Early and Middle Proterozoic evolution of Yukon, Canada[1, 2]

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Abstract: This paper provides a comprehensive synthesis of virtually all units and events of Early and Middle Proterozoic age in the Yukon, spanning ~1 Ga. Early and Middle Proterozoic time was dominated by a series of extensional-basin-forming events punctuated by orogenesis, magmatism, and hydrothermal activity. Basinal deposits include the Wernecke Supergroup (>1.71 Ga), Pinguicula Group (~1.38 Ga), and Mackenzie Mountains Supergroup (1.00–0.78 Ga). Igneous rocks include the Bonnet Plume River Intrusions (1.71 Ga), Slab volcanics (≥1.6 Ga), Hart River sills and volcanics (1.38 Ga), and Bear River (Mackenzie) dykes (1.27 Ga). A voluminous hydrothermal event generated the widespread Wernecke breccias at 1.60 Ga. The Racklan orogeny deformed the Wernecke Supergroup prior to emplacement of the Wernecke Breccia. The Corn Creek orogeny deformed Mackenzie Mountains Supergroup and older rocks prior to deposition of the Windermere Supergroup (<0.78 Ga). Long intervals with scanty rock records extended for as much as 300 Ma and appear to represent periods of crustal stability and subaerial conditions. By the time of Windermere rifting (<0.78 Ga), the supracrust of northwestern Laurentia was a mature, largely denuded orogenic belt with a composite sedimentary–metamorphic–igneous character. New isotopic data include Nd depleted mantle model ages for the Wernecke Supergroup (2.28–2.69 Ga) and Wernecke Breccia (2.36–2.96 Ga), a U–Pb zircon age for a Hart River sill $1381.9^{+5.3}_{-3.7}$ (Ma), detrital U–Pb zircon ages from the basal Pinguicula Group (1841–3078 Ma), detrital muscovite ages from the Mackenzie Mountains Supergroup (1037–2473 Ma), and muscovite ^{40}Ar/^{39}Ar cooling ages from the Wernecke Supergroup (788 ± 8 and 980 ± 4 Ma).

Résumé : Cet article fournit une synthèse complète de presque toutes les unités et les événements du Protérozoïque précoce et moyen du Yukon, couvrant environ un milliard d'années. L'époque du Protérozoïque précoce et moyen était dominée par une série d'événements d'extension, formant des bassins, ponctués par des orogènes, du magmatisme et de l'activité hydrothermale. Les dépôts de bassins comprennent le Supergroupe de Wernecke (>1,71 Ga), le Groupe de Pinguicula (~1,38 Ga) et le Supergroupe des monts Mackenzie (1,00–0,78 Ga). Les roches ignées comprennent les intrusions de Bonnet Plume River (1,71 Ga) les volcaniques de Slab (≥1,6 Ga); les filons couches et les volcaniques de Hart River (1,38 Ga) et les dykes de Bear River (Mackenzie) (1,27 Ga). Un volumineux événement hydrothermal a généré les brèches étendues de Wernecke à 1,60 Ga. L'orogène de Racklan a déformé le Supergroupe de Wernecke avant la mise en place des brèches de Wernecke. L'orogène de Corn Creek a déformé le Supergroupe des monts Mackenzie et des roches plus anciennes avant la déposition du Supergroupe de Windermere (<0,78 Ga). De longs intervalles avec de minces évidences rocheuses s'étendent pour des périodes aussi longues que 300 Ma et semblent représenter des périodes de stabilité crustale et de conditions subaériennes. À l'époque de la distension Windermere (<0,78 Ga), la supercroûte nord-ouest Laurentia était une ceinture orogénique mature, largement dénudée avec un caractère composite sédimentaire-métamorphique-igné. De nouvelles données isotopiques comprennent : les âges du modèle de manteau appauvri en Nd pour le Supergroupe de Wernecke (2,28–2,69 Ga) et la brèche de Wernecke (2,36–2,96 Ga); un âge U–Pb sur zircon pour un filon-couche de Hart River ($1381.9^{+5.3}_{-3.7}$ Ma); des âges U–Pb sur des zircons détritiques du Groupe de Pinguicula de base (1841–3078 Ma); des âges sur de la muscovite détritique du Supergroupe des monts Mackenzie (1037–2473 Ma) et des âges ^{40}Ar/^{39}Ar de refroidissement de la muscovite du Supergroupe de Wernecke (788 ± 8 et 980 ± 4 Ma).

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Introduction

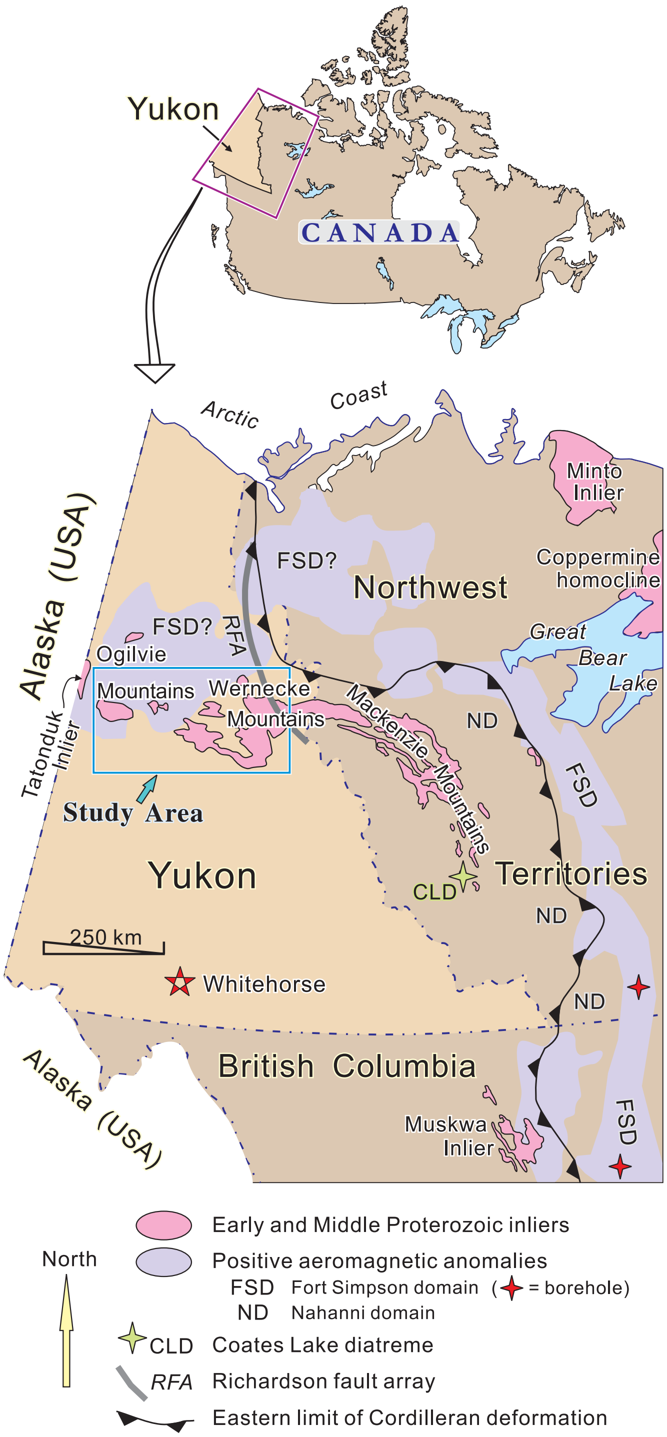
The Cordilleran region of Canada is well known for its Late Proterozoic and Phanerozoic history in which western Laurentia (ancestral North America) was transformed from rifted margin to complex orogen. Its earlier history is less well understood and extends back into Middle and Early Proterozoic times, spanning over a billion years. This "pre-Cordilleran" interval is well preserved in the Yukon (Fig. 1), where the region underwent several cycles of extension and basin formation punctuated by other events, including deformation, metamorphism, exhumation, magmatism, and hydrothermal brecciation. Understanding the chronology of these events is essential to a broader understanding of Laurentia, its role in Precambrian continental interactions, and its transition from craton to modern landmass.

This paper integrates more than a decade of our research on Early and Middle Proterozoic rocks in the Wernecke, Ogilvie, and Mackenzie mountains of northern Yukon Territory with other studies in northwestern Canada (Fig. 1). The rocks are exposed in three main regions known as the Wernecke, Hart River, and Coal Creek inliers, collectively called the study area in this paper (Fig. 2*a*). The inliers occur in structural culminations surrounded by strata of Late Proterozoic to early Tertiary age and lie within the Mackenzie–Ogilvie platform, a paleogeographic division of the Foreland belt characterized by platformal successions of Neoproterozoic and Early Paleozoic age (Norris 1997; Gordey and Anderson 1993). Most of our research has focused on the Wernecke and Hart River inliers. Our studies of Early and Middle Proterozoic rocks in the Coal Creek inlier are cursory. All of these inliers are remote and mountainous and are most easily accessed by helicopter in the summer.

Our research involved 1 : 50 000 scale mapping and complementary geochemical and isotopic studies and was built on a foundation laid by workers dating back to the 1940s. Arguably the most enduring previous contributions to the early history of the region are maps and reports of the Geological Survey of Canada, particularly those by Wheeler (1954), Green (1972), Blusson (1974), Eisbacher (1981), Norris (1984), Bell (1986*a*), and Thompson (1995). Other influential works include a regional geological framework by Young et al. (1979), stratigraphic and structural studies by Delaney (1981) and Eisbacher (1981), descriptions of hematitic breccias by Bell (1986*b*), and seismic interpretations by Dredge Mitchelmore and Cook (1994).

By integrating our findings with previous data and interpretations, we herein provide a comprehensive record of the early geological history of the Yukon. This contribution builds on maps and reports by Abbott (1997), Thorkelson (2000), and Thorkelson et al. (2001*a*, 2001*b*, 2002). Some of the interpretations in this paper are based on limited information and need to be tested by more detailed studies. Nevertheless, this is the first report to integrate information from all the known sedimentary, igneous, and metamorphic units of Early and Middle Proterozoic age in the Yukon and should serve as a basis for additional investigations in North America and worldwide.

Fig. 1. Map of northwestern Canada, showing the study area in the context of main Early and Middle Proterozoic inliers, positive aeromagnetic anomalies, Coates Lake diatreme, and selected boreholes into crystalline basement. The aeromagnetic anomalies are defined by regions with approximately > 60 nT (Brunstein 2002) and have been interpolated in data-poor areas in the Wernecke Mountains.



The unseen basement

All of the rocks described in this paper lie within the supracrust. The underlying crystalline basement is nowhere exposed in the Yukon, and its age and nature are virtually unknown. In a general sense, the basement is likely to be attenuated igneous and metamorphic crust of the Laurentian craton (e.g., Norris and Dyke 1997). The nearest exposures of crystalline basement are located in the Northwest Territories, approximately 750 km to the east of our study area. In the western Northwest Territories, "basement" is commonly regarded as the rocks involved in, or older than, the Wopmay orogen, with a cratonization age range of ca. 1.84–2.0 Ga. Reviews of basement ages and affinities in and around the Cordilleran region are provided in Ross (1991*a*, 2002), Ross et al. (1999), and Ross and Eaton (2002).

Aeromagnetically defined basement provinces in northwestern Canada were identified by Hoffman (1989) and revisited by numerous workers, including Villeneuve et al. (1991), Cook et al. (1992), and Aspler et al. (2003). In most of the Northwest Territories these provinces are reasonably well defined because the overlying sedimentary cover is relatively thin (e.g., Villeneuve et al. 1991; Dredge Mitchelmore and Cook 1994; MacLean and Cook 2004), and the magnetic signature of basement rocks is discernable. Toward the Cordillera, however, where the sedimentary cover is stratigraphically thicker and has been thickened further during orogenesis, the aeromagnetic signal of basement weakens and in many places cannot be reliably distinguished from the supracrustal rocks (Aspler et al. 2003). A broad gap in aeromagnetic coverage exists for much of the study area and along most of the border between the Yukon and the Northwest Territories (Brunstein 2002; Aspler et al. 2003).

Two aeromagnetically defined domains are evident to the east and southeast of the Yukon (Hoffman 1989). One is the Fort Simpson domain (Fig. 1), a positive, curvilinear anomaly that extends from northern Alberta into the Northwest Territories. Flanking this domain on both sides is the Nahanni domain (Hoffman 1989), which is characterized by an aeromagnetic "low." Subsequent interpretations by Aspler et al. (2003) restricted the Nahanni domain to the low west of the Fort Simpson "high." Near Great Bear Lake, the Fort Simpson anomaly appears to widen and curve westward, forming two elliptical magnetic highs that are among the largest in North America (Brunstein 2002) (labelled "FSD?" in Fig. 1). Aspler et al. used a 50 km low-pass filter to show that the Fort Simpson domain may be tracked continuously into these elliptical highs but acknowledged that different geological features may be responsible for this seemingly unbroken geophysical feature. Positive isostatic gravity anomalies generally coincide with the north-trending Fort Simpson aeromagnetic domain and the elliptical magnetic highs. A distinct gravity low separates the north end of the Fort Simpson domain from the elliptical magnetic anomalies to the west, however (Cook et al. 1992, 2004; Geological Survey of Canada 1999). Additional work is required to determine if the elliptical magnetic highs belong to the same geological feature as the Fort Simpson domain or if they represent, for example, a distinct geological basement province.

Granitoid samples from basement within the Fort Simpson domain, taken from the bottom of two oil and gas exploration bore holes (Fig. 1), yielded U–Pb zircon ages of ca. 1.845 Ga. This finding led Villeneuve et al. (1991) to support Hoffman's (1989) interpretation of the Fort Simpson domain as an Early Proterozoic magmatic arc. Villeneuve et al. hypothesized that the Fort Simpson arc may have been derived by anatexis of the Nahanni domain and used isotopic arguments to show that, if this were the case, the Nahanni would have an age >2.1 Ga. The only direct indication of what the Nahanni basement may consist of comes from granitic clasts in the Coates Lake diatreme, 400 km to the southeast of the study area (Fig. 1). The clasts contain zircons with a crystallization age of ca. 1130 Ma and a poorly constrained inheritance age of ~1.6 Ga (Jefferson and Parrish 1989; Mortensen and Colpron 1998).

Regional geological framework

The inliers of Early and Middle Proterozoic rocks in the Yukon (Fig. 2) are exposed in structural culminations that are mainly contractional in origin. Some of the smaller inliers may be simply regarded as the cores of anticlinoria that developed during Late Cretaceous – Paleogene (Laramide) orogenesis (Norris 1984). In the larger (and better studied) exposures, such as the Wernecke inlier (Figs. 2, 3), the rocks display a history of deformation and exhumation extending back to the Early Proterozoic (e.g., Eisbacher 1978). A combination of Proterozoic and Phanerozoic deformational events is therefore necessary to account for the present patterns of distribution.

Young et al. (1979) divided the supracrustal strata of northwestern Laurentia into three unconformity-bounded "sequences" A, B, and C, with the A–B boundary at ca. 1.2 Ga and the B–C boundary at ca. 0.8 Ga (Fig. 2*b*). Subsequent studies have shown greater complexity in the stratigraphic and structural history of the region such that the original sequences have been split into temporally distinct subdivisions. In the Yukon, we recognize three unconformity-bounded successions within sequence A, at least two within sequence B, and at least two within sequence C (Fig. 1*b*) (Gordey and Anderson 1993; Abbott 1997; Thorkelson 2000). Differences in age within a sequence are as great as they are between sequences. Consequently, the sequence A–B–C nomenclature has limited remaining utility in the characterization of Proterozoic Yukon.

Although the Early and Middle Proterozoic sedimentary successions are impressively thick (Fig. 2*b*; cumulative thickness ~22 km), the total duration of basinal sedimentation was probably small in relation to the overall interval of time (~1 Ga). Consequently, the paleogeographic character of northwestern Laurentia in Early to Middle Proterozoic time was probably dominated by subaerially exposed terrain rather than marine basins. During the long hiatuses that are evident among the successions (Fig. 2*b*), the region may have been denuding and supplying sediment to basins beyond the Yukon. In these intervals of subaerial exposure, entire continental successions may have been deposited and eroded, an idea that is supported by Laughton et al. (2005). Indeed, a broad range of geological events occurred between periods of basinal sedimentation. A recent depiction of the paleogeographic environment as a series of long-lived submarine fans (Snyder et al. 2002) overemphasized the importance of sedimentation relative to other processes and lacunas in the geological record (Thorkelson et al. 2003*a*).

Genesis of the supracrustal environment, >1.71 Ga

The long and varied history of the Precambrian supracrustal evolution of the Yukon began, most likely, with atten-

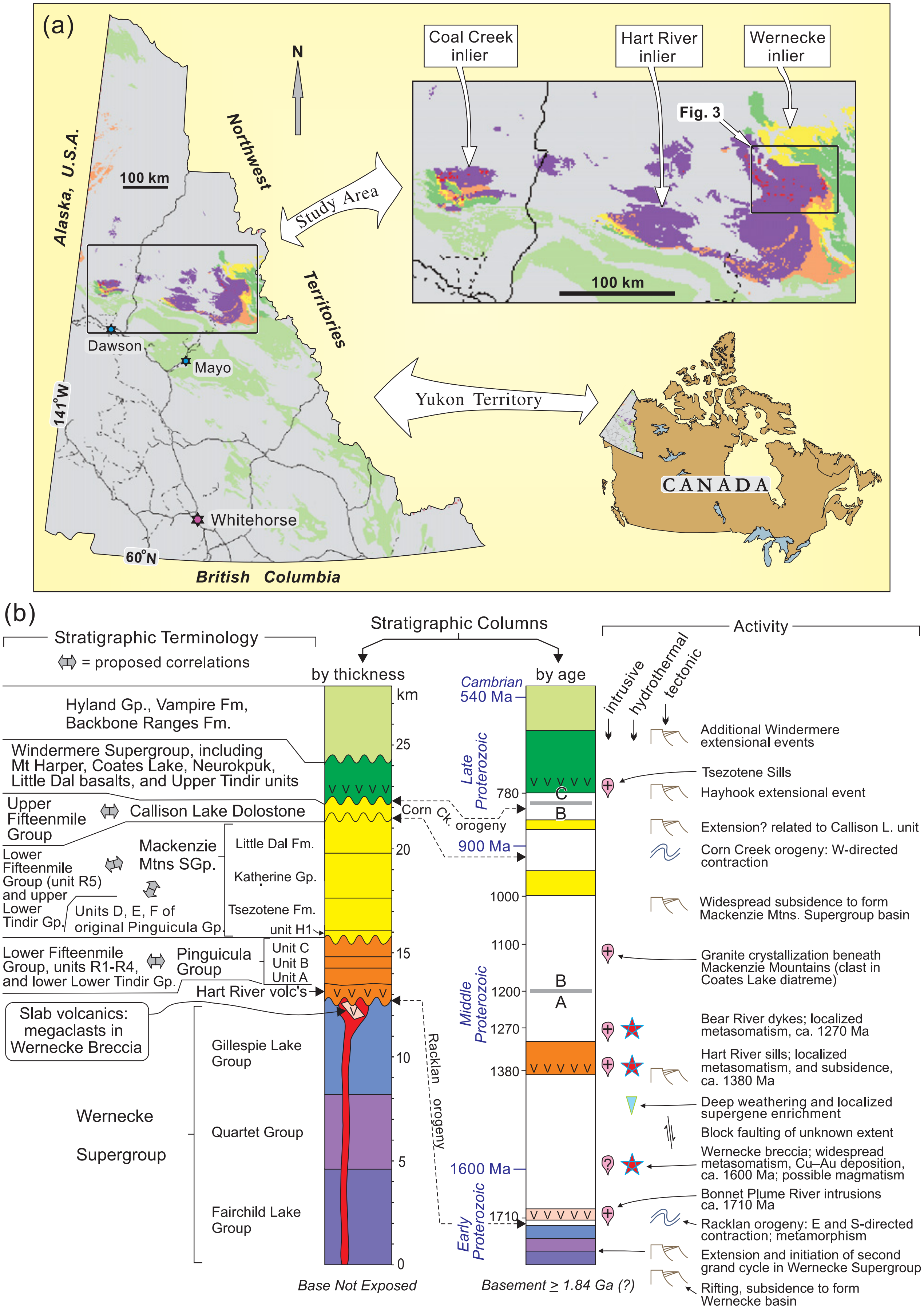

(a)
N
100 km
Alaska, U.S.A.
Northwest
Territories
Study Area
Coal Creek inlier
Hart River inlier
Wernecke inlier
Fig. 3
100 km
Dawson
Mayo
Whitehorse
141°W
60°N
British Columbia
Yukon Territory
CANADA
(b)
Stratigraphic Terminology
= proposed correlations
Stratigraphic Columns
by thickness
by age
Activity
intrusive
hydrothermal
tectonic
km
Hyland Gp., Vampire Fm, Backbone Ranges Fm.
Windermere Supergroup, including Mt Harper, Coates Lake, Neurokpuk, Little Dal basalts, and Upper Tindir units
Upper Fifteenmile Group
Callison Lake Dolostone
Lower Fifteenmile Group (unit R5) and upper Lower Tindir Gp.
Mackenzie Mtns SGp.
Little Dal Fm.
Katherine Gp.
Tsezotene Fm.
unit H1
Units D, E, F of original Pinguicula Gp.
Lower Fifteenmile Group, units R1-R4, and lower Lower Tindir Gp.
Pinguicula Group
Unit C
Unit B
Unit A
Hart River volc's
Slab volcanics: megaclasts in Wernecke Breccia
Wernecke Supergroup
Gillespie Lake Group
Quartet Group
Fairchild Lake Group
25
20
15
10
5
0
Corn Ck. orogeny
Racklan orogeny
Base Not Exposed
Cambrian 540 Ma
Late Proterozoic
780
C
B
900 Ma
1000
1100
B
A
1200
1270
1380
Middle Proterozoic
1600 Ma
1710
Early Proterozoic
Basement ≥ 1.84 Ga (?)
Additional Windermere extensional events
Tsezotene Sills
Hayhook extensional event
Extension? related to Callison L. unit
Corn Creek orogeny: W-directed contraction
Widespread subsidence to form Mackenzie Mtns. Supergroup basin
Granite crystallization beneath Mackenzie Mountains (clast in Coates Lake diatreme)
Bear River dykes; localized metasomatism, ca. 1270 Ma
Hart River sills; localized metasomatism, and subsidence, ca. 1380 Ma
Deep weathering and localized supergene enrichment
Block faulting of unknown extent
Wernecke breccia; widespread metasomatism, Cu–Au deposition, ca. 1600 Ma; possible magmatism
Bonnet Plume River intrusions ca. 1710 Ma
Racklan orogeny: E and S-directed contraction; metamorphism
Extension and initiation of second grand cycle in Wernecke Supergroup
Rifting, subsidence to form Wernecke basin

Fig. 2. (*a*) Map of the Yukon, showing the study area with main Proterozoic assemblages (produced using Gordey and Makepeace 1999) and the location of Fig. 3. Colours are keyed to columns and descriptions in (*b*) except for Wernecke Supergroup, shown as uniform purple in (*a*) but as three colours in (*b*). Black lines on the detailed inset map are roads and major trails. (*b*) Generalized thickness-stratigraphic and time-stratigraphic columns indicating major geological events and proposed correlations between units. Age of the Racklan orogeny is constrained between age of Slab volcanics (shown here as ~1.71 Ga but possibly as young as 1.60 Ga) and the Wernecke Supergroup (>1.71 Ga). Age of Corn Creek orogeny is constrained between age of Mackenzie Mountains and Windermere supergroups (1.00–0.78 Ga), and two possible positions relative to the Upper Fifteenmile Group – Callison Lake dolostone are shown. The upper and lower age limits shown for strata in the "by age" column are generally poorly constrained except where a thick bounding line is used (e.g., base of the Mackenzie Mountains Supergroup). Boundaries between sequences A, B, and C as proposed by Young et al. (1979) are shown in the "by age" column. Fm., Formation; Gp., Group; SGp., Supergroup.

uation of the underlying crystalline basement to form the Wernecke basin. Attenuation seems probable because the oldest known supracrustal succession, the Wernecke Supergroup, is ~13 km thick (Delaney 1981; Thorkelson 2000). A succession of this thickness cannot be envisaged, using modern tectonic analogues, to have been deposited on basement of normal continental thickness. The timing of the attenuation, however, is as uncertain as the age and nature of the underlying basement (virtually unknown). The minimum age of basin development is provided by the Bonnet Plume River Intrusions (1.71 Ga), which crosscut the Wernecke Supergroup (Figs. 2, 4) (Thorkelson et al. 2001*a*). If 10 Ma is regarded as the minimum time required for basin formation and infilling, then 1.72 Ga is the minimum age for the genesis of the Wernecke basin.

The depositional history of the Wernecke Supergroup as described by Delaney (1981) and Thorkelson (2000) may be understood in terms of two main clastic-to-carbonate grand cycles as recorded in three sedimentary groups (Figs. 2, 3). These groups are the Fairchild Lake Group (lowest; ~4.6 km thick), the Quartet Group (~3.4 km thick), and the Gillespie Lake Group (highest; ~4.7 km thick). Although metamorphism of the succession locally reaches greenschist grade, most of it retains enough sedimentary character for its depositional history to be firmly understood. The first grand cycle is recorded by the Fairchild Lake Group, which grades upward from siltstone and minor carbonate into carbonate and shale. This progression reflects general shallowing and sediment starvation of the basin. Inception of the second grand cycle is marked by the black, pyritic shales of the lower Quartet Group which indicate continuing sediment starvation in a deeper water environment. Higher in the Quartet Group, increased sediment influx is indicated by interbedded shale and siltstone, which generally coarsen upward and grade into thick, platformal carbonates of the Gillespie Lake Group, completing the second cycle. Shallow-water to locally emergent conditions were maintained during deposition of the Gillespie Lake Group.

The two grand cycles in the Wernecke succession may be interpreted as two stages of basin subsidence and sediment infilling, with the first cycle representing initial basin development and the second representing deepening of the basin and marine transgression, possibly through extension. The accumulation of >4 km of shallow-water carbonates at the end of the second cycle indicates protracted subsidence of the Wernecke basin. Abrupt facies changes in the Gillespie Lake Group infer that synsedimentary normal faulting played an important role in basin architecture (Thorkelson 2000).

The Wernecke basin has been characterized as a deep marine depocentre that received detritus shed from cratonal areas of Laurentia (Delaney 1981; Dredge Mitchelmore and Cook 1994). The size and shape of the basin are poorly constrained, although paleocurrents from the Wernecke inlier indicate a mainly southward direction of sediment transport from a highland to the north (Delaney 1981). These paleocurrent directions, however, may have been reoriented by subsequent events of deformation, as suggested by Aitken and McMechan (1992) and Schwab et al. (2004), leaving the actual direction of sediment transport and provenance uncertain. Delaney (1981), following Norris (1972), argued that the Wernecke basin margin followed an east–west trend to the north of the Wernecke and Ogilvie mountains, then curved southward in the Northwest Territories to follow the general trend of the Cordilleran orogen (Fig. 1). Seismic and aeromagnetic data were subsequently used to corroborate such an arcuate basin configuration and a westward-dipping ramp (Cook et al. 1991; Cook 1992; Dredge Mitchelmore and Cook 1994). Correlations of strata and seismic stratigraphy (Aitken and McMechan 1992; Cook 1992; Cook and Maclean 1995) suggest that the Wernecke basin may have shallowed dramatically to the east (e.g., Cook and van der Velden 1993; Cook et al. 2004) and continued as a pericratonic sea as far east as the Coppermine homocline (Fig. 1). These interpretations hinge largely on the equivalence of the Wernecke Supergroup with some or all of the Hornby Bay assemblage in the Coppermine homocline (Aitken and McMechan 1992), a speculative correlation that remains problematic due to the younger age (at least in part) of the Hornby Bay assemblage (Thorkelson et al. 2001*a*). MacLean and Cook (2004) attempted to reconcile this problem by proposing a significant hiatus between the lowest units of the Hornby Bay assemblage and the overlying Hornby Bay Group, and then correlating the lower units with the Wernecke Supergroup. The Muskwa assemblage in northern British Columbia (Fig. 1) may also be correlative with the Wernecke Supergroup, as suggested by broadly similar sedimentary and Nd-isotope characteristics (Long and Pratt 2000; Ross et al. 2001). Farther south, no potentially correlative strata have been identified. The Belt–Purcell Supergroup of southeastern British Columbia and the northwestern United States was deposited mainly from 1.47 to 1.37 Ga (Anderson and Davis 1995; Doughty and Chamberlain 1996) and is younger by hundreds of millions of years.

Delaney (1981) suggested that the Wernecke Supergroup formed in response to the Wopmay orogeny, an event of contractional tectonism that was largely complete by 1.84 Ga. This suggestion remains plausible and is broadly supported by Nd depleted-mantle model ages from the Wernecke Super-

Fig. 3. Geology of the northern Wernecke Mountains. (*a*) Generalized geological map and cross sections showing main rock units and structures. (*b*) Cross sections from (*a*), enlarged and annotated. S.L., sea level. (*c*) Sources used for this compilation. (*d*) Legend for map and sections.

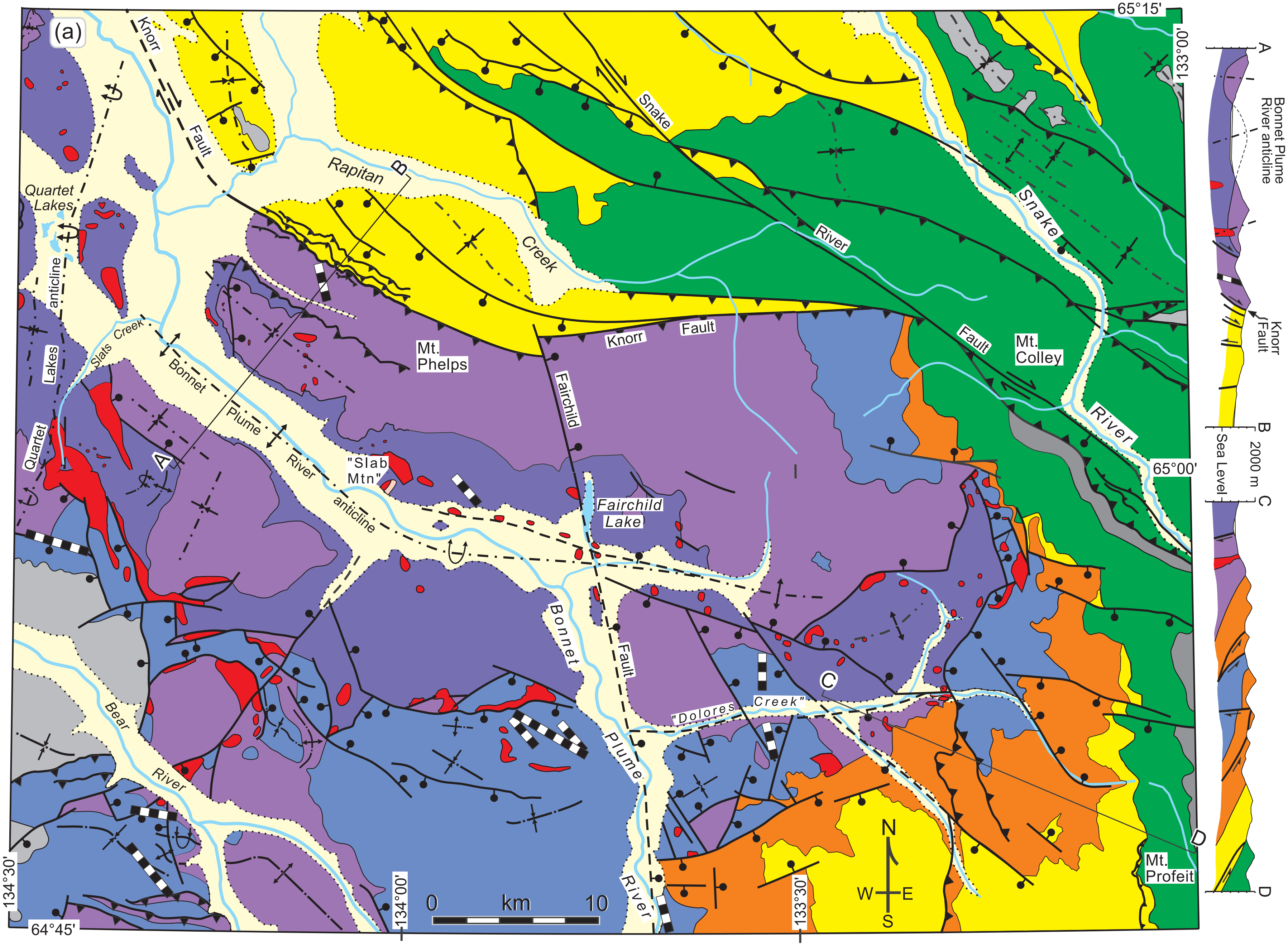
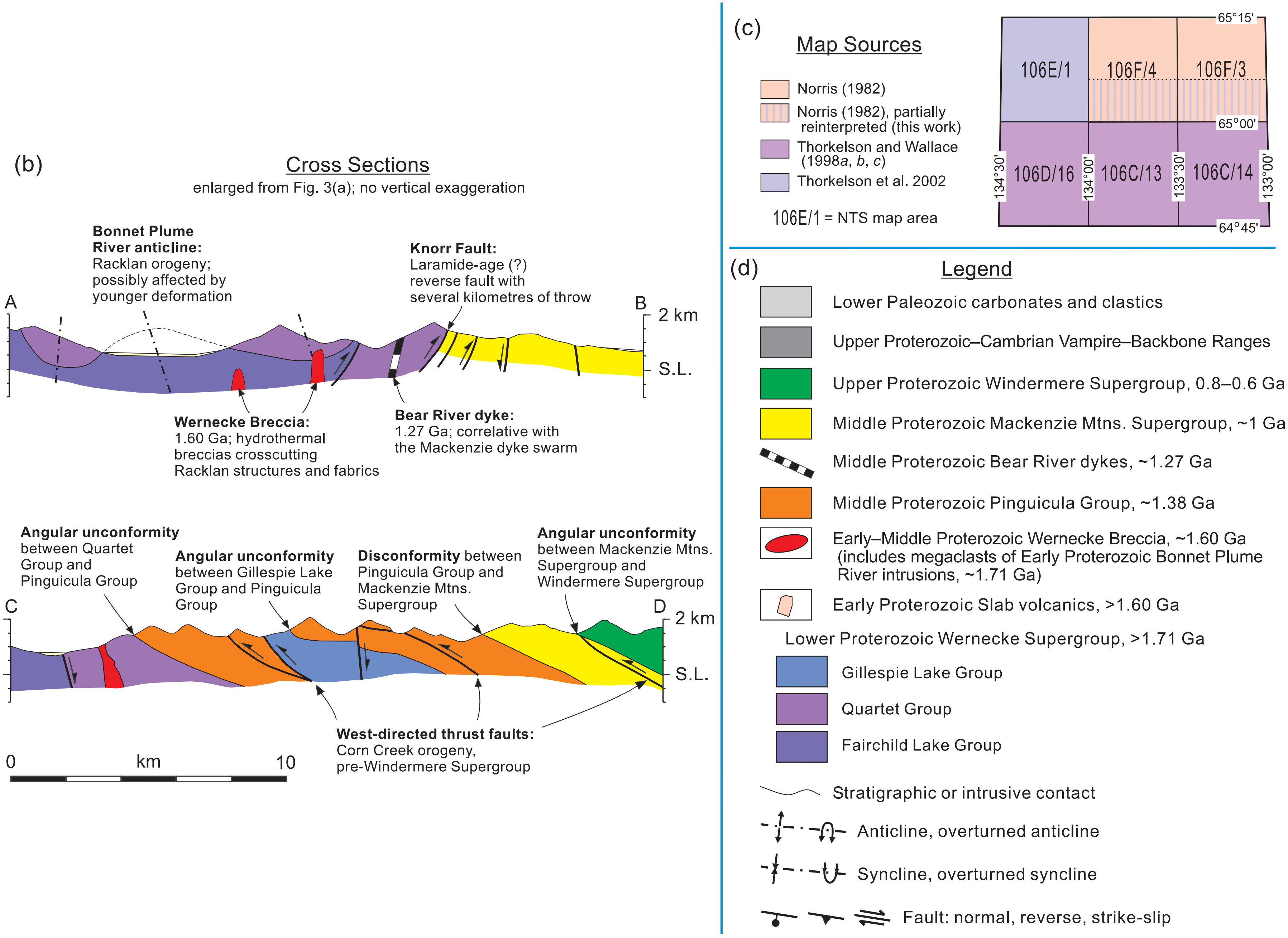


Fig. 3 (*concluded*).



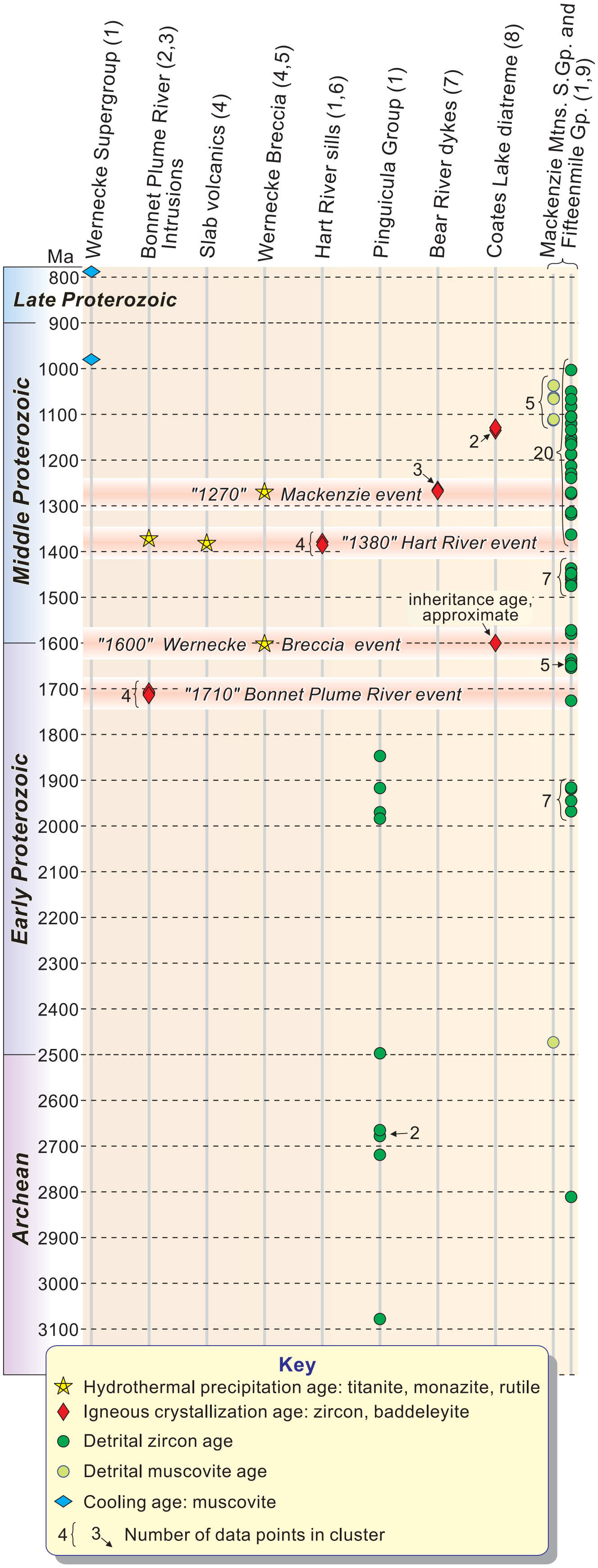


Fig. 4. Summary of geochronology for the Proterozoic of the Yukon and nearby areas of the Northwest Territories. Ages are grouped by interpreted significance (igneous crystallization, cooling, etc.). Main igneous and (or) metasomatic events are indicated by horizontal bands, with approximate ages in Ma. The following sources of data are indicated in parentheses in column headings: 1, this paper; 2, Thorkelson et al. (2001*a*); 3, Thorkelson (2000); 4, Thorkelson et al. (2001*b*); 5, Parrish and Bell (1987); 6, Abbott (1997); 7, Schwab et al. (2004); 8, Jefferson and Parrish (1989) and Mortensen and Colpron (1998); 9, Rainbird et al. (1997).

group of 2.28–2.69 Ga (Table 1) and $\varepsilon_{\mathrm{Nd}}$ values (Fig. 5) that are consistent with derivation in large part from the Wopmay orogen and other rocks of the northwestern Canadian Shield. It is not known whether the Wernecke Supergroup was deposited synchronously with Wopmay orogenesis or at a later time. In either case, basin formation by extension and crustal thinning is implied by the pair of grand cycles evident in the Wernecke stratigraphy. The thickness of the Wernecke Supergroup is consistent with an origin on a passive margin or in a deep intracontinental extensional basin (Thorkelson et al. 2001*a*), and we regard both possibilities as viable options. An intracratonic basin is attractive because of the preponderance of other Early and Middle Proterozoic intracratonic cover sequences on Laurentia (e.g., Ramaekers 1981; Hoffman 1989; Ross et al. 2001; Rainbird et al. 2003). A passive margin is attractive because the intensity of Racklan orogeny (discussed in the next section) is more consistent with collisional tectonics along a continental margin than intracratonic deformation of a more structurally insulated interior basin. Wernecke-correlative basins to the south could have split away from Laurentia during continent-scale Windermere rifting and subsequent lower Paleozoic extension (e.g., Cook et al. 1991; Ross 1991*b*) and may reside entirely on another landmass. The lack of exposure of the lowest Wernecke Supergroup strata (or older supracrustal units that may underlie the Wernecke Supergroup) hinders further interpretations regarding basin origin and evolution.

Early Proterozoic deformation, metamorphism, and magmatism

After deposition, the Wernecke Supergroup was deformed, metamorphosed, uplifted, partially exhumed, intruded by magma, and overlain by volcanics (Figs. 2–4). These processes can be grouped into three main events: (*i*) Racklan orogeny, (*ii*) emplacement of the Bonnet Plume River Intrusions, and (*iii*) eruption of the Slab volcanics. All three occurred prior to 1.60 Ga, when much of the study area was invaded by hydrothermal fluids to form the Wernecke breccias. Eruption of the Slab volcanics occurred after the Racklan orogeny, but the timing of Bonnet Plume River plutonism relative to the volcanism and orogenesis remains uncertain.

Racklan orogeny

In our view, the Racklan orogeny is defined as contractional deformation in the Yukon that occurred prior to hydrothermal brecciation at 1.60 Ga. The meaning of “Racklan orogeny,” however, has evolved with time, from Wheeler’s (1954) original observation of an angular unconformity in the Wernecke

Table 1. Sm–Nd isotope data for Wernecke Supergroup rocks and Wernecke breccias.

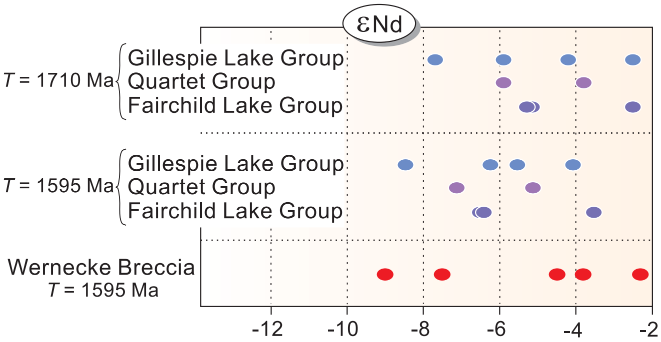
	UTM coordinates												
Sample No.	Northing	Easting	Unit	Notes	Sm (ppm)	Nd (ppm)	$\frac{^{147}Sm}{^{144}Nd}$	$\frac{^{143}Nd}{^{144}Nd}$	Uncertainty[a]	T_{DM}[b]	T (Ma)	$\varepsilon_{Nd}(T)$	ε_{Nd} at 1595 Ma
DT-92-77-1B	7200450	533050	Wernecke Breccia		6.88	36.26	0.1147	0.511583	0.000012	2.42	1595	–3.8	
DT-93-11-3B	7201800	534400	Wernecke Breccia		2.81	13.51	0.1255	0.511507	0.000008	2.85	1595	–7.5	
DT-93-107-2B	7191700	557900	Wernecke Breccia		3.04	13.95	0.1317	0.511841	0.000011	2.45	1595	–2.3	
DT-93-160-3B	7207600	546100	Wernecke Breccia		6.76	32.85	0.1245	0.511419	0.000008	2.96	1595	–9.0	
DT-94-13-1	7199300	571500	Wernecke Breccia	Cu rich	6.44	38.65	0.1008	0.511405	0.000008	2.36	1595	–4.5	
DT-92-55-1B	7203200	532400	Quartet Group	Siliciclastic	8.08	45.83	0.1066	0.511431	0.000008	2.45	1710	–3.8	–5.1
DT-92-27-2	7204050	536300	Quartet Group	Siliciclastic	9.14	49.91	0.1107	0.511372	0.000015	2.64	1710	–5.9	–7.1
DT-92-80-1B	7197300	532050	Gillespie Lake Group	Siliciclastic	7.77	51.06	0.0920	0.511332	0.000008	2.28	1710	–2.5	–4.1
DT-93-129-1B	7195300	555400	Gillespie Lake Group	Siliciclastic	2.69	15.50	0.1049	0.511392	0.000008	2.47	1710	–4.2	–5.6
DT-92-129-2B	7195300	555400	Gillespie Lake Group	Dolostone	2.16	7.44	0.1756	0.512100	0.000013	na	1710	–5.9	–6.2
DT-92-99-1	7192250	545800	Gillespie Lake Group	Siliciclastic	2.40	9.84	0.1473	0.511691	0.000009	na	1710	–7.7	–8.4
DT-92-126-1	7200900	544800	Fairchild Lake Group	Siliciclastic	5.92	29.74	0.1204	0.511508	0.000021	2.69	1710	–5.3	–6.5
DT-92-129-1	7198100	545600	Fairchild Lake Group	Siliciclastic	2.39	11.70	0.1236	0.511691	0.000009	2.48	1710	–2.5	–3.5
DT-92-132-7	7197200	546100	Fairchild Lake Group	Siliciclastic	4.20	24.25	0.1049	0.511347	0.000012	2.53	1710	–5.1	–6.4

Note: Sm–Nd isotope analyses were carried out at the Radiogenic Isotope Facility at the University of Alberta, Edmonton, Alta. The analytical protocols used are the same as those reported by Creaser et al. (1997).

[a] Uncertainty in $^{143}Nd/^{144}Nd$ is expressed as 2 standard errors (2SE).

[b] Depleted-mantle model age uses the formulation of Goldstein et al. (1984). For samples with $^{147}Sm/^{144}Nd > 0.14$, no model age is calculated (na).

Fig. 5. Comparison of ε_{Nd} values from the Wernecke Supergroup with those from the Wernecke Breccia (data in Table 1). Values for Wernecke Supergroup units are shown for youngest possible age of deposition (1710 Ma) and age of Wernecke Breccia formation (1595 Ma).



inlier (Fig. 2) to (*i*) introduction of the term Racklan orogeny by Gabrielse (1967), (*ii*) confirmation and elaboration of Racklan orogeny by Eisbacher (1978, 1981), (*iii*) recognition of an equivalent event in the Coal Creek inlier (Fig. 2) called the Fifteenmile orogeny by Mercier (1989), (*iv*) extension of the term to cover more time and space by Cook (1992), (*v*) restriction of the term to pre-1.60 Ga deformation (two phases) by Thorkelson (2000), (*vi*) recognition of a third phase of pre-1.60 Ga deformation by Brideau et al. (2002), and (*vii*) clarification of the timing of Racklan orogeny relative to eruption of the Slab volcanics by Laughton et al. (2005).

Our reasons for restricting the term Racklan orogeny to pre-1.60 Ga are twofold. First, the angular unconformity observed by Wheeler (1954) lies beneath the Pinguicula Group (Figs. 2, 3), which is crosscut by Hart River sills (ca. 1.38 Ga; Figs. 2, 4) (Abbott 1997; this report). These relations restrict the timing of Racklan orogeny to >1.38 Ga. Second, nearly all of the pre-1.38 Ga structures formed prior to 1.60 Ga, as indicated by crosscutting hydrothermal breccias of that age (Figs. 2, 4) (Thorkelson et al. 2001*b*; Brideau et al. 2002; Laughton et al. 2005). The oldest possible age of the Racklan orogeny is the age of the Wernecke Supergroup, which is constrained to be >1.71 Ga and is likely no older than 1.84 Ga. In Fig. 2*b*, the Racklan orogeny is indicated within a narrow interval between the Slab volcanics and the Wernecke Supergroup. Although the stratigraphically bounded position of the Racklan is shown correctly in Fig. 2*b*, the actual possible age range is greater than shown because the age of the Slab volcanics could be as young as 1.6 Ga, and the age of the Wernecke Supergroup may be as old as 1.84 Ga (or older). The term Fifteenmile orogeny for deformation in the Coal Creek inlier (Mercier 1989) is supplanted by the term Racklan orogeny.

Racklan structures include polyphase folding and fabric development at greenschist metamorphic grades. The folds occur as both major and minor structures in two nearly perpendicular sets. The fabrics include a foliation plus superposed crenulations and kink bands and are best developed in "schist belts" of the Fairchild Lake Group, the lowest stratigraphic unit. The earliest folds, as recognized in the Wernecke inlier (Fig. 2), are north-trending tight folds, generally overturned to the east. The Quartet Lakes anticline is the largest fold of this set identified so far (western side of Fig. 3*a*). Where it involves the Fairchild Lake Group, it is associated with a fine-grained schistosity that is particularly well developed in tight, minor folds. The second phase of folding, as understood from the geology in Figs. 3*a* and 3*b*, produced open to tight, south-verging folds, the most prominent of which is the Bonnet Plume River anticline. How these folds correlate with the crenulations and kink bands has not yet been determined. Large folds in the lower and middle crust imaged by the Slave – Northern Cordillera Lithospheric Evolution (SNORCLE) line of Lithoprobe (Cook et al. 2004) may be Racklan structures.

In the schist belts of the Fairchild Lake Group, which commonly range from 100 m to 2 km wide, the main schistosity was developed at peak temperatures of 450–550 °C, as indicated by porphyroblasts of chloritoid (and locally garnet) in a foliated groundmass of quartz, chlorite, and muscovite (Brideau et al. 2002; Laughton et al. 2005). Subsequent events deformed the rock into a crenulated schist and, in turn, imparted a set of conjugate kink bands but did not result in significant growth of new minerals. The striking three-phase history of these deformed schists contrasts with flanking regions of relatively undeformed and metamorphosed rocks, many of which retain original sedimentary structures. Apparently, the parts of the Wernecke Supergroup which were transformed into schist (because of localized, high strain) became significantly weaker than the surrounding rock once the mica- and chlorite-rich schistosity developed. Subsequent phases of deformation preferentially affected these metamorphosed zones and produced superposed crenulations and kinks, while surrounding rocks with more granular textures behaved more competently and generally did not develop these features. The greenschist metamorphic conditions are more evident in the schists than in neighbouring "unmetamorphosed" rocks because of enhanced new mineral growth induced by deformation.

In summary, the Racklan orogeny can be understood as a series of at least three phases of deformation, the first of which was apparently east directed and is now preserved as east-verging overturned folds and a localized schistose fabric. Subsequent phases formed south-directed folds, crenulations, and superposed kink bands. All of these features led to partial exhumation and erosion of the Wernecke Supergroup followed by eruption of the Slab volcanics and brecciation by hydrothermal fluids, as described later in the paper and justified more fully in Thorkelson et al. (2001*b*) and Laughton et al. (2005). The Racklan orogeny is regarded as the temporal equivalent to the Forward orogeny in the Northwest Territories (Cook and MacLean 1995; Thorkelson et al. 2003*c*; MacLean and Cook 2004). Taken together, these events affected a region >1250 km across, from the western Ogilvie Mountains to the Coppermine homocline (Fig. 1).

Bonnet Plume River Intrusions

Between lithification of the Wernecke Supergroup and invasion of fluids at 1.60 Ga (Figs. 2, 4), mantle-derived magmas of mainly fine-grained dioritic to gabbroic composition crystallized within the Wernecke Supergroup at ca. 1.71 Ga (Thorkelson et al. 2001*a*). The resultant suite of mainly mafic plutons is termed the Bonnet Plume River Intrusions. The suite also includes an occurrence of quartz syenite and one of anorthosite, although hydrothermal alteration during Wernecke Breccia formation may be partly responsible for these unusual compositions (Laznicka and Edwards 1979; Thorkelson 2000). The mafic rocks are generally characterized as continental tholeiites with enriched alkali and alkaline earth element abundances relative to transition metals and rare earth elements, with slight, negative Nb anomalies. An extensional or rift origin was preferred by Thorkelson et al. (2001*a*), but an association with a mantle plume or even a distal volcanic arc cannot be ruled out, geochemically. Samples with modest alteration have ε_{Nd} values, calculated for the time of crystallization, of +0.7 to –1.7, and depleted-mantle model ages of 2.29–2.57 Ga (Thorkelson et al. 2001*a*). Progressive contamination by crust with older Proterozoic or Archean ages appears likely.

The original dimensions and contact relations of these plutons are generally unknown because most have been displaced from their original intrusive positions by massive surges of hydrothermal fluids during the formation of Wernecke Breccia (Figs. 2–4; described later in the paper). As such, most Bonnet Plume River Intrusions occur as clasts and megaclasts in zones of Wernecke Breccia and are commonly altered, veined, mineralized, and locally fragmented. Most of the plutonic clasts range from a few metres to 20 m across, but the largest ones are up to 200 m wide and 1000 m long, with exposed areas of up to 0.2 km^2 (Fig. 3*a*). These large bodies are perhaps more suitably envisaged as intrusions that have been engulfed by brecciating fluids and separated from their host rocks, but not displaced far from their original positions in the crust. In most cases, engulfment appears to be complete in the sense that intrusions cannot be tracked across their bounding breccia zones and into the Wernecke Supergroup host rock. For this reason, the intrusions may have originally resembled small spheres or vertical cylinders rather than long, tabular dykes.

Whether the Bonnet Plume River Intrusions occurred before or after Racklan orogeny is unknown. Few if any of the intrusions are foliated, suggesting a post-kinematic age of emplacement. It is possible, however, that the intrusions are pre-Racklan and did not develop a foliation because they were stronger and less prone to recrystallization than some units in the Wernecke Supergroup. Another possibility is that the intrusions were emplaced during the Racklan orogenic interval but after the main phase of foliation development and metamorphism.

Slab volcanics

The Slab volcanics are an informally named unit that is spectacularly exposed on Slab Mountain (unofficial name) and at other nearby localities in the Wernecke inlier (Fig. 3). The volcanics consist of subaerially deposited, basaltic pahoehoe lava flows and a few intercalated beds of sandstone and conglomerate (Laughton et al. 2002, 2005). Like most of the Bonnet Plume River Intrusions, the Slab volcanics are preserved only as megaclasts within zones of Wernecke Breccia. Their origin on the Earth's surface makes them particularly interesting, however, and tectonically important. The main occurrence on Slab Mountain is a spectacularly exposed block (160 m × 380 m) consisting of about 34 steeply dipping lava flows. The block is contained within a large zone of Wernecke Breccia that is hosted by schist and deformed siltstone of the Fairchild Lake Group.

Where were the volcanics originally deposited? Three general possibilities can be entertained. One is that they were deposited as part of the Wernecke Supergroup. This option is rejected, however, because the Wernecke Supergroup was deposited in a submarine environment, and nowhere have lavas (either submarine or subaerial) been identified, despite excellent exposure throughout the study area. A second possibility is that they were deposited beneath the Wernecke Supergroup. This option could be viable only if the megaclasts were brought upwards through the crust to their present location within the Fairchild Lake Group during breccia emplacement, a scenario that seems highly improbable given the huge size of the main block. It is also considered unlikely because the volcanics are apparently less metamorphosed and deformed than the surrounding multiply deformed Fairchild Lake Group. The third and most probable option is that the volcanics were originally deposited above the Wernecke Supergroup and were brought into their present location by foundering into a zone of Wernecke Breccia (Laughton et al. 2005).

In the preferred model for the origin of Slab volcanics, in which eruption occurred subaerially above the Wernecke Supergroup, volcanism must have post-dated Racklan orogenesis. Specifically, at least 8 km of the Wernecke Supergroup must have been eroded from structural culminations (such as the Bonnet Plume River anticline) prior to volcanism, as argued by Laughton et al. (2005). This timing is necessary because of the immense thickness of strata that would have otherwise overlain the Fairchild Lake Group (the overlying Quartet and Gillespie Lake groups have a combined minimum thickness of approximately 8 km). It is unreasonable to propose that the volcanic megaclasts moved downward for >8 km into the crust, i.e., all the way through the Gillespie Lake and Quartet groups and into the Fairchild Lake Group. Consequently, the most appropriate model for the origin of the volcanics involves (*i*) Racklan orogenesis, resulting in high structural relief, (*ii*) denudation to expose the Fairchild Lake Group in eroded cores of anticlines, (*iii*) eruption of the Slab volcanics, and (*iv*) brecciation and foundering of megaclasts into the upper portions of Wernecke Breccia zones.

The age of the Slab volcanics lies between the maximum possible age of the Racklan orogeny (unconstrained, but probably <1.84 Ga) and Wernecke Breccia (1.60 Ga). Attempts to date the volcanics directly have so far yielded only one isotopic age, a ca. 1.38 Ga U–Pb date on rutile (Thorkelson et al. 2001*b*). This age is regarded as a product of secondary fluid flow through the enclosing breccia zone caused by heat liberated during emplacement of the Hart River sills (described later in the paper) (Thorkelson et al. 2001*b*). The volcanics may be the extrusive equivalents of the Bonnet Plume River Intrusions, but this possibility has not yet been addressed using geochemical and isotopic methods. An alternative is that the Slab volcanics are approximately the same age as Wernecke Breccia and share a common origin. A hint of igneous activity the age of Wernecke Breccia was discovered in the Coates Lake diatreme of the Mackenzie Mountains (Fig. 1). Granitic clasts in the diatreme crystallized at ca. 1130 Ma and contain a component of inherited zircon dated at ~1.6 Ga (Jefferson and Parrish 1989; Mortensen and Colpron 1998).

Brecciation at the Early–Middle Proterozoic boundary

Massive hydrothermal fluids invaded many parts of the study area at the boundary between Early and Middle Proterozoic time (Figs. 2, 4). The fluids expanded rapidly in the upper crust and generated voluminous breccia zones in the country rock (Laznicka and Edwards 1979; Bell 1986*b*; Lane 1990). These breccia zones, known collectively as Wernecke Breccia (formalized in Thorkelson et al. 2001*b*), commonly occur in curvilinear arrays along or parallel to major folds and faults (Fig. 3*a*). Individual breccias range from a few metres across to hundreds of metres wide and 5 km long. As noted earlier in the paper, the main breccia event has been dated by U–Pb titanite at 1.60 Ga and has been demonstrated through field relations to post-date the Racklan orogeny (Thorkelson et al. 2001*b*). The rocks that were brecciated include all units of the Wernecke Supergroup, the Bonnet Plume River Intrusions, and, locally, the Slab volcanics (Hunt et al. 2002). The source of the fluids is uncertain but may have been thermally coupled to possible igneous intrusions at depth (Hitzman et al. 1992) which conceivably could be related to the Slab volcanics. Local mineralization of mainly Cu and Au, alongside hematite and (or) magnetite, has been grouped with other Proterozoic iron-oxide copper–gold occurrences by Hitzman et al. (1992). Venting of brecciated rock and fluid (cf. Browne and Lawless 2001) is considered likely, particularly at Slab Mountain where momentary open space in the breccia zone seems required by the downward-travelled megaclasts of the Slab volcanics. In some breccia zones, faulting appears to have been active concurrently with brecciation. Breccia fragments are cemented together by hematite, quartz, carbonate, chlorite, feldspar, mica, and other minerals. In most cases, clasts and wall rocks were hydrothermally altered, leading to metasomatic growth of secondary minerals including flecks of hematite, octahedra of magnetite, and rhombs of dolomite. In many of the breccias, widely disseminated earthy hematite and local potassic alteration in the breccia clasts resulted in colour changes from original drab hues of gray and brown to striking pink and red. In some breccias, especially those with sodic rather than potassic compositions, clast and matrix colours are generally grey (Laughton et al. 2005).

Although the main phase of brecciation and metasomatism occurred at ca. 1.60 Ga, a subsequent minor hydrothermal event, seemingly related to the Hart River sills, occurred at ca. 1.38 Ga (described earlier in the text). A third pulse of hydrothermal activity at ca. 1.27 Ga, possibly related to the coeval Bear River dykes (described later in the text), led to monazite growth in at least one breccia zone within the Richardson fault array (Fig. 1) (Parrish and Bell 1987; Schwab et al. 2004). Sm–Nd-isotope analysis of breccia samples from the area shown in Fig. 3*a* reveals a wide range of $\varepsilon_{\mathrm{Nd}}(T)$ values, from –2.3 to –9.0 (Table 1; Fig. 5). This range overlaps with, but is slightly larger than, that of the Wernecke Supergroup (at 1595 Ma). These data suggest that the breccia-forming fluids may have been largely in equilibrium with and (or) derived from the Wernecke Supergroup. The slightly more negative part of the breccia range could represent an influence from underlying and possibly more evolved basement, whereas the slightly more positive component could represent a contribution from the Bonnet Plume River Intrusions (Thorkelson et al. 2001*a*) or other less evolved material.

Mineralized breccias at and near the Olympic Dam deposit in South Australia have mineral assemblages and textures very similar to those of the Wernecke breccias. These similarities suggest that breccias in both regions belong to a common hydrothermal province and support models that link Laurentia to Australia in Proterozoic time (Bell and Jefferson 1987; Moores 1991; Thorkelson et al. 2001*b*). We speculate that the large, positive aeromagnetic anomalies beneath northwestern Canada (Fig. 1) may be underplated basalt of Wernecke Breccia age and may coincide with a regional transfer of heat from asthenosphere to lithosphere during mantle plume activity. In this scenario, the heating was sufficient to liberate water from the lower crust by prograde metamorphic reactions and, ultimately, to massive near-surface hydrothermal activity and brecciation. Such a basaltic underplate would be coeval with the Hiltaba–Gawler Range igneous suite in South Australia which is intimately related to breccia formation and mineralization at the Olympic Dam mine (Creaser and Cooper 1993).

Middle Proterozoic sedimentation and magmatism

Erosion and weathering prior to volcanism and sedimentation

The rock record from ca. 1.60 to 1.38 Ga, although scanty, suggests that northwestern Laurentia was largely emergent and tectonically quiescent between Wernecke brecciation and Hart River magmatism – Pinguicula sedimentation (Fig. 2*b*). Weathering, erosion, supergene enrichment, and regolith formation are the only geological processes that can be identified with certainty in the 220 Ma following Wernecke brecciation (Figs. 2*b*, 4). Erosion is evident from the lack of preservation of the Slab volcanics as part of the stratigraphy. Although the volcanics were present at the time of Wernecke Breccia formation, they are nowhere preserved beneath the sub-Pinguicula unconformity (except for down-dropped megaclasts) and were removed entirely prior to deposition of the Pinguicula Group. A weathered horizon and a locally silicic regolith are evident in the upper few metres of older rock (Wernecke Breccia and folded Wernecke Supergroup) beneath the Pinguicula and Lower Fifteenmile groups in both the Wernecke and Coal Creek inliers (Fig. 2). Malachite staining at the base of the weathered zone indicates (weak) supergene oxide mineralization (Thorkelson et al. 2001*b*).

Apparently, no major sedimentary or igneous events took place in the Yukon during this interval, and the only known structural feature that may have been generated is a single brittle shear along the margin of a zone of Wernecke Breccia (“block faulting of unknown extent” in Fig. 2*b*). Even this shear, however, may have developed late in the Wernecke Breccia event. The apparent stability of northwestern Laurentia from 1.60 to 1.38 Ga contrasts markedly with the degree of igneous and tectonic activity recorded in southwestern Canada and the adjacent United States. The Priest River complex hosted granitic plutonism at ca. 1.58 Ga (Doughty et al. 1998), and the Belt–Purcell basin and Moyie sill complex developed largely at ca. 1.47 Ga (Anderson and Davis 1995; Doughty and Chamberlain 1996).

Renewed magmatism and sedimentation at ca. 1.38 Ga

Marine sedimentation (Pinguicula and Lower Fifteenmile groups), marine volcanism (Hart River volcanics), and mafic sill emplacement (Hart River sills) occurred in a new round of extensional tectonism in the Yukon (Figs. 2, 4). The Hart River sills (1.38 Ga) and volcanics are geochemically similar and were correlated by Abbott (1997). All three events (sedimentation, volcanism, and plutonism) appear to have occurred concurrently, as explained later in the paper. In the Wernecke inlier, the 3.5 km thick Pinguicula Group was deposited in a southward-deepening basin with angular unconformity on folded, cleaved, and brecciated rocks of the Wernecke Supergroup.

The Pinguicula Group (Fig. 2), as originally defined by Eisbacher (1981), consisted of clastic and carbonate formations named units A–F. The Pinguicula Group was recently split into two successions by Thorkelson (2000), however. The lower part of the group, comprising units A–C, retained the Pinguicula Group name, whereas the upper part (units D–F) was named the Hematite Creek Group (which has since been correlated with the Mackenzie Mountains Supergroup by Thorkelson et al. 2003*b*). Unit A consists of a basal sandstone that grades up into maroon and green mudstone (detrital zircon geochronology provided later in the paper). Unit B gradationally overlies unit A and consists mainly of orange-weathering micritic dolostone. Unit C overlies unit B and comprises grey micritic carbonate with rare intercalations of black mudstone. The carbonate is commonly overprinted by narrow (1–15 mm wide) en-echelon bands of white sparry dolostone, known as zebra texture. Dividing the original Pinguicula Group was considered necessary because unit A of the Pinguicula is ca. 1.38 Ga, whereas unit E yielded detrital minerals as young as 1.00 Ga (Fig. 4) (Rainbird et al. 1997; Thorkelson 2000). Separating the successions along the unit C–D contact was reasonable because units C and D are separated by an angular unconformity in the Hart River inlier (Abbott 1997). Units R1–R4 of the Lower Fifteenmile Group (Thompson 1995) and the lower part of the Lower Tindir Group (Young 1982) are regarded as correlatives of the revised Pinguicula Group (Abbott 1997).

In the Wernecke inlier, unit A hosts several Hart River sills (and dykes), indicating that the sills are younger than the Pinguicula Group at that locality (geochronology provided later in the paper). In the Hart River inlier, however, the volcanics unconformably overlie the Wernecke Supergroup but underlie (and are therefore older than) the Pinguicula Group. These conflicting age relations may be reconciled if basin formation and magmatism were broadly concurrent but locally diachronous. This idea is supported by stratigraphic relations in the Coal Creek inlier (Fig. 2*a*), where different members of the correlative Fifteenmile Group rest directly on the Wernecke Supergroup (Thompson 1995). Apparently, differential movements of crustal blocks during extensional basin formation led to selective deposition and preservation of Fifteenmile strata. Taken together, relations across the study area record a complex pattern of block faulting, basin development, and localized magmatic activity.

Geochronology of 1.38 Ga rock units

Hart River sills

The ca. 1.38 Ga age of the Hart River igneous event (Fig. 4) was initially established on three sills by M.L. Bevier as recorded by Abbott (1997). These sills, however, were emplaced into the Gillespie Lake Group (Wernecke Supergroup) and did not provide a direct constraint on the overlying Pinguicula Group. In this section, we provide an age determination for a Hart River sill that intrudes unit A of the Pinguicula Group. Sills within the Pinguicula succession were originally identified by Blusson (1974), but their age and importance were not fully understood.

Igneous zircon from the sill (sample TOA-96-6-4-2B) was dated at the Pacific Centre for Isotopic and Geochemical Research at the University of British Columbia. The methodology for sample selection, dissolution, geochemical preparation, and mass spectrometry is described by Mortensen et al. (1995). Procedural blanks were 2–5 pg for Pb and 1 pg for U.[4] The results are shown in a conventional U–Pb concordia plot and a U–Pb isochron diagram in Fig. 6. Errors are given at the 2σ level.

Abundant zircon was recovered from a ~25 kg sample of diorite containing pegmatitic segregations. The zircon comprised clear, pale pink, stubby to elongate square prismatic grains with simple terminations. Four strongly abraded fractions of the coarsest, least magnetic zircon were analyzed (Table 2). The analyses range from 2.0% to 5.1% discordant. Three of the analyses (A, B, and C) define a linear array (Fig. 6) with calculated upper and lower intercept ages (based on a Davis-type regression; probability of fit = 27%) of $1381.9^{+5.3}_{-3.7}$ Ma and 267 Ma, respectively. The upper intercept age is interpreted to give the crystallization age of the rock, and the lower intercept indicates mainly recent Pb loss. Fraction D falls slightly below the calculated regression line and yields a ^{207}Pb/^{206}Pb age of 1387.1 Ma. This fraction is interpreted to have contained a small fraction of older inherited zircon.

Pinguicula Group

Detrital zircons from the basal sandstone of unit A of the Pinguicula Group (sample DT-93-153-1B) were dated at the Geological Survey of Canada laboratories in Ottawa. The sample was a greenish grey, medium-grained sandstone from near the base of unit A. Heavy mineral concentrates were prepared by standard crushing, grinding, Wilfley table, and heavy liquid techniques. Mineral separates were sorted by magnetic susceptibility using a Frantz™ isodynamic separator. Detrital zircon grains were chosen for optical clarity and to represent the range of morphological types in the sample. The single-grain zircon fractions analyzed were very strongly air abraded following the method of Krogh (1982). U–Pb

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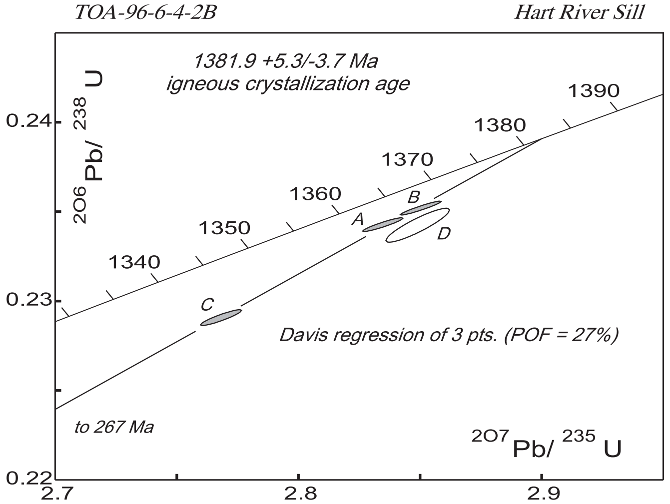
[4] Complete set of U–Pb tabular data and discussion of the error analysis may be purchased from the Depository of Unpublished Data, Document Delivery, CISTI, National Research Council of Canada, Ottawa, ON K1A 0S2, Canada. These data will also be available along side the article on the NRC Research Press Web site.

Fig. 6. U–Pb concordia plot of sample TOA-96-6-4-2B from a Hart River sill that crosscuts unit A of the Pinguicula Group. Data in Table 2. POF, probability of fit.



thermal ionization mass spectrometry (TIMS) analytical methods utilized are outlined in Parrish et al. (1987), and treatment of analytical errors follows Roddick (1987). U–Pb analytical results are presented in Table 3, where errors on the ages are reported at the 2σ level, and displayed in the concordia plot (Fig. 7). U–Pb sample location is displayed in Fig. 7, and Universal Transverse Mercator (UTM) coordinates are provided in Table 3.

The Pinguicula sample contained abundant zircon grains that ranged from subrounded to rounded and pink to colourless. Some of the rounded grains were frosted and had surficial pitting. All of the data are <4% discordant, and most are <1% discordant (Table 3). The analyses plot in two main clusters (Figs. 4, 7) including Early Proterozoic (1984–1841 Ma) and late Archean to earliest Early Proterozoic (2719–2497 Ma) ages, with a single zircon of middle Archean age (3078 Ma). This distribution is broadly similar to that of the Muskwa assemblage (Ross et al. 2001).

The zircons were probably recycled, at least in part, from the underlying Wernecke Supergroup. This provenance is likely because conglomerate that interdigitates with the sampled Pinguicula sandstone locally hosts abundant clasts of Wernecke Breccia, indicating a local sediment source. The possibility also exists, however, that some of the zircons were derived directly from Laurentian crystalline rocks, including plutons associated with the Wopmay orogen and the Slave craton. Future geochronology on detrital minerals of the Wernecke Supergroup will help to address the issue of local versus distal sources for the Pinguicula Group.

Dyking and hydrothermal activity at 1.27 Ga

The Bear River dykes were emplaced into the Wernecke Supergroup at ca. 1270 Ma (Figs. 2, 4) (Thorkelson 2000; Schwab et al. 2004). The age, major and trace element geochemistry, and Nd isotopic compositions of these dykes compare favourably with those of the Mackenzie dyke swarm and suggest that the Bear River dykes are a western subset of the Mackenzie swarm (Schwab et al. 2004). The Mackenzie dykes form the Earth's largest giant radiating dyke swarm, which transects much of the Canadian Shield as a consequence

Table 2. U–Pb analytical data for zircons from sample TOA-96-6-4-2B (dyke 1; UTM coordinates 7160000 m N, 573900 m E), a Hart River sill crosscutting Pinguicula Group unit A.

Sample description[a]	Wt. (mg)	U content (ppm)	Pb[b] content (ppm)	$\frac{^{206}Pb}{^{204}Pb}$ (meas.)[c]	Total common Pb (pg)	^{208}Pb (%)[b]	$\frac{^{206}Pb}{^{238}U}$ (±% 1σ)[d]	$\frac{^{207}Pb}{^{235}U}$ (±% 1σ)[d]	$\frac{^{207}Pb}{^{206}Pb}$ (±% 1σ)[d]	$\frac{^{206}Pb}{^{238}U}$ age (Ma; ±% 2σ)[e]	$\frac{^{207}Pb}{^{206}Pb}$ age (Ma; ±% 2σ)[e]
A: N5, +134	0.041	509	157	31 260	10	28.8	0.23428 (0.08)	2.8347 (0.15)	0.08776 (0.07)	1356.9 (2.0)	1377.1 (2.9)
B: N5, +134	0.043	615	178	42 330	9	23.7	0.23520 (0.08)	2.8503 (0.15)	0.08789 (0.07)	1361.7 (2.0)	1380.1 (2.8)
C: N5, +134	0.032	398	114	85 270	2	24.8	0.22908 (0.10)	2.7682 (0.15)	0.08764 (0.08)	1329.7 (2.3)	1374.6 (2.9)
D: N5, +134	0.036	350	103	27 420	7	25.4	0.23424 (0.20)	2.8491 (0.23)	0.08821 (0.10)	1356.7 (5.0)	1387.1 (3.9)

[a] N2 and M2 denote nonmagnetic and magnetic at indicated number of degrees side slope on Frantz isodynamic magnetic separator; grain size is given in micrometres (µm).
[b] Radiogenic Pb, corrected for blank, initial common Pb (from Stacey and Kramer 1975), and spike.
[c] Corrected for spike and fractionation.
[d] Corrected for blank Pb and U and common Pb.
[e] Decay constants from Steiger and Jäger 1977; errors assigned using the numerical error propagation method of Roddick (1987).

Table 3. U–Pb TIMS analytical data for sample DT-93-153-1B from the Basal Pinguicula Group (Z3386; UTM zone 8, 7191000 m N, 571400 m E).

								Isotopic ratios[f]			Ages (Ma)[g]			
Fraction[a]	Description[b]	Wt. (µg)	U (ppm)	Pb (ppm)[c]	$\frac{^{206}\mathrm{Pb}}{^{205}\mathrm{Pb}}$[d]	Pb (pg)[e]	$\frac{^{208}\mathrm{Pb}}{^{206}\mathrm{Pb}}$	$\frac{^{206}\mathrm{Pb}}{^{238}\mathrm{U}}$	$\frac{^{207}\mathrm{Pb}}{^{235}\mathrm{U}}$	$\frac{^{207}\mathrm{Pb}}{^{206}\mathrm{Pb}}$	$\frac{^{206}\mathrm{Pb}}{^{238}\mathrm{U}}$	$\frac{^{207}\mathrm{Pb}}{^{225}\mathrm{U}}$	$\frac{^{207}\mathrm{Pb}}{^{206}\mathrm{Pb}}$	% Disc.[h]
B (90)	Co, Clr, fdIn, Ro, NM5°	3	152	55	2050	5	0.10	0.3440 (0.12)	5.568 (0.22)	0.11737 (0.17)	1906.1 (4.1)	1911.1 (3.7)	1917 (6)	0.6
C (90)	Co, Clr, fIn, NM5°	3	133	50	1811	0	0.09	0.3586 (0.11)	6.026 (0.14)	0.12187 (0.09)	1975.5 (3.9)	1979.6 (2.5)	1984 (3)	0.5
D (100)	Co, Clr, fdIn, Ro, NM5°	4	58	30	781	8	0.15	0.4579 (0.16)	10.351 (0.18)	0.16394 (0.12)	2430.3 (6.3)	2466.6 (3.3)	2497 (4)	3.2
E (90)	Co, Clr, NM5°	3	182	68	1697	8	0.12	0.3501 (0.15)	5.835 (0.16)	0.12090 (0.07)	1934.9 (4.9)	1951.7 (2.8)	1970 (3)	2.0
F (80)	Co, Clr, SRo, NM5°	2	233	84	1223	7	0.15	0.3300 (0.13)	5.121 (0.17)	0.11257 (0.10)	1838.2 (4.3)	1839.6 (2.8)	1841 (4)	0.2
G (120)	pPk, Clr, Ro, NM5°	8	14	8	613	5	0.19	0.5256 (0.25)	13.580 (0.26)	0.18738 (0.12)	2723.0 (10.9)	2720.9 (4.9)	2719 (4)	–0.2
H (90)	pPk, Clr, Sro, NM5°	3	142	108	3831	4	0.20	0.6152 (0.32)	19.824 (0.11)	0.23372 (0.31)	3090.7 (15.7)	3082.8 (2.1)	3078 (10)	–0.5
I (90)	pPk, Clr, fdIn, Ro, NM5°	3	241	87	778	19	0.15	0.3310 (0.19)	5.153 (0.74)	0.11291 (0.66)	1843.3 (6.1)	1844.9 (13.0)	1847 (24)	0.2
J (80)	Co, Clr, Ro, NM5°	2	71	46	435	11	0.28	0.5152 (0.17)	12.876 (0.23)	0.18127 (0.15)	2678.7 (7.5)	2670.6 (4.4)	2665 (5)	–0.7
K (100)	Pk, Clr, Ro, NM5°	4	140	74	1447	12	0.05	0.4948 (0.11)	12.466 (0.17)	0.18272 (0.12)	2591.5 (4.5)	2640.2 (3.3)	2678 (4)	3.9

[a] All zircon fractions are single grains strongly abraded following the method of Krogh (1982). Numbers in parentheses refer to the size of the zircon in micrometres (µm).

[b] Clr, clear; Co, colourless; fdIn, fluid inclusions; fIn, few inclusions; NM5°, nonmagnetic at 1.8 Å, 5° SS; Pk, pink; pPk, pale pink; Ro, rounded; Sro, subrounded.

[c] Radiogenic Pb.

[d] Measured ratio, corrected for spike and fractionation.

[e] Total common Pb in analysis corrected for fractionation and spike.

[f] Corrected for blank Pb and U and common Pb; errors in parentheses are quoted as 1σ in percent.

[g] Corrected for blank and common Pb; errors in parentheses are quoted as 2σ in Ma.

[h] Discordance.

of a mantle plume centred beneath Victoria Island (Ernst et al. 1995). The Bear River dykes strike mainly northwest, in contrast to a northeast strike expected from radiating swarm geometry. The difference in orientation may be a result of extensive post-Mackenzie deformation in the northern Cordillera, as summarized by Norris (1997) and Thorkelson (2000).

The former presence of a Mackenzie-age basaltic volcanic succession in the Yukon seems likely as a volcanic expression of the Bear River dykes. Such a succession would correlate with the Coppermine River Group volcanics near Great Bear Lake (Fig. 1) (Baragar 1969) and the subsurface Tweed Lake basalts north of the Mackenzie Mountains (Fig. 1) (Sevigny et al. 1991). However, such volcanics are nowhere exposed in the Yukon and, if they were erupted, were apparently eroded prior to deposition of the Mackenzie Mountains Supergroup.

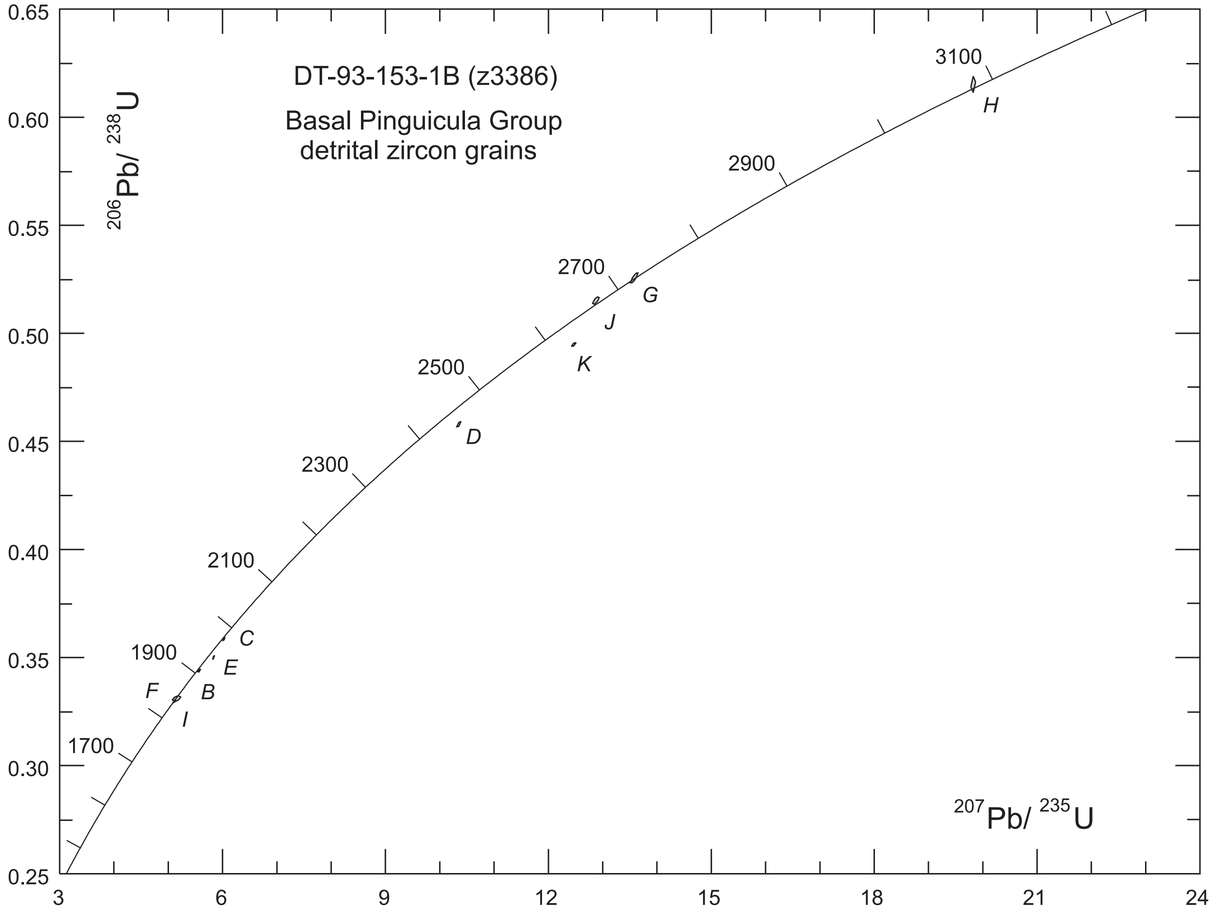
The Mackenzie (Bear River) igneous event (Fig. 4) appears to have generated secondary hydrothermal activity in a zone of Wernecke Breccia (Nor mineral occurrence) in the Richardson fault array, north of the Wernecke inlier (Fig. 1). The hydrothermal activity is evident from monazite from the breccia dated at ca. 1270 Ma (Parrish and Bell 1987). The age of this activity may be shared by U–Th–K–Cu metasomatism (undated) evident in a Bear River dyke in the Wernecke inlier (Schwab and Thorkelson 2001). If the Nor breccia reflects Bear River dykes at depth, and these dykes are included in the Mackenzie swarm, then the size of the swarm is ~50% larger than previously demonstrated.

Mackenzie Mountains Supergroup

Deposition of the Mackenzie Mountains Supergroup (Aitken and McMechan 1992) and correlative strata (Fig. 2) occurred throughout much of northwestern Laurentia in the late Middle Proterozoic and (or) early Late Proterozoic (Rainbird et al. 1997). The Mackenzie Mountains Supergroup in the Northwest Territories and the northern Wernecke inlier of the Yukon were described by Aitken et al. (1982) and Norris (1982) and revised by Jefferson and Parrish (1989). The succession consists of the following four main stratigraphic units, from bottom to top: unit H1, the Tsezotene Formation, the Katherine Group, and the Little Dal Group. Unit H1 is a poorly exposed carbonate with an uncertain connection to the other units. The overlying Tsezotene and Katherine groups represent a coarsening-upward succession of mainly shallow marine and fluvial(?) siltstone, with dolostone interbeds. The Little Dal Group reflects a return to quiescent, locally evaporitic basinal conditions, consists largely of grey-weathering carbonate with a member of gypsum, and is capped by the Little Dal basalts. These basalts have since been correlated with the 779 Ma Tsezotene sills (Dudás and Lustwerk 1997) and tentatively reassigned to the overlying Coates Lake – Windermere succession by Thorkelson (2000). Perhaps some of the underlying Little Dal Group, which contains unconformities of at least local significance, is also of Windermere age and affinity. As emphasized by Jefferson and Parrish, a significant hiatus within the Little Dal Group, as suggested in the paleomagnetic study of Morris and Park (1981), remains possible.

Recent stratigraphic correlations in and near the study area are favourable to a more complete understanding of the

Fig. 7. U–Pb concordia plot of nine detrital zircon grains from the basal sandstone of the Pinguicula Group. Data in Table 3.



Mackenzie Mountains Supergroup. In the central part of the Wernecke inlier, strata originally defined as the upper part of the Pinguicula Group were separated from it (as described previously) and renamed the Hematite Creek Group by Thorkelson (2000). Subsequent inspection of the Mackenzie Mountains Supergroup sensu stricto during regional mapping by Thorkelson et al. (2002) led to a more certain correlation between this succession and the Mackenzie Mountains Supergroup, confirming suggestions by Rainbird et al. (1997) and previous workers. As such, the term Hematite Creek Group is now abandoned and the strata are herein assigned to the Mackenzie Mountains Supergroup. Specifically, unit D is clearly correlative, at least in part, with the Tsezotene Formation, and unit E is equivalent to the Katherine Group. The Tsezotene and Katherine units may be regarded as different facies of the same succession, as implied by Aitken et al. (1982) and argued by Thorkelson (2000) for Pinguicula units D and E. Working farther west, Abbott (1997) identified strata that he correlated with the "upper" Pinguicula Group in the Hart River inlier, and those strata are herein regarded as part of the Mackenzie Mountains Supergroup. Abbott also correlated unit R5 of the Lower Fifteenmile Group in the Coal Creek inlier (Fig. 2) (Thompson and Roots 1982; Thompson 1995) and the upper part of the Lower Tindir Group in the Tatonduk inlier (Fig. 1) (Young 1982) with this succession.

These correlations are exciting because, in several localities in the Yukon, the stratigraphic base of the Mackenzie Mountains Supergroup (and correlatives) is well exposed, unlike sections in the Northwest Territories where no base has been identified. Consequently, a more complete, composite section for the supergroup can be constructed once correlations have been firmly established. Unit H1, for example, may correlate with a grey- to orange-weathering dolostone succession that is prominently exposed near the base of the succession in the central Wernecke inlier (Eisbacher 1981; Thorkelson 2000).

Influx of "Grenville" detritus

One of the most interesting discoveries regarding Proterozoic evolution of Laurentia is the discovery of Grenville-aged detrital zircons (ca. 1.00–1.25 Ga) in the Mackenzie Mountains Supergroup and most other "sequence B" strata (Fig. 4) (Rainbird et al. 1992, 1996). As demonstrated by Rainbird et al. (1997), these strata appear to have been deposited by continent-wide braided river systems originating from the Grenville orogen. The zircon dates also provide a maximum age for the Mackenzie Mountains Supergroup, i.e., <1.0 Ga. Whether the age of the Mackenzie succession is 1.0–0.9 Ga (Middle Proterozoic), as suggested in this paper, or <900 Ma (upper Proterozoic), as described by Rainbird et al. (1997), remains an open question.

In this report, we provide additional support for this hypothesis using ages of detrital muscovite grains from the Mackenzie Mountains Supergroup in the central Wernecke inlier. The sampled strata, which are described in more detail in Thorkelson (2000), belong to what was originally called unit D of the Pinguicula Group by Eisbacher (1981) and the Hematite Creek Group of Thorkelson. In this paper, we consider the sampled strata to belong to the Tsezotene Formation of the Mackenzie Mountains Supergroup.

Six grains of white mica separated from a sample of dark, fine-grained micaceous wacke were analyzed at the Geochronology Laboratory of the Geological Survey of Canada, Ottawa, Ontario. Analytical details are provided in Table 4, and the age spectra and interpreted ages are shown in Fig. 8 and plotted in Fig. 4. The methodology is given in Appendix A. Five of the six grains yielded ages ranging from 1037 ± 10 to 1113 ± 11 Ma. The sixth grain is much older, with an age of 2473 ± 27 Ma. The five younger dates fall into the Grenville age range of the zircons dated by Rainbird et al. (1997) and support the idea of a source region on the opposite side of the continent, although the possibility of a cryptic intra-Cordilleran source (cf. Ross and Villeneuve 1997) remains plausible.

Upper Fifteenmile Group – Callison Lake dolostone

In the Coal Creek and Hart River inliers (Fig. 2), nearly 1 km of dolostone with minor siltstone sits unconformably above the Mackenzie Mountains Supergroup and correlative rocks (Thompson and Roots 1982; Abbott 1997) and unconformably beneath the Mount Harper Group (Fig. 2*b*; Windermere Supergroup) (Mustard and Roots 1997). In the Coal Creek inlier, the strata are called the Upper Fifteenmile Group. In the Hart River inlier, where only the lower half of the succession occurs as a massive carbonate layer, it is called the Callison Lake dolostone (Abbott 1997). How these strata fit into the broader Proterozoic framework is uncertain. Presently we do not know if they are close in age and affinity to the Mackenzie Mountains Supergroup or the Windermere Supergroup, or if they form a temporally distinct stratigraphic package peculiar to the western part of the study area, as shown in Fig. 2*b*.

Middle to Late Proterozoic orogeny and exhumation

Corn Creek orogeny

Eisbacher (1978, 1981) recognized that the Mackenzie Mountains Supergroup (upper Pinguicula Group of Eisbacher 1981) (Fig. 2) was involved in contractional deformation prior to deposition of the Windermere Supergroup in the central part of the Wernecke inlier, southwest of Mt. Profeit (Fig. 3). Thorkelson (2000) identified several additional structures in that region and called the event Corn Creek orogeny. The structures comprise west-verging thrust faults and folds, some of which are overturned and clearly truncated by the basal Windermere conglomerate (Fig. 19 of Thorkelson 2000). Motion along one of the faults thrust Wernecke Supergroup strata over the Pinguicula Group (Fig. 3*b*). These structures occur near the dextral Snake River fault (Fig. 3*a*), which has a long-lived tectonic history (Norris 1982) and may have played a role in the origin of the Corn Creek orogeny (Eisbacher 1981). Whether the Corn Creek orogeny occurred before or after deposition of the Upper Fifteenmile Group and Callison Lake dolostone is uncertain. The close association of the Coates Lake Group with the Windermere Supergroup suggests to us that the Coates Lake succession was preceded by the Corn Creek orogeny.

The Corn Creek family of structures may extend into other parts of the Wernecke and northern Mackenzie mountains but may be difficult to identify in the absence of an exposed Mackenzie Mountains Supergroup – Windermere Supergroup contact. In the Northwest Territories, contractional deformation between deposition of the Mackenzie Mountains and Windermere supergroups has been recognized locally (with southeast vergence), but the regional sub-Windermere unconformity is more generally related to extensional tectonism and widespread bevelling of an uptilted land surface (reviewed by Jefferson and Parrish 1989). The cause of the regional sub-Windermere unconformity is most appropriately termed the "Hayhook extensional event" (Fig. 2*b*) (Jefferson and Parrish 1989) rather than "Hayhook orogeny" (Young et al. 1979). In the Wernecke inlier, east-trending normal faults of the Hayhook extensional event crosscut the older contractional structures of the Corn Creek orogeny and control thicknesses in the basal Windermere conglomerate. Eisbacher (1981) suggested that some of the Corn Creek contraction may have occurred in conjunction with basal Windermere sedimentation, although the normal syn-Windermere cross-faults identified by Thorkelson (2000) imply that the Corn Creek event occurred earlier.

Exhumation revealed by cooling

The Corn Creek orogeny may have caused significant uplift and exhumation of the Wernecke inlier. We entertain this possibility because of the similarity in timing between Corn Creek deformation and cooling ages of metamorphic white mica from the Wernecke Supergroup. Two white mica samples analyzed by the ^{40}Ar–^{39}Ar method yielded plateau ages of 788 ± 8 and 980 ± 4 Ma (Fig. 4; details later in this section; methodology in Appendix A). These dates lie between the minimum and maximum ages possible for the Mackenzie Mountains Supergroup, i.e., within the possible age range of the Corn Creek orogeny. These results are broadly supported by four preliminary ^{40}Ar–^{39}Ar ages ranging from 0.8 to 1.2 Ga obtained on muscovites that grew in veins associated with 1.60 Ga Wernecke Breccia (J.A. Hunt, personal communication, 2004).

The white micas were taken from schist of the Fairchild Lake Group in the Wernecke inlier near Slab Mountain (Fig. 3*a*) and grew synkinematically with the first identifiable phase of Racklan deformation (Thorkelson 2000; Brideau et al. 2002). The schist was formed and exhumed in the Early Proterozoic prior to Wernecke Breccia emplacement (1.60 Ga), as discussed previously. Both samples of white mica show a saddle or U-shape characteristic of excess argon at low ^{39}Ar release and an older age component at high temperature (Fig. 9). Our fundamental interpretation is that these spectra are dominated by one or two Middle to Late Proterozoic thermal events that almost completely reset the argon systematics. The older ages at high ^{39}Ar release record memory of an Early Proterozoic age. Sample DT-95-22-1 steps up to ~1.8 Ga, perhaps recording the time of primary mica growth at or before ~1.8 Ga. Sample DT-95-2B reaches a maximum age of ~1.4 Ga, reflecting either the original age of the mica or, more likely, partial argon loss from this part of the mineral.

At the time of Wernecke brecciation, the muscovites in the

Table 4. Ar–Ar data for detrital muscovites from the Tsezotene Formation of the Mackenzie Mountains Supergroup (sample DT-95-11-1B; J^f = 0.01944 ± 0.00019; Z4539; 64.8466°N, 133.1149°E).

Laser power[a]	Volume ^{39}Ar ($\times10^{-11}$ cm^3)	$\frac{^{36}\text{Ar}}{^{39}\text{Ar}}$	$\frac{^{37}\text{Ar}}{^{39}\text{Ar}}$	$\frac{^{38}\text{Ar}}{^{39}\text{Ar}}$	$\frac{^{40}\text{Ar}}{^{39}\text{Ar}}$	% Atm. ^{40}Ar[b]	$\frac{^{40}\text{Ar}^*}{^{39}\text{Ar}}$	f_{39} (%)[c]	Apparent age (Ma)[d]
Aliquot A									
2.0[e]	0.4032	0.0107±0.0032	0.141±0.018	0.002±0.003	146.266±0.910	2.2	143.101±1.112	44.8	2399.83±10.31
3.0	0.4497	0.0086±0.0019	0.132±0.008	0.003±0.002	154.748±2.066	1.6	152.213±2.046	50.0	2482.41±18.13
4.0	0.0464	0.0596±0.0242	1.280±0.174	0.012±0.007	158.553±19.378	11.1	140.947±19.742	5.2	2379.74±185.13
Aliquot B									
2.0	1.1433	0.0026±0.0010	0.032±0.012	0.004±0.001	40.750±0.269	1.9	39.992±0.308	13.0	1037.66±6.08
45.0[e]	0.1829	0.0288±0.0049	0.256±0.024	0.004±0.003	46.844±1.233	18.1	38.345±1.120	2.1	1004.87±22.51
Aliquot C									
45.0	1.1454	0.0020±0.0009	0.046±0.003	0.002±0.001	41.901±0.314	1.4	41.316±0.318	15.3	1063.59±6.18
Aliquot D									
45.0	1.2527	0.0034±0.0009	0.039±0.005	0.002±0.001	44.929±0.403	2.3	43.917±0.418	19.8	1113.50±7.91
Aliquot E									
45.0	0.9580	0.0079±0.0012	0.063±0.007	0.004±0.001	46.163±0.771	5.1	43.822±0.790	18.9	1111.68±14.97
Aliquot F									
45.0	3.2117	0.0008±0.0003	0.020±0.002	0.002±0.000	41.691±0.356	0.6	41.443±0.360	78.1	1066.07±7.00

Note: All uncertainties quoted at the 2σ level. *, radiogenic.

[a] As measured in percentage of full nominal power (10 W).

[b] Percent of total blank and interference corrected ^{40}Ar of non-radiogenic origin.

[c] Fraction of ^{39}Ar as percentage of total run.

[d] Errors are analytical only and do not reflect error in irradiation parameter *J*.

[e] Heating step not included in the final age calculation.

[f] Nominal *J*, referenced to FCT–SAN = 28.03 Ma (Renne et al. 1998).

Fig. 8. Release spectra for ^{40}Ar–^{39}Ar muscovite ages of six detrital white mica grains from the Tsezotene Formation of the Mackenzie Mountains Supergroup. Data in Table 4.

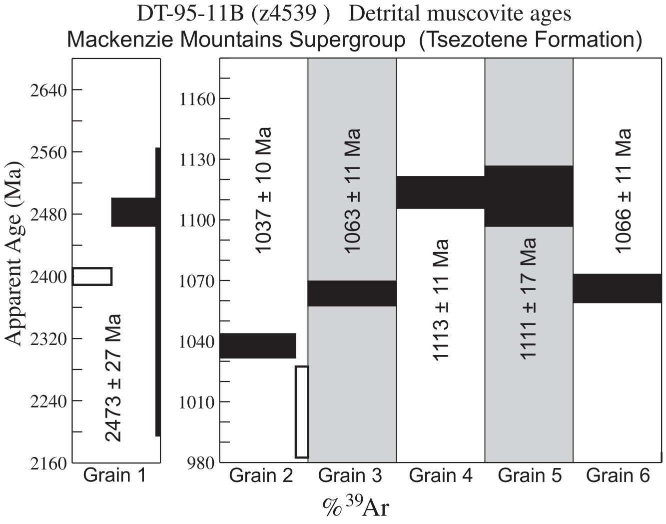
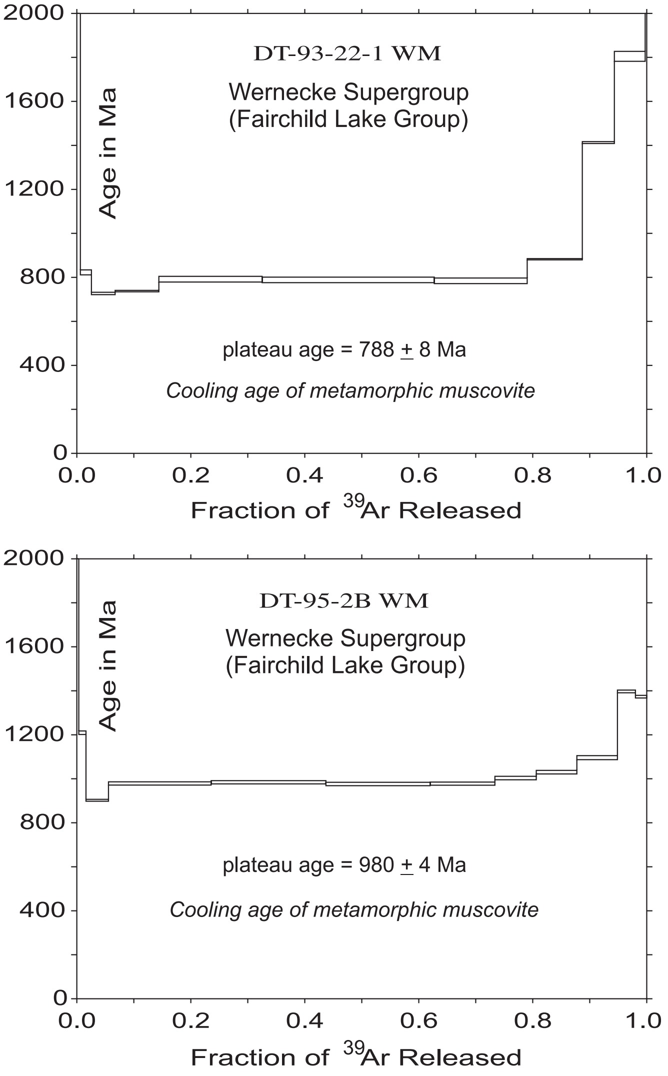


Fig. 9. Release spectra for ^{40}Ar–^{39}Ar ages of two multigrain white mica fractions from schist of the Fairchild Lake Group (Wernecke Supergroup) along the Bonnet Plume River anticline (Fig. 3). Data in Table 4.



sample area must have been close to the surface, probably at depths <1 km, and the system would have been closed to diffusion of Ar. If the schist had remained near the surface, the muscovite cooling ages would likely have yielded Early Proterozoic plateau ages. The Early Proterozoic memory (~1.8 Ga) of sample DT-95-22-1 may reflect the approximate age of original metamorphism and mica growth during the Racklan orogeny.

The resetting of the mica ages is interpreted as a result of reheating of the Fairchild Lake Group schists by reburial of the Wernecke Supergroup during deposition of the Pinguicula Group, Mackenzie Mountains Supergroup, and possible volcanism associated with Bear River dyking. As shown in Fig. 2*b*, these strata could have buried the underlying Wernecke Supergroup by 10 km. At a moderate geothermal gradient of 30 °C/km, the schist of the Fairchild Lake Group could have been reheated to 300 °C, close to that required to reset the isotopic system. Moreover, the geothermal gradient in the upper crust may have been higher than 30 °C/km through addition of heat from the 1.27 Ga Mackenzie (Bear River) igneous event and, over a longer interval of time, by radiogenic heating from potassically altered zones of Wernecke Breccia and country rock. The Middle to Late Proterozoic plateau ages reflect a second interval of exhumation and cooling of the Wernecke Supergroup, possibly related to the Corn Creek orogeny. The difference of >190 Ma between the plateau ages of the two samples remains unexplained but could reflect variable or incomplete argon loss during burial and reheating.

Conclusions

Chronology of events

Proterozoic rocks in the Yukon provide a valuable record of the supracrustal history of northwestern Laurentia. This record spans at least 1.2 billion years, more than a quarter of Earth history, and reveals a region that evolved toward a complex orogen in a series of diverse geological events. These events include crustal extension, mountain building, mafic magmatism, and hydrothermal brecciation. They are typically separated from one another by long intervals of time for which there is little or no geological information. Thus, despite advances in our understanding of individual events and how some of these events relate to one another, much of the geological history has no discernable record. During these lacunas, northwestern Laurentia is likely to have varied from an eroding highland to a low-relief continental platform undergoing little geological activity. In some of these intervals, however, strata may have been deposited and entirely eroded, leaving little or no trace of their existence.

The main Early Proterozoic to early Late Proterozoic geological events in the Yukon (and nearby regions) are provided in chronologic order in this section. This listing generally follows the time-stratigraphic column of Fig. 2 but is more strict in its treatment of age constraints and durations of events (see caption for Fig. 2). The description of events ends prior to Late Proterozoic extension and deposition of the Coates Lake Group and Windermere Supergroup. The significant figures of numerical ages given as follows respect the amounts of uncertainty in relevant isotopic age determinations:

(1) ~1.84 Ga — Suspected time of the last major tectonic and magmatic events of northwestern cratonal Laurentia based on the geology of the Wopmay orogen and the Fort Simpson domain.

(2) >1.71 Ga — Suspected attenuation of northwestern Laurentia, forming a passive margin or intracratonic basin. The adjoining landmass, if one existed, remains unidentified. Deposition of the Wernecke Supergroup as two clastic to carbonate grand cycles in two main phases of extension.

(3) >1.60 Ga — Possible age range for (*i*) the Racklan orogeny, and (*ii*) eruption of the Slab volcanics. The Racklan orogeny inverted the Wernecke basin and involved at least two phases of folding and three phases of cleavage and fabric development. The Slab volcanics post-date Racklan orogenesis but pre-date Wernecke Breccia.

(4) 1.71 Ga — Emplacement and cooling of the Bonnet Plume River Intrusions, probably in response to extension. Also a possible age of the Slab volcanics.

(5) 1.60 Ga — Surges of hydrothermal solutions to form Wernecke Breccia and related iron-oxide copper–gold occurrences (probable link to South Australia). Incorporation of Slab volanics and Bonnet Plume River Intrusions as megaclasts in breccia. Minor magmatism beneath the Mackenzie Mountains and (?) elsewhere occurred at approximately 1.6 Ga. Also a possible age of the Slab volcanics.

(6) 1.60–1.38 Ga — Block faulting of unknown but probably modest extent. Regional weathering and regolith formation.

(7) 1.38 Ga — Extension leading to (*i*) emplacement of the Hart River sills and eruption of the Hart River volcanics, (*ii*) hydrothermal alteration in previously established zones of Wernecke Breccia, and (*iii*) concurrent deposition of unit A of the Pinguicula Group. Probable inception of Lower Fifteenmile Group (units R1–R4) and lower Lower Tindir Group deposition.

(8) 1.38–1.27 Ga — Possible age range of units B and C of the Pinguicula Group, Lower Fifteenmile Group (units R1–R4), and lower Lower Tindir Group, although a depositional age close to 1.38 Ga for all units is considered likely.

(9) 1.27 Ga — Age of the Mackenzie igneous event throughout much of northern Laurentia. In the Yukon, this event comprises (*i*) emplacement of Bear River dykes, and (*ii*) hydrothermal alteration in previously established zones of Wernecke Breccia.

(10) 1.1 Ga — Crystallization of granite of unknown extent beneath the Mackenzie Mountains.

(11) 1.00–0.78 Ga — Age-range of (*i*) regional extension and deposition of strata as described in the following text, and (*ii*) the Corn Creek orogeny. The Katherine and Little Dal groups and the Callison Lake dolostone are constrained to this interval. The age range almost certainly applies to the underlying Tsezotene Formation, which is gradational over a large stratigraphic interval with the Katherine Group. By proposed correlations, this interval also applies to unit R5 of the Lower Fifteenmile Group. Unit H1, which is poorly exposed and has been provisionally included at the base of the Mackenzie Mountains Supergroup, may also belong to this interval but its affinity and maximum age are, to date, unconstrained. The Corn Creek orogeny may be of local rather than regional extent. It is younger than the Tsezotene and Katherine successions but may be older than the Callison Lake dolostone, Upper Fifteenmile Group, and part of the Little Dal Group, particularly the Little Dal basalts.

Geological environments through time

The paleogeography of northwestern Laurentia prior to the time of Wernecke basin formation is unknown. Western Laurentia may have been connected to another cratonal landmass in the Proterozoic such as Australia (Bell and Jefferson 1987), Antarctica (Moores 1991; Dalziel 1991), Siberia (Sears and Price 2000), or South China (Li et al. 1995), or nothing at all as inferred for the Yukon region by Delaney (1981). The timing and location of continental juxtaposition remain controversial (Karlstrom et al. 2001; Burrett and Berry 2000; Wingate et al. 2002). Australia is an attractive option for Early Proterozoic time because the Burramundi orogeny in Australia (Page and Williams 1988; Solomon and Groves 1994; Compston 1995; Page et al. 2000) and Wopmay orogeny and Fort Simpson magmatism in northwestern Laurentia (Hoffman 1989; Villeneuve et al. 1991; Bowring and Grotzinger 1992) share similar intervals spanning 1.84–1.97 Ga and may be orogenic counterparts of one another.

Visualizing the earliest supracrustal environment, the Wernecke basin, can be done in three general ways. (1) The basin may have developed as an interior basin within a cratonal landmass. The Wernecke basin is presently located on the pre-accretionary margin of Laurentia, so this option requires that Laurentia was previously connected to another continent. (2) The basin may have been part of a newly formed passive margin that developed in response to separation of Laurentia from another continent. In this case, the other continent could have drifted far away from Laurentia and need not be the same landmass that lay alongside Laurentia in Middle to Late Proterozoic time, as discussed later in the paper. (3) The basin may have been a long-lived depocentre (such as a passive margin) that lay along the western shore of Laurentia since, and perhaps before, the time of the Wopmay orogeny (Delaney 1981).

Prior to 1.71 Ga, the Wernecke Supergroup filled the basin in two grand cycles consisting of siliciclastic sediments capped by carbonates. The 4.6 km thickness of carbonate at the end of the second grand cycle (Gillespie Lake Group) indicates a protracted history of gradual but deep subsidence. Abrupt contrasts in sedimentary facies of this succession infer concurrent faulting. These features are more consistent with a recently formed passive margin or intracratonic basin (options 1 and 2) than a preexisting, deep-water, less tectonically active basin (option 3).

Uplift, metamorphism, polyphase deformation (Racklan orogeny), and partial exhumation of the Wernecke basin reshaped northwestern Laurentia. Folding was directed first toward the east and then to the south, and extended into the Northwest Territories where it is called the Forward orogeny. At least 8 km of structural relief developed and, concurrent to subsequent erosion, produced a denuded landscape with the more metamorphosed strata of the lowest unit, Fairchild Lake Group, exposed in the cores of structural culminations flanked by the Quartet Group and Gillespie Lake Group. The mafic Slab volcanics were locally deposited on the eroded succession, probably as subaerial shield volcanoes or plateau lavas. The Bonnet Plume River Intrusions may also have been emplaced at this time. Two scenarios for the Racklan orogeny were proposed by Thorkelson et al. (2001*b*). In one, the deformation resulted from far-field stresses generated in concurrent accretionary events along the southern margin of Laurentia (Yavapai and Mazatzal orogenies; Condie 1992). In the other, which is most suitable if the Wernecke basin were part of a passive margin, the Racklan represents an inboard part of a more nearby collision with a terrane or another continent, perhaps Australia (possibly related to the Isan or Kimban orogenies).

At 1.60 Ga, the upper crust was brecciated and metasomatically altered by hydrothermal solutions to form the Wernecke breccias. Following the general model of Browne and Lawless (2001), the land surface was probably breached by repeated, widespread hydrothermal eruptions, producing maar-like pits surrounded by mounds and fields of rocky debris ejected from the upper crust. Significant open space formed in vent areas as material was expelled. Megaclasts of rock from the uppermost crust (including the Slab volcanics) tumbled into the vents and became surrounded by a coarse matrix of other clasts and streaming, Fe-rich hydrothermal solutions. Cycles of hydrothermal precipitation, rebrecciation, and metasomatism occurred by repeated surges of fluids, and the zones were eventually cemented by crystallization of silicate, carbonate, and Fe-oxide minerals, with localized concentrations of Cu and Au in disseminations and veins. The near-identical age and petrology of the Wernecke breccias with similar breccias in South Australia strongly suggest that Australia lay alongside northwestern Laurentia at this time.

Northwestern Laurentia probably remained above sea level for much of the next 220 Ma, apparently in a state of crustal stability. In contrast, central to southern Laurentia experienced a major extensional event only 130 Ma after Wernecke brecciation to form the Belt–Purcell basin. This basin received detritus probably sourced from South Australia (Ross et al. 1992), a relation that may indicate a southward transit of Australia relative to Laurentia between 1.60 and 1.47 Ga, or a major southeastward-flowing river system at ca. 1.47 Ga. Minor block faulting occurred, followed by weathering and local supergene enrichment of Wernecke Breccia mineral occurrences.

Crustal stability was broken by a major extensional event at 1.38 Ga, leading to mafic volcanism, sill emplacement (Hart River magmatism), and concurrent deposition of the Pinguicula Group and related successions. The Pinguicula deposystem probably extended into the Northwest Territories to form the Dismal Lakes succession (Cook and MacLean 1995). Broadly correlative strata and mafic sills and granites (1.37 Ga) also occurred toward the end of Belt Supergroup sedimentation in Idaho (Doughty and Chamberlain 1996) and, together with the Pinguicula – Hart River event, may point to an interval of regional extension in western North America.

Pinguicula deposition was probably over by 1.27 Ga when much of Laurentia, including the northwestern corner, was intruded by mafic dykes generated by a mantle plume in the Canadian Arctic (Mackenzie igneous event; Bear River dykes; Schwab et al. 2004). Mafic volcanism probably accompanied the dyking. The cessation of Pinguicula deposition probably occurred by gentle crustal uplift possibly related to the igneous activity. The possibility that a continent separated from Laurentia during the Pinguicula event cannot be discounted, although the generally fine grained siliciclastic to carbonate composition of the preserved deposits implies that such rifting was most likely focused along an axis far away, probably to the north or west.

Except for the possibility of mafic volcanism at 1.27 Ga and minor plutonism at 1.1 Ga, northwestern Laurentia was apparently a low-lying, uneventful terrain for 300 Ma after Pinguicula deposition. Shortly after 1.00 Ga, a new round of extension occurred, and the Mackenzie Mountains deposystem formed in shallow marine and fluvial environments and received detritus derived largely from the Grenville orogen. Similar deposits were laid down throughout much of northern Laurentia and could be a manifestation of continental separation, with an axis of rifting located farther west. Carbonate strata of the Upper Fifteenmile Group and related strata were deposited somewhat later, probably in a gently subsiding basin related to reactivation of long-lived basement structures (Roots and Thompson 1988). The Corn Creek orogeny locally deformed some of these rocks and may have been associated with strike-slip motion along the nearby Snake River fault. The deformation may be regarded as a “late Grenville” event that led to termination of Mackenzie Mountains sedimentation prior to Hayhook extension and genesis of the Cordilleran-long, Late Proterozoic Windermere rift basins.

The Proterozoic rocks of the Yukon are presently exposed as inliers within the Cordilleran fold and thrust belt, and their configuration has been strongly influenced by Phanerozoic extensional, transcurrent, and especially contractional deformation, largely of Cretaceous to Eocene age (Norris 1972, 1997; Eisbacher 1981; Cecile et al. 1982; Cecile 1984; Thompson 1995; Abbott 1997; Lane 1998). The crust of northern Yukon continues to be seismically active and is involved in both contractional and transcurrent deformation related to modern tectonics along the west coast of British Columbia and Alaska (Mazzotti and Hyndman 2002).

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Appendix A: ^{40}Ar/^{39}Ar methods of white mica analysis

Geological Survey of Canada: detrital mica from the Mackenzie Mountains Supergroup

Detrital muscovite was separated from sample DT-95-11B and analyzed by laser ^{40}Ar/^{39}Ar step-heating and total fusion methods at the Geochronology Laboratory of the Geological Survey of Canada in Ottawa, Ontario. Individual mineral separates were loaded into aluminum foil packets along with a single grain of Fish Canyon Tuff Sanidine (FCT–SAN) to act as flux monitor (apparent age = 28.03 ± 0.28 Ma; Renne et al. 1998). The sample packets were arranged radially inside an aluminum can. The samples were then irradiated for 12 h at the research reactor of McMaster University in a fast neutron flux of ~3×10^{16} neutrons/cm^2.

Upon return from the reactor, individual grains were split into several aliquots and loaded into separate 1.5 mm diameter holes in a copper planchet. Heating of individual sample aliquots in steps of increasing temperature was achieved using a Merchantek MIR10 10W CO$_2$ laser equipped with a 2 mm × 2 mm flat-field lens. The released Ar gas was cleaned over getters for 10 min, and then analyzed isotopically using the secondary electron multiplier system of a VG3600 gas source mass spectrometer; details of data collection protocols can be found in Villeneuve and MacIntyre (1997) and Villeneuve et al. (2000). Error analysis on individual steps follows methods of Roddick (1988). Because of small gas amounts released, samples were either step heated in two to three steps or, for most grains, completely fused, making the results roughly equivalent to K–Ar analysis (Table 4).

Neutron flux gradients throughout the sample canister were evaluated by analyzing the sanidine flux monitors included with each sample packet and interpolating a linear fit against calculated J factor and sample position. The error on individual J factor values is conservatively estimated at ±1.0% All errors are quoted at the 2σ level of uncertainty.

Table A1. Summary of Ar–Ar ages on muscovites from schist of the Fairchild Lake Group.

	UTM coordinates						
Sample No.	Northing	Easting	Mineral	Integrated age (Ma)	Plateau age (Ma)	Plateau information	Comments
DT-93-22-1	7208150	552800	White mica	954±5	788±8	Three fractions; 65% ^{39}Ar release; MSWD = 0.1	Saddle shape; max. high-temp. age = 1805±23 Ma
DT-95-2B	7208350	546200	White mica	1039±4	980±4	Four fractions; 68% ^{39}Ar release; MSWD = 0.2	Saddle shape; max. high-temp. age = 1397±6 Ma

Note: Samples run against standard Mmhb-1 with an age of 513.9 Ma and processed using the constants of Steiger and Jäger (1977). All errors quoted to ±1σ. MSWD, mean square of weighted deviates.

University of Alaska: micas from schist of the Fairchild Lake Group

For ^{40}Ar/^{39}Ar analysis, samples DT-93-22-1 and DT-95-2B were crushed and the samples were washed in deionized water and sieved, and the 100–250 µm fraction of white mica was selected. The samples were wrapped in aluminum foil and arranged in one of two levels, labeled top and bottom, within aluminum cans of 2.5 cm diameter and 4.5 cm height. Samples of hornblende MMHb-1 with an age of 513.9 Ma were used to monitor the neutron flux. The samples were irradiated for 70 MW·h in position 5c of the uranium-enriched research reactor of McMaster University in Hamilton, Ontario, Canada.

Upon their return from the reactor, the samples (5–20 crystals) and monitors were loaded into 2 mm diameter holes in a copper tray that was then loaded in an ultra-high-vacuum extraction line. The monitors were fused, and the samples were step-heated using a 6 W argon-ion laser using the technique described in York et al. (1981) and Layer et al. (1987). Argon purification was achieved using a liquid nitrogen cold trap and a SAES Zr–Al getter at 400 °C. The samples were then analyzed in a VG3600 mass spectrometer at the Geophysical Institute, University of Alaska, Fairbanks, Alaska. The argon isotopes measured were corrected for system blank and mass discrimination, as well as calcium, potassium, and chlorine interference reactions following procedures outlined in McDougall and Harrison (1988). For each sample, a plateau age was determined from three or more consecutive fractions whose ages are within 2σ of each other and total more than 50% of gas release. A summary of all the ^{40}Ar/^{39}Ar results is given in Table A1, with all ages quoted to the ±1σ level and calculated using the constants of Steiger and Jäger (1977), and the analytical details are given in Table A2.

Table A2. Details of Ar–Ar age determinations for white micas from schist of the Fairchild Lake Group.

Laser power (mW)	Cum. ^{39}Ar	$\frac{^{40}Ar}{^{39}Ar}$ measured	$\frac{^{37}Ar}{^{39}Ar}$ measured	$\frac{^{36}Ar}{^{39}Ar}$ measured	% Atm. ^{40}Ar	Ca/K	Cl/K	$\frac{^{40}Ar*}{^{39}Ar_K}$ measured	Age (Ma)
DT-93-22-1 WM (weighted average of *J* from standards = 0.008270 ± 0.000024)									
100	0.001	821.180±123.311	0.2031±0.6576	0.1074±0.0605	3.86	0.373±1.207	0.06421±0.04069	789.524±119.825	3642.1±237.4
150	0.002	430.003±33.353	0.1942±0.1833	0.0630±0.0281	4.32	0.356±0.336	0.03084±0.01054	411.436±32.954	2674.0±111.7
200	0.003	440.647±39.908	0.4599±0.2119	0.0976±0.0223	6.54	0.844±0.389	0.03773±0.01112	411.943±37.816	2675.7±128.0
300	0.006	278.404±10.029	0.2926±0.0669	0.0268±0.0096	2.84	0.537±0.123	0.01132±0.00274	270.522±10.151	2119.3±46.8
450	0.026	71.423±1.136	0.2958±0.0109	0.0053±0.0014	2.15	0.543±0.020	0.00645±0.00067	69.877±1.190	822.8±11.3
600	0.067	59.998±0.492	0.1786±0.0121	0.0000±0.0006	–0.04	0.328±0.022	0.00248±0.00035	59.997±0.527	726.9±5.3
800	0.144	61.299±0.300	0.1412±0.0039	0.0007±0.0005	0.32	0.259±0.007	0.00169±0.00013	61.077±0.336	737.6±3.3
1000	**0.325**	**66.681±1.339**	**0.0771±0.0019**	**0.0002±0.0002**	**0.06**	**0.141±0.004**	**0.00071±0.00009**	**66.614±1.341**	**791.7±12.9**
1500	**0.627**	**66.413±1.311**	**0.0933±0.0015**	**0.0004±0.0001**	**0.14**	**0.171±0.003**	**0.00066±0.00007**	**66.293±1.312**	**788.6±12.6**
2000	**0.790**	**66.111±1.315**	**0.1477±0.0028**	**0.0008±0.0002**	**0.36**	**0.271±0.005**	**0.00100±0.00014**	**65.853±1.316**	**784.4±12.7**
2500	0.887	76.127±0.293	0.1799±0.0034	–0.0006±0.0004	–0.24	0.330±0.006	0.00109±0.00019	76.291±0.314	882.5±2.9
3500	0.943	144.032±0.598	0.3947±0.0094	0.0013±0.0007	0.25	0.724±0.017	0.00463±0.00043	143.679±0.636	1412.8±4.3
6000	0.997	210.201±4.120	0.2365±0.0067	0.0079±0.0010	1.10	0.434±0.012	0.00564±0.00040	207.892±4.130	1804.7±22.7
9000	1.000	1419.964±51.193	0.2637±0.0819	0.0552±0.0103	1.15	0.484±0.150	0.04153±0.00527	1403.885±50.713	4572.4±60.0
Integrated		84.617±0.568	0.1452±0.0013	0.0013±0.0001	0.45	0.266±0.002	0.00178±0.00006	84.212±0.569	953.5±5.5
DT-95-2B WM (weighted average of *J* from standards = 0.008270 ± 0.000024)									
150	0.001	2350.510±208.035	–0.1580±0.0942	0.0317±0.0407	0.40	–0.290±0.173	0.21982±0.02698	2340.861±207.512	5436.6±152.1
200	0.002	492.567±32.538	–0.0199±0.0544	0.0050±0.0275	0.30	–0.037±0.100	0.16859±0.01427	491.056±33.443	2925.4±98.6
300	0.004	517.573±17.764	–0.0094±0.0324	0.0206±0.0136	1.18	–0.017±0.059	0.06234±0.00645	511.459±18.007	2984.6±51.4
450	0.016	116.404±0.848	0.0144±0.0045	0.0031±0.0022	0.78	0.026±0.008	0.00375±0.00078	115.470±1.059	1209.4±8.1
600	0.056	78.360±0.428	0.0055±0.0015	–0.0005±0.0007	–0.18	0.010±0.003	0.00106±0.00024	78.469±0.473	902.3±4.3
800	**0.236**	**87.180±0.855**	**0.0031±0.0002**	**0.0001±0.0002**	**0.04**	**0.006±0.000**	**0.00070±0.00008**	**87.117±0.856**	**978.9±7.4**
1000	**0.437**	**87.716±0.852**	**0.0043±0.0002**	**–0.0001±0.0002**	**–0.05**	**0.008±0.000**	**0.00068±0.00008**	**87.727±0.854**	**984.2±7.4**
1200	**0.620**	**86.815±0.899**	**0.0053±0.0002**	**–0.0001±0.0002**	**–0.03**	**0.010±0.000**	**0.00053±0.00008**	**86.814±0.900**	**976.2±7.8**
1400	**0.734**	**87.105±0.832**	**0.0062±0.0005**	**0.0002±0.0003**	**0.07**	**0.011±0.001**	**0.00057±0.00009**	**87.016±0.836**	**978.0±7.3**
1600	0.806	89.866±0.922	0.0067±0.0005	–0.0004±0.0004	–0.13	0.012±0.001	0.00051±0.00019	89.958±0.930	1003.3±8.0
2000	0.877	92.972±0.950	0.0060±0.0008	–0.0006±0.0003	–0.18	0.011±0.001	0.00068±0.00011	93.110±0.954	1030.1±8.0
2500	0.949	101.082±1.107	0.0042±0.0008	–0.0001±0.0004	–0.03	0.008±0.001	0.00069±0.00013	101.080±1.113	1096.1±9.0
3500	0.980	140.965±0.859	0.0061±0.0017	–0.0015±0.0008	–0.32	0.011±0.003	0.00144±0.00029	141.394±0.893	1397.1±6.1
7000	1.000	137.159±0.795	0.0028±0.0025	–0.0028±0.0009	–0.61	0.005±0.005	0.00126±0.00028	137.966±0.842	1373.4±5.9
Integrated		94.152±0.328	0.0048±0.0002	–0.0001±0.0001	–0.03	0.009±0.000	0.00114±0.00004	94.152±0.329	1038.9±3.6

Note: Bold rows indicate fractions used in calculation of plateau ages. Cum., cumulative; Atm., atmospheric; Ar_K, potassium derived ^{39}Ar.