships and instead interprets these contacts northwest of Lapie fault as stratigraphic. An unconformity under the Platy Siltstone, so it sits on the Pass Peak formation, is consistent with the interpretation in Tay River map area of an unconformity beneath the Platy Siltstone (Gordey, in press).
Not mentioned in this section is a long fault on the 1:250,000 map (northwest of the Lapie fault) which places Black Slate formation against Askin group. It would correspond to the lower thrust but makes more sense as a stratigraphic contact.

the northeast, which would represent the direction of transport.

Earlier it was shown that the Seagull-Porcupine thrust is Jurassic or Early Cretaceous. If the contact between the gneiss and schist is indeed a fault, the movement was accomplished after, or during, the metamorphism. The truncation of the postulated thrust by Nisutlin batholith is not clear. Instead, the thrust may cut and carry the quartz monzonite over metamorphic rocks (see Fig. 50). If so, the conclusions concerning movement need to be modified further because Nisutlin batholith has early Late Cretaceous K-Ar cooling ages of 88.1, 91.2 and 96.0 Ma. In this case displacement on the thrust presumably coincides with, or may postdate these cooling ages and may be mid-Cretaceous.

Seagull thrust

The Seagull thrust (Fig. 53) does not crop out where it follows Seagull Creek, but its northwestward extension between the Lapie and Groundhog faults is well exposed on ridges. The thrust has an upper and lower strand which merge locally along Seagull Creek. Elsewhere the two strands are separated by Siluro-Devonian carbonate rocks of the Barite Mountain formation (Fig. 60). The lower strand brings the Barite Mountain formation above felsic volcanics of the Mississippian Seagull group and the upper strand places Cambro-Ordovician phyllite of the Groundhog formation on Askin group carbonate. Where the strands merge the Groundhog formation is thrust directly over the Seagull group. Stratigraphic throw across the lower strand is between 500 and 1000 metres and across the upper it may be 1000 metres.

On the map (Fig. 53) both strands of the Seagull thrust diverge toward the northwest. East of Lapie River 1000 metres of the Porcupine formation lies between them. The two strands are folded over the Porcupine anticline and probably join the twin Porcupine thrusts which have similar relations and which juxtapose the same rocks (Cross-section D). The Seagull thrust are therefore considered to be the southern continuation of the Porcupine thrusts and the sole of the Porcupine-Seagull thrust sheet.

Seagull thrust-2 (Fig. 53) is probably the extension of the Seagull thrust across the valley of McConnell River. Although they cannot be connected on the ground they have the same juxtaposition. Seagull thrust-3 is a lower strand of Seagull-2.

A profound change of facies is seen in the Seagull group across the Seagull fault. In the footwall the unit is represented by thick felsic volcanics, but in the hanging wall those volcanics are replaced by relatively thin Black Slate. The Seagull thrust has telescoped these facies an unknown amount. The thrust may be localized by this facies transition in places. In particular the footwall of the Seagull thrust along Seagull Creek may follow this transition. If the transition was originally fault controlled and related to Mississippian normal faults the Seagull thrust may follow such an older structure along Seagull Creek.

Groundhog faults

The Groundhog faults comprise a dozen steeply dipping faults that cut the Cloutier and Porcupine-Seagull thrust sheets and which are referred to by number for convenience (Fig. 53). Most of the faults can be traced for only a few kilometres and had minor displacement which allowed parts of the Seagull thrust sheet to move with respect to each other. Eight of the Groundhog faults trend northeast, but two important breaks are perpendicular to this. Of the northeast trending structures, faults -1, -2 and -6 had left-handed or northwest side up movement, while Groundhog 5 and 7 show the opposite sense. faults 3 and 4, the northwest trending structures, raised the block between them about one kilometre (Cross-section G) allowing the flank northeast of fault-3, and that southwest of fault-4, to drop.

Relations in the small blocks between Groundhog-1 and -2 and between -2 and -10 are interesting (Cross-sections F and G). An open anticline of Askin and Seagull group strata is overthrust by phyllite of the Kechika Group. The thrust, also folded, is probably the upper strand of the Seagull thrust. The anticlinally folded beds and thrust are absent southeast of fault-10 illustrating that although these tears are small, they developed different structures on both sides. In the wedge-shaped block between faults -1 and -2 the Seagull thrust also reappears and is evidently cut by both faults.

The small klippe between the Seagull thrust and Groundhog-3 is a cap of Kechika Group strata resting subhorizontally above slate of the Seagull group. The stratigraphic throw is about 1000 metres. The sole of this klippe is the next thrust above the Seagull thrust and the klippe may be part of the McConnell thrust sheet.

Groundhog faults -11 and -12 are steeply dipping faults on which the northeast side was dropped relative to the southwest. Offset on the order of 100 metres is indicated by displacement of the thick diabase sill that caps ridges east of Lapie Lakes.

The Groundhog faults appear to be a set because they are spatially linked, and are thought to be genetically related. The northeast trending faults most likely developed as tear faults during thrusting. Some, such as Groundhog-1 and -2, are tears that cut the Seagull thrust; they may root in the Cloutier thrust. Groundhog-3 and -4 are not tears, but steep reverse faults, interpreted to merge with a thrust below the Seagull thrust, most probably the Cloutier thrust. Groundhog-9 and -10 could be tears within the Seagull thrust sheet: Groundhog-5 and -8 faults cut the Seagull and Pass Peak thrust sheets and Groundhog-6 and -7 are tears confined to the Pass Peak thrust sheet. Together the Groundhog faults are considered expressions of the general compressive deformation and whereas some (faults -6 and -7) moved early with Pass Peak thrust, the most significant movement occurred after the Seagull thrust, during movement of the Cloutier thrust sheet (Groundhog faults -1, -2, -3 and -4).

Ragged Peak thrust

Gordey (1981) described and named this fault, which in its type area superposes the Kechika Group on the Hoole formation with about 2.5 kilometres of stratigraphic throw. The fault is followed northwestward from its type area in the same footwall glide zone, but it cuts upward across the Kechika Group into Askin group beds in its hanging wall. It is offset by the Starr fault from the Porcupine thrust on the north limb of the Twin Lakes anticline (Cross-section O). In this area the Ragged Peak thrust is likely the continuation of the Porcupine thrust[60].

West of Lonely Creek the Ragged Peak thrust is truncated by the McNeil fault-2 (Fig. 53, 61). Southeast of the Lonely Creek fault the Ragged Peak thrust continues, placing the Kechika Group first above the Hoole formation and then against progressively older strata southeastward. Finally, at its southeastern end, the Ragged Peak thrust terminates against the Ings River fault.

In its type area the Ragged Peak thrust dips gently southwest and is folded over a small syncline - anticline pair (Cross-section Q, cross-section sheets), but to the northwest the fault dips more steeply, perhaps averaging 60 or 70 degrees, and east of Lonely Creek it dips moderately southwest.

Southwest of McNeil fault-2 the Mount Vermilion thrust is considered the extension of the Ragged Peak thrust (Cross-sections N, O). This implies a minimum cross-strike extent of six kilometres. As the Ragged Peak is the continuation of the Porcupine thrust, movement of several tens of kilometres is possible.

Displacement on the Ragged Peak thrust presumably occurred in the Late Jurassic or Early Cretaceous, more likely the latter, in the same interval that the other thrusts moved. Within this time span the Ragged Peak thrust may be relatively young.

Ings River fault

The relatively short northwest trending fault east of, and parallel to, the Lonely Creek fault is named the Ings River fault (Fig. 53). It follows one side of Ings River and was traced about twelve kilometres southward from its termination against the Ragged Peak thrust. The fault may extend to the Liard River thrust. Recognition is based on the interruption of strata across it as observed on the ground, but the fault cuts Kechika Group strata on both sides and displacement is comparatively small. Ings River fault dips steeply and is probably a tear fault in the Ragged Peak thrust sheet: the sense of movement is unknown.

Unnamed thrusts east of Pass Peak

Lower Cambrian limestone of the White Creek formation is imbricated with Late Proterozoic quartzite and siltstone about eight kilometres east-southeast of Pass Peak. The three thrusts that accomplish this repetition have relatively minor stratigraphic throw of a few hundred metres. They emphasize that this stratigraphic interval is particularly susceptible to deformation by transposition on thrusts as seen also above the Liard thrust, the McConnell thrust and the Big Salmon thrusts. The three thrusts are considered internal to the Seagull thrust sheet and probably merge with the Seagull thrust down dip. They show that the style of deformation of the Lower Cambrian strata in the Seagull thrust sheet is the same as the McConnell-Pass Peak thrust sheet.

Cloutier thrust sheet

Introduction

The Cloutier thrust is considered the next important fault below the Porcupine-Seagull thrust and strata between these two faults constitute the Cloutier thrust sheet (Fig. 62). Generally the sheet carries beds like those of the Porcupine-Seagull thrust sheet. The Seagull group is represented by Black Slate with abundant Felsic Volcanics; the Askin group by the thinner and more variable Barite formation, with a thick sequence of graptolitic rocks in the Platy Siltstone; the Kechika Group includes slate and phyllite with few volcanics referred to the Cloutier formation. Lower Cambrian rocks are represented by the McConnell and White Creek formations of the Ketza River area.

Where exposed, the hanging wall of the Cloutier thrust generally carries lower parts of the Askin group. The footwall of the thrust is in the Seagull group or Hoole formation. The Cloutier thrust is thought to be folded over the same structures as the Porcupine thrust, but it cuts the latter near its leading edge and splays from the Cloutier thrust also cut the Porcupine thrust. Such steeply dipping splays of the Cloutier fault include the McNeil and Ram faults as well as the Ketza fault.

[60] Reviewer's comment:
The description of the Ragged Peak thrust northwestward from its type area seems to fit with the 1:250,000 scale geological map, but a different fault (i.e. one that does not fit this description) has been "connected" to the Ragged Peak thrust in the simplified sketch of Fig 53, and on cross-section O. The description (and geological map 105G) indicate that the Ragged Peak thrust continues into what is called the Cloutier thrust in later sections.

The Twin Lakes fault is considered a tear on the Cloutier thrust that cuts the Cloutier, Porcupine and McConnell thrust sheets. On average the Cloutier thrust sheet is about 2.5 kilometres thick.

Cloutier thrust

The Cloutier thrust consists of three en échelon reverse faults parallel to, and about ten kilometres southwest of, the Tintina fault. The faults are well defined by generally good exposures and are numbered for reference. Cloutier thrust-1 (as labelled on map sheet 5; 105F/9) is traceable from Lapie River to eight kilometres southeast of Ketza River with consistent relations placing strata generally low in the Askin group above beds of the Seagull group or Hoole formation, a throw of about two kilometres. The thrust dips southwest at steep to moderate angles as indicated by the orientation of beds in the fault walls and by the fault trace. Dips are steeper on the southeast than on the northwest. At its northwest end Cloutier thrust-1 nearly merges with the Lower Porcupine thrust, where it is cut by the Lapie fault. The faulted continuation of these merged thrusts may be the Magundy fault, which doubles the Askin group along most of the fault. Cloutier thrust-1 ends in the Hoole formation and presumably doubles these beds for some distance, but merges with Cloutier thrust. Cloutier thrust-2 is about one kilometre southwest of thrust-1 and starts on the ridge between Cloutier Creek and Ketza River in Seagull group strata. It increases stratigraphic separation at Ketza River where it has the same juxtaposition relations and throw as thrust-1. The fault dips steeply southwest. Cloutier thrust-3 begins ten kilometres farther southeast. It is about one kilometre southwest of thrust-2 and has the same throw and relations as the other thrusts.thrust-3 truncates the Twin Lakes fault and cuts the Porcupine thrust where it is folded by the Twin Lakes anticline. thrust-3 also cuts the Ragged Peak thrust[61].

Unlike the Porcupine thrust, the Cloutier thrust remains buried across the strike and may be folded beneath the Porcupine anticline and syncline and other folds (as shown in cross-sections C. D, E, H, J, cross-section sheets), or passes undeformed beneath these structures. If the thrust is folded, it lies about a kilometre below the surface within structures such as the Ketza High and Porcupine anticline, sub-parallel to the Porcupine-Seagull thrust. The Ram thrusts are considered splays of Cloutier thrust in the core of Porcupine anticline (Cross-sections F, G, cross-section sheets). If the Cloutier thrust exists under the Porcupine anticline, its hanging wall is probably below the Askin group there, and most likely in the Kechika or Ketza groups as the fault should be stratigraphically at or below the level where it surfaces northeast of Porcupine syncline. As the Ketza group is exposed in the Ketza High, that structure probably brings the Cloutier thrust to within 500 metres of the surface. The Ketza faults are considered to be steep splays of the Cloutier thrust and tear faults in that thrust sheet.

The change in facies across the Cloutier thrust (for example, thin Barite Mountain formation thrust over thick Seagull group) is as profound as that across the Porcupine and McConnell thrusts and movement must have been considerable and on the same scale as the two higher thrusts. No measure of the displacement is known, but stratigraphic throw is more than 2 kilometres and the cross-strike overlap of the fault is 25 kilometres in some cross-sections. Dip-slip movement may have been on the order of ten kilometres.

The Cloutier thrust probably merges with the St Cyr thrust down dip and is interpreted to do so more rapidly in southeastern than northwestern parts. For instance near Hoole River (cross-section F) the Cloutier thrust probably dips steeply to join the St Cyr thrust ten kilometres down dip. Instead, under the Porcupine syncline, it is flatter and is interpreted to splay 25 kilometres down dip (cross-sections H, J). This implies greater movement on Cloutier fault in northwestern than southeastern parts.

There is no evidence to independently limit the time of movement on the Cloutier thrust, but as it truncates the Porcupine thrust its last movement postdates that fault. Similarly Cloutier thrust predates the St Cyr thrust.

McNeil fault

The McNeil fault is separated into two segments (here called McNeil-1 and -2 for reference, and to emphasize that these two, though speculated to be the same, may be unrelated; Fig. 62). McNeil fault-1 is west of the Starr Creek fault and McNeil-2 lies to the east. McNeil fault-1 truncates the southwest side of the Twin Lakes syncline and anticline (Fig. 53) and repeats the Porcupine thrust. The northeast side has dropped relative to the southwest by about two kilometres judging from the offset on the Porcupine thrust. Its straight trace implies that McNeil fault-1 dips steeply. The Twin Lakes fault offsets it only about one kilometre, suggesting that both faults are essentially vertical and that movement on the Twin Lakes fault was largely dip-slip with minor strike-slip.

McNeil fault-2 juxtaposes the Porcupine thrust sheet next to Seagull group strata in the footwall of the Porcupine thrust; the same combination seen on McNeil

[61] Reviewer's comment:
The continuation of the Cloutier thrust as shown in Figure 62 is problematical southeast of Starr Creek fault. Figure 62 does not match precisely with the distribution of geological units as portrayed on sheet 2 (105G). As shown on the geological map, Cloutier fault-3 would merge with the Ragged Peak thrust. Also, what is figured (Fig. 62) as the southeasternmost part of the Cloutier fault (southeast of the bend of the Ragged Peak thrust) is on the map, one of a family of steep anastomozing faults with relations incompatable with a thrust.

fault-1. Displacement of the two breaks across the Starr Creek fault is four kilometres, suggesting that the McNeil faults dip more moderately (southwest?), than where cut by the Twin Lakes fault and/or that displacement on the Starr Creek fault involved considerable strike-slip. McNeil fault-2 is traced southeastward at least to Lonely Creek through the Indigo Lake area where Gordey (1981a,b) mapped, but did not name it. Gordey (1981a,b) considered that the fault dips steeply and in cross-sections he showed at least three kilometres of northeast-side-down normal movement. McNeil fault-2 truncates the Ragged Peak thrust, the name Gordey (1981a,b) used for the Porcupine thrust in his area. These relations are the same as seen for McNeil fault-1 and implies equivalence of the two.

The McNeil faults moved after the Porcupine thrust sheet was emplaced, but before thrusting ceased in the region. They are offset by the Twin Lakes and Starr Creek tear faults which were active when the Cloutier thrust moved. Therefore the McNeil faults probably had reverse movement. They are most likely steep subsidiaries of the Cloutier thrust, which they may join at depth and in this interpretation are back-thrusts in the hanging wall. This explanation is used for the McNeil faults in the cross-sections.

Ram thrusts

The Ram thrusts are a pair of faults at the heads of Ram Creek, Cloutier Creek and McConnell River on the south limb of the Porcupine anticline (Fig. 28). The faults are traceable for 25 kilometres along strike and dip gently southwest, judging from the orientation of beds in the fault walls. For most of their length the faults are well exposed and clearly defined by repetition within the Askin group, or by superposition of that unit on the Seagull group. Stratigraphic throw is on the order of 500 to 1000 metres. The faults enclose a wedge of the Kechika and Askin groups within the Seagull group, merge to the southeast, and gradually lose stratigraphic offset to the northwest. The wedge of Askin group beds between the faults is folded into an open anticline, which probably formed while the thrusts moved.

The Ram thrusts are considered to be splays of the Cloutier or St Cyr thrusts (cross-sections F, G, H) and largely confined to the St Cyr and Cloutier thrust sheets. They formed when these sheets moved. In this interpretation the Ram thrusts are backlimb faults to the Porcupine anticline. They should then be younger than the Porcupine-Seagull thrust and could conceivably cut those faults or may be deflected by, or merge with, them up-dip. There is no evidence that the Ram thrusts are folded by Porcupine anticline.

Ketza High and Ketza faults[62]

Cambrian and older beds near the upper reaches of Ketza River are exposed in a structural high formed by northeast and northwest trending faults. These have lifted the roughly rectangular Ketza block so that it lies structurally above its surroundings. The block is bounded by eight faults on its northwest, southeast and northeast sides; its southwest side is not broken. In effect it is a large flap with one side unfaulted, like an upward opened door hinged on the southwest. Subsidiary faults involved in the culmination are roughly parallel to the main structures. For reference the faults are numbered the Ketza 1 to -8 and the White Creek fault (Fig. 62 and sheet 5 (map 105F/9).

Ketza fault-1 has a throw of 200 metres on Mount Misery and loses displacement gradually to the north. To the south its displacement is at least 600 metres. This fault terminates against the Twin Lakes fault, but its northeast-side down displacement continues at about 500 metres to its southern end. Ketza fault-2 is parallel to fault-1 with the same sense, and about 600 or 700 metres, of throw. Near their northern end faults -1 and -2 juxtapose different thicknesses of the Askin group. West of fault-2 and east of fault-1 only 100 metres of the Askin group occurs, but between the two, about 300 metres of these beds are present. This may be evidence that faults 1 and 2 are Paleozoic, and that they controlled facies during Askin group deposition.

Ketza fault-3 is a steeply dipping fault that cuts Lower Cambrian rocks and which dropped the south side an unknown, but presumably small amount. fault-4 is apparently its offset continuation, but it seems to have the opposite sense, more displacement, perhaps 500 metres, and cuts younger rocks than fault-3. Faults -5 and -6 are a pair of steeply dipping faults with the northeast sides dropped about 100 or 200 metres and involving only Lower Cambrian strata. Ketza fault-7 parallels this pair, and has the same sense of displacement. All three faults die out along strike by losing throw, and are not replaced by other structures. fault-8 bounds the Ketza High on the northwest and juxtaposes the felsic volcanics of the Seagull group next to Kechika Group and Askin group beds. The throw is 800 metres or more.

Faults -1, -2 and -8 juxtapose different facies or thicknesses of the same units and may have moved in the Paleozoic as well as more recently. Faults -3, -5, -6 and -7 show no such evidence. Fault-4 cuts the continuation of the Porcupine thrust and is probably Cretaceous. The White Creek fault also cuts thrust faults and must have Mesozoic or later movement. Older displacement can not be demonstrated on either of them. Apparently the Ketza high was raised by Siluro-Devonian and Devono-Mississippian faults and by

[62] Alternative interpretations of this structure are Abbott (1986) and Fonseca (1998).

steeply dipping faults, some of which cut the Seagull thrust. Together they combine to elevate the area structurally.

The Ketza High is unique in the project area because it is bounded by a roughly orthogonal fault set, structurally elevated zone along the trend of the Porcupine Anticline. It is most likely the southwest end of the broad Porcupine upwarp rather than an isolated feature. Alternately the Ketza High may be localized above a Cretaceous quartz monzonite intrusion that is not unroofed. In this interpretation it represents the raised and broken roof above the intrusion. This is unlikely because the rocks are not thermally metamorphosed or altered and no dykes are present to suggest an intrusion below.

The Ketza faults appear to be a set of steeply dipping faults, with apparently conflicting evidence for their time of movement and no obvious relationship to the main through going thrusts. Some apparently predate the thrusts and others are later. Nevertheless the faults may be steep splays of the Cloutier thrust that have displaced parts of the Cloutier thrust sheet differently. Faults -1, -2, -5, -6 and -7 are thought to belong in this group and are shown with this interpretation in Cross-sections K and L. Because of their orientation, faults -3, -4 and -8 may be steeply dipping tears in the Cloutier thrust sheet.

Twin Lakes fault

The Twin Lakes fault is an important tear at the east central edge of Quiet Lake map area; it cuts several large thrusts and juxtaposes a variety of Paleozoic rocks (Fig. 62). Although it follows a valley and is not exposed, its locus is closely defined by discontinuity of map-units and structures. The fault is traced northeastward from Mount Hogg for about 25 kilometres and, judging from its straight trace, it is probably vertical or nearly so. It cuts the McConnell thrust, the probable extension of the Seagull thrust and the Porcupine thrust, and offsets these about three kilometres in an apparent left-lateral sense so that the east side moved down and/or northward relative to the western side. Structures on both sides dip generally southwest, but near its northern end the Twin Lakes fault juxtaposes the Twin Lakes syncline and Anticline.

An important difference is seen across the fault in the locus of the hanging-wall glide zone of the McConnell thrust. East of the fault this thrust carries the Kechika Group, west of it the Ketza group. Eastward across the Twin Lakes fault the McConnell thrust is stepped up-section, implying greater northeastward transport on the thrust east of the Twin Lakes fault than west of it. No such change is seen with the Seagull thrust.

Across the Twin Lakes fault the Seagull thrust is matched with the Mount Vermilion thrust and the Seagull thrust-3 is equated with the lower Mount Vermilion thrust.

The Twin Lakes fault is most disruptive with the greatest difference of structures in its northern parts up to its terminus on the Cloutier thrust. In the south it ends imperceptibly and strata are traced around it. The fault may have formed in response to irregularities in the footwall glide zone of its sole, the Cloutier thrust, or may result from unevenness in the strata carried on the thrust. It was likely propagated from the base upward, toward the south, as movement on the sole continued and displacement of structures on both sides increased. In the Twin Lakes fault, as in the Lapie fault, southward and upward propagation involved cutting thrusts that had ceased moving. Alternately the Twin Lakes fault may have started as a tear in the Porcupine-Seagull thrust sheet propagated downward when the Cloutier thrust moved.

Starr Creek fault

The Starr Creek fault (Figs. 53, 62) is a companion to the Twin Lakes fault in west central Finlayson Lake map area. Like the latter it terminates on the Cloutier thrust and is traced for ten kilometres southward. It ends in the felsic volcanics of the Seagull group. The fault truncates Paleozoic strata, the Porcupine thrust and the Twin Lakes Anticline. It is exposed on several ridges and its location is closely defined. As for the Twin Lakes fault, apparent movement was left lateral and about three kilometres, as measured by displacement of the McNeil River fault. Twin Lakes Anticline west of the fault is replaced on the east by a faulted anticline.

Lonely Creek fault

The Lonely Creek fault (Figs. 49, 52, 53), in southwestern Finlayson Lake map area, is traced for 20 kilometres north-northwestward from Mud Lake along the valley of Lonely Creek, a headwater tributary of Liard River. The fault is not exposed and its existence is inferred from differences across the valley. It has apparent left-lateral displacement of eight kilometres as measured by the offset of the McConnell-Liard thrust, and assuming a moderate dip this represents about three kilometres relative downward motion of the eastern side of the fault. The sense and amount of movement matches that for the Twin Lakes fault. Lonely Creek fault also truncates the Mount Vermilion thrust and the Askin and Seagull group strata in its hanging wall, juxtaposing them next to a wide swath of limy phyllite of the Kechika Group. The orientation is not known, but the fault most likely dips steeply. To the north the Lonely Creek fault terminates on the steep dipping McNeil fault. If it is a tear in the Cloutier thrust sheet above the McNeil fault the Lonely Creek fault may extend five kilometres below the surface. Nothing is known about the southern end of the Lonely Creek fault as outcrop is lacking. Structures on opposite sides of the

Lonely Creek fault are not different as they are on either side of the Lapie fault. The beds in the thrust sheets, and the thrusts themselves, dip moderately southwest on opposite sides of Lonely Creek fault.

The McConnell thrust steps northeastward across the Twin Lakes fault and again across the Lonely Creek fault (Fig. 52). This implies that maximum overlap on the McConnell thrust system is in the segment referred to as the Liard thrust and that the McConnell thrust loses displacement northwest from Lonely Creek fault.

Lonely Creek fault truncates the Mount Vermilion thrust (Fig. 53), which brings the Kechika and Askin groups above the Seagull group, but that thrust has not been recognized east of the Lonely Creek fault. It may be present there somewhere in the wide swath of Ram formation which seems too thick to be an unrepeated stratigraphic sequence. If so this hypothetical thrust is in roughly the same hanging-wall glide zone as, but lower in the section in the footwall glide zone than the Mount Vermilion thrust. If the Mount Vermilion thrust is the repetition of the Ragged Peak thrust across McNeil fault-2, it does not recur east of the Lonely Creek fault, because the Ragged Peak thrust and McNeil faults have merged there.

Several important stratigraphic differences across Lonely Creek fault may have localized this break. The Kechika Group loses its large volcanic component across the fault. On the east is shaly limestone of the Ram formation and on the west are roughly time equivalent basalt and non-calcareous slate of the Cloutier and Groundhog formations. The Askin group, thick and well developed west of it, does not reappear on the east. Similarly the Seagull group which has a thick volcanic component on the west side is not exposed again on the east. Further the Triassic strata, which are extensive west of the Lonely Creek fault, are not seen across it. The Kechika Group contains volcanic rocks on the west side but these strata do not recur east of the fault (nor do they to the southeast in Wolf Lake map area). Although the distribution of volcanics may have been naturally irregular, it suggests that the Lonely Creek fault may have controlled facies from the Cambrian through the Mississippian. This feature (possibly an older fault) was reactivated during the Mesozoic deformation .

Changes across the Lonely Creek fault suggest that, if it indeed follows an ancestral fault, that structure probably maintained the east side positive with respect to the west during much of its history. Shaly limestone was deposited on the east, while submarine volcanics erupted on the west in the Cambro-Ordovician. Shallow water carbonate and quartz sandstone were laid down west of the fault, while no deposition occurred to the east in the Silurian and Early Devonian. In the Late Devonian and Mississippian slate and thick felsic volcanics accumulated to the west, while no record of deposition is seen on the east. In the Upper Triassic a thick sequence of silty limestone was deposited on the west, while no record of deposition is seen to the east[63].

The Lonely Creek fault differs from the Lapie and Twin Lakes faults. These last are northeast trending tears, that moved during thrusting and folding, generating profound differences in the structure across them. They are not localized on a pre existing feature. Lonely Creek fault may have a long pre-thrusting history; its north-northwest trend may be inherited from a Paleozoic fault that had the opposite movement of the younger fault and that effected profound changes in depositional patterns across it. The Lonely Creek fault lacks the spectacular contrast of structures across it that is seen on the Lapie and Twin Lakes faults[64].

St Cyr thrust sheet

Introduction

The St Cyr thrust sheet, between the St Cyr and Cloutier thrusts, is the lowest of the four fault slices in the autochthonous rocks southwest of Tintina fault (Figs. 49, 63). The sheet carries higher thrust slices piggyback and has folded and faulted them while it moved.

Integral faults of the St Cyr thrust sheet include its sole, the St Cyr thrust. Important northeast- directed splays like the McNeil and Groundhog-3 faults, southwest-verging branches like the Rose Lake-Fox Creek and Bacon Creek thrusts and Groundhog #4 fault, and several large tears like the Lapie, Twin Lakes and Starr Creek faults. Folds like the Porcupine Anticline-Ketza High, Porcupine syncline and the Twin Lakes folds are genetically part of the St Cyr thrust sheet although they are expressed in the higher Porcupine-Seagull thrust sheet. The St Cyr thrust is interpreted as the surface on which the large folds die out and with which other thrusts merge; it is also the base to the large tear faults. Only a five-kilometre-wide strip of the St Cyr thrust sheet is exposed and the geometry of the slab is inferred from that of the higher slices to be fairly regular and to dip gently southwest. St Cyr fault generally places the Kechika Group over the Hoole formation with stratigraphic separation of two or three kilometres. Its amount of northeast-directed movement is uncertain, but thought to be several tens of kilometres.

In the St Cyr thrust sheet the Kechika Group is represented by the Ram formation, the Askin group by a relatively thin Barite Mountain formation and the Seagull group by Black Slate and Cherty Tuff with

[63] Alternatively, the effects of structural repetition (structural level) combined with modern erosion could have dictated the absence of these beds.
[64] Reviewer disagrees. The contrast in the structural pattern and structural level across Lonely Creek fault are profound.

minor volcanics. The Hoole formation is in the same facies as elsewhere. The facies resemble those in the Cloutier thrust sheet, but the difference in facies across the St Cyr thrust is greater than that across the other thrusts. Many Paleozoic units are difficult to match across the fault. This may be accounted for by greater telescoping across the fault than across others, originally more profound and rapid facies variation, or a combination of both. The St Cyr thrust marks the boundary between Cassiar platform and Selwyn basin facies in Ordovician through Devonian strata.

St Cyr thrust

The St Cyr thrust (Fig. 63) immediately southwest of the Tintina fault was first mapped and named (as St Cyr fault) by Wheeler, Green and Roddick (1960). This report modifies how the name is applied. Their St Cyr fault coincides with that of this report only southeast of Ketza River. To the northwest their mapped trace corresponds with the fault referred to in this report as the Cloutier fault (lower splay). Also, Green and Roddick (1960) did not map the St Cyr fault northwest of Lapie River as is done here. St Cyr fault itself is not exposed, but its locus is well defined and its relations are clear, as exposures of the affected strata are generally excellent.

The St Cyr fault is a remarkably continuous structure, traced from the north-central edge of Quiet Lake map area to south-central Finlayson Lake map area, a distance of 140 kilometres. It can be followed northwest into Tay River map area, for another 40 kilometres at least. Along its length the fault juxtaposes profoundly different, time equivalent, Lower Paleozoic strata, so that fairly characteristic stratigraphic units of the Pelly Mountains are on the southwest side whereas their shaly equivalents are on the northeast. Although stratigraphic ties exist across the fault in the Lower Paleozoic rocks, these are more speculative than those across more southwesterly faults. The profound difference in the lower and middle Paleozoic strata across the fault is not seen in the upper Paleozoic and Triassic beds, which are similar on opposite sides.

Near Lapie River the St Cyr fault dips steeply southwest and appears to be a thrust with older strata above younger, mainly the Kechika Group in its Ram facies above the Cherty Tuff of the Seagull group or above siltstone of Hoole formation. Near Starr Creek, at the Quiet Lake-Finlayson Lake map areas boundary, this relationship is reversed. The fault is vertical and theHoole formation is southwest of older lower and middle Paleozoic beds, so that the offset appears to be strike-slip. Further southeast strata that are roughly time-equivalent, though of different facies, occur on opposite sides. The stratigraphic separation is about 2-1/2 kilometres near Lapie River with the southwest side up relative to the northeast. The separation is about the same, but with the opposite sense near Starr Creek and decreases southeast of there.

For 60 kilometres, from Cloutier Creek southeast to Ings River, the St Cyr fault is bounded on the north by a 100 to 400 metres thick, rusty orange weathering slate (Fig. 32d, e)[65]. In this segment the southern side comprises several stratigraphic units.. From northwest to southeast they are the Hoole formation, the Black Slate and parts of the Askin group. Dextral movement could be indicated by the fact that on its south side the fault cuts up section to the northwest while stratigraphic level remains constant on the northern side.

The change along the fault from apparently straightforward thrust relations on the northwest to more complex relations on the southeast occurs at Cloutier Creek. There a splay (St Cyr fault-2; see sheets 4, 8 and 9 for maps 105F/9, /15 and /16) joins the main fault on its southern side, cutting out the Kechika and Askin groups south of St Cyr fault (Fig. 63). At the same place a second southern splay (St Cyr fault-4) cuts out a panel of the Seagull group and Hoole formation. A third branch (St Cyr fault-3) on the northern side removes the upper Paleozoic beds in the hanging wall of the St Cyr thrust. The St Cyr thrust is therefore a hybrid, a simple thrust at the northwest, which merges with several branches along strike (and down dip?) that change the stratigraphic juxtaposition. The three branches dip steeply; faults 2 and 4 possibly had relatively minor dip-slip on the order of one kilometre, but the slip on fault-3 was considerably more, though unknown.

At its southeast end the St Cyr thrust merges with two of the Hoole faults, so that the panel between the Hoole and St Cyr fault is cut out. A single fault that separates the Harvey and Kechika groups continues southeastward, between the Tintina fault and Black River batholith and is intruded by the batholith on the southeast.

The St Cyr fault is offset one or two kilometres by the Lapie fault. On the northwest the St Cyr dips steeply southwest and consistently superposes the Kechika Group on the Seagull group or Starr formation. The relations are those of a thrust, like the Magundy fault (northwestern continuation of Cloutier fault), with stratigraphic separation of two or three kilometres. In Tay River map area the hanging wall cuts gradually up-section to the northwest, but the footwall does the reverse. Omission of strata thus decreases to zero about nine kilometres northwest of the Quiet Lake and Tay River map areas boundary. Still farther northwest the footwall cuts down-section gradually so that the fault places younger rocks on older, omitting part of the stratigraphic sequence. There the fault no longer looks like a simple thrust, but appears as one developed in folded or faulted strata, or through strike-slip. The

[65] a subunit of the Ankeritic Slate formation

transition is gradual and because the fault cuts obliquely across a syncline the "thrust in folded strata" interpretation seems preferable.

Movement along the St Cyr thrust was probably more than the two or three kilometres of stratigraphic separation across it. Differential sliding of about four kilometres is indicated by offset of the Porcupine Creek syncline across Lapie fault. The distance from the St Cyr to the Fox Creek thrust is about ten kilometres less than that from the St Cyr to the Rose Lake thrust. If the Fox Creek and Rose Lake thrusts are the same and if they root in St Cyr thrust, this indicates that amount of northeasterward differential slip (i.e. on either side of Lapie fault) on the St Cyr thrust. Although, the amount of overlap on the St Cyr is unknown, the profound change in facies of the Paleozoic rocks across it suggests considerable telescoping.

Minimum dip-slip on the St Cyr thrust is given by the total slip on its splays which include the Hoole, McNeil and secondary St Cyr faults. About 3 kilometres were probably taken up on McNeil fault-2. A further four kilometres on St Cyr faults 2 and 4 are seen in another cross-section. As the Cloutier thrust is itself thought to be a splay of the St Cyr, it's dip-slip although uncertain, but possibly ten kilometres, may be added. The total dip-slip on the splays is 33 kilometres.

Rose Lake fault

The Rose Lake fault (Figs.52, 53) crosses the Canol Road near Rose Lake. It dips northeast at 20 to 30° under the Lapie syncline and places Proterozoic carbonate rocks above stratigraphically higher slate and quartzite with throw of about 900 metres. It is interpreted as part of the St Cyr thrust sheet.

On the northwest near the Lapie fault, the Rose Lake fault is well defined and its relations are clear, but it is difficult to follow southwest of Rose Lake and the fault is probably truncated by Nisutlin batholith near the mouth of Upper Sheep Creek. The fault does not reappear beyond this intrusion and presumably dies southeastward.

The Rose Lake fault cuts up-section to the southwest in the hanging wall and footwall indicating that it has the same regionally anomalous vergence as the Fox Creek and Bacon Creek thrusts. Therefore the fault cannot be the folded continuation of the Porcupine-Seagull thrust, but must be deeper. It is interpreted to flatten and merge with the St Cyr thrust down dip. If this is correct it must cut the Porcupine-Seagull and Cloutier thrusts (Cross-sections, C, E, H).

Dip-slip was about 1500 metres as estimated from the cross-sections, the same order of magnitude as movement on the Fox Creek and Bacon thrusts and much less than that postulated for the northeast directed thrusts.

Fox Creek thrusts

Four, branched, northeast-dipping thrusts near the headwaters of Fox Creek and northwest of the Lapie fault are collectively called the Fox Creek thrusts (Figs. 49, 53). They are individually numbered to simplify description (see map 105F/14). The faults dip northeast, with the strata they cut, and repeat Lower Cambrian or Proterozoic rocks. Exposures are good and the faults are clearly defined. At its northwest end, near the Fox Mountain stock the main fault "thrust-l" doubles the gneiss. Without the intervening Lower Cambrian beds it cannot be followed northwest although it may continue some distance. Toward the southeast thrust-1 increases stratigraphic throw to between 500 and 1000 metres. Dip-slip on the Fox Creek thrusts collectively may be two kilometres. About one kilometre movement is indicated in Cross-section A on each of Fox Creek-1 and -2.

Unlike most other thrusts the Fox Creek thrusts cut up-section to the southwest in the hanging-wall and footwall and are therefore southwest directed. They are apparently not the continuation of the Cloutier thrust folded under the Porcupine syncline. The Fox Creek thrusts may be unrelated to the general northeast transport and reflect a southwest directed event, that followed the general northeast transport. More likely they are mechanically connected to the northeast verging thrusts. They are thought to merge with the Cloutier or St Cyr fault at depth and are considered coeval with one of them. This interpretation is depicted in cross-sections A and E.

The Fox Creek thrusts are continuous with the Rose Lake thrust across Lapie fault and the two are thought to be equivalent. Both dip northeast and have the same anomalous vergence although they carry different strata at the surface. The Rose Lake thrust cuts about 500 metres stratigraphically lower than the Fox Creek thrusts. The Rose Lake and Fox Creek thrusts are considered interdependent with the Lapie fault and with the St Cyr thrust.

Bacon Creek thrust

The Bacon Creek thrust (Fig. 63) is about five kilometres west of Barite Mountain. It dips northeast at moderate angles and brings Eocambrian or Proterozoic slate and siltstone above Cambro-Ordovician strata, a throw of about 800 metres. The fault lies between the Fox Creek and Porcupine thrust and is traced for about ten kilometres. It is cut on the southeast by the Bacon Creek stock, a probable mid-Cretaceous quartz monzonite.

The Bacon Creek thrust cuts up-section southwestward in the footwall indicating the same southwest vergence seen also for the Fox Creek and Rose Lake faults, but opposite that of most other thrusts. Therefore the Bacon Creek thrust is considered a part of the Fox Creek, Rose Lake thrust system of southwesterly-directed faults, which are thought to be mechanically connected with the Lapie fault and the St Cyr thrust. Probably the Bacon Creek thrust merges down-dip with the St Cyr thrust (Cross-section A).

Dip-slip movement on the Bacon Creek thrust is probably small, perhaps one kilometre. Although it does not cut the Porcupine thrust in outcrop this relationship is postulated in the cross-section.

Porcupine syncline

Porcupine syncline (Fig. 53) is a broad, open flexure extending about 55 kilometres southeast from the Lapie fault and about 45 kilometres to the northwest. This structure forms a continuous depression 100 kilometres long with an amplitude near 2.5 kilometres and wave length of 15 kilometres. On the map-scale the Porcupine syncline bends thrust faults and the main rock units. It folds the Porcupine thrust and strata in its hanging wall and footwall, but the fold may die downward and the Cloutier and/or St Cyr thrusts may pass undeformed beneath it (Cross-sections C, E).

In the Porcupine syncline faults within the Askin group are irregularly oriented (Fig. 64). Some are longitudinal and others transverse to the fold. The movement was dominantly or exclusively dip-slip. Other faults are oblique to the Porcupine syncline and may have strike-slip. The faults are thought to be compressional and to have jostled the various blocks of competent strata so as to accommodate the fold. The faults tend to end rapidly in the incompetent beds. Décollement or adjustment by sliding has occurred at the Kechika-Askin groups contact (Figs.64, 65).

On smaller scales bedding in competent rocks, like those of the Askin group also outlines Porcupine syncline, as does cleavage or foliation in very fissile or slaty beds, such as those of the Kechika Group. Small-scale transposition folds contained by the foliation show the same reorientation around the fold. However, slate of the Seagull group is not penetratively deformed like that of the Kechika Group and lacks the minor folds and cleavage. In the Porcupine syncline, beneath the Porcupine thrust, bedding in the Seagull group outlines the fold and is bent regularly around it. Cleavage that is locally present is steep-dipping, not re-oriented and probably related to development of the syncline.

During folding the Porcupine thrust may have been locally reactivated as a décollement where the Askin group rests on the Seagull group. These units have important competence contrast and probably behaved disharmoniously as did the stratigraphic contact between the Kechika and Askin groups.

Porcupine anticline and Lapie syncline

The Porcupine anticline and Lapie syncline (Fig. 53) are companion folds southwest of Porcupine syncline, although they are not as continuous, or as clearly outlined, as that structure. Unlike Porcupine syncline, their northwest extent ends at the Lapie fault. Across that fault the three folds are replaced by one. The folds are symmetrical with a wave length of about 15 kilometres and an amplitude of 2.5 kilometres. Porcupine a is defined by folded beds and by the folded Porcupine-Seagull thrust near the Lapie fault (Cross-sections D, E, F, G and H). It runs parallel to the syncline between the Porcupine and Seagull thrusts. Near McConnell River the anticline broadens to become the Ketza High (Fig. 53).

Lapie syncline is outlined near Lapie fault by bedding-dip reversals and by the opposite dip of the Seagull and Rose Lake thrusts on the two limbs of the fold (Fig. 53). The syncline trends southeast, diverging from the Porcupine Anticline, and gradually loses definition on the northern flank of Nisutlin batholith.

The Porcupine syncline and Anticline and the Lapie syncline are late structures, formed after the thrusts that they fold. They are considered to be folds in the Cloutier and Porcupine thrust sheets and may die downward above the St Cyr thrust, as shown in Cross-sections C and E. The folds are thought to be linked genetically to movement on the St Cyr thrust. They are mechanically connected with displacement on the Fox Creek-Rose Lake thrusts, and with slip on the Lapie fault. St Cyr thrust probably moved while the folds formed. The Lapie fault allowed different parts of the St Cyr thrust sheet to fold independently.

Twin Lakes syncline and Anticline

Twin Lakes syncline west of the Twin Lakes fault in east central Quiet Lake map area is a 15-kilometre-long, northwest trending, open fold nearly 5 kilometres across (Fig. 53). It is outlined by the Porcupine thrust and by strata above and below it. The thrust carries the Kechika and Askin groups above slate and volcanics of the Seagull group (Fig. 66). The whole is folded into a broad syncline (Cross-section M) that is roughly on trend with, and probably the continuation of, the Porcupine syncline, though cut off from it by faults in Cloutier Creek and Ketza River. Askin group strata in the core of the syncline are broken by three small northwest trending, near vertical faults which have dropped the axial block of the syncline with respect to the flank roughly 500 metres. The syncline thus has an irregular trace like parts of the Porcupine syncline.

Twin Lakes Anticline replaces the syncline between the Twin Lakes and Starr Creek faults. It is about eight kilometres across and six kilometres long and is also defined by the folded Porcupine thrust and its hanging wall and footwall beds. Kechika Group basalt and slate is folded over Felsic Volcanics and Black Slate of the Seagull group which rests in turn on a core of sandy dolostone of the Askin group.

Lapie fault

The Lapie fault, dips steeply and trends northeast along part of Lapie River cutting Proterozoic to Triassic strata along its 30 kilometre length (Figs. 53, 63). It also cuts the Seagull, Porcupine and Cloutier thrusts and several smaller splays. The structure on opposite sides differs drastically (compare cross-sections C with E) and the shape of the Porcupine thrust sheet changes across the fault. On the southeast side of the fault the thrust sheet is folded in an open anticline-syncline pair, about 35 kilometres across (i.e. maximum distance between St Cyr and Rose Lake faults). Northwest of Lapie fault the sheet is bent into an open syncline, about 25 kilometres across (i.e. maximum distance between St Cyr and southwesternmost Fox Creek faults). The Lapie fault cuts the folds roughly at right angles.

At both ends the Lapie fault terminates against late faults. On the northeast it ends at the Tintoona fault and on the southwest at the Pony Creek fault. The Lapie fault has little displacement northeast of the St Cyr thrust or southwest of the Fox Creek-Rose Lake thrusts. Lapie fault is thought to be related to the general compressional deformation and to have formed as a tear in the St Cyr thrust sheet above the St Cyr and Fox Creek-Rose Lake thrusts. This interpretation is adopted in the cross-sections. The fault probably was extended to the Pony Creek and Tintoona faults during later movement when those faults were active.

If the interpretation of the Lapie fault as a tear rooted on the St Cyr-Fox Creek-Rose Lake thrusts is correct, the Lapie fault moved when these thrusts were active and after most or all movement on the Porcupine-Seagull and Cloutier thrusts. By further extrapolation the Lapie fault was coeval with, and an integral part of, displacement on the St Cyr-Fox Creek-Rose Lake thrusts and relatively late in the thrust-history of this region. If the Lapie fault is a tear in the St Cyr thrust sheet it extends about three kilometres below sea level under the Porcupine syncline.

The mechanical need for the Lapie fault is probably to be found in the northwestward narrowing of the zone between the St Cyr and Fox Creek-Rose Lake thrusts, but the reason for the location of the fault is open to speculation. The fault does not follow a profound facies boundary in the rocks it cuts and, in fact it truncates important facies boundaries at right angles. In the Cloutier thrust sheet an important change is seen in Seagull group strata roughly across the Lapie fault. Southeast of Lapie fault Felsic Volcanics dominate the unit, whereas northwest of it the Black Slate dominates. This change in single unit within a single thrust sheet was probably insufficient to have localized the fault. The Porcupine syncline and Anticline and Lapie fault may reflect a change in the topography of the basement below the St Cyr thrust. This basement may rise toward the northeast roughly below the Porcupine anticline[66].

The Lapie fault is thought to have accommodated relatively minor differential sliding on the St Cyr thrust. Furthermore if the Fox Creek thrusts and Rose Lake fault are equivalent and earlier structures, their trace was offset by five or ten kilometres (Fig. 53)[67]. This implies greater tightening of the Porcupine anticline in the northwest than southeast of the Lapie fault[68]. The up to ten kilometres difference in sliding across Lapie fault indicates the order of magnitude of transport on the St Cyr thrust. It may be tens of kilometres.

Minor Structures in the autochthonous rocks

Slaty units and metamorphosed parts of the autochthonous succession in the project area are generally deformed internally by small scale folds. By contrast the competent rocks generally lack such mesoscopic features. The penetrative flaser fabrics that characterize the transported rocks are absent in the autochthonous beds. Most strain in the autochthonous beds was concentrated on thrust faults, and the intervening strata escaped severe deformation.

Minor structures include small scale folds, and a crenulation cleavage on which the folds are transposed to various degrees. The folds are mostly sub-isoclinal and some tens of centimetres across (Fig. 67a, b, c), with planar limbs, sharp hinges, and apical angles of 30° to 45°. A crenulation cleavage, roughly axial planar to the folds on which the limbs are sheared, is commonly well developed. Generally the cleavage is a set of fractures spaced a few centimetres to several millimetres apart (Fig. 67d, e), but where the rocks are metamorphosed to greenschist or above, micas are recrystallized along it, and the cleavage is a foliation or schistosity. A lineation that results from the intersection of the cleavage with bedding is common in slate, and a crinkle lineation is seen in the phyllitic rocks (Fig. 32a). The cleavage and folds are locally deformed by newer folds, which are invariably more open than the first set (Fig. 67f).

The minor structures are developed throughout the Big Salmon Complex in all rock units. Even the normally competent beds of the Askin group are internally deformed there by structures like those of the

[66] On the Lapie fault there is only apparent dextral offset,of about 2.5 km of both Porcupine syncline and St Cyr fault.
[67] The determination of 5-10 km of differential slip is unclear. The Fox Creek - Rose Lake faults are nearly continuous (although the former has several strands; Figure 53 and Sheet 1 (map 105F)).
[68] This greater tightening is not observed. In fact, the amount of shortening northwest of Lapie fault seems less than southeast of the fault. Northwest of the fault there are fewer thrusts and folds, and in general bedding attitudes are shallower than southeast of the fault.

slaty rocks elsewhere (Fig.67e, g). Around the margins of the complex, in a zone about ten kilometres wide on the northeast side of the Big Salmon and Nisutlin batholiths, the stratigraphically low units, including the Ketza and Kechika groups show this deformation (Fig.67a, b). Farther northeast the small scale folds are restricted to slaty parts of these groups (Fig. 67d) and slate of higher units like the Seagull group lacks minor folds.

Lineation and the axes of minor folds generally trend northwest, and plunge northwest or southeast at shallow angles. They are parallel to large folds and the trace of thrust faults. Cleavage dips generally southwest, but it is folded and follows the large structures. For example, the cleavage in the Groundhog formation, above the Porcupine thrusts, outlines the Porcupine syncline and anticline. The cleavage orientation conforms to that of the thrusts.

In the narrow zone of steeply dipping faults on the southwest side of the Tintina fault low units of the Harvey group, the Canyon and Danger formations, are commonly strongly deformed by small scale folds (Fig. 67f), but the higher units lack these minor structures.

Broadly the internal deformation of the autochthonous beds is most penetrative and pervasive in the Big Salmon Complex and falls off to the northeast. In each thrust sheet the lower slaty units are more strongly deformed than higher, equally incompetent beds, but the same stratigraphic unit is deformed equally intensely in the exposed parts of different thrust sheets. Away from the Big Salmon Complex the degree of penetrative internal deformation therefore increases downward in individual thrust sheets, but apparently not in the entire structural stack.

Steeply dipping faults southwest of Tintina fault

Introduction

Southwest of Tintina fault is a ten-kilometre-wide zone of steeply dipping faults and fault panels distinguished by nearly vertical bedding, conflicting fault relations and by stratigraphy different from most of the Pelly Mountains elsewhere (Figs. 32d, e, 63). This faulted zone, broadly between the Cloutier and Tintina faults, includes the Hoole, Kumquatly, Tintoona, Mount Ross and St Cyr faults[69]. It narrows gradually from northwest to southeast: the distance between the Tintina and Cloutier faults declines from eleven kilometres near Lapie River to six kilometres at the head of Ings River. Individual faults within the zone also trend toward the Tintina fault at acute angles. An example, the St Cyr fault, is six kilometres southwest of the Tintina near Lapie River and less than two kilometres from it, near the Ings River. The Kumquatly fault merges with the Tintina fault to the southeast. As the zone narrows to the southeast some of its faults lose stratigraphic separation and die out. Strata of the Harvey group dominate the steep zone, and although these rocks can be correlated with the Kechika, Askin and Seagull groups, the connections are more tenuous than correlations across other faults to the southwest.

Faults within the zone merge and splay along strike so that this strip of faults is characterized by long, narrow, lozenge-shaped fault-bounded lenses arranged en échelon on the southwestern side of Tintina fault. Individual segments of apparently continuous structures display widely different relations at different places. Some faults are thrusts that dip steeply, but others place younger on older beds and are thrusts formed after the strata were folded and/or faulted; still others may be strike-slip fractures.

On the northwest, beds in the faulted zone dip steeply southwestward but are right-way-up; to the southeast dips are near-vertical and some fault panels have overturned beds. Narrowing of the zone and steepening of the dip toward the southeast suggests that the entire zone may be rotated, more in the southeast than in the northwest.

The only faults traced into the steeply dipping zone are thrusts, and the anomalous relations of the structures in that zone probably resulted from thrusts that followed folding or earlier thrusts The anomalous faults may also have had strike-slip movement related to dextral translation on the Tintina fault. Distinction between these alternatives is difficult; both are possible. Alternatively some of the faults may be normal faults related to graben formation along the Tintina Trench in the Late Tertiary, or be older normal faults lacking physiographic expression. The absence of fault slivers that introduce strata from the northeastern side of Tintina fault in this zone argues against strike-slip faults. Lack of topographic relief on the faults negates significant normal dip-slip; such faults are young

[69] As the author notes, this is a complicated anastomozing zone of steeply dipping faults. The interpretations he presents are driven by three assumptions: 1) that a fault in this zone with a southwest dip (albeit steep), and with beds on the southwest side younger than those on the northeast is a "thrust", 2) that steeply-dipping faults with beds younger on the northeast side than the southwest are "strike-slip", and 3) that faults with both relationships along their length have both "thrust" and "strike-slip" offset.

These presumptions may be incorrect in some cases. There seems no reason to suggest, for example, that a steep southwest-dipping fault with younger beds on the southwest could NOT be strike-slip. This would "allow" the St Cyr fault to be entirely strike-slip, eliminating the mechanical difficulties of linking dominantly thrust and dominantly strike-slip motion along a single structure. Profoundly different broad-scale interpreations might follow. For example, a strike-slip interpretation need not imply that the St Cyr fault is listric at depth, and that it is the basal detachment on which other Pelly Mountain structures to the southwest root. Also, a strike-slip interpretation suggests that the profoundly different stratigraphy of the Harvy Group northeaast of St Cyr fault is a result of strike-slip offset of these strata from a different location, rather than thrust telecoping of a facies boundary.

enough to retain physiographic expression (viz. Tintina Trench).

If some of the steep faults with anomalous relations are thrusts that followed folding or faulting, this argues for the existence of the Tintina fault or some other structural discontinuity at its locus when these thrusts formed. This in turn favours the idea that thrusting and transcurrent movement were not separate events, but that they overlapped in time as they did in place, and that one may have led to the other.

Thrusting is dated roughly as Late Jurassic and/or Early Cretaceous, whereas the transcurrent movement is timed equally imprecisely, as Late Cretaceous and/or Tertiary[70]. Initiation of the transcurrent movement may have been roughly mid-Cretaceous, about the time the last thrust sheet south of the Tintina moved. Indeed the transcurrent fault may be localized where the thrust front was when strike-slip began[71]. The transition from northeast directed thrusts to dextral strike-slip on a northwest trending fault requires a relatively large change in the orientation of the stress field, from a northeast-directed principal stress during thrusting, to one that is more northerly directed for strike-slip.

Tintoona fault

This fault, named for its proximity to the Tintina Trench, angles diagonally from the Kumquatly to the St Cyr fault. It can be traced for 90 kilometres from its join with the St Cyr fault in Tay River map area to where it merges with the Kumquatly near Ketza River. The fault is well defined by good exposures and is a sharp topographic break on most ridges.

Over most of its length the Tintoona fault lies within Siliceous Slate formation. Beds on both sides are correlated with Pelly Mountains strata although the facies differ significantly from other parts of the Seagull group. Although the small stratigraphic throw and similarity of strata on opposite sides implies slight telescoping, the length of the lineament and consistent relations across it suggest important movement. This fault may be a thrust that accommodated later strike-slip movement.

Relations between the Tintoona and Lapie faults show that the Tintoona is younger than the thrusts to the southwest. The Tintoona fault truncates, and is therefore younger than, the Lapie fault. Because the Lapie fault was active while the St Cyr thrust may have still been moving, the Tintoona is evidently younger than both. This does not help to clarify whether the Tintoona is a thrust, strike-slip fault or both.

Kumquatly fault

Between the Tintina and St Cyr faults is the Kumquatly fault named for the campsite of that name at the canyon on lower Lapie River (Fig. 63). This fault is traceable from near Halfmoon Lake in the Tintina Trench (central Finlayson Lake map area) to the north central edge of Quiet Lake map area. Near Halfmoon Lake the fault merges with the Tintina fault and 150 kilometres northwest in Tay River map area it joins the St Cyr fault.

A sharp break can be seen on ridges crossed by the fault and it is well defined by good outcrop. Although the slaty rocks on both sides are penetratively transposed and folded these structures are cut by the fault with little or no shearing along the margins. The fault is a narrow zone, only a few metres wide at most, beyond which there are no penetrative effects.

Consistently the fault juxtaposes black graphitic slate of the Danger and Siliceous Slate formations with the Ankeritic Slate formation. Although the rocks are thought to be partly time-equivalent, the relations between them are unknown. Unlike breaks to the southwest the Kumquatly fault lacks thrust relations anywhere along its length. It is essentially vertical and the same units remain in each fault wall. The fault obliquely truncates the Danger anticline, but otherwise cuts no large structures. The southeastern half of the fault is isolated and no other structures join it, but on the northwest several smaller faults merge with it and the Tintoona fault angles into it near Ketza River.

The nature, timing and amount of movement on the Kumquatly fault are unknown. No offset strata can be matched across it to demonstrate strike-slip and the fault is not obviously a thrust. It is a 150-kilometre-long, straight, vertical fracture that joins the Tintina, a dextral strike-slip fault, to the St Cyr fault, a steep reverse fault 'with northeast directed movement. The Kumquatly fault may have displacement of the same sense as either or both faults and may be a thrust with some dextral slip. It juxtaposes strata that are difficult to correlate with strata farther southwest, but the telescoping across the fault is probably less profound than across the St Cyr fault. The stratigraphy is equally unlike that of the Anvil Range. A match for the distinctive strata between the Kumquatly and Tintina faults should be sought outside the map area, perhaps to the northwest in Glenlyon map area.

A fault that parallels the Kumquatly, about one kilometre to the northeast, separates black sooty and black siliceous slate. This fault is a splay or a subsidiary

[70] It is now established that offset along Tintina fault of about 400-430 km is latest Cretacesous to Tertiary (post 67 Ma) (Gabrielse et al, 2006), not mid-Cretaceous (i.e. 100 Ma

[71] Strike-slip (latest Cretaceous-Paleogene, post 67 Ma) was a few tens of millions of years after thrusting (Early to mid-Cretaceous) as summarized in Gabrielse et al, 2006.

of the Kumquatly fault thought to follow a stratigraphic contact. The displacement is unknown.

Possibly the Kumquatly fault is transitional between the thrusts and the strike-slip fault in movement and time. It may have accommodated moderate telescoping of facies and been steepened to accept early strike-slip of the Tintina fault system during the early Late Cretaceous.

Hoole faults

Three or four kilometres southwest of the Tintina fault in western Finlayson Lake map area are the Hoole faults (numbered Hoole-1 to Hoole-4 on Fig. 63). They are near-vertical structures in the set of steep faults adjoining the Tintina fault. The faults merge and splay and are probably closely related and hence treated together.

Hoole-1 can be followed for 90 kilometres from Ings River northwest to near the Ketza River across the boundary between Quiet Lake and Finlayson Lake map areas.

Over the last 15 kilometres near its southeast end fault-1 loses separation rapidly and dies in the Kechika Group. There it also cuts, or merges with, the Ragged Peak thrust which loses separation in the same direction and over a similar strike distance.

Although it may have strike-slip movement fault-1 behaves like other thrusts in the region following strata for long distances along strike and migrating from unit to unit gradually. It also consistently places younger rocks south of older beds and does not introduce drastically different rocks. It is therefore perhaps best considered as a fault with reverse and dextral movement and may be transitional between the thrusts and the Tintina fault. If the transition from reverse to strike-slip motion is gradual in space so that some faults have characteristics of both, the strike-slip probably followed the thrusting directly without a break in time. The change probably occurred about the mid-Cretaceous or early in Late Cretaceous time. This is dated indirectly as there are no closely limiting stratigraphic data. In cross-sections M, O, Q and S, the Hoole-1 fault is drawn as a splay of St Cyr fault, with between one and two kilometres of dip-slip.

Hoole -2, a northern splay of fault-1, brings a near-vertical panel of the Askin group southwest of an equally steeply dipping slice of the Black Slate - a relatively minor stratigraphic throw of about 500 metres. The fault is 25 kilometres long and ends on the southeast where it merges with Hoole-3. Hoole-3 has stratigraphic relations and separation similar to fault-2, but on its southern side it cuts obliquely across a northwest trending anticline outlined by the Askin group between Hoole3 and -4. Hoole-4 joins Hoole-3 at both ends and cuts obliquely across the strata on its northern side. Stratigraphic separation on this fault is also relatively small, but variable. Hoole-1, -2, -3 and -4 may have had some strike-slip, but they are considered steep splays above the St Cyr thrust with reverse dip-slip of about five kilometres on each of Hoole-3 and -4 and somewhat less on Hoole-1 and -2. They are splays on which movement along the sole fault of the St Cyr thrust is distributed. Their combined movement (± 16 kilometres) must be accommodated on the St Cyr thrust elsewhere down-dip and along strike to the northwest.

Tintina fault system

Tintina fault

Within the map area the name Tintina fault is applied to a single profound lithologic break at the mountain front southeast of Tintina Trench as was done by Wheeler et al. (1960). This break is considered to be the main fracture of the Tintina fault system that extends from east central Alaska to the British Columbia-Yukon border. Outside the project area the name is used in a broader sense, perhaps for several, or a number of, parallel and en échelon faults instead of one continuous break. In Tay River map area the same structure was named the Buttle Creek fault (Tempelman-Kluit, 1972), but because that name fails to emphasize the relative importance of this fault it is not used here.

The fault is at the base of the spectacular front of the Pelly Mountains which rise as an abrupt, straight wall nearly 1000 metres above the floor of Tintina Valley, a kilometre or less from the fault. The mountain front is probably the scarp of the fault. Southeastward in central Finlayson Lake map area the fault follows a straight valley, about two kilometres wide, and the mountains rise sharply on both sides, though they are generally higher on the southwestern side.

On the southwest side in both map areas the Tintina fault offsets beds of the Harvey group for 180 kilometres. There is more variety on the northeast side, which in Quiet Lake map area is formed by quartz feldspar porphyry of possible Cretaceous or Tertiary age, and in Finlayson Lake map area by Proterozoic or older gneiss, last metamorphosed in the Cretaceous. The fault cuts two different plutons of Cretaceous quartz monzonite on opposing sides in southeastern Finlayson Lake map area. Gneiss northeast of the fault resembles that of the flanks of the Nisutlin and Big Salmon batholiths, but the Harvey group of the southwest side has no counterpart in the project area on the northern side. The Late Tertiary or Quaternary volcanics northeast of the fault are restricted to Tintina Valley and probably filled a fault depression.

Generally the fault is covered by overburden, but in Finlayson Lake map area it is exposed in a small canyon of a northeast flowing tributary to Hoole River 19 kilometres northwest of Halfmoon Lake. There the fault (or a shear zone parallel to it) is a 30-metre-wide vertical zone of gouge and sheared black graphitic slate. The slate is disrupted by innumerable, irregularly spaced slip surfaces a few millimetres or centimetres apart, but its most remarkable feature is its commonplace appearance which gives little clue to the enormity of movement. Rocks in the fault walls are not sheared; gneiss only half a kilometre from the fault looks unaffected by the fault and lacks obvious minor structures. The slate on the other side is penetratively cleaved and folded, but the minor structures predate the fault.

Although the Tintina fault has remarkable physiographic expression that can be accounted for only by youthful (Late Tertiary) relative uplift of the southwest side, it also marks a profound bedrock change, which can not have been accomplished by uplift alone. Broadly, the metamorphic rocks northeast of the fault in Finlayson Lake map area can be matched on the southwest side with rocks between Macmillan River and the Yukon-Alaska border or farther northwest. These are between 350 and 650 kilometres away. Similarly, the Paleozoic strata of the Pelly Mountains southwest of the fault system compare best with beds northeast of the Rocky Mountain Trench in northeastern British Columbia near Tuchodi Lakes, some 500 kilometres to the southeast. They have no match northeast of the fault less than 350 kilometres away. On this basis the Tintina fault has a dextral strike-slip of between 350 and 650 kilometres.

The Tintina fault is probably the main strike-slip structure of the Tintina fault System in the project area, but others like the Kumquatly, Tintoona and Hoole faults are associated structures on the southwest side. The Grew Creek and Danger Creek faults are similarly related northeastern breaks. Unlike the Tintina fault, however, each of those faults juxtaposes kindred rocks, whereas the Tintina brings gneissic and granitoid rocks against unrelated types.

Near parallelism of the Tintina with thrust faults and folds on its southwest side and the constancy of the southern fault wall in the same rock unit for 180 kilometres implies that an earlier northwest trending inhomogeneity may have controlled folds, thrusts and the location of the Tintina fault System. This inhomogeneity may be the facies transition from the Seagull, Askin and Kechika groups on the southwest to the Harvey group on the northwest. This profound facies change in the Paleozoic strata is a northwest trending stratigraphic discontinuity now generally less than ten kilometres southwest of Tintina fault. The facies change itself is most likely positioned on a Paleozoic normal? fault. In regional terms the transition southwest of Tintina fault, when restored is possibly the boundary between Cassiar platform and Kechika Trough

In summary, the Tintina is localized near the most profound facies boundary in the Paleozoic rocks. This boundary may itself be controlled by Paleozoic faults. The thrust front was perhaps stalled at this ancient fault controlled boundary when strike-slip began[72]. Hints of old structures which may have localized the northeastern side of Tintina fault are lacking in the metamorphic rocks of Finlayson Lake map area. Metamorphism may have destroyed this evidence if it existed, or the postulated fundamental break may not be expressed at surface in the allochthonous rocks but may be restricted to the underlying autochthonous strata. Instead the northeastern side of the Tintina in the project area is behind (southwest of) the metamorphic culmination of the Mink complex. Generally the trace of Tintina fault lies close to the boundary between involved and passive basement, except opposite Cassiar platform (Fig. 68).

Evidence for the age of the strike-slip on Tintina fault within the project area is limited, and complicated by the problem that the strike-slip was on the same surface as the later dip-slip which retains physiographic expression. Best regional evidence suggests that the Tintina fault is a Late Cretaceous and/or Early Tertiary strike-slip fault along which a late Miocene or Pliocene graben, Tintina Trench, is located Roddick (1967) was the first to document the dextral sense and magnitude (420 km) of displacement of the fault and he presented evidence to show that most of the movement was Late Cretaceous. He postulated some early Paleozoic displacement on the fault. Corroboration of the amount and timing of movement is given in Tempelman-Kluit (1970, and 1972a).

Reconstructions of the distribution of facies in Lower Cambrian, Cambro-Ordovician, Siluro-Devonian and Upper Triassic strata and of mid-Cretaceous plutonic and metamorphic rocks (Fig. 68) show that these strata are each displaced equally by roughly 450 kilometres across the fault. The thrust front of allochthonous rocks is displaced by 500 kilometres (Fig. 74) on the Tintina fault and the boundary between passive and involved basement as mapped in Figure 74 is offset by 390 kilometres. This suggests that all the right lateral displacement postdates the mid-Cretaceous (the plutonic rocks).

The evidence for Pliocene normal movement along Tintina fault, discussed in Tempelman-Kluit (1980), is largely physiographic and is based on the comparative youth of fault scarps and the age of some sedimentary rocks within the trench.

[72] :It is now well established that strike-slip (latest Cretaceous-Paleogene, post 67 Ma) was a few tens of millions of years after thrusting (Early to mid-Cretaceous) as summarized in Gabrielse et al, 2006

Grew Creek fault

The Grew Creek fault, recognized and named in Tay River map area (Tempelman-Kluit, 1972) extends into northern Quiet Lake map area. It separates probable Tertiary sedimentary strata on the northeast from quartz feldspar porphyry possibly of similar age. On the west side of Lapie River the fault occupies a covered interval, about 20 metres wide, between shattered outcrops of acid volcanics and gently south dipping beds of immature sandstone. Its straight trace implies a steep dip and the northeast side is dropped an undetermined amount relative to the southwest. Last movement is considered to be Tertiary. Eocene beds on the northeast side are folded and tilted near the fault; this final displacement may have been dip-slip. On the southeast the Grew Creek fault probably merges with the Tintina fault, but exposures are poor and the trace of the fault is speculative southeast of Lapie River.

Klippen of Nisutlin and Anvil allochthons southwest of Tintina fault

Introduction

Disconnected subhorizontal slices of the allochthonous rocks are preserved in structurally low, topographically high areas southwest of Tintina fault (Figs. 49, 52). They include the Dunite Mountain, Mendocina Creek, Quiet Lake, St Cyr and McNeil klippen; altogether there are 14 separate structural outliers, ranging from one or two square kilometres, to nearly 1000 square kilometres. The klippen are probably erosional remnants of originally extensive thrust sheets placed above the imbricated autochthonous strata. Although the maximum preserved thickness of the transported rocks is slightly more than one kilometre, the total thickness and number of thrust sheets remains unknown.

The klippen comprise Nisutlin and Anvil allochthons thrust over Paleozoic and Mesozoic autochthonous beds. Generally rocks of the klippen are more metamorphosed and more penetratively deformed than the strata below. The transported rocks record an episode of Triassic and Lower Jurassic strain and metamorphism prior to thrusting. The basal thrust of the klippen is a sharp contact with localized strain. In this respect the geometry of the base is like the thrusts in the autochthonous succession and unlike the strain surfaces within the transported rocks where ductile deformation is distributed through the entire sequence.

Southwestern klippen lie above the Nasina formation and northeastern outliers above the Hoole formation, as the basal thrust of the klippen apparently cut up-section toward the northeast in the footwall. The klippen are remnants of one or more thrust sheets that originated southwest of Teslin Suture. Some were transported northeast at least 115 kilometres.

St Cyr klippe

St Cyr klippe (Figs. 49, 52) extends 70 kilometres northwestward from the southeast corner of Quiet Lake map area. It occupies a 15-kilometre-wide strip southwest of Nisutlin batholith and is the structurally highest preserved part of St Cyr syncline. The rocks and relations are well exposed on and near Mount St Cyr, but outcrops southeast of Nisutlin River are sparser and relations there are less clear. St Cyr klippe includes a lower slice of Nisutlin assemblage and an upper slice of the Anvil assemblage, but the two sheets are discontinuous, so that only the higher is represented in the northwest part of the klippe.

Near Mount St Cyr, on the north side of the klippe, basalt of Anvil assemblage lies directly on Black Slate of the Seagull group in the top of the McConnell thrust sheet. The basalt is a resistant greenstone to amphibolite in which primary structures and textures are destroyed and disrupted by shear zones up to a metre wide and spaced tens of metres apart and subparallel with the base of St Cyr klippe. Irregularly oriented small faults also cut the basalt. Serpentinite lenses, several hundred metres thick and several kilometres long, are enclosed by the basalt. These are oriented generally parallel with the foliation of the rocks and the base of the klippe. The fault is also roughly parallel with bedding in the autochthonous strata, although on the map-scale it cuts out units.

The contact zone between the overlying transported rocks and the autochthonous is abrupt, about a metre thick, like thrust-fault contacts in the autochthonous beds. Relations at the base of the basalt south of Tower Peak are not exposed, but the basalt is interpreted to lie above graphitic siltstone of the Nasina formation. This assumes that the thrust beneath the St Cyr klippe cuts up-section northeastward in the footwall and northeastward transport of the klippe.

Along much of its northeast edge St Cyr klippe is intruded by the Nisutlin batholith whose contact dips moderately southwest under the klippe. The basalt and serpentinite are thermally metamorphosed near the granitic contact. A large xenolith of serpentinite that caps the ridge between Big Creek and Nisutlin River is presumably an outlier of St Cyr klippe.

Along the southwest edge of St Cyr klippe a slice of metamorphosed rocks of Nisutlin assemblage lies between the autochthonous strata of the Nasina and Hogg formations, and the Anvil assemblage. The basal surface is the same thrust that floors the basaltic and ultramafic strata to the northeast. The contact between the two allochthonous assemblages is not exposed, but is inferred to parallel the metamorphic strain fabric in

Nisutlin assemblage. It may be a splay of the basal thrust, but is more probably an older contact on which the two assemblages were brought together, before being thrust into place.

St Cyr klippe continues into Teslin and Wolf Lake map areas south of the project area, but the extent of the klippe there is unknown. Northwestward along St Cyr syncline the klippe terminates up the plunge 10 kilometres west of Mount St Cyr, but it reappears 40 kilometres farther northwest, at Dunite Mountain. The Dunite Mountain klippen are probably outliers of the St Cyr klippe along the syncline. Similarly the McNeil klippen are outliers of the St Cyr klippe across strike on the northern side of Big Salmon-Nisutlin Arch. The maximum preserved thickness of the St Cyr klippe is about 1200 metres as measured from cross-sections.

McNeil klippen

Eight subhorizontal klippen of metamorphosed rocks in the southwest corner of Finlayson Lake map area and adjacent Quiet Lake map area are collectively called the McNeil klippen and numbered for description (Fig. 52). They were first recognized by Wheeler, Green and Roddick (1960a,b) and they are described in some detail elsewhere (Tempelman-Kluit, 1979a; Gordey, 1981a,b). The klippen place muscovite-quartz blastomylonite of Nisutlin assemblage onto undeformed unmetamorph-osed strata of the Seagull group or Hoole formation. The penetrative fabric of the cataclastic rocks dips gently, and conforms roughly with the basal thrust, in the same way the fault follows beds in the autochthonous strata for long distances although it cuts them on the map scale. The klippen are up to 500 metres thick and are remnants of a sheet originally less than 2 kilometres thick. The degree of shearing in the klippen varies and immature conglomerate, the protolith for some (or all?) of the mylonite is preserved locally within the klippen.

The basal contact of the klippen is commonly exposed as an abrupt break without imbrication. Like the contact beneath the St Cyr klippe it is a thrust that postdates the penetrative strain and metamorphism of the rocks in the transported slice.

Strata of the McNeil klippen are immature sandstone that may be upper Paleozoic and Triassic. They were sheared and metamorphosed in the Triassic and thrust into place about the Late Jurassic. The depositional age of the rocks is unknown because no fossils were collected from them. The same rocks in Tay River map area contain Triassic conodonts (Tempelman-Kluit, 1979b). The penetrative strain is dated by two K-Ar determinations on muscovite from McNeil klippen which gave Lower Triassic ages of 230 and 226 Ma (Fig. 49; Appendix 3). Elsewhere in the allochthonous beds the penetrative fabric has given a range of ages, some as young as Lower Jurassic.

Sandstone preserved near the base of the McNeil klippen 1 and 7 can be traced laterally into mylonite within the klippen over distances of a few tens of metres. The sandstone also contains clasts and grains of mylonite. It was apparently deposited when cataclastic rocks were available for erosion, but before shearing had stopped, as it is itself sheared. The depositional age of the rocks may therefore coincide roughly with their Early Triassic time of metamorphism and ductile strain given by the K-Ar ages.

The youngest autochthonous rocks below the McNeil klippen are Upper Triassic, but the Hogg formation underlies klippe-1. The allochthonous rocks are assumed to be thrust northeast from Teslin Suture Zone, the nearest region where similar strained rocks are found. From Teslin Suture to McNeil klippe-6 is 115 kilometres measured perpendicular to the Teslin Suture and this is considered the minimum overlap of the transported rocks over the autochthonous ones.

The McNeil, St Cyr and Dunite Mountain klippen are considered outliers of a single allochthonous sheet, because they have the same relations to the autochthonous beds and because they can be joined together reasonably in cross-section. This single allochthonous sheet is traceable along strike for 125 kilometres and across trend for 55 kilometres; the maximum preserved thickness in this project area is about one kilometre. Its area compares to that of the Seagull-Porcupine thrust sheet, the second lowest slice within the autochthonous beds (Fig. 52).

If the McNeil, Dunite Mountain and St Cyr klippen overlie the same thrust, that fault cuts up-section northeastward in the footwall and remains in roughly the same strata along trend. This is compatible with the apparent northeastward transport of the allochthonous beds. In the hanging wall the fault cuts a structural succession, not a stratigraphic sequence. It cuts this succession from southeast to northwest across the postulated transport direction. This may reflect irregularities in the strained structural succession or may indicate a northwestward component to the transport.

Dunite Mountain klippen

Two slices of serpentinized peridotite cap adjacent peaks astride the boundary between Quiet Lake and Lake Laberge map areas. They are called the Dunite Mountain klippen after Dunite Mountain, which was named by Roddick during 1960 reconnaissance mapping. Each klippe covers about 20 square kilometres and 300 metres of serpentinized peridotite is preserved as subhorizontal slabs above graphitic quartzite of the Nasina formation (Figs. 26 a, b). The basal contact is sharp and subparallel with bedding in the strata below. Serpentinite is generally massive, but displays a flaser or fish-scale texture locally.

The Dunite Mountain klippen are structurally disconnected and erosionally preserved remnants of part of the Anvil assemblage, like the serpentinite lenses in St Cyr klippe. They are about the same size as the largest serpentinite in St Cyr klippe. The serpentinite is probably upper Paleozoic and was probably derived from the Teslin Suture Zone, about 15 kilometres to the west, where slices of serpentinized ultramafic rocks are interfoliated with other representatives of the Anvil assemblage.

Mendocina klippen

Two disconnected remnants of a slice of quartz muscovite schist, the Mendocina klippen, form a west-southwest trending synform near upper Mendocina Creek (Figs.49, 52). The synform is 1 or 2 kilometres wide and is traceable for 16 kilometres to the west edge of Quiet Lake map area. The schist is a part of Nisutlin assemblage with a strong flaser fabric which dips steeply or vertical. The schist lies above graphitic siltstone, quartzite and limestone of the Nasina formation; the bedding in these rocks also dips steeply into the syncline.

The Mendocina klippen and the autochthonous beds below are folded between the Quiet Lake batholith and Dycer Creek stock. All the strata are thermally metamorphosed so that the contact between the transported rocks and the Nasina formation is a sharp, "normal" looking boundary without signs of dislocation or shearing.

Quiet Lake klippe

On the low ridge between Quiet Lake and Nisutlin River are massive greenstone, amphibolite and greenschist, which are considered part of the Anvil assemblage. If this correlation is valid the rocks are part of another structural outlier, referred to as the Quiet Lake klippe. The klippe is probably floored by a subhorizontal thrust and probably cut on the east by the Scurvy fault, but as the contacts are not exposed these relations are hypothetical. Quiet Lake klippe is thought to be thrust over the Nasina formation, the nearest autochthonous rocks exposed, and thought also to be the continuation of St Cyr klippe dropped about 600 metres across the Scurvy fault (Cross-section K).

Minor structures in the allochthonous rocks

Minor structures of the transported cataclastic rocks northeast of Tintina fault area were described in detail by Erdmer (1982). Those of the McNeil klippen in southwestern Finlayson Lake map area were discussed by Gordey (1981a,b). A summary follows.

The transported rocks range from protomylonite to mylonite, blastomylonite and mylonite schist. Their penetrative cataclastic fabrics dominate other features and prevail on every scale. The fabrics include a strong foliation, colour and compositional lamination, strong lineation - generally a quartz rodding, and small scale folds transposed on locally well- developed crenulation foliation. These fabrics are developed to different degrees in different parts of the transported sequence. Marked variation in the planar fabrics is seen from place to place, along and across strike. Although it has not escaped strain, the fabrics are not as well developed in the Anvil assemblage as in the more siliceous Nisutlin and Simpson assemblages.

Generally one foliation dominates at any place so that the rocks have a single schistosity which coincides with the colour and compositional lamination. The foliation varies from a fine flaser fabric to a coarse schistosity or gneissosity (Fig. 41). In mylonite and protomylonite the foliation is extremely closely spaced and defined by parallel layers of crushed and recrystallized minerals that bend around the few larger, less broken grains. By contrast the blastomylonite or mylonite schist has a schistosity that reflects form-oriented recrystallized micas and quartz. In mylonite, grains (commonly about 0.005 mm across) are strung out and result in a very fine and intense fabric. In blastomylonite the mica plates are commonly 0.05 millimetres thick and two to five times as large in the plane of the fabric; quartz grains are somewhat larger, leading to a coarse fabric. In amphibolite the foliation is given by strongly oriented actinolite needles about 0.01 millimetres thick. The same range in grain size and consequent variation in coarseness of the fabric is seen as in the more siliceous types, and the rocks range to amphibole gneiss.

Colour lamination or streaking is compositional and results from variation in the proportion of micas and quartz-feldspar matrix between laminae (Fig. 41). Quartzo-feldspathic laminae are generally thicker than the micaceous layers. In mylonite the laminae are thinner than a millimetre, but in the mylonite schist the layering may be two or three millimetres thick. The laminae are closely parallel, and in detail, laterally discontinuous. They coalesce and split in characteristic anastomosing patterns along their length. Colour lamination is seen mainly in the coarser grained varieties of amphibolite; it results from segregation of feldspar and amphibole into separate laminae one or two millimetres thick.

Form-oriented, elongate quartz grains in the blastomylonite and mylonite schist also give the rocks a spectacular and common rodding lineation. Quartz grains are ovoids whose axes may average in the ratio 1: 2: 4. The lineation is strong in quartz-rich layers, and an accompanying crinkle lineation is seen locally in micaceous parts. Lineation formed by preferred orientation of acicular actinolite is present locally in the amphibolites. Lineation generally trends 70 or 80° east of north and is subhorizontal.

The planar fabrics are commonly folded over small-scale folds, which range from open and upright to isoclinal and recumbent flexures. The two fold types are not separate phases, but gradational. They are considered the arrested stages of a deformation continuum carried to various degrees of completion in different places. Open folds have limb angles near 90°, sharp hinges, amplitudes of several centimetres and steeply dipping axial planes along which recrystallization is rare. Tighter folds are progressively more recumbent, with rounder axes and amplitudes of many centimetres or tens of centimetres. These folds are generally transposed on a newer crenulation foliation which has cut the limbs and localized recrystallization. In places the folds are isoclinal and the new foliation is sub-parallel with the older. The vergence of the subisoclinal folds is generally to the northeast and their axial planes dip southwest, more steeply than the deformed fluxion structure. Axes are subhorizontal and trend eastward. The small-scale folds seem to be haphazardly distributed; few are seen in large areas and they are common in other, more restricted, parts. Locally the open folds are superposed on the subisoclinal types, but generally one style is seen at one place. Folds are more common in the mylonite schist and blastomylonite than in mylonite or protomylonite. Open folds are commonest in micaceous blastomylonite wheras tighter folds are seen in the quartz-rich varieties. Folds of either type are rare in the amphibolite.

Structures northeast of Tintina fault

Introduction

The project area northeast of the Tintina fault is dominated by the allochthonous rocks (Fig. 49) having the same penetrative cataclastic fabrics seen southwest of the Tintina. Cross-sections emphasize the strikingly low structural relief, generally less than two kilometres, which contrasts with that of the autochthonous rocks (Fig. 50). Two important contacts, the base of the Anvil assemblage and the top of the gneissic rocks, traceable through most of the area, outline the structure. Foliation and flaser fabrics are subhorizontal or dip gently; they are broken by younger steep faults, folded around broad metamorphic-plutonic domes and deformed into wide, open synforms which trend in various directions. The structural lows expose the highest slices in the succession such as the Money, Finlayson, North and Wolverine klippen. Down-faulted blocks, like the Simpson Range, similarly contain high structural elements. By contrast the domes, like the Mink High, introduce the lowest strata and may indeed expose upwarped crystalline basement beneath the transported, sheared rocks. The domes surround roughly equant Late Cretaceous metamorphic-plutonic cores and the structural lows are the irregularly oriented depressions between them. The three allochthonous assemblages were most probably stacked above each other before being thrust into place as a single sheet. Alternately they may be separate thrust sheets, emplaced above each other during the general northeastward thrusting Because they are not imbricated with the autochthonous strata, and because their fabric and metamorphism contrasts with that of the in-place beds, this is unlikely. The maximum thickness of the transported rocks is estimated as two kilometres.[73]

On the northeast the allochthonous, sheared and metamorphosed rocks are thrust over the essentially unmetamorphosed, unsheared Nasina formation with an abrupt contact; the relations are postulated to be the same as southwest of Tintina fault (Fig. 51).

The autochthonous rocks northeast of Tintina fault in Finlayson Lake map area are not imbricated by thrust faults like those southwest of Tintina fault. Instead they are bowed into folds, arched over granitic intrusions, and broken by steep faults that accommodated relatively minor shortening. Fortin Arch, the main structural feature of these rocks, is a five-kilometre-wide northwest-trending block raised between two steep-dipping faults. It exposes hornfelsed Proterozoic slate and siltstone around the oval shaped Fortin stock.

Discussion

Fabrics, minor structures and textures in the strained allochthonous rocks reflect arrested development in a period of concurrent and repeated strain and metamorphism. Mineral textures demonstrate the relative timing and dominance of two opposing processes, one of grain destruction by shear, the other of grain growth by metamorphism. The rocks are commonly made up of metamorphic minerals (Fig. 41d, e) without original detrital grains or volcanic or plutonic crystals. Yet these metamorphic minerals are themselves bent, crushed and strained in most thin sections, showing that cataclasis postdates metamorphic mineral growth. In the same rocks the crushed groundmass is recrystallized and healed in places and in other rocks metamorphic minerals are grown across the fluxion structure. Metamorphic regrowth therefore also followed at least some of the strain. These ambivalent relations are common and show that the mechanical deformation and metamorphism were not separate events, but that they proceeded a pace; in some instances one outlasted the other and elsewhere the other prevailed.

[73] There is considerable geological information for this area. Many references are cited in Murphy et al. (2006).

Variation in the grain size of the deformed metamorphic rocks reflects which of the two opposing processes dominated the last stages. Where grain size is comparatively large, recrystallization outlasted strain, and where the rocks are finer grained recrystallization is not as advanced, or cataclasis dominated the final history.

K-Ar ages of biotite and muscovite in the ductile deformed rocks from in and outside the project area range from 160 to 230 Ma (Fig. 49). The ages show no recognized patterns in terms of mineral dated, geographic location or position in the structural "stack". In the project area metamorphic muscovite from Simpson Allochthon gave 183 ± 7 Ma and 201 ± 7 Ma, while metamorphic micas from Nisutlin Allochthon gave 226 ± 8 and 230 ± 8 Ma. Micas from samples of equivalent rocks in Yukon Crystalline Terrane south and west of Dawson, about 400 kilometres west of the project area, gave ages in the same range, namely 160, 161, 168, 175, 178, 181, 182, 187, and 202 Ma. The two dates of 137 and 138 Ma are thought to be partly reset (Tempelman-Kluit and Wanless, 1975). The micas may have grown and cooled over a long period, which may argue for a similarly lengthy interval of metamorphism and ductile deformation. Alternately the range in the mica ages may show that the minerals contain significant and variable amounts of inherited radiogenic argon. If inherited argon is the explanation, the youngest age, 160 Ma, is presumably closest to the time of deformation. Because the dated micas are the products of metamorphic growth the temperature should have been sufficient to drive off earlier radiogenic argon, so that large amounts of inherited gas are unlikely. The age range is therefore thought to reflect cooling through Late Triassic and Early Jurassic time. It is debatable if the spread in cooling ages also argues for long-lived cataclasis and metamorphism. By itself the long range of the cooling ages does not signify protracted ductile strain and metamorphism, but fossil evidence from an unsheared precursor of some of the deformed rocks does support that conclusion.

Immature sandstone, with abundant detrital grains of mylonite and blastomylonite (Fig. 42), occurs as slightly sheared lenses within mylonite of the Nisutlin assemblage on both sides of Tintina fault near the preserved leading edge of the transported strata (Tempelman-Kluit, 1979). It contains Upper Triassic conodonts at one place near Faro. Elsewhere its detrital muscovite gave a K-Ar date of 174 ± 5 Ma (Fig. 49). By Upper Triassic-Lower Jurassic time ductile deformation and metamorphism must have been fully underway, as some of its products were available already for erosion and were incorporated in the sandstone. Because this immature clastic rock was itself the protolith for more mylonite, ductile deformation also followed the Upper Triassic. The conodonts date deposition that was preceded, and followed, by ductile strain. They show that the K-Ar age determinations of 210 to 200 Ma date at least a real event. Yet those K-Ar ages are not distinguishable from the others, and when all are taken together they indicate long-lived ductile deformation and metamorphism.

The flaser fabric and schistosity are commonly deformed (Fig. 41b) and locally so strongly folded and transposed, that the new crenulation foliation rivals the planar fabric it deforms. This suggests that there may be rocks in which the newer fabric has obliterated the older. In such cases the newer is the first preserved, not the first, foliation. This raises the question of whether the first apparent foliation, is the first, second or nth such structure. It also raises the problem of correlating the fabric at one place with that at another, because in the final stage each looks like its precursor.

Synchronization and the long period of metamorphism and cataclasis, and the presence of deformed fluxion structure imply that some of the rocks may be polycataclastic and polymetamorphic. Thus the rocks at one place were strained first. Deformation jumped, after a time, to another place while neocrystallization began at the first. The focus of strain moved again and again at irregular intervals, followed each time by metamorphic regrowth. Eventually the process returned to rocks affected earlier, perhaps several or even many times during the deforming interval. Probably, all rocks were not deformed simultaneously, equally intensely, or equally often. The size of the strained zone, and the duration and rate of strain probably varied.

Campbell, Mink, Finlayson and Grass faults[74]

Four northwest- and northeast-trending faults are inferred to cut the allochthonous rocks northeast of Tintina fault (Fig. 49), but because outcrop is generally sparse, their extent and even existence are uncertain.. Gently dipping thrusts and steep faults with different orientations were also considered, but were rejected as less satisfactory in explaining the distribution of strata than those postulated here. Aeromagnetic patterns were used to help locate and interpret the faults. The breaks are named the Campbell, Mink, Finlayson and Grass faults after nearby topographic features.

Campbell fault is required to account for juxtaposition of the Anvil assemblage against sheared conglomerate of the Nisutlin assemblage roughly across Pelly River. The fault is fairly closely located on the south by exposures along Big Campbell Creek where porphyritic quartz monzonite is next to Anvil assemblage greenstones, but less certainly northward.

[74] Campbell and Grass faults are no longer recognized . Short segments of Mink and Finlayson faults persist on more recent maps. See Murphy et al. (2001, 2002, 2006).

The fault may not extend to the north edge of Finlayson Lake map area as shown, but may end 10 or 20 kilometres farther south. As assumed, the fault trends north-northwest and dips steeply. Offset of the Anvil assemblage requires that its western side fell 600 metres relative to its eastern.

Mink fault is presumed to separate the Mink plug (Fig. 46) from ultramafic rocks to the north. The ultramafic rocks are themselves not exposed but are postulated from aeromagnetic data. The fault is extended eastward to end on the Campbell fault. It may not extend as far west as Tintina fault as shown on the map, but may terminate 20 or 30 kilometres east of there. Mink fault has dropped its northern side about half a kilometre relative to its southern.

Finlayson fault is as uncertain as the others. Its central section seems necessary to explain juxtaposition of greenstone of the Anvil assemblage with gneisses on northern and southern sides, but the extension west to Tintina fault is hypothetical with little outcrop control. Continuing the fault eastward to near Finlayson River may explain why the thick limestone of Nisutlin assemblage ends against Anvil assemblage greenstones. As hypothesized the northern side has dropped fully two kilometres relative to the southern.

Grass fault is defined by good outcrops on the ridge just south of Grass Lakes and by the change in orientation of foliation across it. It is not in the same hypothetical class as the other three. This break is essentially the fractured axis of a late antiform that folds the fabric of the cataclastic rocks and is considered the partner of the Grass Synform, which is parallel and ten kilometres northwest. The fault is vertical and its southern side is dropped about 600 metres.

The steep faults cut the penetrative fabrics which are as young as Early Jurassic, and they are concurrent with, or younger than, emplacement of the allochthonous rocks during the Late Jurassic or Early Cretaceous. The faults are therefore Cretaceous or Tertiary. They may be thrust faults related to shortening of the Selwyn Mountains, but their trend argues against this. Otherwise they could be connected with movement on Tintina fault as their mapped terminations on that break imply. Those terminations, however, are largely hypothetical. The faults probably are not coeval with Late Tertiary normal movement on Tintina Trench, because they lack the topographic expression expected of such youthful features. In cross-sections the faults are shown as listric normal faults on the décollement (Fig. 50). If this interpretation is valid they are coeval with late stages of shortening far to the northeast in the Selwyn and Mackenzie mountains and Early Tertiary.

Simpson block, Simpson faults and Simpson arch

The Simpson Range in the southeast corner of Finlayson Lake map area is a down-faulted block, bounded by steep faults, in which the three allochthonous assemblages are folded over an open antiform, Simpson arch (Fig. 49). The faults on the west, north and east sides are are well defined by good outcrops of contrasting rocks on either side so that the locus of the faults is fairly certain. Their straight traces imply steep dips. The interior of the block has dropped one or two kilometres relative to its surroundings. Dip-slip is roughly two kilometres. On the southwest the block abuts the Tintina fault. The steep faults are late structures, perhaps synchronous with strike-slip on the Tintina.

Simpson arch is an open, northwest-trending antiform that folds the allochthonous rocks and their flaser fabrics. It may be coeval with movement on the Tintina and with the fall of the faulted 'block'. Simpson arch and the faulted Simpson block can be followed southeastward into adjacent Watson Lake map area (Gabrielse, 1967) a distance of 70 kilometres.

The Simpson block could be the surface expression of a depression in the detachment on which strata northeast of the Tintina fault slid during strike-slip. This detachment may be the base of continental crust or a higher décollement, like that on which shortening under the Selwyn Mountains was accomplished.

Structural history summary: - northeast of Tintina fault

Structural development northeast of Tintina fault resembles that on the southwest and has similar constraints. The main difference is the extensive flat-topped gneiss (Mink complex), not found to the southwest. The allochthonous rocks include the same lithologically distinctive late Paleozoic and early Mesozoic assemblages as southwest of the Tintina fault. These were deformed under ductile conditions, and metamorphosed and structurally assembled remote from where they now rest during Triassic and Early Jurassic time. The pre-assembled stack was thrust northeastward over autochthonous Proterozoic and Paleozoic strata which resemble those southwest of Tintina fault during the (?)Late Jurassic. In the (?)Cretaceous the autochthonous strata were deformed, but less foreshortened than southwest of the Tintina. This telescoping, and that farther northeast in the Selwyn and Mackenzie mountains was transmitted to a basal detachment.

Gneiss and schist of the Mink complex may represent crystalline North American basement[75]. They lie directly below the allochthonous rocks over large areas. They have no direct equivalent southwest of Tintina fault in the project area, but resemble the Big Salmon Complex. The gneiss may include Late Proterozoic or Eocambrian cover strata.

Steeply dipping faults that displace the allochthonous rocks and the gneiss below are assumed to be listric normal faults that branch from a deep detachment and may be synchronous with Early Tertiary foreshortening in the Mackenzie Mountains.

MINERAL OCCURRENCES

Introduction

The Pelly Mountains host several types of mineral showings including silver-lead veins, tungsten and molybdenite skarn, massive sulphide, bedded barite, and asbestos deposits. Most of the known occurrences can be fitted within the stratigraphic and structural framework developed in this report. The region is well endowed and showings are widespread (Fig. 69) and some are restricted to certain rock units (Fig. 70). Most occurrences are in the autochthonous beds southwest of the Tintina fault, but the transported rocks and the late tectonic metamorphic and plutonic suites also hold metal concentrations. The deposits have proven uneconomic to date.

In the autochthonous strata, silver-lead veins are the commonest showings. They occupy a spectrum of lithologic and stratigraphic niches and a range of structural levels, but most are in the Ketza and Askin groups, southwest of the Tintina fault. The veins are probably Late Cretaceous, but some may have originated with the Mississippian volcanics. The "in-place" strata also host bedded barite and conformable sulphides. These are restricted to the Seagull group and may be exhalative deposits genetically linked with the felsic volcanic and subvolcanic rocks. They are most likely Mississippian. Concentrations of rare earth elements are found in dykes of this suite east of McConnell River.

The transported metamorphic rocks have showings of asbestos in serpentinite of the Anvil assemblage and massive sulphides in the Nisutlin allochthon. The asbestos is on both sides of the Tintina, but the massive sulphides are restricted to the northeast side. The asbestos most likely formed after the ductile deformation that characterizes the host. The massive sulphides are older than the Early Jurassic as they are metamorphosed and have the ductile fabric of the enclosing strata.

[75] - now known to represent deformed Devono-Mississippian intrusive rocks of Yukon-Tanana terrane (Murphy et al, 2006).

Skarns with tungsten or molybdenum are known in and near the Big Salmon and Mink complexes, close to the mid-Cretaceous intrusives. Anomalous uranium is associated with the molybdenite occurrences.

Coal is interbedded with Cretaceous and Eocene fluvial sandstone in Tintina Trench deposited during or following strike-slip. Interestingly, these occurrences are higher grade than their Eocene age suggests, implying that they are metamorphosed.

Chronologically the stratabound sulphides and bedded barite in the Seagull group are the oldest, i.e., Mississippian. The conformable sulphides in the transported beds are perhaps Triassic[76]; the skarns are Cretaceous, and the silver veins may be Late Cretaceous.

Silver-lead-zinc veins

Silver-lead-zinc veins are concentrated near the headwaters of Ketza River, between Seagull and Groundhog creeks and in south central Finlayson Lake map area. At Ketza River they are fillings of irregular fractures and breccia zones in the Cloutier thrust sheet where it is arched over the Ketza High: Mount Misery (052: all mineral occurrence numbers listed in Deklerk and Traynor, 2005; Appendix 4); Key3 (53); Lap 10 (054); Hoey (055); Stump (056); K18 (Ketzakey 057[77]). The fractures have no preferred orientation, but generally dip steeply; they are breaks on which minor movement was accommodated. Sulphides occupy vein within, or oblique to enclosing structures. Competent beds, mostly dolostone and quartzite included in the Askin group, enclose the most persistent and thickest veins, but Seagull group slate and Kechika Group phyllite also carry mineralization. Several veins with replacements are in Ketza group limestone (Kay (13); Oxo (15)). Galena is the main sulphide, but tetrahedrite, pyrite and sphalerite are seen as well. Siderite and quartz form the gangue. Most veins are small; the largest contain in the order of 10,000 to 100,000 tonnes grading near 10% Pb with about 500 gm/t Ag. The ratio of silver to lead of Yukon silver-lead occurrences is traditionally compared to that in the Keno Hill district. There the ratio is near 4:l (oz/t Ag:%Pb) or 125:l (gm/t Ag:%Pb). In the Pelly Mountains the silver content is generally lower; this ratio is near 2:l or 60:l respectively

Veins near Groundhog Creek are in the Barite formation and in the Cloutier and Seagull thrust sheets (20: Grayling; 24: Groundhog). They are as irregular, and about the same size, as those near Ketza River, and their mineralogy is similar. Those in southern Finlayson

[76] : conformable sulphides are now known to be largelly Late Devonian to Early Mississippian in age (Murphy et al, 2006).

[77] The Ketzakey occurrence was developed by Canamax Resources and mined from 1988 until 1990. See Stroshein (1996) and Fonseca (1998).

Lake map area (4:Tintina) are in the McConnell formation, low in the McConnell thrust sheet and close to the south edge of Tintina stock. At Tintina about 90,000 tonnes with 620 gm/t Ag, 6% Pb and 10% Zn are indicated.

The variety of ages, lithology and structural levels in the host rocks implies that these factors are incidental to the mineralization. Other features must determine metal localization. Because no thrusts or other large faults are mineralized, and some veins transect foliation in the autochthonous strata, and because the structural level exerts no control, the mineralization must post-date the main deformation. Perhaps it coincided with the wane of Cretaceous plutonism, or is younger still.

Stratabound sulphide and barite deposits

The MM (10) is an intriguing massive sulphide in black slate with intercalated felsic tuff of the Seagull group. It is structurally high in the Seagull thrust sheet, below the Seagull thrust's twin imbricates, just north of Nisutlin batholith. Its vertical zoning and other preserved primary features show that it is a proximal exhalative massive sulphide[78]. It is small and similar nearby occurrences in the same rocks (CPA (11); Box (19)) are also limited in size and grade.

Fine grained, bedded barite with well developed, thin lamination is found just south of Tintina fault in black slate low in the Seagull group (Dirk (54)). Though no sulphides are associated, the barite grade and size are encouraging. Veins of coarsely crystalline barite, without sulphides, are known in the Askin group, notably on Barite Mountain (32). This barite may be mobilized into late fractures, perhaps from the bedded occurrences in the Seagull group.

Skarn occurrences in the Big Salmon and Mink complexes

Skarn is developed commonly where the Ketza group is metamorphosed next to mid-Cretaceous granite, as in the Big Salmon Complex (Ham (27); Eva (31); Obvious (57); Hidden (72) and Ayduck (73)). Most skarns contain diopside, tremolite, quartz and garnet. Scheelite is locally disseminated through it, but also fills narrow fractures. Chalcopyrite and pyrrhotite are disseminated in some skarns. The Risby (28), a scheelite skarn, contains a possible 310,000 tonnes at 1.02% W03. Stormy (9) has molybdenite and scheelite, probably 13,500 tonnes of 0.73% Mo and about 15,300 tonnes of 1.5% WO_3. Molly (1) is a skarn in the Askin group, next to Nisutlin batholith, in which interesting concentrations of uranium are known.

[78] See Murphy (1998); Hunt (2002).

Northeast of the Tintina are skarns in limy strata high in the Mink complex (Myda (43); Fog (52); Howdee (56)) and in amphibolitic rocks of the Anvil assemblage (Boot (55). Though the stratigraphic affiliation of the host is uncertain, the intrusives near them are the same mid-Cretaceous granites as those associated with the skarns southwest of the Tintina.

Other mineral occurrences in the allochthonous rocks

Asbestos is found at Tower Peak (7) and on Dunite Mountain in serpentinized peridotite that represents the Anvil assemblage. Northeast of the Tintina are two small, similar occurrences (Bot (27); Pup (28)). In each place asbestos occupies irregularly oriented fractures, up to a few centimetres wide. The fractures and mineralization postdate the strain fabric in these rocks and may be related to fluid movement after or during late stages of thrusting. The host rocks are probably Late Paleozoic and the fabric Early Jurassic.

Pyritic massive sulphides form lenses in the ductile deformed and metamorphosed Nisutlin assemblage northeast of Tintina fault (Hoo (9); E1 (10); Pack (21); Fyre (22); Fetish (49)[79]).Though they are stratum confined, they are small and laterally discontinuous. This suggests that they are structurally dismembered remnants of originally larger deposits. They are probably pre-metamorphic concentrations of volcanic exhalative sulphides, sheared and recrystallized during deformation of their host.

Coal in Tintina Trench

Sub-bituminous coal is interbedded with the Eocene immature fluvial, coarse clastic rocks in Tintina Trench near Ross River (Lapie River (41); Whiskey Lake (42)). As far as known the deposits are small and their host rocks are preserved in restricted fault blocks.. Nevertheless their potential seems good and inadequately tested.

Property Descriptions

Appendix 4 contains descriptions of most showings.. Occurrences are named and described following the scheme of Yukon Geology and Exploration 1979-80 (INAC, 1981)[80].

[79] Additional deposits were discovered in this area (Schulze, 1993; Hunt, 2002). The Fetish occurrence is close to the portal of the Wolverine deposit (Bradshaw et al., 2001, 2003).

[80] This publication follows a numbering scheme that was adopted for Yukon Minfile (Deklerk and Traynor, 2005). The property decriptions in Appendix 4 were written before this publication and may contain information not captured in it.

DISCUSSION[81]

Amount and rate of shortening

Foreshortening across the autochthonous strata southwest of Tintina fault can be estimated by adding the overlap on the four main thrusts. Unfortunately this slip is only roughly known for some of the faults and the overall shortening is therefore uncertain. Allowing 15 to 30 kilometres on the McConnell thrust, 25 to 40 kilometres on the Porcupine-Seagull system, 10 to15 kilometres on the Cloutier thrust and 30 to 50 kilometres on the St Cyr thrust gives totals between 80 and 135 kilometres. Additional shortening on other faults and folds is comparatively small. For example, shortening over the Porcupine syncline-anticline pair is between five and ten kilometres. Such further telescoping may be near 10 or 20 per cent and therefore falls within the broad limits given above.

The erosionally preserved overlap of transported rocks indicates that the thrust that superposes the allochthonous beds onto autochthonous strata has a minimum slip of 115 kilometres. This overlap[82], the distance from the northeast edge of Teslin Suture Zone to the McNeil klippen, is within the limits given above for the total overlap on the four other thrusts. It implies that range may be valid.

Two of the main thrusts of central Quiet Lake map area, the Cloutier and McConnell, are absent northwest of Lapie River. The McConnell thrust may have extended that far before erosion, but the Cloutier thrust dies out and merges with the St Cyr thrust. The preserved shortening northwest of Lapie fault is therefore less than on the opposite side by 20 to 40 kilometres; the original difference may have been similar. Reduced shortening in the northwest relative to the southeast is supported by narrowing of the zone of metamorphic rocks southwest of the imbricated cover strata. If these metamorphic rocks are mobilized basement stripped of most autochthonous cover (see elsewhere), less imbrication of the cover and less shortening are expected where this belt is narrower.

The shortening rate may be approximated by averaging total displacement (80-135 km) over the available time (Late Jurassic and Early Cretaceous; ~50 Ma). This gives rates between 1.5 and 2.7 mm/yr. Movement probably occurred spasmodically instead of uniformly, so that the average rate is at best only a guide. The shortening rate was used to estimate the life of the main thrust sheets. Much higher shortening rates are possible if the displacement is averaged over a shorter time.

The shortening velocity of the McConnell thrust compares to 3 mm/yr. in the southern Rocky Mountains (30 km in 10 Ma between Early Campanian and Paleocene; Wheeler et al., 1974) and to a more general rate of 4.5 mm/yr, for 200 kilometres of telescoping in the southern Rocky Mountains (Price and Mountjoy, 1970) in an interval of 45 Ma (from Albian through Paleocene). However, these rates are more than an order of magnitude slower than those projected for the Alps by Trümpy (1973). He considered that only 8 Ma were available for the 300 kilometres of shortening of the Alpine supracrustal strata. This converts to 4 cm/yr. At such a rate telescoping the Pelly Mountains could have been completed in 2.5 Ma.

The general direction of shortening is northeastward, as indicated by the vergence of many thrusts which cut up-section in that direction. An average direction of imbrication, given by the N55°W trend of the structures on surface, is N35°E. Traces of the oldest thrusts that emplaced the allochthonous strata are subparallel to the younger thrust faults in the autochthonous sequence. Even the metamorphic welts trend generally northwest. This implies the direction of shortening remained constant.

Reduced shortening in the northwest compared to the southeast implies that the thrust sheets were rotated into position. If shortening averaged 100 kilometres in southwest, and 70 kilometres in northwest parts of the project area, the sheets pivoted about a point roughly 300 kilometres northwest of the project area. Their transport path described an arc and was not perpendicular to the present thrust trend. The N45°E tangent to the arc most closely represents the average direction of shortening.

Sequence, age and duration of deformation

Structures in the autochthonous rocks developed in a sequence that is indicated by mutually cross-cutting relations. Generally the four main thrusts moved successively from southwest to northeast and their up-dip splays and tear faults were active when the respective sole thrusts moved (Fig. 71). Each splay moved in sequence from the southwest to the northeast, and generally moved progressively to higher stratigraphic levels.. Folds in the thrust sheets developed when the associated tear faults moved, generally during displacement on a thrust below the sheet in which the folds are expressed. The two granite-cored arches, with their metamorphic welts and the structural depression between them, formed after the higher thrusts. Strike-slip on the Tintina followed, perhaps directly after the lowest thrusts. The entire sequence may have lasted 40

[81] The reasoning and postulations in this last chapter reflect deductions and geological paradigms of the mid 1980s. This chapter has been edited for clarity only. Arguments used and conclusions differ significantly from most modern tectonic interpretations.

[82] Sketch map to illustrate this interpretation is missing.

or 50 Ma beginning perhaps, in the Latest Jurassic and continuing through the whole Early Cretaceous period.

The earliest thrust faults in the autochthonous succession may have been the Moose and Upper Sheep Creek thrusts. The Big Salmon thrust faults, are probable imbricates of the Moose thrust, and therefore were synchronous with it. Northeast sliding was next transferred to the Liard-McConnell-MM-Pass Peak thrusts, which accommodated between 10 and 25 kilometres of probable transport. Displacement took possibly between one and ten million years and the fault probably propagated from southwest to northeast. The Mount Hogg faults probably slipped during this time and in sequence, beginning with fault-8 and followed in succession by faults -7, -6, -5, -4 and -1 in that order. As the sole fault moved. the folds between the splays grew, although the folds probably initiated the deformation and localized the thrusts. The Mount Hogg tears are synchronous so that fault-11 moved with fault-8, fault-10 with 4 and fault-9 with the leading edge of the McConnell thrust. Imbrication of the Ketza group above the McConnell and Liard thrusts also occurred at this time.

Next to move was the Porcupine-Seagull thrust with its extensions, the Ragged Peak and Mount Vermilion thrusts. The time interval and southwest-to-northeast propagation rate was probably similar to that of the McConnell thrust. Offset of 25 kilometres or more is indicated by displaced facies in the Seagull group. The Groundhog faults moved with this thrust, beginning with Groundhog-12 and -11 and followed by -4 and -3; tears including Groundhog -1, -2, -5, -7, -8, -9 and -10 coincide with them. Repetition of the Ketza group by imbrication above the Seagull thrust also occurred at this time.

When the Porcupine-Seagull thrust locked, slip was transferred down to the Cloutier and subsidiary Ram thrusts. At this time the Twin Lakes and Starr Creek faults propagated from their base in the Cloutier thrust, up through the Cloutier and Porcupine-Seagull thrust sheets. The Twin Lakes folds also formed and the Porcupine syncline grew. Ketza High rose as its bounding faults slipped. Ketza fault may be the first up-dip splay of the Cloutier thrust and Ketza-6, -5, -2, -3, -1 and -4 followed. The northeast directed splays, including the McNeil faults, developed next. Northeast slip on the Cloutier thrust decreased northeast of (beyond) these branches.

Movement on the St Cyr thrust, latest and lowest of the exposed faults, may have overlapped with the last movements of the Cloutier thrust system, before becoming the main slip surface. Lapie fault and the southwest directed Bacon Creek, Fox Creek and Rose Lake thrusts developed with it. These allowed differential sliding on the St Cyr thrust, accounting for fold differences across Lapie fault. Now the Porcupine and Lapie folds reached their final form and Cloutier syncline was completed. The Twin Lakes folds and Ketza High, initiated earlier, may have been accentuated. Judging from the profound change of facies across it, the St Cyr thrust sheet moved as far as any in the region, possibly 50 kilometres. Nearly 20 kilometres of northeast transport near the leading edge was distributed on several important splays. The St Cyr faults -2, -3 and -4 and the Hoole faults are thought to be its branches.

If Scurvy fault is a southwest directed branch of the St Cyr thrust, they probably moved simultaneously. If connected to a lower thrust it may be younger.

Metamorphism and plutonism probably began during the deformation and continued with it. Cooling dates of the plutonic rocks (i.e., their K-Ar ages) cluster around 90 Ma (roughly mid-Cretaceous; Fig. 72). It is a late stage in the plutonic-metamorphic history, but do not give an idea of its duration. Metamorphism and intrusion may have occurred generally behind, or southwest of, the active thrusts because it apparently affected previously formed structures. Thus the Big Salmon-Nisutlin and Quiet Lake arches and their intrusions arched the Moose, McConnell, Upper Sheep Creek, and Porcupine-Seagull thrusts. In contrast the St Cyr thrust may be younger than those batholiths. If so, some exposed intrusions must be only the decapitated upper parts of intrusions whose roots are farther southwest.

The Pony Creek and Gray Creek faults are thought to be normal faults related to the rise of the big Salmon, Nisutlin and Quiet Lake intrusions. They are probably synchronous with late stages of batholith emplacement.

Transcurrent movement on the Tintina fault (as well as theTintoona and Kumquatly faults) may have followed directly after the St Cyr thrust. They may be transpressive structures related to the thrusts.

Constraints on the timing of deformation are imprecise, indirect and few. Thrusts in the autochthonous rocks probably formed after emplacement of the allochthons. The penetrative fabric of the transported sheets predates their emplacement as well as the deformation of the autochthonous sequence. The flaser texture is dated isotopically at several places in the project area as Triassic to Middle Jurassic (Fig. 49). At the other end of the time range the thrusts southwest of the Tintina Trench had all moved before slip on the Tintina fault. Compressive deformation was at least partly completed by the Early Eocene, when immature sandstone was deposited in fault blocks along this dextral fault. The thrusts therefore moved in the Upper Jurassic and/or Cretaceous. Some thrusts predate the granitic intrusions, which cooled over 30 million years in the Late Cretaceous. A lamprophyre dyke in the top of the Cloutier thrust sheet, dated isotopically as 112 Ma, lacks the fabric of the rocks it invades. It is assumed to have cooled quickly and to immediately postdate the Cloutier thrust.

The duration of thrusting and shortening in the autochthonous rocks are similarly approximate. thrusting may have lasted for as little as 10 Ma and for as long as 50 Ma. If the time during which intrusions cooled (measured by the range in K-Ar dates) is related to the duration of shortening; a 30 Ma span is indicated.

Depth of detachment

Folds, thrust faults and tears probably die out at depth as well as along strike, but the depth to which surface structures extend is uncertain. Both the thickness and thickness variations of rock units are poorly known. Cross-sections are therefore more schematic downward. In constructing them, it was postulated that the thrust faults are upward splays of a single basal detachment, beneath which the strata are not structurally offset. Folds and tear-faults are similarly assumed to be confined to the hanging wall of the detachment without expression below it. If this is valid the basal detachment is five to eight kilometres below sea level, as shown in the cross-sections. The Paleozoic succession is thus an internally imbricated parautochthonous slice, less than ten kilometres thick. It is structurally overlapped by allochthonous, ductile deformed and metamorphosed rocks. Both lie above a basal detachment that separates them from lower autochthonous rocks and possibly crystalline basement.

Tintina fault at depth[83]

The Tintina fault disrupts the imbricated rocks above the décollement on which Pelly Mountains strata are shortened.Whether it extends below that surface into the basement, or is confined to the detached imbricated slab of stratified rocks, is unknown. Like the thrusts, it may merge with the décollement instead of cutting it. If it postdates all shortening, on both its sides, it might cut and displace cover and basement and disrupt the décollement on which telescoping was accomplished. In this instance the fault probably extends to, and merges with, the base of the continental crust or some higher detachment (Option 1, Fig. 72). If shortening and strike-slip were concurrent the Tintina probably merges with this detachment (Option 2, Fig. 72). Thirdly, if some shortening followed strike-slip the fault is cut and displaced by the detachment (Option 3, Fig. 72). The timing of shortening and strike-slip are not closely dated but on present evidence seems to be concurrent. Shortening in the Selwyn and Mackenzie mountains, northeast of Tintina fault, occurred in the same interval as the strike-slip on Tintina fault, the Late Cretaceous and Early Tertiary.

If Tintina faulting was concurrent with northeastward contraction, the first option is eliminated. Furthermore it suggests that the strike-slip may be restricted to the imbricated slab above the detachment, like a giant tear in that slice. In this case the fault extends to between eight and ten kilometres of the surface and merges with the décollement, but had no effect below. If the Tintina moved while it was carried piggy-back in the detached slab and was deformed internally, its long straight trace seems improbable. However, shortening in the detached slab occurred only near its leading edge. When Tintina fault was active the thrust front has moved far to the northeast. At the locus of Tintina telescoping of the slab was complete and the trace of strike-slip could therefore have been unaffected.

The third alternative, that the Tintina fault offset cover and basement before telescoping in the Selwyn and Mackenzie mountains would result in severing of the Tintina fault by the detachment on which northeastward shortening in the Selwyn and Mackenzie was transmitted to that in the Pelly Mountains. The Tintina's trace in the footwall of the detachment must then be displaced southwest of that in the hanging wall, a distance equivalent to the total shortening in the Mackenzie and Selwyn mountains. The footwall trace would be difficult to locate, because its top is eight to ten kilometres below the surface. If present below the detachment, it might control later faults and be transmitted upward to the surface because it is a deep crustal break[84]. Teslin lineament, between 100 and 130 kilometres southwest of Tintina fault, is a northwest trending fracture zone could have had such an origin.

The different rates of movement suggest a way to resolve these alternatives. The second (i.e., dextral tear in detached slab) is favoured by this reasoning. Shortening in the Selwyn and Mackenzie mountains and strike-slip on Tintina fault probably occurred at different rates. The shortening, at least 53 kilometres according to Gordey (1981) and Cecile et al. (1981), may have proceeded at 1 or 2 mm/yr. For example roughly 200 kilometres of shortening in the southern Rocky Mountains (Price and Mountjoy, 1970), and the smaller telescoping in the northern Rockies, were accomplished in an interval variously estimated at 50 to 100 Ma. Rates of one or two millimetres annually, require 53 or 26 Ma respectively, for the shortening in the Selwyn and Mackenzie mountains. Tintina strike-slip was probably 1 or 2 cm/yr, an order of magnitude faster than shortening. It requires between 45 and 23 Ma to complete the observed 450 kilometres of slip on the

[83] A seismic refraction and more detailed vibroseismic reflection survey was conducted along the North ans South Canol roads in 1996 and 1998. A key finding was that the Tintina fault extends near-vertically through the entire crust (Snyder et al., 2005).

[84] Magneto-telluric studies during the Slave-Northern Cordillera Lithospheric Experiment shed light on this fundamental crustal break. See Ledo et al. (2002).

Tintina at these rates. The relatively long but finite periods, 26-53 Ma for shortening and 23-45 Ma for strike-slip, compared with the available time of 45 Ma in the Late Cretaceous and Early Tertiary, suggest time overlap, even if they were not entirely synchronous and even if they did not occupy the entire interval available. Concurrence makes the second alternative the most plausible.

Detachment and the Mink complex

In northern Finlayson Lake map area the contact between the transported and autochthonous rocks is sharp, as it is in Quiet Lake and southern Finlayson Lake map areas. It superposes sheared and metamorphosed rocks atop undeformed and unmetamorphosed strata. The difference in deformation and metamorphic state between the superposed and subjacent rocks is accentuated by their lithologic differences. This contrast is not seen in central Finlayson Lake map area, where the allochthonous rocks lie above similarly deformed and metamorphosed gneiss and schist with quartz monzonite intrusions - the Mink complex. The contact between them is gradational and not marked by an abrupt metamorphic difference. The gneiss and schist resemble the Big Salmon Complex, but do not match other autochthonous strata, nor do they compare to other allochthonous rocks. The augen gneiss is without equivalent in the project area. The gneiss and schist may be crystalline basement that rose through, or was stripped of, most of its autochthonous cover. It is now overlain directly by the allochthonous rocks (Fig. 73a, b). Alternately it may be a new lower slice of allochthonous rocks not seen elsewhere, above the same autochthonous strata (Fig. 73c).

Gneiss and schist of the Mink complex are intimately ssociated with quartz monzonite plugs, which are thought to be more mobilized parts of the gneiss and to have been intruded with it. Though no direct data are available, these plugs are presumed to be mid-Cretaceous because they resemble dated plutons of that age nearby. Radiometrically dated gneiss, with relations like those of the Mink complex, and possibly its equivalent displaced on the Tintina fault, is found 550 kilometres to the northwest on the opposite side of, and some distance from, the Tintina in the Fifty-mile batholith. Biotite from the batholith gave a K-Ar date of 97.6 Ma and a U/Pb zircon date indicates original crystallization at 375 Ma (Devonian; Tempelman-Kluit and Wanless, 1980). Aleinikoff et al. (1981) dated zircons from a similar gneiss in Alaska,. Their data imply that the gneiss intruded during the Late Devonian (345 Ma) and that it incorporates a Proterozoic protolith (ca 2.3 Ga); its micas give a K-Ar age of 110 Ma and indicate that the last thermal episode is Cretaceous. These dates indicate that, although the gneiss is an old rock, its final history is younger than the fabrics and emplacement of the allochthonous rocks. This favours its origin as basement mobilized into, or with, its covering strata. If the Mink complex is upwarped or domed crystalline basement along with its attenuated cover, its rise relative to the time of shortening in the Selwyn and Mackenzie mountains is crucial to interpretations about the basal detachment of shortening. Did the crystalline rocks rise diapirically after shortening, so that the detachment was warped passively after it ceased to be active? (Fig. 73a), or did the basement rise while detachment continued? If the latter, did décollement proceed at the top of the crystalline rocks (Fig. 73a) or was it transferred to a lower surface so that the crystalline upwarp is cut off below? (Fig. 73b). Because the contact between the gneiss and the allochthonous rocks is gradational and metamorphically "healed" the option of the gneiss as a distinctive new allochthonous slice (Fig. 73c) seems improbable. For this reason the possibility of important detachment along the contact after the metamorphism, as required by the option shown in Figure 73a, is also unlikely. The second alternative, wherein the gneiss is considered to be basement intruded into the base of the allochthonous rocks and detached at depth, is therefore favoured.

The question "Why is the gneiss penetratively deformed, whereas imbricated cover strata only 30 kilometres to the northeast escaped this strain?" remains. Both are overridden by the same allochthonous rocks. Perhaps the thickness of the overthrust slice differed drastically in the two places. The gneiss fabrics reflect flow and mobilization at considerable depth. Presumably the 10-15 kilometres thick forearc system rested above (see following discussion: "Allochthonous rocks as subduction complex"). By contrast the autochthonous cover was overridden by the thinner leading edge of the obducted forearc basin. In the northeast the cover strata may have escaped penetrative strain, because the allochthonous sheet was thin at its leading edge. For the same reason the northeastern supracrustal rocks were not involved in basement mobilization.

The preferred explanation is that the Mink complex represents a Devonian core complex of North American crust. The Paleozoic cover was tectonically stripped from it during Late Devonian time (345-375 Ma by U/Pb zircon). In the Jurassic the allochthonous rocks were thrust across both the core complex and the detached cover that is now exposed northeast of it. In this interpretation the penetrative fabric of the Mink complex owes its origin to Devonian extension whereas that of the allochthonous rocks originated during Mesozoic subduction. The allochthons now rest on the gneiss as a result of Jurassic over-thrusting and the whole stack was metamorphosed and mobilized in the Cretaceous (the 97.6 Ma date). The timing of postulated extension coincides with the onset of Earn Group deposition and may provide a mechanism and source for the clastic rocks. The time slightly predates extrusion of the Early Mississippian felsic volcanics which may be an expression of this extension.

This "Paleozoic core complex" interpretation also accounts for Devonian gneiss elsewhere in the Cordillera (e.g., augen gneiss bodies in Yukon-Tanana,Fifty-mile batholith, Mount Fowler batholith region; Okulitch et al, 1975). Furthermore, the apparent tectonic stripping of cover from the Big Salmon Complex (Note “mobilized and passive basement” on Fig. 51) may be the product of Devonian tectonic denudation instead of Jurassic compression. Evidence for a Devonian age of the fabrics in the Big Salmon Complex is lacking, but there it may simply be more effectively masked by the Cretaceous event than in the Mink complex.

The allochthonous rocks as a subduction complex of the Lewes River arc

Transported rocks exposed as klippen are considered to be the fossil mélange of deep parts of a subduction zone because their tectonic mixing, ductile strain fabrics, metamorphism and lithologic variety are difficult to envisage in another setting. The postulated subduction zone dipped westward beneath the Mesozoic volcanic rocks of the Lewes River arc, which is now accreted to North America by northeastward overthrusting (Tempelman-Kluit, 1979; his Fig. 20).

To speculate further (Figs. 74 and 75), the consistent stacking order of the three allochthonous assemblages probably reflects their original lateral position and/or the sequence of assembly in the subduction zone. If so, the protolith of Simpson assemblage formed southwest of the Anvil assemblage and was subducted earlier. Similarly the Anvil assemblage formed southwest of the Nisutlin assemblage. The absence of structural mixing of the assemblages implies that they formed remote from each other.

Previously the Simpson assemblage was interpreted as the sheared plutonic root of arc volcanics in the arc massif (Tempelman-Kluit, 1979). They are, however, too old (± 350 Ma) to be part of the Lewes River arc, which occurred between 220 and 160 Ma, and this interpretation is no longer tenable.

Anvil assemblage probably originated as Paleozoic oceanic or transitional crust east of the Lewes River arc. It became the floor of the forearc basin and its equivalent may include all or part of the Cache Creek Group exposed in Laberge area, east of the forearc basin fill (Laberge Group). Such basins, built on oceanic crust trapped next to an arc massif, were termed residual by Seely (1979). The ophiolite was sheared and dismembered during subduction, before or during down-plating of the Nisutlin assemblage. A less likely alternative is that the Anvil assemblage was oceanic crust that originated far northeast of the arc, and was inserted into the subduction complex during down-plating[85].

The Nisutlin assemblage is the relic of a mainly clastic sedimentary sequence, probably deposited largely as trench-fill during subduction. The rocks were penetratively deformed and structurally dismembered while oceanic crust was down-plated immediately beneath. Intermediate volcaniclastic rocks intercalated with the Nisutlin assemblage may represent distal arc products. The "synorogenic clastic rocks", which contain metamorphosed and plastically deformed debris and are themselves sheared, were probably eroded from elevated parts of the trench-slope break and redeposited in the trench. An analogue of such a positive trench slope break in the Sunda Arc is the island Nias (Karig et al., 1979). Part of the Nisutlin assemblage may have been laid down far northeast of the arc near the outer edge of the forearc basin, but carried into the subduction zone when the oceanic crust on which it was deposited was consumed (Tempelman-Kluit, 1979).

If the allochthonous rocks represent a fossil subduction complex that dipped southwest, and which grew by underplating at its base, the age of deformation and metamorphism should decrease downward in the structural stack. Thus Simpson Allochthon should have been sheared and deformed first and the Anvil and Nisutlin progressively later. The long range of K-Ar cooling dates shows that deformation and metamorphism of these rocks occurred over a substantial interval (*ca*. 60 Ma, between 220 and 160 Ma ago), but they show no systematic change. Perhaps the cooling dates do not reflect the last time of strain, but only termination of the latest metamorphism. If so, metamorphism and strain are coeval only in the broadest sense. Shearing may also have recurred at different levels in the structural stack over a period of time, keeping the entire strained zone active throughout down plating, instead of progressing downward through the complex.

Original thickness and extent of allochthonous rocks

A clue to the thickness of the allochthonous rocks at emplacement comes from the late tectonic granites, and these constrain estimates of the size of the obducted arc system. The highest parts of the Nisutlin, Big Salmon and Quiet Lake batholiths are mesozonal and probably rose to within five kilometres of the paleo-surface. Plutonic rocks under St Cyr syncline may have cooled at greater depth, as shown in cross-sections. As the top

[85] For a modern interpretation, see Nelson et al (2006); Piercey et al. (2006).

of Nisutlin batholith coincides with the base of allochthonous strata, some five kilometres of transported rocks must have been present during cooling. But the transported rocks were thrust into place before than the intrusions cooled. The allochthonous rocks were therefore probably thicker when emplaced, than when they were intruded. Erosion following intrusion about 90 Ma ago, has removed the five kilometres of batholith cover and roughly one kilometre of autochthonous beds, a total of 6 kilometres. If erosion in the preceding 50 Ma was at the same rate, a further three or four kilometres of strata must have been present originally. This makes for an original thickness of eight or nine kilometres of transported rocks about the Late Jurassic above the locus of Nisutlin batholith before it was intruded.

Gneissic rocks northeast of the Tintina fault are catazonal and also indicate that the tectonic stack had an original thickness of about ten kilometres. The top of the gneiss is about one kilometre below the base of the allochthonous mylonite schist, but the gneiss, which locally melted to produce a muscovite biotite granite, probably cooled about twelve kilometres below surface (Wyllie, 1977) so that 11 kilometres of allochthonous rocks were present originally and about eight or nine kilometres have been eroded during the roughly 150 Ma since their emplacement.

These thicknesses are consistent with the idea (discussed above) that the allochthonous rocks are the plastically deformed sole of the obducted Lewes River arc (Figs. 74, 75). At the trench-slope break, the arc may have been 10 or 15 kilometres thick; farther southwest it was thicker. Presumably Nisutlin batholith lay roughly below the trench-slope break of the obducted arc system.

If the allochthonous rocks are the sole of an over-thrust arc system whose trench-slope break lay roughly above the present locus of Nisutlin batholith, the leading edge of the transported rocks (when emplaced) was probably 50 to 100 kilometres farther northeast. For instance, about 100 kilometres is measured between the trench-slope break and trench in the Sunda Arc (Karig et al., 1979) and roughly 50 kilometres between the Peru-Chile trench and its slope break (Seely, 1979); these may be analogous to the Lewes River arc. The transported rocks therefore extended as a northeast-thinning wedge from Nisutlin batholith, northeastward at least as far as the present Tintina fault.

Mobilized and passive basement and speculation about crustal thickness

Autochthonous crystalline rocks and autochthonous unmetamorphosed strata occur separately on both sides of Tintina fault. A line can be drawn between areas of schist and gneiss with a thin (perhaps thinned) cover, and areas underlain by imbricated cover strata. This line is the boundary between basement that was actively involved in late stages of the deformation, and basement that remained passive beneath its imbricated, structurally repeated, autochthonous cover. Southwest of Tintina fault this boundary (Figs. 46, 49, 51) extends from the northeast side of the Fox batholith, along the northeast side of metamorphosed rocks, and thence along the northeast side of Nisutlin batholith. Southeastward and outside the project area, it continues between the Marker Lake and Cassiar batholiths. Northwestward from the Fox batholith and outside the map area (Fig. 76) this line separates Glenlyon batholith from strata on its northeast side. Northeast of Tintina fault the boundary follows the northern side of the Mink complex to Wolverine Lake and thence, outside the project area, through the Simpson Range to Tintina fault In most places the boundary is sharp, but locally basement on its northeast side has been mobilized to various degrees. For example, the Mount Billings batholith with its metamorphic mantle may represent a relatively isolated basement upwarp. Southwest of the boundary the total basement may have been involved, with limited mobility on the northeast side.

The boundary between mobilized and uninvolved basement perhaps marks the limit to which thick imbricated cover was thrust northeastward, mechanically torn from its basement ahead of the allochthonous strata. In effect allochthonous cover replaced the original autochthonous cover, above the same basement. While the cover was stripped from its basement southwest of this line, it was imbricated and structurally thickened on the northeast side, being transported as thrust sheets two or three kilometres thick (Fig. 51).Tectonic stripping and imbrication took place in the Early Cretaceous (and Late Jurassic), whereas basement was reactivated in the Late Cretaceous, when shortening of the cover rocks had ceased. The thickness of the autochthonous succession (Lower Cambrian through the Triassic age) averages eight kilometres.The thickness of the fore-arc prism that displaced this succession was presumably at least eight kilometres. Interestingly the boundary between passive and involved basement lies beneath where the trench-slope break (Fig. 75) of the obducted arc system is projected. At this point the thickness of the arc may have been ten kilometres.

Assuming that the schist and gneiss represents mobilized basement stripped of its supracrustal rocks, the distance from Teslin Suture to the northeast edge of stripping (i.e., the boundary between mobilized and passive basement; Figs. 49, 51) is the minimum shortening that must have been accommodated by the autochthonous cover. This distance varies from 40 kilometres to 85 kilometres on the southern side of Tintina fault in the project area and increases from northwest to southeast. This is less than the limits for telescoping determined from the thrusts themselves (i.e., 80 to 135 km) and thus the difference, something like 30 to 50 kilometres of supracrustal shortening, is unaccounted for by crustal stripping. This 30 to 50 kilometres of crust may have under-plated the arc (i.e., under the forearc and beneath the zone where basement was stripped of autochthonous cover). It thickened the crust there and buoyed the obducted arc leading to its erosion (Fig. 76a). It also promoted melting and mobilization at its base.

Accepting that the "unaccounted-for crust" was added mainly to the overridden, but partly to the imbricated zone, we can estimate the present thickness of crust by making two assumptions. First: before deformation and thrusting the crust under the 8 kilometres of autochthonous cover was 27 kilometres thick (i.e., normal continental thickness of crust and cover of 35 km (Brune, 1969)). Second: the zones with mobilized and passive basement are now in isostatic equilibrium with each other so that their respective crustal columns balance. The sum of the original thickness of crust, and that added from the unaccounted part, together with the thickness of present cover should therefore be equal in both zones. In cross-section (Fig. 76b) the 30 (to 50 km) wide zone of 27-kilometre-thick "unaccounted-for" crust is redistributed under the 80-kilometre-wide mobilized zone (its entire width) and under a 30-kilometre-wide strip of the imbricated zone (the width of the zone where small intrusions are found northeast of the active-passive basement boundary). Neglecting relatively small density differencesbetween cover and continental crust, this can be expressed as two simultaneous equations (Fig. 76b).

(1)Present columns of balance[86]

Original thickness of crust (27 km) + thickness of crust added to mobilized zone from unaccounted crust during deformation (y) + present thickness of cover above mobilized zone (Ø) = Original thickness of crust in imbricated zone (27) + thickness of crust added to imbricated zone from unaccounted crust during deformation (x) + present thickness of cover inimbricated zone.

(2) In cross-section the amount of unaccounted crust is only redistributed, but not lost. 30 kilometres (x) + 80 kilometres (y) = 30 x 27 (width x thickness of the strip of unaccounted crust).

Solving the two equations for 30 and 50 kilometres wide zones of unaccounted crust gives thickness of crust added beneath the mobilized zone of 9.5 kilometres and 14.5 kilometres, respectively, and under the imbricated zone of 1.5 and 6.5 kilometres, respectively. This gives present columns 36.5 and 41.5 kilometres thick under each zone for 30 and 50 kilometres of unaccounted crust, respectively.

Both alternatives can satisfy the dual requirement of extensive crustal anatexis under the mobilized zone, while the allochthonous forearc (perhaps 10 kilometres thick) was in place, and little melting under the imbricated cover zone next to it. Both also provide relative buoyancy to raise the mobilized zone and erode to its basement under the obducted forearc, without eroding the imbricated autochthonous zone to basement. The thinner alternative implies that both regions are now nearly in balance with continental crust elsewhere, whereas the second option implies disequilibrium. Because the average elevation of the surface is more than a kilometre above sea level, balance has probably not been attained and a crustal root five to eight kilometres deeper than normal crustal thickness is possible, given the density ratio between crust and mantle. This favours the alternative in which a 50 kilometre wide strip of crust was redistributed.

Schuiling (1973) calculated that for a linear mountain chain initial crustal thickening of 10 kilometres results eventually in erosion of 20 kilometres so that thickening ultimately effects the opposite (i.e., a significant reduction of the crust). He calculated a total of three kilometres of uplift resulting from expansion, anatexis and phase transitions due to the excess radiogenic heat produced initially by the ten kilometres of crust. This, he concluded, in turn leads to 17 kilometres of compensatory isostatic rise and erosion at the surface. In the project area the zone of mobilized basement is the originally thickened belt. Part of a forearc basin, roughly 15 kilometres thick, was added above it to replace an original 8 kilometres of supracrustal cover that was mechanically stripped. At the same time basement was thickened 10 or 15 kilometres (Fig. 77). The net addition of crust in the mobilized zone was 17 or 22 kilometres. Using Schuiling's (1973) assumptions the isostatic compensation would be 20 and 26 kilometres respectively, for initial thickening of 17 or 22 kilometres, so that ultimately 20 or 26 kilometres should be removed erosionally. To date the 15 kilometres thick forearc and about 1 kilometre of mobilized basement have been removed and further erosion of between 4 and 10 kilometres is expected from this data.

Crustal redistribution calculations such as those done above are difficult for the area northeast of Tintina fault because the data are incomplete. Northeast of Tintina fault the stripped basement is 35 kilometres wide and the minimum shortening in the autochthonous strata of the Mackenzie Mountains far to the northeast is 53 kilometres (Gordey, 1981b). Unfortunately neither measurement is the total. Reconstruction before movement of the Tintina fault suggests that a further 130 kilometres of basement may be stripped between

[86] The critical reader commends this attempt to balance crust but points out that multiple unconstrained assumptions renders the result meaningless. The assumptions include: present thickness of cover, conservation of crust in arc-zone, and character and the thickness of redistributed crust.

the fault and Teslin Suture (i.e., the width of the Yukon cataclastic complex southwest of Dawson). Gordey's (1981) estimate of shortening does not include telescoping in the Selwyn Mountains[87], which may be as much as that in the Mackenzie Mountains. Total basement stripped may be 165 kilometres wide and supracrustal shortening is more than 53 kilometres. Because this area may be isostatically balanced with that on the opposite side of the fault a similar crustal thickness is expected.

Compression, gravity flow or gravity gliding

The thrust-dominated deformation in the supracrustal strata ahead of a plutonic-metamorphic core seen in the Pelly Mountains is found in many mountain belts including the Rocky Mountains (Price and Mountjoy, 1970; Thompson, 1981), the Appalachians (Williams, 1979; Hatcher, 1981) and the Alps (Debelmas and Kerckhove, 1973; Clar, 1973). Three mechanisms are postulated to effect imbrication of the cover. These were reviewed by Bally (1981) and by Cooper (1981). The three schemes are *gravity gliding*, where thrust sheets slide down an inclined surface; *gravity spreading*, where outward flow of thrust sheets is induced by a diapirically emplaced load; and *tectonic compression*, in which thrust sheets are pushed up an inclined surface. A fourth possibility was proposed by Hatcher (1981). Struck by the association of thrust belts with metamorphic cores he hinted that thermal energy from the cores is dissipated mechanically in propelling the thrust sheets.

Diverticulation, where in the lowest thrust sheets represent the originally highest and/or most distal part of a stratigraphic sequence and successively higher sheets derive from progressively lower in the stratigraphy (Lemoine, 1973), has not been recognized in the Pelly Mountains. It is considered characteristic of gravity gliding nappes. Because the plutonic-metamorphic core of the Pelly Mountains rose after many of the thrusts moved, diapiric rise of the culmination could not have driven the thrusts by gravity spreading. This contradiction in timing also seems to be a problem for Hatcher's (1981) thermal energy mechanism.

Pelly Mountains deformation is considered the product of tectonic compression. The mountains can be divided into three zones: an external (northeastern) belt of imbricated supracrustal rocks; a central zone of metamorphic and plutonic rocks (mobilized basement stripped of its cover); and an internal (southwestern) strip of plastically deformed strata. The plastically deformed zone may represent the base of an arc system, allochthonous with respect to the first two zones. Deformation in the external supracrustals results from compression by the overriding Lewes River arc system and its continental massif (the Stikine block; see Tempelman-Kluit, 1979; his Fig. 20f). The overthrust arc system must have been sufficiently massive to generate a wide annular zone of thin-skinned effects.

Overriding lasted long enough to deform the cover and to metamorphose and partly melt the core zone. Compression between the accreted Lewes River arc and ancient North America followed subduction of the oceanic lithosphere that separated them, in a west dipping Benioff Zone at the face of the arc. Over-thrusting of the arc onto ancient North America began about the Middle Jurassic. By the Late Jurassic subduction of the North American plate beneath the arc had ceased, and the northeast-verging subduction zone had "stepped down" into the autochthonous cover strata, stripping basement and imbricating cover ahead of it.What had begun as ‘B subduction’ in the Triassic, when oceanic lithosphere was under thrust, graduated to ‘A subduction’ (see Bally, 1981) as continental crust was down-plated beneath the accreted arc. Down plated continental crust, too buoyant to be consumed, accumulated under the face of the accreted arc during the Early Cretaceous. Metamorphism, partial melting and mobilization thickened the zone of continental crust. The thickened core was mobilized and cooled about the early Late Cretaceous. Shortening in the foreland continued after southwestern parts of the metamorphic-plutonic welt had cooled. This last telescoping of the cover led to crustal down plating into (and northeast of) the cooling metamorphic and plutonic welt, extending its life and increasing its width. During the Late Cretaceous dextral strike-slip of about 450 kilometres occurred on the Tintina fault. Crustal telescoping and transcurrent movement ceased by the Paleocene. Compression was then taken up entirely in the outboard (i.e., southwestern) subduction zone that dipped eastward, beneath the newly generated collision orogen.

The effective boundary between the paleo-Pacific and North American plates changed its locus with time. During oceanic downplating in the Late Triassic and Early and Middle Jurassic the boundary was at the top of the downgoing North American slab (Tempelman-Kluit, 1979; his Figs. 20d, e). In the Pelly Mountains this was the tectonic base of the Nisutlin assemblage and its extension, the Teslin Suture Zone. With overthrusting of the arc in the Late Jurassic, this boundary stepped downward into what had until then been the North American plate. The décollement between cover and basement effectively became the subduction zone and the northeasternmost listric thrust of this detachment at any time was the subduction surface. More and more of the original North America plate became mechanically linked with the paleo-Pacific element as the thrust front moved northeast.

[87] See Gordey (in press).

Viewed from this perspective the Tintina fault is probably confined to the detached cover and does not extend into the basement. If this break was a transform within the paleo-Pacific plate it must terminate on the décollement, which was the boundary between the Pacific and North American slabs when the Tintina moved. If the Tintina separated the old Pacific and North American plates it must merge downward with the décollement, the slip surface between the two plates during Tintina movement. In the first instance the décollement had progressed under the Tintina and surfaced northeast of it; in the second case the detachment extended to, but not northeast of, the fault.

Relative movement between Ancient North American and Paleo-Pacific plates

If the décollement and the Tintina fault acted as the boundary between the ancient North American and paleo-Pacific plates, their displacement together gives the relative movement of the two plates. During the Late Cretaceous the Tintina slipped 450 kilometres in a dextral sense, while at least 50 kilometres of shortening occurred in the Mackenzie Mountains and perhaps a similar amount in the Selwyn Mountains. In this 45 Ma span the paleo-Pacific plate moved northwest 450 kilometres and northeast about 100 kilometres relative to the North American plate. The combined relative movement is 460 kilometres toward N33°W (Fig. 78). This represents average convergence at 1.02 cm/yr. In the 50 Ma spanning the Late Jurassic and Early Cretaceous, during obduction of the Lewes River arc and during telescoping of cover in the Pelly Mountains, only the relative motion of about 100 kilometres on the décollement is indicated. The relative displacement of the two plates in the Early Cretaceous was therefore much smaller and almost at right angles to motion in the Late Cretaceous. However if the relative motion was similar about 500 kilometres of dextral movement is unrecognized on one or more faults that were early (i.e., Lower Cretaceous) analogues of the Tintina fault. The best candidate for such slip is Teslin lineament - the extension of the Kutcho-Pinchi-Findlay fault system southwest of the project area. The Teslin lineament is a straight valley extending about 400 kilometres southeastward from near Carmacks, to the Thibert and Kutcho faults (Gabrielse et al. 2006). No large lateral displacement has been demonstrated on it. The Kutcho fault, however, may have moved 175 kilometres right laterally. Monger and Irving (1980) considered that rocks as young as Lower Cretaceous are displaced 13° northward relative to North America, implying about 800 kilometres of movement on faults like the Kutcho-Teslin lineament.

Along the Teslin River, Teslin lineament juxtaposes Laberge and Cache Creek group strata; fault blocks containing the Tantalus formation lie along the lineament. This implies that if there was movement along the lineament it postdated the Lower Jurassic (the age of the Laberge Group) and may be synchronous with the Tantalus formation (i.e., Upper Jurassic-Lower Cretaceous).

Relations at the northwest terminus of Teslin lineament are obscure. The lineament ends in plastically deformed rocks of the Yukon cataclastic complex (Fig. 74). The lineament may merge with the ductile zone that was the basal decollement during motion. At its northwest end the Teslin lineament is presumably eroded deeply enough to expose its base in plastically deformed rocks. It shows that the fault does not cut that zone. It implies that the Yukon cataclastic complex was the base to the Teslin lineament in the Late Jurassic-Early Cretaceous. In the same way the cover-basement detachment may have been the base of the Tintina fault during the Late Cretaceous.

The present velocity of the Pacific plate relative to the North American (5.3 cm/yr, Le Pichon, 1968) differs drastically from that deduced above for the décollement during the Cretaceous (ca 1 cm/yr) although the direction of relative motion is the same (i.e., northwestward). Assuming that the velocity of relative motion has not changed drastically since the Cretaceous, about 4 cm/yr of displacement must have occurred on other faults during this time. Possibly displacement on the Tintina and Teslin lineament was greater than the roughly 500 kilometres each, considered above. To match the present rate these faults would each have to have slipped about 2500 kilometres. This is untenable for the Tintina and improbable for the Teslin. Another explanation for the 4 cm/yr discrepancy is that estimates for the duration of strike-slip are too long by a factor of five. This is also unlikely. The most plausible explanation seems to be that the 4 cm/yr of displacement were accomplished outside the system considered so far. The main dextral slip between the North American and Pacific plates must have been localized southwest of the Lewes River arc by the Late Jurassic, perhaps on ancestral Queen Charlotte and Fairweather faults. This means that during the Late Jurassic and Cretaceous, while the Lewes River arc was obducted and while cover strata were shortened, the bulk of the intraplate motion was already outside the collision belt. Through the Triassic and Early Jurassic relative movement between the two plates was accomplished perhaps entirely on the subduction system now represented by the Teslin Suture Zone. But about the Middle Jurassic, when the Lewes River arc joined North America, most of the intraplate motion must have shifted outboard to the Queen Charlotte fault system and to a subduction zone dipping northeast under the Coast plutonic belt (Fig. 79). About 1/5 of the intraplate movement was accomplished on the Teslin lineament and the Tintina fault and their décollements. When motion on the Tintina ceased and shortening in the

Mackenzie Mountains was complete, intraplate motion was taken up entirely by the Queen Charlotte fault. During the Late Jurassic and Early Cretaceous the Intermontane, Coast Plutonic and Insular belts, between the Queen Charlotte fault and Teslin lineament, acted as a partly slipping 'clutch' between the Pacific and North American plates, a clutch more closely linked with North America than the Pacific. In the Late Cretaceous this slipping clutch was enlarged to encompass the zone between the Queen Charlotte and Tintina faults, so that a part of the Omineca Belt had become part of the clutch.

If the bulk of the strike-slip between the North American and Pacific plates was taken up outside the collision zone most of the convergence was perhaps also accomplished there. Assuming that convergence was one fifth of the strike-slip, in the same ratio as shortening and strike-slip in the Pelly Mountains, about 1000 kilometres of Pacific crust must have been consumed eastward beneath the Coast Mountains in the Late Jurassic and Cretaceous.

SUMMARY AND CONCLUSIONS

The Pelly Mountains include an autochthonous succession of mainly Paleozoic, marine platform to shelf strata that represent a southwestern shelf facies of Paleozoic North America. Allochthonous sheared metamorphic rocks were thrust over them from the southwest about the Late Jurassic. During the Cretaceous and Early Tertiary the rocks were imbricated by thrust faults, folded, regionally metamorphosed and intruded by granite batholiths. Finally, they were displaced 450 km on a dextral strike-slip fault.

The "in-place" succession, between six and eight kilometres thick, is a depositional record for Late Proterozoic or Eocambrian through Triassic time, with breaks in Middle and Late Cambrian, Early Devonian, Late Devonian, Pennsylvanian, and Early to Middle Triassic time. Proterozoic, Eocambrian and Lower Cambrian beds are conformable and constitute the Ketza group. They include a lower shale and sandstone, and an upper, argillaceous carbonate, deposited in shallow water outboard (southwest) of the North American carbonate platform edge. The carbonate is generally argillaceous, but contains prominent, laterally discontinuous Archeocyathid buildups, several hundred metres thick, and lenses of clean limestone. The Kechika Group is an Upper Cambrian and Ordovician phyllite and basalt deposited unconformably on the older beds in deep water. From northeast to southwest the group changes from shaly limestone to phyllite with locally voluminous intercalated volcanics. The volcanics (locally 500 metres thick) are submarine alkaline basalt with restricted distribution, erupted from local centres.

The Askin group, conformable above the Kechika, consists of a lower dolomitic siltstone with an upper dolostone and sandstone. It represents gradual upward shoaling and formed partly on tidal flats during the Late Silurian and Devonian. The upper part includes four laterally equivalent formations with rapid facies and thickness variation. It rests unconformably on the lower siltstone in some places, bevelling blocks that wer rotated by faulting, possibly during Late Silurian time. Volcanic rocks interstratified with the carbonate are the product of local submarine eruptions similar to those associated with the Kechika Group.

The Seagull group is trangressive and represents a general deepening of depositional conditions about Late Devonian time. It is dominated by black slate and alkaline felsic volcanic rocks, but locally includes greywacke and cherty tuff, a lateral equivalent of the volcanics. Facies variation is rapid and may be controlled by coeval faults. Locally the Seagull group unconformably rests on the Askin group. Faults by which blocks were rotated to permit erosional bevelling probably slipped about the Late Devonian.

A thin Permian siltstone, the Starr formation, disconformably rests on the Seagull group, and is itself overlain disconformably by an Upper Triassic siltstone, the Hoole formation. Both indicate a clastic marine shelf environment. Their present distribution is more restricted than that of the older groups.

The Harvey group, a thick slate unit immediately southwest of Tintina fault, is considered to be the northeastern basinal equivalent of the Ketza, Kechika, Askin and Seagull groups. It represents part of Selwyn basin, whereas the Ketza, Kechika, Askin and Seagull groups constitute Cassiar platform. St Cyr fault, the present boundary between them, may follow an older break that controlled facies during deposition.

The Pelly Mountains stratigraphic succession correlates with similar strata in the Cassiar Mountains. The succession also matches that in the northern Rocky Mountains on the northeast side of the Tintina-Northern Rocky Mountain fault system. This similarity is the basis for measurements of the displacement on that system.

The allochthonous rocks are thrust over the "in-place" beds as three lithologically distinct slices, each several hundred metres thick. Generally only one or two slices are seen at one place, but where all are present the three are in a consistent stacking order without structural mixing between them. Lowest is the Nisutlin Allochthon, the remnant of a sequence of immature clastic and volcanic rocks; the middle unit the Anvil Allochthon comprises sheared basalt and dismembered ultramafic rocks, and the uppermost, the Simpson Allochthon, is a sheared granodiorite and quartz monzonite, now granodiorite gneiss and schist. They are Mesozoic, late Paleozoic and mid-Paleozoic, respectively.

Lithologic units within the allochthonous slices are discontinuous, and each slice is internally dismembered. Rocks in each assemblage vary from place to place and the internal stratigraphy of the slices is unknown; depositional relations between strata from one slice to those of another and to other rocks are speculation. The allochthonous rocks have a penetrative flaser fabric by which they are transposed in various degrees to rocks that range from protoclastic types through mylonite to mylonite schist and gneiss. Fabric is best developed and most homogeneous in the siliceous rocks, and least penetrative in the mafic middle slice. The foliation is itself transposed, locally so strongly that the second fabric rivals the first, suggesting that the rocks were deformed repeatedly. Foliation in the rocks is generally subhorizontal or gently inclined. The rocks are retrogressively metamorphosed to greenschist facies, but locally include remnants of eclogite, which indicates metamorphism at high pressure and moderate temperature. Metamorphism and ductile deformation were simultaneous as indicated by the fabrics and are dated, by fossils, as Late Triassic, and isotopically, as Late Triassic and Early Jurassic.

Allochthonous rocks are preserved as klippen: the farthest displaced is 115 kilometres northeast of their root in the Teslin fault, here considered a suture. They show two different relations to the autochthonous rocks. The northeasternmost klippen rest on the Upper Triassic Hoole formation with a sharp subhorizontal contact roughly parallel to bedding in the footwall and to the foliation in the hanging wall. In more southwesterly exposures the contact is also sharp, but the metamorphic contrast is absent. There the rocks were metamorphosed and deformed in the Cretaceous following emplacement of the allochthonous slices, so that metamorphic contrast is masked.

Four northeast-directed thrusts, each carrying a sheet two or three kilometres thick, imbricated the autochthonous beds southwest of the Tintina fault. They moved in the Late Jurassic and Early Cretaceous, and their cumulative shortening may approach 100 kilometres. Some thrusts are folded into large open anticlines and synclines, and broken by tear faults and by newer thrusts. The newer folds and faults postdate the deformed thrust; they reflect movement on lower thrusts. Such relations indicate that the thrusts formed in order from southwest to northeast, and each carries, and locally deforms, earlier thrust sheets "piggyback" style. The thrusts are imbricated above a detachment, perhaps eight or ten kilometres deep, on which the deformed sequence slid northeastward over basement.

Between Tintina fault and the imbricated thrust zone is a ten-kilometre-wide strip with steeply dipping faults subparallel to the thrusts and the Tintina. These faults have conflicting relations; in places they may be thrusts, but elsewhere they could be transcurrent. They may be transitional between the reverse movement of the thrusts and dextral slip of Tintina fault.

"In place" and transported strata were regionally metamorphosed in parts of the project area about the late Early Cretaceous to produce the Big Salmon and Mink complexes. As unmetamorphosed Eocambrian and Early Paleozoic strata can be followed into Big Salmon Complex, its parent is known. The precursor for the Mink complex is not clear, but it and the Big Salmon Complex may be a core complex from which this cover was stripped by Devonian extension. Metamorphism occurred after the southwestern thrusts, but probably during movement on the more northeastern thrusts. Metamorphism reached lower amphibolite facies of the Barrovian facies series, and promoted partial melting. Several concordant intrusions (Big Salmon, Nisutlin, Quiet Lake batholiths) with five-kilometre-thick, - lit-par-lit migmatite margins formed and cooled, and are enclosed by the rocks from which they formed. The largest parts of the batholiths are still rooted in their parent rocks, but parts were mobilized so that they intrude their cover. The large batholiths rose five to ten kilometres, with their mantling schist, bowing the rocks and structures in their roofs and forming two northwest trending culminations, the Big Salmon-Nisutlin and Quiet Lake arches. St Cyr syncline is the structural depression between them. The culminations were produced about the early Late Cretaceous, while the granitic rocks cooled.

The boundary between the metamorphosed, extensively mobilized rocks and unmetamorphosed strata with a few small intrusions, is an abrupt northwest trending line. Southwest of it basement was extensively involved in the deformation, but on the northeast the basement remained passive. Where basement was mobilized the autochthonous cover is thinned tectonically; where it was passive the cover is imbricated and structurally repeated on large thrusts. Metamorphic culminations and some of the intrusions rose while shortening continued to the northeast. Some culminations and intrusions are therefore probably sliced by the basal detachment.

Tintina fault moved during the Late Cretaceous and Early Tertiary, following the south western thrusts and after most granitic intrusions had cooled. Dextral strike-slip has displaced the older rocks equally by about 450 kilometres. Strike-slip coincided with possible additional shortening of the strata in the Mackenzie Mountains. Tintina fault is thought to be confined to the detached, imbricated, eight to ten kilometres thick slab, and to merge with the basal detachment at depth. It probably behaved as a giant tear in the detached slice. Generally the Tintina fault is close to the boundary between involved and passive basement, but in the project area it is far to the northeast, close to the platform-basin transition. That transition may itself follow Paleozoic faults.

Emplacement of the allochthonous rocks, imbrication of the autochthonous cover, regional metamorphism, melting and mobilization, strike-slip on the Tintina, and shortening of the Selwyn and Mackenzie mountains, apparently occurred continuously, without important breaks, and with overlaps in time. They spanned the Late Jurassic, Cretaceous and Early Tertiary. They are thought to be the products of one oblique collision in which a western Mesozoic volcanic arc, built on a continental block, slid northward along, and thrust northeastward over, the ancient North American margin, while the Pacific Plate was subducted under its western edge. The allochthonous rocks are the subduction system of the arc. Autochthonous cover was imbricated in response to the overriding arc. Metamorphism, partial melting and granitic intrusion are responses to loading and down-buckling of the North American crust by the arc. Strike-slip on the Tintina reflects oblique motion between the accreted arc and North America, and shortening in the Selwyn and Mackenzie mountains is the expression of continued obduction. This model probably applies to most of the western margin of North America.

None of the mineral occurrences in the Pelly Mountains are currently economic, but many are prospective. The four main types are silver-lead-zinc veins, volcanic and exhalative zinc-lead masses, tungsten or molybdenum skarns and metamorphosed stratabound sulphides. Most silver veins are in the autochthonous beds southwest of the Tintina fault. They are in all rock units and at various structural levels and evidently postdate the main deformation. Two or three sulphide concentrations in Seagull group volcanics are in central Quiet Lake map area. They may be Mississippian proximal exhalites genetically linked with bedded barite also found in these strata. Tungsten or molybdenum skarns are spatially associated with the mid-Cretaceous granites. Southwest of the Tintina fault they are in the metamorphosed Ketza group, but on the northeast most are near the top of the Mink complex. The metamorphosed stratabound sulphides are restricted to the Nisutlin assemblage northeast of the Tintina. Because these showings have the same fabric and metamorphic grade as their host, mineralization must be pre-Early Jurassic.

REFERENCES

1. from original document

Abbott, J.G.

1977: Structure and stratigraphy of the Mount Hundere Area, southeastern Yukon; unpublished M.A.Sc. thesis, Queens University, 111 p.

1981a: A new geological map of Mt. Hundere and the area north; *in* Yukon Geology and Exploration 1979-80: Department of Indian and Northern Affairs, Whitehorse, p. 45-50.

1981b: Geology of the Seagull tin district; *in* Yukon Geology and Exploration 1979-80: Department of Indian and Northern Affairs, Whitehorse, p. 32-44.

Adams, C.J.

1981: Uplift rates and thermal structures in the Alpine Fault Zone and Alpine Schists, southern Alps, NewZealand; *in* "Thrust and Nappe Tectonics", K.R. McClay and N.J. Price, eds., Geological Society, London, Special Publication 9, p. 211-222.

Aleinikoff, J.N., Dusel-Bacon, C., Foster, H.L., and Futa, K.

1981: Proterozoic zircon from augen gneiss, Yukon-Tanana Upland, east central Alaska; Geology, 9, p. 469-473.

Bally, A.W.

1981: Thoughts on the tectonics of folded belts; in "thrust and Nappe Tectonics", K.R. McClay and .J. Price, eds., GSLSP 9, p. 13-32.

Bamber, E.W., Taylor G.C., and Proctor, R.M.

1968: Carbonaceous and Permian stratigraphy of northwestern British Columbia; Geological Survey of Canada Paper 68-15.

Bassett, H.G.

1961: Devonian stratigraphy, central Mackenzie River, Yukon, Northwest Territories, Canada; *in* Geology of the Arctic; v. 1, p. 481-498.

Blusson, S.L.

1966: Geology, Frances Lake, Yukon Territory; Geological Survey of Canada, Map 6-1966.

Bostock, H.S.

1936: Carmacks District, Yukon; Geological Survey of Canada, Memoir 189.

1948a: Physiography of the Canadian Cordillera, with special reference to the area north of the fifty-fith parallel; Geological Survey of Canada Memoir 247, 106 p.

1948b: Preliminary map, McQuesten Yukon Territory; Geological Survey of Canada Paper 48-25.

1957: Yukon Territory, Selected Field Reports of the Geological Survey of Canada 1898 to 1933; Geological Survey of Canada Memoir 284, 650 p.

1966: Notes on Glaciation in central Yukon Territory; Geological Survey of Canada Paper 65-36.

Brune, J.N.

1969: Surface waves and crustal structure; in "The Earth's Crust and Upper Mantle" P.J. Hart, ed.; Geophysical Monograph 13, American Geophysical Union, Washington, p. 230-242.

Cairnes, D.D.

1917: Scroggie, Barker, Thistle and Kirkman creeks, Yukon Territory; Geological Survey of Canada, Memoir 97, 47 p.

Campbell, R.B.

1967: Reconnaissance geology of Glenlyon map area, Yukon Territory; Geological Survey of Canada, Memoir 352, 92 p.

Campbell, R.B., Mountjoy, E.W., and Young, F.G.

1973: Geology of McBride map area, British Columbia; Geological Survey of Canada, Paper 72-35.

Cecile, M.P.

1982: The Lower Paleozoic Misty Creek Embayment, Selwyn basin, Yukon and Northwest Territories; Geological Survey of Canada, Bulletin 335, 78 p.

Cecile, M.P., Cook, D.G., and Snowdon, L.R.

1981: Plateau overthrust and its hydrocarbon potential, Mackenzie Mountains, Northwest Territories; Geological Survey of Canada, Paper 82-1A, p. 89-94.

Chapman, R.M., Weber, F.R., and Taber, B.

1971: Preliminary geologic map of the Livengood Quadrangle, Alaska; U. S. Geological Survey, Open File Map 483.

Chronic, F.J.

1979: Geology of the Guano-Guayes rare earth element bearing skarn property, Pelly Mountains, Yukon Territory; unpublished M.Sc. thesis, University of British Columbia, 122 p.

Churkin, M. Jr., and Brabb, E.E.

1965: Ordovician, Silurian and Devonian biostratigraphy of East-central Alaska; Bulletin American Association of Petroleum Geologists, v. 79, no. 6, p. 172-185.

Clar, E.
1973: Review of the structure of the eastern Alps; *in* "Gravity and Tectonics", K.A. de Jong and R. Scholten, eds.; Wiley Interscience, p. 253-270.

Cooper, M.A.
1981: The internal geometry of Nappes: criteria for models of emplacement; *in* "Thrust and Nappe Tectonics", K.R. McClay and N.J. Price, eds.; Geological Society of London, Special Publication 9, p. 225-234.

Coutts, R..C.
1980: Yukon: Places and names: Gray’s Publishing Limited, Sidney, British Columbia, Canada, 294 p.

Curray, J.R., Moore, D.G., Lawver, L.A., Emmel, F.J., Raitt, R.W., Henry, M., and Kieckhefer, R.
1979: Tectonics of the Andaman Sea and Burma; American Association of Petroleum Geologists, Memoir 29, p. 189-198.

Dawson, G.M.
1888: Report on an exploration in the Yukon District, N.W.T., and adjacent northern portion of British Columbia 1887; Geological and National Historical Survey, Canadian Annual Report 111, Part l P, B.

Debelmas, J., and Kerckhove, C.
1973: Large Gravity Nappes in the French-Swiss Alps: - in "Gravity and Tectonics", K.A. de Jong and R. Scholten, eds.; Wiley Interscience, p. 189-200.

Department of Indian and Northern Affairs
1981: Yukon Geology and Exploration 1979-80; Geology Section, DIAND, Whitehorse, Y.T., 364 p.

Erdmer, P.
1981: Comparative studies of cataclastic and allochthonous rocks in McQuesten, Laberge and Finlayson Lake map areas; Yukon Geology and Exploration 1979-80; Department of Indian and-Northern Affairs, Whitehorse, p. 60-64.

1982: Nature and significance of the metamorphic minerals and structures of cataclastic allochthonous rocks in the White Mountains, Last Peak and Fire Lake areas,Yukon Territory; unpublished Ph.D thesis, Queens University, 254 p.

Findlay, D.C.
1967: The mineral industry of Yukon Territory and southwestern District of Mackenzie, 1966; Geological Survey of Canada, Paper 6740.

1968: The mineral industry of Yukon Territory and southwestern District of Mackenzie, 1967; Geological Survey of Canada, Paper 68-68.

1969: The mineral industry of Yukon Territory and southwestern District of Mackenzie, 1968; Geological Survey of Canada, Paper 69-55.

Fradkin, P.L.
1977: The first and forgotten pipeline; Audubon magazine, vol. 79, no. 6, p. 59-79.

Fritz, W.H.
1972: Lower Cambrian trilobites from the Sekwi Formation type section, Mackenzie Mountains, northwestern Canada; Geological Survey of Canada, Bulletin 212.

1976: Broad correlations of some Lower and Middle Cambrian strata in the North American Cordillera; Geological Survey of Canada, Paper 75-1A, p. 533-540.

1980: Two new formations in the lower Cambrian Atan Group, Cassiar Mountains, north-central British Columbia; *in* Current Research, Part B, Geological Survey of Canada, Paper 80-1B, p. 217-225.

Fyles, J.T.
1964: geology of the Duncan Lake Area, Lardeau District, British Columbia; British Columbia Department of Mines and Petroleum Resources, Bulletin 49.

Gabrielse, H.G.
1963: McDame map area, Cassiar District, British Columbia; Geological Survey of Canada, Memoir 319, 138 p.

1967: Geology, Watson Lake, Yukon Terri-tory; Geological Survey of Canada, Map 19-1966.

1985: Major transcurrent displacements along the northern Rocky Mountain Trench and related lineaments in north central British Columbia; Geological Society of America, Bulletin 96, p. 1-14

Gabrielse, H.G., and Wheeler, J.O.
1961: Tectonic framework of southern Yukon and northwestern British Columbia; Geological Survey of Canada, Paper 60-24.

Gibson, D.W.
1971: Triassic stratigraphy of the Sikanni Chief River-Pine Pass Region, Rocky Mountain Foothills, Northeastern British Columbia; Geological Survey of Canada, Paper 70-31.

1975: Triassic rocks of the Rocky Mountain Foothills and Front Ranges of Northeastern British Columbia and West central Alberta; Geological Survey of Canada, Bulletin 247, 61 p.

Gordey, S.P.
1977: Stratigraphy, structure and tectonic evolution of the Southern Pelly Mountains in the Indigo Lake Area, Yukon Territory; unpublished Ph.D thesis, Queens University, 200p.

1981a: Stratigraphy, structure and tectonic evolution of Southern Pelly Mountains in the Indigo Lake Area, Yukon Territory; Geological Survey of Canada, Bulletin 318, 44 p.

1981b: Structure section across south central Mackenzie Mountains, Northwest Territories; Geological Surveyof Canada, Open File 809.

Gordey, S.P., Gabrielse, H., and Orchard, M.J.
1982: Stratigraphy and structure of Sylvester Allochton, Southwest McDame map area, Northern British Columbia; *in* Current Research, Part B, Geological Survey of Canada, Paper 82-1B, p. 101-106.

Green, L.H.
1965: The mineral industry of Yukon Territory and southwestern District of Mackenzie, 1964; GeologicalSurvey of Canada, Paper 65-19.

1966: The mineral industry of Yukon Territory and southwestern District of Mackenzie, 1965; Geological Survey of Canada, Paper 66-31.

1972: Geology of Nash Creek, Larsen Creek and Dawson map areas, Yukon Territory; Geological Survey of Canada, Memoir 364, 157 p.

Green, L.H., and Godwin, C.I.
1963: The mining industry of Yukon Territory and southwestern District of Mackenzie, 1962; Geological Survey of Canada, Paper 63-38.

1964: The mineral industry of Yukon Territory and southwestern District of Mackenzie, NWT, 1963; Geological Survey of Canada, Paper 64-36.

Green, L.H,. and Roddick, J.A.
1967: Geology, Nahanni, Yukon Territory and District of Mackenzie, Geological Survey of Canada, Map 14-1961.

Hare, F.K., and Thomas, M.K.
1974: Climate Canada; Wiley Publishers of Canada, 256 p.

Hatcher, R.D.
1981: Thrusts and Nappes in the North American Appalachian Orogen; *in* "thrust and Nappe Tectonics", K.R. McClay and N.J. Price, eds.; Geological Society of London, Special Paper 9, p. 491-499.

Hughes, J.D., and Long, D.G.F.
1980: Geology and coal resource potential of Early Tertiary strata along Tintina Trench, Yukon Territory; Geological Survey of Canada, Paper 79-32, 21 p.

Hughes, O.L., Campbell, R.B., Muller, J.E., and Wheeler, J.O.
1969: Glacial limits and flow patterns, Yukon Territory south of 65 degrees North latitude; Geological Survey of Canada, Paper 68-34, 9 p.

Hultén, E.
1968: Flora of Alaska and neighboring Territories; Stanford University Press, 1008 p.

Irving, E., Monger, J.W.H., and Yole, R.W.
1980: New paleomagnetic evidence for displaced terranes in British Columbia, - in "The continental crust and its mineral deposits", D.W. Strangway, ed.; Geological Association of Canada, Special Paper 20, p. 441-456.

Jackson, D.E., and Lenz, A.C.
1962: Zonation of Ordovician and Silurian graptolites of Northern Yukon, Canada; Bulletin American Association of Petroleum Geologists, vol. 40, p. 30-45.

James, N. P.
1979: Facies models 10- Shallowing Upward Sequences in Carbonates in "Facies Models", R.G. Walker, ed.; Geoscience Canada, Reprint Series 1, p. 109-119.

Johnston, I .R.
1936: A reconnaissance of Pelly River between Macmillan River and Hoole Canyon, Yukon; Geological Survey of Canada, Memoir 200, 19 p.

Karig, D.E., Suparka, S., Moore, G.F., and Hehanussa, P.E.
1979: Structure and Cenozoic evolution of the Sunda Arc in the central Sumatra region; American Association of Petroleum Geologists, Memoir 29, p. 223-237.

Kendrew, W.G., and Kerr, D.
1955: The climate of British Columbia and the Yukon Territory; Queen's Printer.

Kindle, E.D.
1946: Geological reconnaissance along the Canol Road, from Teslin River to Macmillan Pass, Yukon; Geological Survey of Canada, Paper 45-21.

Lees, E.J.
1936: Geology of Teslin-Quiet Lake area, Yukon; Geological Survey of Canada, Memoir 203, 30 p.

Lemoine, M.
1973: About gravity gliding tectonics in the western Alps; in "Gravity and Tectonics", K.A. de Jong and R. Scholten, eds., Wiley Interscience, p. 201-216.

Le Pichon, X.
1968: Seafloor spreading and continental drift; Journal of Geophysical Research, vol. 73, p. 3661-3697.

McConnell, R.G.
1898: Big Salmon River, Teslin River and Lake, Nisutlin River and route from Teslin Lake to Telegraph Creek; *Reprinted in*: Bostock (1957), p. 11-17.

Mertie, J.B. Jr.
1937: The Yukon-Tanana region, Alaska; United States Geological Survey, Bulletin 872, 276 p.

Money, A., and East, B.
1975: This was the North; General Publishing Co. Ltd., Toronto, 244 p.

Monger, J.W.H., and Irving, E.
1980: Northward displacement of north-central British Columbia; Nature, vol. 285, p. 289-294.

Morin, J.A.
1977: Ag-Pb-Zn mineralization in the MM deposit and associated Mississippian felsic volcanic rocks inthe St Cyr Range, Pelly Mountains; Department of Indian and Northern Affairs, Mineral Industry Report 1976, p. 83-97.

Mortensen, J.K.
1979: Stratigraphic, structural and tectonic setting of an Upper Devonian-Mississippian volcanic, sedimentary sequence and associated base metal deposits on the Pelly Mountains, southeastern Yukon Territory; unpublished M.Sc. thesis, University of British Columbia, 122 p.

Mountjoy, E.W.
1967: Triassic stratigraphy of Northern Yukon Territory; Geological Survey of Canada, Paper 66-19.

Mulligan, R.
1963: Geology of Teslin map area, Yukon Territory (105C); Geological Survey of Canada, Memoir 326, 96 p.

Norford, B.S.
1962: The Silurian fauna of the Sandpile Group of Northern British Columbia; Geological Survey of Canada, Bulletin 78

1964: Reconnaissance of the Ordovician and Silurian rocks of Northern Yukon Territory; Geological Survey ofCanada, Paper 63-39.

Okulitch, A.V., Wanless, R.K. And Loveridge, W.D.
1975: Devonian plutonism in south-central British Columbia; Canadian Journal of Earth Sciences, vol. 12, no. 10, p. 1760-1769.

Okulitch, V.J., and Greggs, R.G.
1958: Archeocyathid localities in Washington, British Columbia and the Yukon Territory; Journal of Paleontology, vol. 32, p. 617-663.

Oswald, E.T., and Senyk, J.P.
1977: Ecoregions of Yukon Territory; Fisheries and Environment Canada, 115 p.

Perry, D.G., Boucot, A.J., and Gabrielse, H.
1981: Late Early Devonian brachiopods from the Mount Lloyd area, northeastern British Columbia; Geological Survey of Canada, Bulletin 300, p. 15-40.

Pike, W.
1896: Through the subarctic forest; Edward Arnold Publisher; *Reprinted by*: Arno Press, New York 1967, 295 p.

Poole, W.H., Roddick, J.A., and Green, L.H.
1960: Geology, Wolf Lake, Yukon Territory; Geological Survey of Canada, Map 10-1960.

Porsild, A.E.
1951: Botany of southwestern Yukon adjacent to the Canol Road; Natural History Museum of Canada, Bulletin 121, 399 p.

Prest, V.K.
1969: Retreat of Wisconsin and Recent ice in North America; Geological Survey of Canada, Map 1257A.

Price, R.A., and Mountjoy, E.W.
1970: Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca rivers - Progress Report; Geological Association of Canada, Special Paper 6, p. 7-25.

Rand, A.L.
1945: Mammal investigations on the Canol Road, Yukon and Northwest Territories, 1944; Natural History Museum of Canada, Bulletin 99, 52 p

Read, B.C.
1976: Lower Cambrian stratigraphy of Pelly Mountains, central Yukon Territory; unpublished M.Sc. thesis, University of Calgary, 146 p.

1980: Lower Cambrian archeocyathid buildups, Pelly Mountains, Yukon; Geological Survey of Canada, Paper 78-18, 54 p.

Roddick, J.A.
1967: The Tintina Trench; Journal of Geology, vol. 75, p. 23-33.

Roddick, J.A., and Green, L.H.
1961a: Tay River, Yukon; Geological Survey of Canada, Map 13-1961.

1961b: Sheldon Lake, Yukon; Geological Survey of Canada, Map 12-1961.

Schuiling, R.D.
1973: Active role of continents in tectonic evolution - Geothermal models; *in* Gravity and Tectonics, K.A. de Jong and R. Scholten, eds.; Wiley Interscience, p. 35-47.

Seely, D.R.
1979: The evolution of structural highs bordering major forearc basins; American Association of Petroleum Geologists, Memoir 29, p. 245-260.

Skinner, R.
1961: Mineral industry of Yukon Territory and southwestern District of Mackenzie; Geological Survey of Canada, Paper 61-23.

1962: Mineral industry of Yukon Territory and southwestern District of Mackenzie; Geological Survey of Canada, Paper 62-27.

Spang, J.H., and Brown, S.P.
1981: Dynamic analysis of a small imbricate thrust and related structures, Front Ranges, southern Canadian Rocky Mountains; *in* Thrust and Nappe Tectonics", K.R. McClay and N.J. Price, eds.; Geological Society of London, Special Paper 9, p. 143-149.

Taylor, G.C., and Stott, D.F.
1973: Tuchodi Lakes map area, British Columbia; Geological Survey of Canada, Memoir 373.

Tempelman-Kluit, D.J.
1974: Geology, Quiet Lake, Finlayson Lake, Sheldon Lake and Tay River map areas, Yukon Territory (105F, G, J and K); Geological Survey of Canada, Open file 212.

1977: Geology of Quiet Lake (105F) and Finlayson Lake (105G) map areas, Yukon Territory; Geological Survey of Canada, Open file 486.

1979: Transported cataclasite, ophiolite and granodiorite in Yukon: Evidence of arc-continent collision; Geological Survey of Canada, Paper 79-14.

Tempelman-Kluit, D.J., Abbott, G., and Read, B.
1974: Stratigraphy and structure of the Pelly Mountains; Geological Survey of Canada, Paper 74-1A, p. 43-44.

Tempelman-Kluit, D.J., Abbott, G., Gordey, S.P., and Read, B.
1975: Stratigraphic and structural studies in the Pelly Mountains, Yukon Territory; Geological Survey of Canada, Paper 75-1A, p. 45-48.

Tempelman-Kluit, D.J., Gordey, S.P., and Read, B.
1976: Stratigraphic and structural studies in the Pelly Mountains, Yukon Territory; Geological Survey of Canada, Paper 76-1A, p. 97-106.

Tempelman-Kluit, D.J., and Wanless, R.K.
1975: Potassium argon age determinations of metamorphic and plutonic rocks in the Yukon Crystalline Terrane; Canadian Journal of Earth Sciences, vol. 12, no. 11, p. 1895-1909.

1980: Zircon ages for Pelly Gneiss and Klotassin Granodiorite in western Yukon; Canadian Journal of Earth Sciences, vol. 17, no. 3, p. 297-306.

Thompson, R. I .
1981: The nature and significance of large "blind" thrusts within the northern Rocky Mountains of Canada; *in* Thrust and Nappe Tectonics", K.R. McClay and N.J. Price, eds.; Geological Society of London, Special Paper 9, p. 449-462.

Trumpy, R.
1973: The timing of orogenic events in the central Alps; in "Gravity and Tectonics", K. de Jong and R. Scholten, eds.; Wiley Interscience, p. 229-251.

Wellman, H.W.
1969: Wrench (transcurrent) fault systems; - in The Earth's Crust and Upper Mantle, P.J. Hart, ed.; Geophysical Monograph 13, American Geophysical Union, p. 544-549.

Wheeler, J.O.
1961: Whitehorse map area; Geological Survey of Canada Memoir 312.

Wheeler, J.O., Charlesworth, H.A.K., Monger, J.W.H., Muller, J.E., Price, R.A., Reesor, J.E., Roddick, J.A., and Simony, P.S.
1974: Western Canada; - *in* Mesozoic-Cenozoic Orogenic Belts - data for orogenic studies, Geological Society of London, Special Paper, 809 p.

Wheeler, J.O., Green, L.H., and Roddick, J.A.
1960a: Quiet Lake, Yukon Territory; Geological Survey of Canada, Map 7-1960.

1960b: Finlayson Lake, Yukon Territory; Geological Survey of Canada, Map 8-1960.

Williams, H.
1979: Appalachian Orogen in Canada; Canadian Journal of Earth Sciences, Tuzo Wilson Volume, v. 16, p. 792-807.

Wilson , C .
1970: Campbell of the Yukon; Macmillan of Canada, Toronto, 189 p.

Wyllie, P.J.
1977: Crustal anatexis: an experimental review; Tectonophysics, vol. 43, p. 41-71.

REFERENCES

2. post 1985, these refer to footnotes in text)

Abbott, J.G.
1986: Epigenetic mineral deposits of the Ketza-Seagull district, Yukon; *in* Yukon Geology, Volume 1, Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 56-66.

Abbott, J.G., Gordey, S.P. and Tempelman-Kluit, D.J.
1986: Setting of stratiform, sediment-hosted lead-zinc deposits in Yukon and northeastern British Columbia; *in*: Mineral Deposits of the Northern Cordillera, J. Morin (ed.), Canadian Institute of Mining and Metallurgy, Special volume 37, p. 1-18.

Bond, J.D. and Kennedy, K.E.
2005: Late Wisconsinan McConnell ice-flow and sediment distribution patterns in the Pelly Mountains, Yukon; *in* Yukon Exploration and Geology 2004, D.S. Emond, L.L. Lewis and G.D. Bradshaw (eds.), Yukon Geological Survey, p. 67-82.

Bond, J.D., Murphy, D.C., Colpron, M., Gordey, S.P., Plouffe, A., Roots, C., Lipovsky, P.S., Stronghill, G. and Abbott, J.G.
2002: Digital compilation of bedrock geology and till geochemistry, northern Finlayson Lake map area, southeastern Yukon (105G); Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File, 2002-7(D) *and* Geological Survey of Canada, Open File 4243.

Bond, J.D. and Plouffe, A.
2002: Finlayson Lake Targeted Geoscience Initiative (southeastern Yukon), Part 2: Quaternary geology and till geochemistry; *in* Yukon Exploration and Geology 2001, D.S. Emond, L.H. Weston and L.L. Lewis (eds.), Exploration and Geological Services Division, Yukon Region, Department of Indian and Northern Affairs, p. 209-228 plus 10 fold-out maps.

Bradshaw, G.D., Tucker, T.L., Peter, J.M., Paradis, S., and Rowins, S.M.
2001: Geology of the Wolverine polymetallic volcanic-hosted massive sulphide deposit, Finlayson Lake district, Yukon Territory, Canada: *in* Yukon Exploration and Geology 2000, Emond, D.S. and Weston, L.H., eds., Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 269-287.

Bradshaw, G.D., Rowins, S.M., Peter, J.M., and Taylor, B.E.
2003: Genesis of the Wolverine deposit, Finlayson Lake district, Yukon: a transitional style of polymetallic massive sulphide mineralization in an ancient continental margin setting: The Gangue, v. 79, p. 1-8.

Breitsprecher, K., and Mortensen, J.K. (compilers)
2004: Yukonage 2004: A database of isotopic age determinations for rock units from Yukon Territory. Yukon Geological Survey, 2004 (1) CD-ROM

Burke, M.
2000: Yukon mining and exploration overview 1999; *in*: Yukon Exploration and Geology 1999, D.S. Emond and L.H. Weston (eds.), Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 2-32.

Colpron, M., Nelson, J. L. and Murphy, D.C.
2006: A tectonostratigraphic framework for the pericratonic terranes of the northern Cordillera; *in* Paleozoic evolution and metallogeny of pericratonic terranes at the ancient Pacific margin of North America, Canadian and Alaskan Cordillera; Colpron, M. and Nelson, J. L. (eds.) Geological Association of Canada, Special Paper 45, p. 1-24.

Colpron, M. and Nelson, J. L.
2009: A Palaeozoic Northwest Passage: incursion of Caledonian, Baltican and Siberian terranes into eastern Panthalassa, and the early evolution of the North American Cordillera; *in* Earth Accretionary Systems in Space and Time. P.A. Cawood and A. Kroner (eds.), Geological Society, London, Special Publication 318, p.273-307, doi:10.1144/SP318.10.

Creaser, R., Goodwin-Bell, J.S., and Erdmer, P.,
1999: Geochemical and Nd isotopic constraints for the origin of eclogite protoliths, northern Cordillera: implications for the Paleozoic tectonic evolution of the Yukon-Tanana terrane. Canadian Journal of Earth Sciences, v. 36, p. 1697-1709.

Deklerk, R. and Traynor, S. (compilers)
2005: Yukon MINFILE 2005. Yukon Geological Survey, CD-ROM

Devine, F.D., Carr, S.D., Murphy, D.C., Davis, W.J., Smith, S., and Villeneuve, M.
2006: Geochronological and geochemical constraints on the origin of the Klatsa metamorphic complex: Implications for Early Mississippian high-pressure metamorphism within Yukon-Tanana terrane: *in* Paleozoic evolution and metallogeny of pericratonic terranes at the ancient Pacific margin of North America, Canadian and Alaskan Cordillera: Colpron, M. and Nelson, J.L. (eds.). Geological Association of Canada, Special Paper. 45, p. 107-130.

Devine, F.D., Murphy, D.C., Kennedy, R., Tizzard, A., and Carr, S.D.
2004 Geological setting of retrogressed eclogite and jade in the southern Campbell Range: Preliminary structure and stratigraphy, Frances Lake area (NTS 105H), southeastern Yukon; *in* Yukon Exploration and Geology 2003, Emond, D.S. and Lewis, L.L., eds., Yukon Geological Survey, p. 89-105.

Duk-Rodkin, A.
1999: Glacial limits map of Yukon Territory: Geological Survey of Canada, Open File 3694, *also* Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Geoscience Map 1999-2,
scale 1:1 000 000.

Erdmer, P.
1985: An examination of the cataclastic fabrics and structures of parts of Nisutlin, Anvil and Simpson allochthons, central Yukon: test of the arc-continent collision model. Journal of Structural Geology, v. 7, p. 57-72.

1987: Blueschist and eclogite in mylonitic allochthons, Ross River and Watson Lake areas, southeastern Yukon: Canadian Journal of Earth Sciences, v. 24, p. 1439-1449.

Erdmer, P., Ghent, E.D., Archibald, D.A., and Stout, M.Z.
1998: Paleozoic and Mesozoic high-pressure metamorphism at the margin of ancestral North America in central Yukon: Geological Society of America Bulletin, v. 110, p. 615-629.

Fonseca, A.
1998: Structural evolution of the Ketza River gold deposit; *in*: Yukon Exploration and Geology 1997, Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 77-81.

Gabrielse, H.
1998: Geology of Cry Lake and Dease Lake map area, north-central British Columbia; Geological Survey of Canada, Bulletin 504, 147 p.

Gabrielse, H., Murphy, D.C. and Mortensen, J.K.
2006: Cretaceous and Cenozoic dextral orogen-parallel displacements, magmatism , and paleogeography, north-central Canadian Cordillera, *in* Paleogeography of the Northern American Cordillera: Evidence for and against large-scale displacements; Haggart, J.W., Enkin, R.J. and Monger, J.W.H., eds., Geological Association of Canada, Special Paper 46, p. 255-276.

Gabrielse, H. and Yorath, C.J.,
1991. Tectono-physiograpic belts. Chapter 1 *in*: Geology of the Cordillera Orogen in Canada; H. Gabrielse and C.J. Yorath (ed.): Geological Survey of Canada, Geology of Canada, no 4, p. 3-11 (also Geological Society of America, The Geology of North America, v. G-2).

Gangloff, R.A., May, K.C. and Storer, J.E.
2004: An early late Cretaceous dinosaur tracksite in central Yukon Territory, Canada. Ichnos, v. 11, p. 299-309.

Goodfellow, W.D., Cecile, M.P. and Leybourne, M.I.
1995: Geochemistry, petrogenesis and tectonic setting of lower Paleozoic alkalic and potassic volcanic rocks, northern Canadian Cordilleran miogeocline: Canadian Journal of Earth Sciemces, v. 32, p. 1236-1254.

Gordey, S.P.
In press: Evolution of the Selwyn Basin region, Sheldon Lake (105J) and Tay River (105K) map areas, central Yukon Territory: Geological Survey of Canada, Bulletin.

Gordey, S.P. and Anderson, R.G.
1993: Evolution of the northern Cordilleran miogeocline, Nahanni map area (105 I), Yukon and Northwest Territories: Geological Survey of Canada, Memoir 428, 213 p.

Gordey, S.P., and Makepeace, A.J.
2001: Bedrock geology, Yukon Territory; Geological Survey of Canada, Open File 3754 *and* Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File 2001-1, scale 1:1 000 000.

2003: Yukon Digital Geology (version 2): Yukon Geological Survey, Open File 2003-9 *and* Geological Survey of Canada, Open File 1749.

Gordey, S.P. and Stevens, R.A.
1994: Geology, Teslin map area, Yukon Terriory (105C); Geological Survey of Canada, Open File 2884.

Grant, S.L.
1997: Geochemical, radiogenic tracer isotopic and U-Pb geochronological studies of Yukon Tanana terrane rocks from the Money klippe, southeastern Yukon, Canada: unpublished M.Sc. thesis, University of Alberta, Edmonton, Alberta, 177 p.

Hart, C.J.R. and Villeneuve, M.
1999: Geochronology of Neogene alkaline volcanic rocks (Miles Canyon basalt), southern Yukon Territory, Canada: the relative effectiveness of laser $^{40}Ar/^{39}Ar$ and K-Ar geochronology; Canadian Journal of Earth Sciences, v. 36, p. 1495-1507.

Hunt, J.A.
2002: Volcanic-associated massive sulphide (VMS) mineralization in the Yukon-Tanana terrane and coeval strata of the North American miogeocline, in the Yukon and adjacent areas: Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Bulletin 12, 107 p.

Jackson, L. E., Gordey, S. P., Armstrong, R. L., **and Harakal, J.**
1986: Bimodal Paleogene volcanics near Tintina fault, east-central Yukon and their possible relationship to placer gold; *in* Yukon Geology, Volume 1, Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 139-147.

Jackson, L.E. , Ward, B, Duk-Rodkin, A. and Hughes, O.L.
1991: The last Cordilleran ice sheet in southern Yukon Territory; Geographie physique at Quaternaire, vol. 45, p. 341-354.

Ledo, J., Jones, A.G. and Ferguson, I.J.
2002: Electromagnetic image of a strike slip fault: the Tintina fault, northern Canadian Cordillera; Geophysical Research Letters, vol. 29, no. 1, p. 661-664.

Ledo, J., Jones, A.G., Ferguson, I.J. and Wolynec, L.
2004: Lithospheric structure of the Yukon, northern Canadian Cordillera, obtained from magnetotelluric data; Journal of Geophysical Research, vol. 109, no. B4, 15 p.

Long, D.G.F., Lowet, G. and Sweet, A.R.
2001: Age and setting of dinosaur trackways, Ross River area, Yukon Territory (105F/15); *in*: Yukon Exploration and Geology 2000, D.S. Emond and L.H. Weston (eds.), Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 181-198.

Lowey, G.W. and Lowey, J.F.
1986: Geology of Spencer Creek (105B/1 and Daughney Lake (105B/2) Rancheria District, southeast Yukon; Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, Open File 1986-1.

Mihalynuk, M.G., Nelson, J.L. and Friedman, R.
1998: Regional geology and mineralization of the Big Salmon Complex (104N NE and 104O NW); *in* Geological Fieldwork, British Columbia Department of Energy, Mines and Petroleum Resources, Paper 1999-1, p. 6-1 – 6-18.

Monger, J.W.H., Wheeler, J.O., Tipper, H.W., Gabrielse, H., Harms, T.A., Struik, L.C., Campbell, R.B., Dodds, C.J., Gehrels, G.E., and O'Brien, J.
1991: Cordilleran terranes. Upper to Devonian to Middle Jurassic Assemblages: *in* Gabrielse, H., and Yorath, C.J., eds., Geology of the Cordilleran Orogen in Canada, Geology of Canada No. 4: Geological Survey of Canada, p. 281-327.

Morin, J.A. (Editor)
1986: Mineral deposits of northern Cordillera; Canadian Institute of Mining and Mtallurgy, Special volume 37, 378 p.

Morrow, D.W.
1999: Lower Paleozoic stratigraphy of northern Yukon Territory and northwestern District of Mackenzie; Geological Survey of Canada, Bulletin 538, 202 p.

Mortensen, J.K.
1983: Age and evolution of the Yukon-Tanana Terrane, southeastern Yukon Territory: unpublished Ph.D thesis, University of California, Santa Barbara, 155 p.

1990: Geology and U-Pb geochronology of the Klondike District, west-central Yukon: Canadian Journal of Earth Sciences, v. 27, p. 903-914.

1992a: Pre-mid-Mesozoic tectonic evolution of the Yukon-Tanana Terrane, Yukon and Alaska: Tectonics, v. 11, p. 836-853.

1992b: New U-Pb ages for the Slide Mountain Terrane in southeastern Yukon Territory: *in* Radiogenic Age and Isotopic Studies: Report 5. Geological Survey of Canada, Paper 91-2, p. 167-173.

Mortensen, J.K., and Jilson, G.A.
1985: Evolution of the Yukon-Tanana terrane: evidence from southeastern Yukon Territory: Geology, v. 13, p. 806-810.

Murphy, D.C.
1988: Geology of Gravel Creek and Irvine Lake map areas (105B/10, B/11); Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File 1988-1.

1998: Stratigraphic framework for syngenetic mineral occurrences, Yukon-Tanana terrane south of Finlayson Lake: a progress report: *in* Yukon Exploration and Geology 1997: Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 51-58.

Murphy, D.C., Colpron, M., Gordey, S.P., Roots, C.F., Abbott, G., and Lipovsky, P.S.

2001: Preliminary bedrock geological map of northern Finlayson Lake area (NTS 105G), Yukon Territory: Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File 2001-33, 1:100 000 scale.

Murphy, D.C., Colpron, M., Roots, C.F., Gordey, S.P., and Abbott, J.G.

2002: Finlayson Lake Targeted Geoscience Initiative (southeastern Yukon), Part 1: Bedrock geology: *in* Yukon Exploration and Geology 2001, Emond, D.S., Weston, L.H., and Lewis, L.L., eds., Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, p. 189-207.

Murphy, D.C., Mortensen, J.K., Piercey, S.J., Orchard, M.J. and Gehrels, G.E.

2006: Mid-Paleozoic to early Mesozoic tectonostratigraphic evolution of Yukon-Tanana and Slide Mountain terranes and overlap assemblages, Finlayason Lake massive sulphide district, southeastern Yukon; *in* Colpron, M. and Nelson, J.L., eds., Paleozoic evolution and metallogeny of pericratonic terranes at the Ancient Pacific Margin of North America, Canadian and Alaskan Cordillera: Geological Association of Canada, Special Paper 45, p. 75-105.

Murphy, D.C. and Piercey, S.J.

1999: Geological map of Wolverine Lake area (105G/8), Pelly Mountains, southeastern Yukon: Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, Open File 1999-3.

2000a: Syn-mineralization faults and their re-activation, Finlayson Lake massive sulphide district, Yukon-Tanana Terrane, southeastern Yukon: *in* Yukon Exploration and Geology 1999, Emond, D.S., and Weston, L.H., eds., Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 55-66.

Nelson, J.L.

1993: The Sylvester allochthon: Upper Paleozoic marginal-basin and island-arc terranes in northern British Columbia: Canadian Journal of Earth Sciences, v. 30, p. 631-643.

Nelson, J.L., Paradis, S., Christensen, J., and Gabites, J.

2002: Canadian Cordilleran Mississippi Valley-type deposits: A case for Devonian-Mississippian back-arc hydrothermal origin: Economic Geology, v. 97, p. 1013-1036.

Nelson, J.L., Colpron, M., Piercey, S.J., Dusel-Bacon, C., Murphy, D.C., and Roots, C.F.

2006: Paleozoic tectonic and metallogenic evolution of pericratonic terranes in Yukon, northern British Columbia and eastern Alaska: *in* Colpron, M., Nelson, and Thompson, R.I., eds., Paleozoic evolution and metallogeny of pericratonic terranes at the ancient Pacific margin of North America, Canadian and Alaskan Cordillera: Geological Association of Canada, Special Paper 45, p. 323-360.

.

Piercey, S.J., Murphy, D.C., Mortensen, J.K., and Creaser, R.

2004 Mid-Paleozoic initiation of the northern Cordilleran marginal back-arc basin: geologic, geochemical and neodymium isotope evidence from the oldest mafic magmatic rocks in Yukon-Tanana terrane, Finlayson Lake district, southeast Yukon, Canada: Geological Society of America Bulletin, v. 116, p. 1087-1106.

Piercey, S. J., Nelson, J. L., Colpron, M., Dusel-Bacon, C., Simard, R.-L. and Roots, C.F.

2006: Paleozoic magmatic, tectonic and crustal growth history of the ancient Pacific margin of North America: Geochemical constraints from Yukon-Tanana terrane of northern British Columbia, Yukon and Alaska; *in* Colpron, M. and Nelson, J. L. (eds.), Paleozoic evolution and metallogeny of pericratonic terranes at the ancient Pacific margin of North America, Canadian and Alaskan Cordillera; Geological Association of Canada, Special Paper 45, p. 281-322.

Pigage, L.C.

2001: Geological map of Anvil District (NTS 105K/2, 3, 5, 6, 7, 11), Central Yukon: Exploration and Geological Services Division, Yukon Region, Indian and Northern Affairs Canada, Open File 2001-31, 1:100 000 scale.

2004: Bedrock geology compilation of the Anvil District (parts of 105K/2, 3, 5, 6, 7, and 11), central Yukon: Yukon Geological Survey, Bulletin 15, 103 p.

Plint, H.E., and Gordon, T.M.

1997: The Slide Mountain Terrane and the structural evolution of the Finlayson Lake fault zone, southeastern Yukon: Canadian Journal of Earth Sciences, vol. 34, p. 105-126.

Pope, M.C. and Sears, J.W.

1997: Cassiar platform, north-central British Columbia: A miogeoclinal fragment from Idaho; Geology, vol. 25, p. 515-518, doi:10.1130/0091-7613

Roots, C.F.
1988: Cambro-Ordovician volcanic rocks in eastern Dawson map area, Ogilvie Mountains, Yukon; *in* Yukon Geology Vol. 2, Exploration and Geological Services, Indian and Northern Affairs Canada , p. 81-87.

Roots, C.F., Harms, T.A., Simard, R-L., Orchard, M.J. and Heaman, L.
2002: Constraints on the age of the Klinkit assemblage east of Teslin Lake, northern British Columbia; *in* Geological Survey of Canada, Current Research 2002-A7, 11p.

Schulze, H.C.
1996: Summary of the Kudz Ze Kayah project, volcanic-hosted massive sulphide deposit, Yukon Territory; *in* Yukon Exploration and Geology 1995, Exploration and Geological Services, Indian and Northern Affairs Canada , p. 29-32.

Smith, C.A.S., Miekle, J.C. and Roots, C.F. (eds.)
2004. Ecoregions of the Yukon, Biophysical properties of Yukon landscapes; Agriculture and Agri-Foods Canada, PARC Technical Bulletin No. 04-01, Summerland, British Columbia, 313 p.

Snyder, D.B., Roberts, B.J. and Gordey, S.P.
2005. Contrasting seismic characteristics of three major faults in northwestern Canada; Canadian Journal of Earth Sciences, v. 42, p. 1223-1237.

Stevens, R.A.
1992. Regional geology, fabric and structure of the Teslin suture zone in Teslin map area, Yukon Territory; *in* Current Research, Geological Survey of Canada, Paper 92-1A, p. 287-296.

1994. Geology of the Teslin suture zone in parts of Lebarge (105E/1), Quiet Lake (105F/4) and Teslin (105C/11, 13, 14) map areas, Yukon Territory; Geological Survey of Canada, Open File 2768.

Stevens, R.A., Mortensen, J.K. and Hunt, P.A.
1993. U-Pb and ^{40}Ar-^{39}Ar geochronology of plutonic rocks from the Teslin suture zone, Yukon Territory; *in* Radiogenic Age and Isotopic Studies: Report 7; Geological Survey of Canada, Paper 93-2, p. 83-90.

Stroshein, R.
1996. Geology and gold deposits at Ketza River, Yukon Territory: a progress report; *in* Yukon Exploration and Geology 1995, Exploration and Geological Services Division, Yukon, Indian and Northern Affairs Canada, p. 43-48.

Yukon Government, Bureau of Statistics
2006: http://www.eco.gov.yk.ca/stats/archives.html
File:demographic/pop122006.pdf (accessed Nov. 2012)