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Geology and Neoarchean tectonic setting of the Central Hearne supracrustal belt, Western Churchill Province, Nunavut, Canada

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Abstract

The Central Hearne supracrustal belt, one of the largest Neoarchean "greenstone" terranes in the Canadian Shield, occurs in the Hearne domain of the Western Churchill Province, one of the largest, poorly known fragments of Archean crust on Earth. The belt contains two isotopically juvenile volcano–plutonic assemblages (I: ∼2710–2690 Ma, and II: ∼2685–2680 Ma), separated in time by localised, ∼2690 Ma, greenschist-facies deformation (D1), and overlain by ∼2680 Ma Archean siliciclastic and chemical sedimentary rocks. Extensive, penetrative, greenschist-facies regional deformation (D2) occurred at ∼2680 Ma, with amphibolite-facies metamorphism localised in the aureoles of isotopically juvenile synkinematic plutons. In many respects, the Central Hearne supracrustal belt is similar to other Neoarchean "greenstone" belts that have been interpreted in terms of arc-subduction systems, e.g. the Abitibi greenstone belt of the Superior Province. However, the principal tectonic characteristics of the Central Hearne supracrustal belt include: (i) location in an anomalously wide (>225 km) swath of penecontemporaneous juvenile crust that extends across much of the Hearne domain; (ii) close, primary intercalation of contemporaneous volcanic rocks of MORB-like and arc-like geochemical signatures, coupled with highly discontinuous volcanic map units; (iii) abundant intermediate to felsic volcanic rocks that do not represent a localised, laterally extensive volcanic arc edifice; and (iv) the development of isolated, independent, felsic volcanic centres throughout the magmatic history of the belt. These features are incompatible with oceanic arc or plateau models. We propose that the early history (assemblage I) of the Central Hearne supracrustal belt may be analogous with a modified extensional, suprasubduction "infant arc" model, such as that described for the earliest (Eocene) phase of construction of the Izu–Marianas–Bonin and Tonga arc-trench systems of the Southwest Pacific Ocean. The later history (assemblage II) may reflect attempted initiation of classical subduction and arc construction.

Keywords: Western Churchill; Neoarchean; Central Hearne; Geology; Tectonics

1. Introduction

One of the largest, poorly known fragments of Archean crust in the world is exposed in the western Churchill Province of the Canadian Shield (Fig. 1). Divided into Rae and Hearne domains by the Snowbird tectonic zone (Hoffman, 1988; cf. Hanmer et al., 1995), the Western Churchill Province is bordered to the northwest and southeast by the ∼2.0–1.9 Ga Thelon and ∼1.9–1.8 Ga Trans-Hudson orogens, respectively. Despite episodic reworking by deforma-

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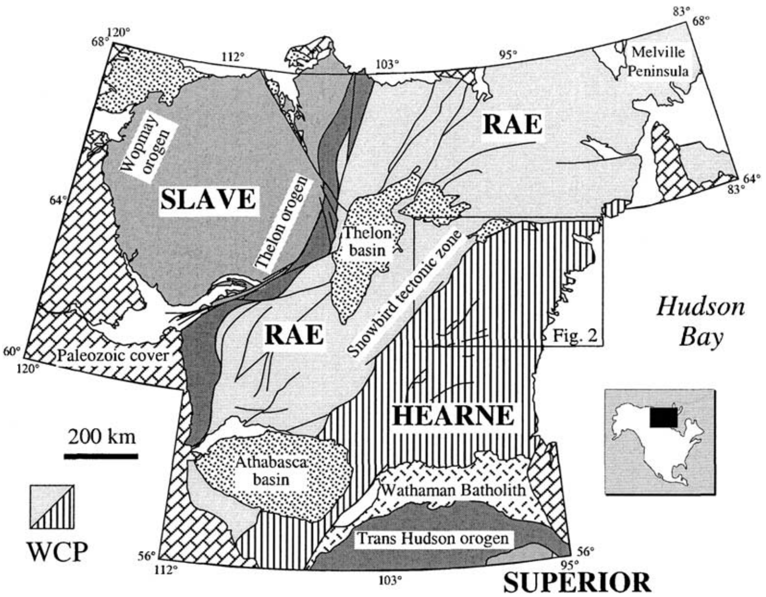


Fig. 1. Major tectonic components of the Western Canadian Shield to show the Western Churchill Province (WCP) divided into Rae and Hearne domains by the Snowbird tectonic zone (STZ), and flanked by the Slave craton. The location of Fig. 2 is indicated.

tion, metamorphism and magmatism, it has long been appreciated that vast areas of the Western Churchill Province are underlain by well-preserved Archean supracrustal rocks (e.g. Wright, 1967 and references therein; e.g. Fig. 2). However, although this represents one of the larger "granite-greenstone" terrains of the Canadian Shield, its conspicuous absence from recent global compilations (e.g. de Wit and Ashwal, 1997) underscores how little work has been undertaken there.

In the Hearne domain (Fig. 2), extensive volcanosedimentary supracrustal remnants, historically referred to as the "*Ennadai-Rankin greenstone belt*" (e.g. Wright, 1967; Aspler and Chiarenzelli, 1996), are distributed in a ∼700 km long swath from northern Saskatchewan to Hudson Bay. Limited continuity of these remnants and sparse geochronological data led Miller and Tella (1995) to subdivide the supracrustal rocks into an older, ∼2690 Ma, Kaminak greenstone belt to the southeast, flanked by younger, ∼2660 Ma belts to the north and west. Although further geochronological study has not validated the original basis for this zonation (see Davis et al., 2000), the general concept of distinct subdomains within the Hearne domain, each with its characteristic Neoarchean to Paleoproterozoic tectonometamorphic history and geophysical signature, is gaining ground (Hanmer and Relf, 2000; Berman et al., 2000; Davis et al., 2000; Jones et al., 2002). Moreover, preliminary dating in the Rankin Inlet area suggests that the stratigraphy there may be significantly younger than in the rest of the *Ennadai-Rankin greenstone belt* (Tella et al., 1995), and that the term is no longer appropriate.

In this contribution we focus on Neoarchean supracrustal and associated plutonic rocks that extend across the central part of the Hearne domain, from southwest of the Henik Lakes, via Kaminak Lake, to Tavani, which we refer to as the Central Hearne supracrustal belt (Figs. 2 and 3). In brief, the supracrustal belt, ∼350 km long by as much as 115 km wide, was deposited, deformed and metamorphosed at greenschist facies, and intruded by synvolcanic to syntectonic plutons, all between ∼2710 and ∼2680 Ma (Table 1; Davis et al., 2004). The belt is affected to the north and south by Paleoproterozoic metamorphism that culminates distally in middle amphibolite-facies conditions (Davidson, 1970a, 1970b). Basement to the supracrustal rocks has not been recognised. However, the supracrustal rocks are themselves unconformably overlain by locally preserved, coarse siliciclastic deposits of late Neoarchean to early Paleoproterozoic age (Montgomery Group, Spi Group, Wilson River

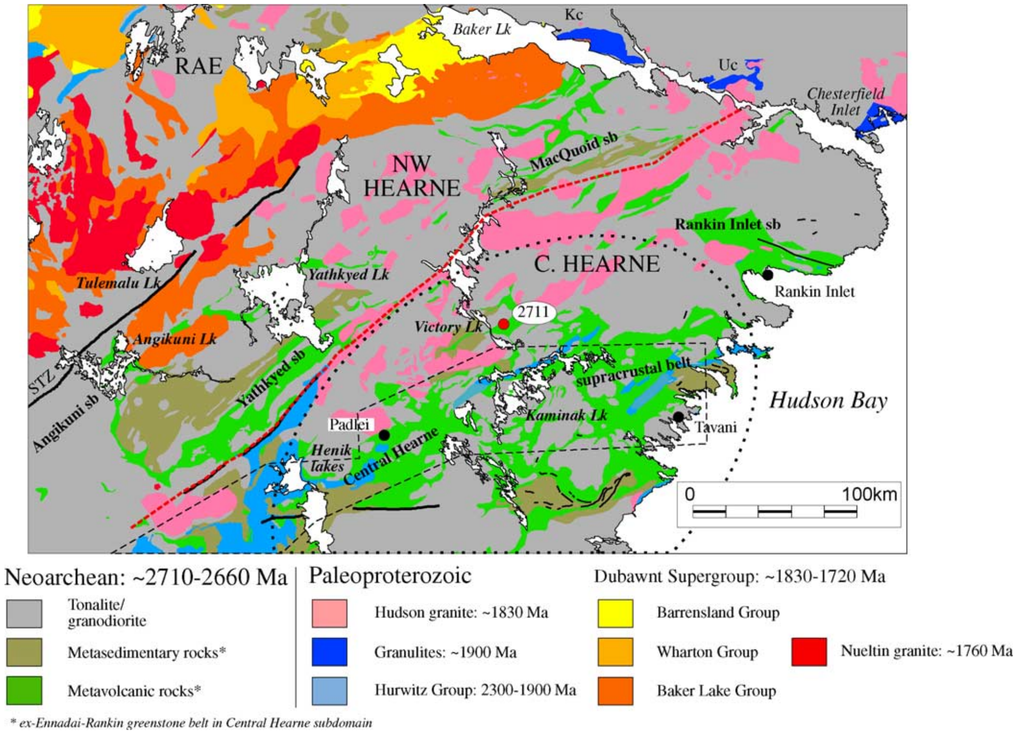


Fig. 2. Major lithological units of the central and northwestern subdomains of the Hearne domain. Black dotted line encompasses the approximate extent of Kaminak dykes. Location of Fig. 3 is indicated by the dashed-line box. Continuous heavy black lines are faults; fine black lines south of Kaminak Lake represent iron formation, dashed red line is the approximate boundary between northwestern and Central Hearne subdomains. The location of 2711 Ma volcanic rocks at Victory Lake is indicated. Kc, Kramanituar Complex; Uc, Uvauk Complex; sb, supracrustal belt; STZ, Snowbird tectonic zone. Hurwitz Group equivalent in the Rae domain is the Amer Group. After Paul (2001).

conglomerate; Rainbird et al., 2003; Fig. 3), as well as the regionally extensive deposits of the Paleoproterozoic Hurwitz Group (Table 1; Aspler and Chiarenzelli, 1996, 1997a; Aspler et al., 2001).

Historically, the Central Hearne supracrustal belt has been studied in three geographically defined segments, from southwest to northeast: Henik, Kaminak, and Tavani (Fig. 3). Detailed descriptions of the Henik (Eade, 1974; Aspler et al., 2000a; Aspler and Chiarenzelli, 1996, 1997a; Aspler et al., 1999a, 1999b, 2000a) and Tavani segments (Park and Ralser, 1992) have been presented elsewhere (see below and Table 1). The principal aim of this paper is to summarise the results of new fieldwork and supporting geoscience undertaken in the Kaminak segment, as part of the Western Churchill NATMAP Project (see Hanmer and Relf, 2000). These data will then be used as a bridge to synthesise the geology of the entire belt, thereby providing a geological framework to companion papers that focus on new geochronological (Davis et al., 2004) and petrological studies (Sandeman et al., 2004a, 2004b; Cousens et al., 2004). Finally, we will briefly consider the geological contribution to understanding the tectonic setting of the Central Hearne supracrustal belt.

2. Kaminak segment

Neoarchean supracrustal rocks of the Kaminak segment have been collectively defined as the Kaminak Group (Davidson, 1970a). New geological mapping of the Kaminak Group and associated granitoid plutons (Hanmer et al., 1998a, 1998b, 1998c; Irwin et al., 1998) builds on the previous work of Davidson (1970a, 1970b), Bell (1971), Relf (1995) and Irwin (1994, 1995, 1996, 1997). Although the new work has resulted in a more detailed lithological subdivision, a

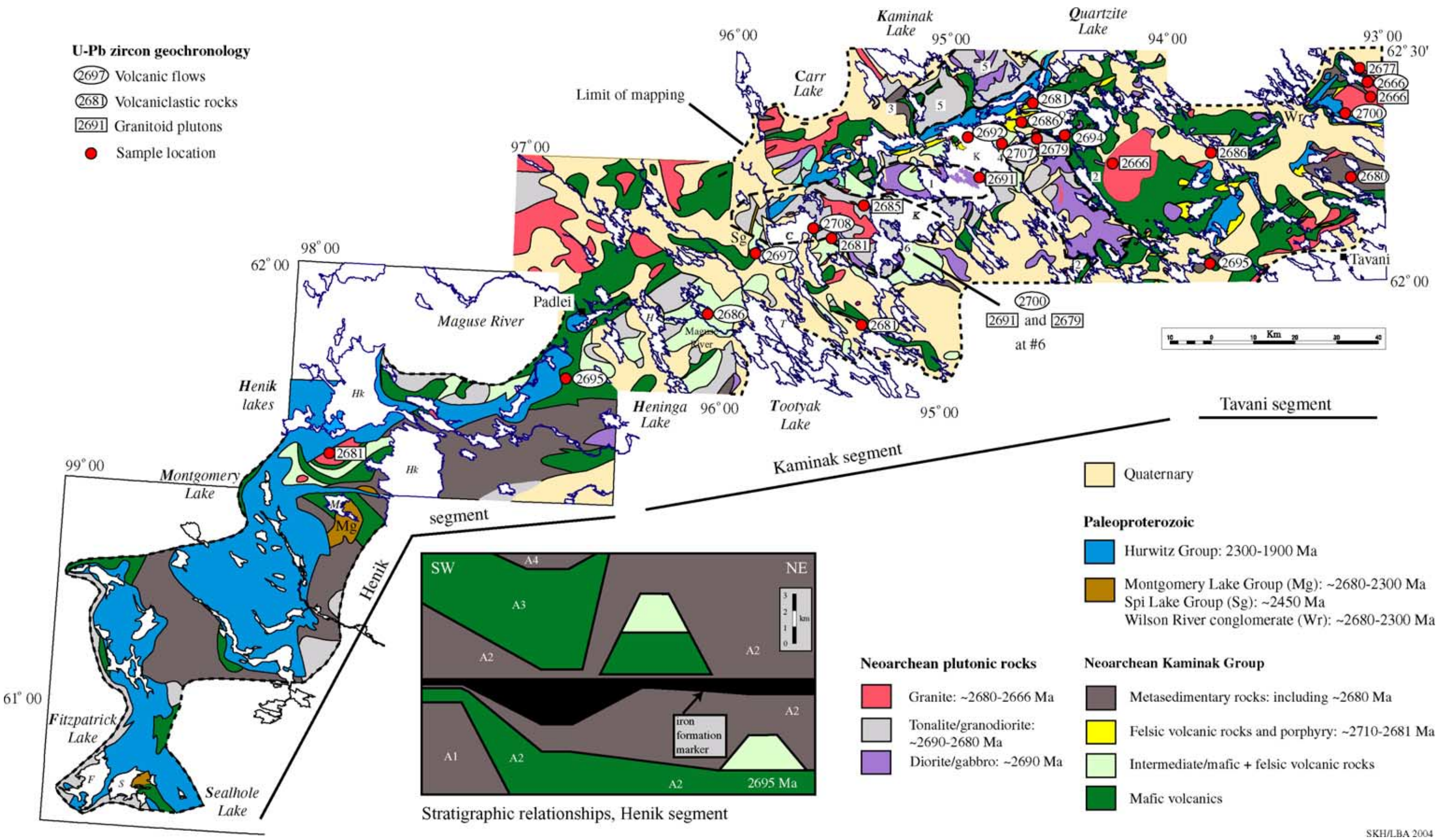


Fig. 3. Schematic lithological representation of the Central Hearne supracrustal belt. Henik segment after Aspler et al. (2000a), Aspler et al. (1997, 1999b, 2000a); Tavani segment after Park and Ralser (1992); Kaminak segment after Irwin et al. (1998) and Hanmer et al. (1998c). Curved dashed black lines in the central part of the Kaminak segment mark the interpolated boundaries of major plutons discussed in the text. Geochronological sample locations are indicated, and ages are given in millions of years. Numbers correspond to locations referred to in the text. Letters in lakes correspond to the lake names located just outside of the geology for clarity. Inset stratigraphic relations in the Henik segment are from Aspler and Chiarenzelli (1996) and Aspler et al. (1999a). See text for details.

Table 1
Geological histories of the segments of the Central Hearne supracrustal belt

	Henik	Kaminak	Tavani
Deformation	1830 Ma: thrusting?	1830 Ma: thrusting?	
~2300–1900 Ma	Hurwitz Group	Hurwitz Group	Hurwitz Group
	Unconformity Montgomery Lake Group	Unconformity ~2450 Ma: Spi Lake Group ~2450 Ma: Kaminak dykes	Unconformity 2450 Ma: Kaminak dykes Wilson River conglomerate
~2670–60 Ma	Unconformity	Unconformity 2666 Ma: group three post-kinematic granite pluton	Unconformity 666 Ma: Localised D3 folding; group 3 post-kinematic porphyry
Deformation	Localised D2 2681 Ma: group 2 late synkinematic granite	Regional D2 2679 Ma: group 2 synkinematic tonalite-granodiorite plutons	Regional D2 2677 Ma: group 2 synkinematic granite plutons
Assemblage II ~2685–2680 Ma	Metasedimentary rocks with BIF and predominantly mafic volcanism	~2680 Ma: southern metasedimentary rocks with BIF 2686–2681 Ma: younger mafic to felsic volcanic succession including major felsic volcanic centre; group 2 tonalite-granodiorite plutons	≤2680 Ma: metasedimentary rocks "*Tagiulik formation*" 2684 Ma: group 2 tonalite pluton
Deformation		D1	D1
Assemblage I ~2710–2690 Ma	2695 Ma: mafic to felsic volcanic succession	2711–2691 Ma: mafic to felsic volcanic succession, group 1 subvolcanic diorite and tonalite plutons	2700 Ma: mafic to felsic volcanic succession, including major felsic centre

Precise ages, errors and sources of data are given in the text.

coherent stratigraphic framework for this segment of the supracrustal belt is not apparent. At the local scale, the map pattern is complex and difficult to trace on the ground; stratigraphic units are laterally discontinuous and cut by voluminous dioritic to granitic plutons, all of which collectively impede the tracing and/or correlation of map units along strike (Hanmer et al., 1998c; Irwin et al., 1998). However, a gross geographical distribution of the volcanic associations can be discerned.

A well exposed, east-northeast trending corridor of predominantly greenschist-facies metavolcanic rocks, from Heninga Lake to Quartzite Lake, is bordered on either side by poorly exposed metasedimentary rocks of similar metamorphic grade (Figs. 2 and 3). It is primarily composed of (i) intermediate, mostly volcaniclastic rocks, (ii) a mixed "bimodal" association of felsic and mafic volcanic flows, breccias and arenites, and (iii) a large (20 km), dioritic, subvolcanic centre (#1 in Fig. 3). These rocks are bounded to the north, east and south by principally mafic, pillowed to massive flows, and layered mafic rocks. The mafic, intermediate and mixed volcanic associations are the volumetrically predominant rock types, in decreasing order. Geochronological data (U-Pb zircon; see Mortensen and Thorpe, 1987; Patterson and Heaman, 1990; Davis et al., 2004) show that subordinate felsic volcanism spanned 30 My, from ~2710 to 2680 Ma. In the lower part of the stratigraphy, felsic volcanic rocks occur as rare welded tuff (2708 +5/−2 Ma), as abundant interbedded layers within the mixed felsic–mafic association (2707 ± 4, 2697 ± 1 Ma), as map-scale units on the north side of Kaminak Lake (2692 ± 1 Ma), or as thin (<5 m), isolated, interbedded flows within the mafic lavas (2700 ± 1, 2694 ± 2 Ma; Fig. 3). However, in the upper part of the stratigraphy, in addition to 2686 +7/−3 Ma, ~50 m thick units of mixed flows and tuffs interbedded with mafic flows east of Heninga Lake, the felsic volcanic rocks form a large (10 km) volcanic centre (2686 +4/−2, 2681 ± 3 Ma), between Kaminak and Quartzite Lakes, that is contemporaneous with voluminous tonalite-granodiorite plutons (2682.2 +5/−2 Ma, 2679 ± 3 Ma; Fig. 3). This it appears that felsic volcanism was coeval with deposition of the entire volcanic succession (Table 1).

In the vicinity of Tootyak Lake (Fig. 3), very poorly exposed metasedimentary rocks on the south side of the volcanic rocks are fine grained arenites with ≤2681 Ma (Davis et al., 2004) volcaniclastic layers and extensive iron formations (Figs. 2 and 3). They are quite distinct from the homogeneous semipelites on the north side of the supracrustal belt that contain an isolated, 50–100 m thick, horizon of 2711 ± 2 Ma, quartz-phyric felsic and pillowed mafic volcanic flows whose contacts are not exposed (Victory Lake, Fig. 2; Aspler et al., 2000b, 2000c; Davis et al., 2004), as well as scattered amphibolite layers of unknown origin (see also Davidson, 1970a, 1970b).

The geochronological (Davis et al., 2004) and geochemical data (Sandeman et al., 2004a, 2004b) indicate that Neoarchean magmatic rocks in the Kaminak segment can be divided into at least two lithostratigraphic assemblages of volcanic rocks (I–II) and three groups of plutonic rocks, punctuated by two or three deformation events (D1–D3?; Table 1). However, because of lithological similarities and the absence of cross-cutting structural relationships in the field, it is generally not possible to attribute rocks to the assemblages, or foliations to deformation events, other than at the dated sample sites (Davis et al., 2004).

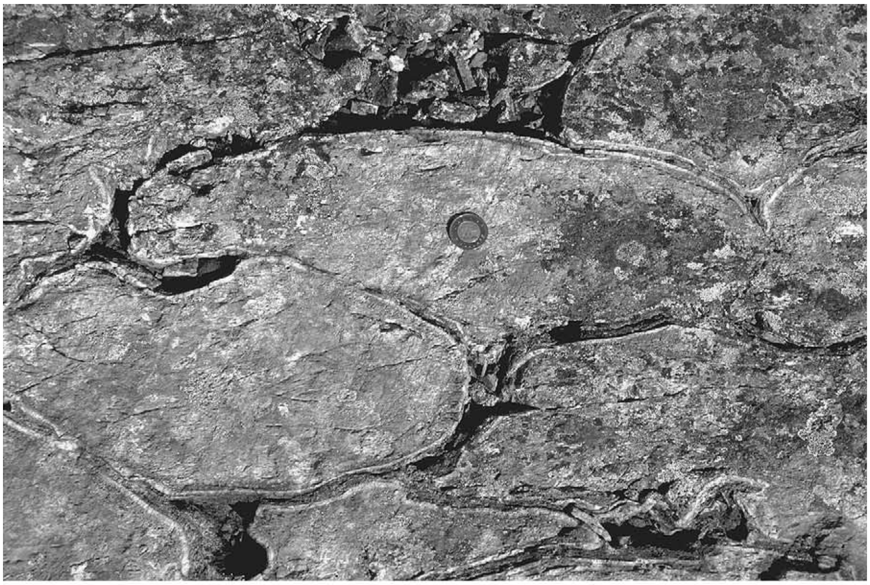
2.1. Volcanic rocks (assemblages I and II)

Most of the mafic flows are massive, fine grained, and generally featureless. However, pillowed flows occur throughout the volcanic pile (Fig. 4A). In places, stratigraphic younging is readily determined from pillow shapes, but unambiguous younging determinations are regionally sparse. Fine-grained mafic rocks, layered on a 1–5 cm scale, are enigmatic. At two localities, (#2 in Fig. 3), the layering is demonstrably the product of tectonic transposition of pillows and their selvages, suggestive of localised shear zones (Hanmer et al., 1998b). However, in the absence of strain markers, similar rocks elsewhere could be foliated mafic arenites with primary layering (Fig. 4B).

Thick packages of volcanic breccia, conglomerate and arenite of bulk intermediate composition occur throughout the volcanic succession (Fig. 4C). Poorly sorted, monomicitic and polymictic conglomerate and breccia contain subangular to rounded clasts (<1 m) of mafic to felsic volcanic rocks in a fine-grained, intermediate to mafic, volcanic arenite matrix. Rare clasts of massive leucodiorite to tonalite were probably derived from subvolcanic intrusions. The absence of obvious bedding, the coarse grain size and the overall textural immaturity, combined with local source-rock types in the clasts, suggest that these rocks were deposited as debris flows derived from nearby volcanic edifices.

Felsic volcanic rocks that form relatively thin (<50 m) layers and lenses within the mafic and intermediate rocks are part of assemblage I (∼2710–2690 Ma), except for a 2686 Ma tuff interbedded with pillow lavas east of Heninga Lake and the large 2686–2681 Ma felsic centre between Kaminak and Quartzite Lakes (Fig. 3), which are both part of assemblage II (Table 1). The felsic centre comprises banded quartz- and feldspar-phyric rhyolites that overlie the northern margin of a 10 km by 5 km, quartz-feldspar porphyry dome. The dome and the rhyolites are mantled by a carapace of felsic volcanic breccia and autobreccia composed of angular rhyolitic fragments (2–20 cm) in a matrix of similar composition (Fig. 4D). This in turn is overlain by matrix-supported volcanic breccias and conglomerates, interpreted as locally derived debris flows (Fig. 4E).

A mixed volcanic association, with a relatively high proportion of felsic versus intermediate to mafic volcanic rocks (1:2–1:3), outcrops discontinuously throughout much of the Kaminak segment (“intermediate/mafic + felsic volcanic” unit in Fig. 3). In general, the intermediate to mafic rocks range from pillow lavas, massive flows, and gabbroic sills, to volcanic breccias and arenites. Interspersed within them are 5–500 m thick horizons of felsic volcanic rocks, similar to those of the assemblage II volcanic centre west of Quartzite Lake. Available geochronological data (2704 and 2697 Ma) place the association in assemblage I (Table 1). The 2711 Ma mafic–felsic volcanic rocks, isolated within the northern metasedimentary rocks at Victory Lake (Fig. 2, Table 1), superficially resemble this mixed association (Aspler et al., 2000c) and are interpreted as part of the same assemblage.



(A)

(B)

Fig. 4. A–H are from the Kaminak segment, I is from the Tavani segment. (A) Well preserved mafic pillows with selvages. Stratigraphic younging is towards the top of the photo. (B) Enigmatic, layered mafic rock, possibly a foliated mafic arenite. (C) Foliated, framework-supported, volcanic, polymictic breccia with elongate mafic to felsic fragments. (D) Well preserved, matrix supported, felsic, volcanic breccia with angular quartz porphyry clasts. (E) Well preserved, framework-supported, volcanic, polymictic breccia with felsic and mafic fragments. (F) Thin bedded, mudstone–sandstone sedimentary succession. (G) Well preserved, laminated banded iron formation with tuffaceous layers (light grey). (H) Well preserved, cross-cutting intrusive relations within the central Kaminak intrusive suite. Commonly oikocrystic hornblende diorite is cut by mafic (dark, right and top centre) and leucodioritic sheets (light, left of centre), all cut by narrow, commonly plagioclase-phyric diorite sheets (parallel to long side of photo). Oikocrysts and phenocrysts are not present in this photo. (I) Well preserved laminated arenite with delicate ball-and-slump sedimentary structures.

(C)

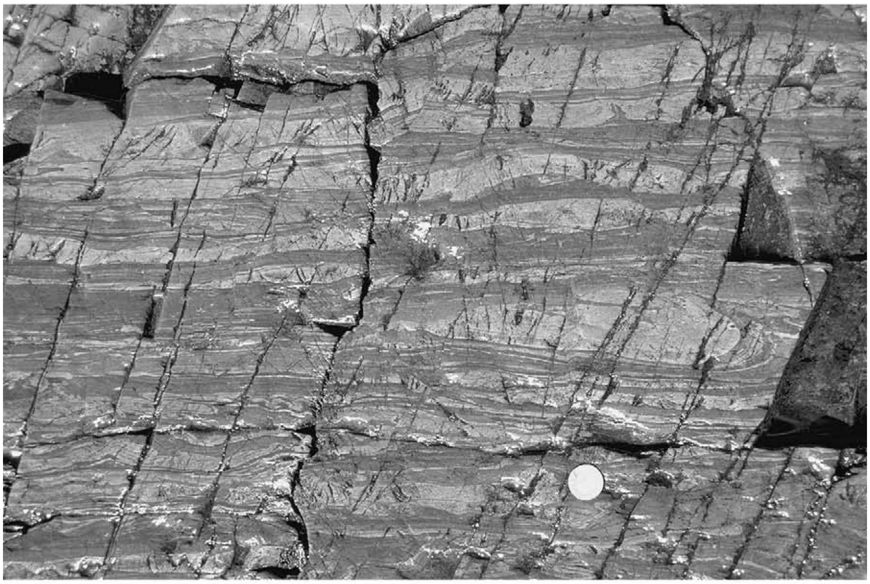
(D)

Fig. 4. (*Contiuned*)

2.2. Sedimentary rocks

Although large volumes of clastic sedimentary rocks flank the volcanic corridor, the best exposed examples occur at the top of, and/or overlie, volcanic rocks of assemblage II (Table 1). East of the felsic dome at Quartzite Lake (Fig. 3), interbedded polymictic conglomerate, sandstone and sandstone/siltstone rhythmite define a thick (>500 m) section that stratigraphically overlies the dome (Relf, 1995; Hanmer et al., 1998a; Irwin et al., 1998). Conglomerate is predominant in the lowermost 150 m of the section.

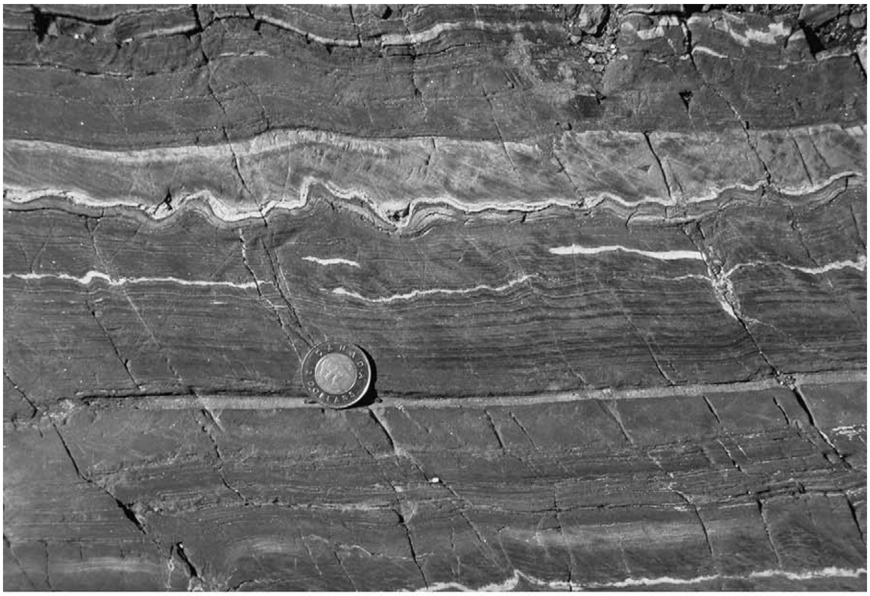
(E)



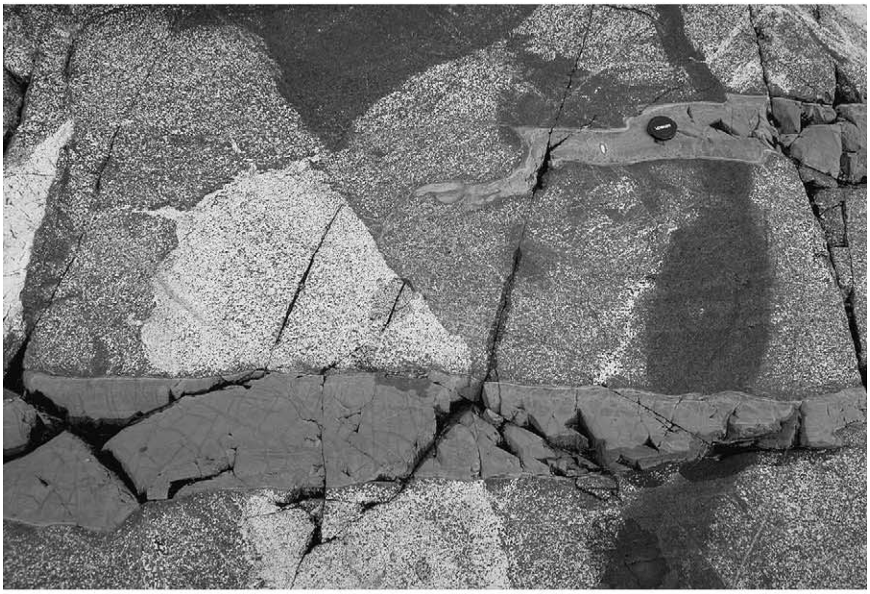
(F)

Fig. 4. (*Contiuned*)

Normal grading is common, but reverse grading and low-angle crossbedding also are preserved. Bedding is well defined by sandy lenses and interlayers. Rhythmically interlayered mudstone and fine-grained sandstone occur in units up to 50 m thick. Mudstone interlayers range in thickness from a few millimetres to 10 cm, and define intervals of well developed wavy and lenticular bedding that are separated by tabular, massive sandstone beds (Fig. 4F). Some sandstone layers are normally graded, with flames or ball-and-pillow structures. We interpret the conglomerates as part of a submarine channel system, and the



(G)



(H)

Fig. 4. (*Contiuned*)

rhythmites as shallow-water turbidites. An increase in the proportion of rhythmite up-section suggests progressive shallowing of the depositional basin.

Normally graded, lithic arenite-siltstone rhythmites, similar to those at Quartzite Lake, outcrop just north of Kaminak Lake (#3 in Fig. 3). The section has an apparent thickness greater than 3 km, but stratigraphic reversals (Irwin et al., 1998) indicate that it has been thickened by unspecified structural repetition. Davidson (1970a) suggested that these sedimentary rocks overlie the volcanic rocks, both structurally and stratigraphically, and are part of the poorly exposed expanse of ho-

(l)

Fig. 4. (*Contiuned*).

mogeneous semipelite with mafic layers that extends across regional strike to just north of Victory Lake (Fig. 2; Hanmer et al., 1998a, 1998b). However, field relationships between the sediments and the volcanic rocks remain unconstrained due to lack of outcrop.

By using linear positive anomalies in the regional aeromagnetic field, Davidson (1970b; see also Geological Survey of Canada, 1987) was able to trace extensive, east–west-striking, banded iron formation units in the poorly exposed terrain south of Kaminak Lake to the coast of Hudson Bay (Figs. 2 and 4G). The magnetite–hematite–chert (jasper) iron formation is interbedded with argillaceous, clastic sedimentary rocks within thick sequences of mafic to intermediate volcanic and volcaniclastic rocks. Oxide layers in the iron formation exhibit very thin, rhythmic, parallel lamination, deformed about syn-sedimentary folds and cut by clastic dykes, features indicative of slope instability. Deposition appears to have occurred in a quiet water setting, below storm wave base (>100 m). The presence of intercalated felsic volcaniclastic sediments, tuffs and magmatic dykes indicate proximity to active volcanic centres. Strictly speaking, detrital ~2680 Ma zircons in a tuffacaeous layer, comparable in age with the youngest volcanic rocks and associated ~2680 Ma granitoid plutons of assemblage II (see below), represent a maximum age for the deposition of these sedimentary rocks (Fig. 3, Table 1; Davis et al., 2004). However, jasper fragments, similar to that in the iron formation, occur in felsic volcanic breccias and tuffs east of Heninga Lake, which are dated at 2686 Ma (Fig. 3; Davis et al., 2004), suggesting that some iron formation formed relatively early with respect to assemblage II.

2.3. Plutonic rocks (groups 1–3)

Neoarchean plutonic rocks intruded into the Kaminak Group include (i) group 1 synvolcanic diorites, (ii) group 2 tonalites, granodiorites and granites, including syntectonic plutons, and (iii) group 3 post-kinematic granites (Hanmer et al., 1998a, 1998c; Irwin et al., 1998; Sandeman et al., 2004b). In addition, a deeply weathered, ~2660 Ma, post-kinematic, carbonatite-bearing alkalic complex at the northeast end of Kaminak Lake has been described in some detail elsewhere (#4 in Fig. 3; Davidson, 1970a, 1970c; Cavell et al., 1992).

The central Kaminak intrusive suite (2691 ± 1 Ma, Davis et al., 2004; #1 in Fig. 3) is a large (20 km), group 1 plutonic complex comprising at least three texturally distinct units (Fig. 4H): (a) hornblende oikocrystic diorite, (b) medium- to fine-grained leucodiorite, (c) plagioclase-phyric diorite, as well as intimately admixed components of all three. Locally, the diorite units both intrude and grade into adjacent hypabyssal and extrusive units of the Kaminak Group. Xenoliths and cross-cutting dykes (Fig. 4H), both similar in composition to intermediate volcanic components of the Kaminak Group, suggest that the central Kaminak intrusive suite is a major subvolcanic centre.

Historically, variably foliated tonalite and granodiorite rocks from the Kaminak segment have been collectively referred to as the "*Kaminak batholith*" (Davidson, 1970a). However, we identify at least three distinct, group 2 composite plutons: Kaminak, Ferguson and Carr. The Kaminak pluton (2679 +3/−2 Ma, Davis et al., 2004; Fig. 3) post-dates the assemblage II volcanic succession (Table 1). It ranges in composition from granodiorite to gabbro, but is predominantly composed of medium-grained, hornblende-biotite granodiorite to tonalite. The pluton has a disruptive, net-vein contact along its northeastern margin, where it incorporates rafts and blocks of volcanic rocks of the Kaminak Group. In contrast, wallrock along the western contact of the pluton is characterised by hornfelsed, highly-deformed, banded metavolcanic rocks. There, variably strained veins emanating from the pluton suggest that the immediately adjacent wallrocks were deformed during pluton emplacement. To the south and east, tonalite of the Kaminak pluton intrudes quartz diorite, diorite and subordinate gabbro. Phases in the undated Ferguson pluton (#5 in Fig. 3) range from minor gabbro, through quartz diorite, to predominantly tonalite and locally granodiorite, and contain inclusions of mafic supracrustal rocks. The pluton is extensively altered and plutonic textures are poorly preserved.

The Carr pluton (Fig. 3) is a composite body, cored by a 2685 +2/−1 Ma (Davis et al., 2004) homogeneous biotite-hornblende granite to granodiorite, with subordinate diorite to monzonite. The northern and southern margins of the pluton are clearly intrusive. The eastern margin comprises a series of margin-parallel, schlieren zones alternating with inclusion-free tonalite and granodiorite that intercalates with the central Kaminak intrusive suite over a distance of ∼4 km. The granodiorite and wallrocks are cut by a 2681 ± 3 Ma monzonite body, time equivalent to the later stages of assemblage II volcanism (Table 1).

2.4. Structure

Within the Kaminak Group, compositional layering varies from centimetre-scale banding to panels of homogeneous pillow lava, tens of metres to kilometres thick. Most outcrop-scale lithological variation is interpreted as bedding (S0), or synvolcanic microgabbroic sills. Except for local outcrops of non-deformed pillowed flows, felsic lavas and silicified volcanic rocks, the supracrustal lithologies exhibit a variably developed layer-parallel foliation. Regionally, this fabric is defined by lower greenschist-facies chlorite-albite ± sericite mineral assemblages, except adjacent to the larger granitoid plutons where the foliation both encloses, and is overgrown by, garnet and staurolite, or contains hornblende as a fabric-forming mineral. The foliation is axial planar to rare, small-scale (<50 cm) intrafolial folds of S0. As will become apparent below, this foliation is an S2 fabric (Table 1). It is steeply dipping to vertical and carries a variably developed, steeply pitching, L2 extension lineation, which is colinear with a locally developed crenulation (F3?) that affects S2. The S2 foliation is deformed by widespread, but isolated structures that, with very few exceptions, are only developed at a small scale (<50 cm). These comprise isolated, asymmetrical folds, and an oblique, axial–planar crenulation cleavage (S3?). Because these post-S2 structures are neither penetrative nor extensive, they cannot be correlated between widely separated outcrops. Strictly speaking, they represent a D3 event, but we cannot exclude the possibility that they developed late during D2, or that they represent several generations of structures formed at different times. These structures are Neoarchean in age because: (i) the geographical distribution of the amphibolite-facies fabric-forming metamorphic assemblages and the metamorphic minerals the foliation encloses demonstrate that the foliation developed synchronously with emplacement of ∼2680 Ma, late group 2 plutons; (ii) the supracrustal rocks contain ∼2670–2655 Ma metamorphic monazite and titanite and are cut by a post-kinematic, 2666 ± 1 Ma, non-deformed, K-feldspar megacrystic granite (Relf, 1995; Davis et al., 2004; Fig. 3), as well as by pristine members of the north–northeast trending Kaminak mafic dyke swarm (Davidson, 1970a) dated at ~2450 Ma (Heaman, 1994); and (iii) they are unconformably overlain by the post ~2450 Ma Hurwitz Group (Table 1; Hanmer et al., 1998b; Irwin et al., 1998; Aspler et al., 2001).

Apart from the group 1 central Kaminak intrusive suite and main body of the group 2 Carr pluton, both of which appear to be synvolcanic with respect to assemblage I and early assemblage II, respectively (Table 1), voluminous tonalitic to granitic group 2 intrusions were emplaced relatively late during regional deformation of the Kaminak Group. For example, a hornblende isograd is spatially related to the 2679 Ma Kaminak pluton, and to an inferred, buried pluton east of Quartzite Lake (Relf, 1995; Irwin et al., 1998). In both cases, hornblende is a fabric-forming phase in the regional S2 foliation, which tends to follow the general shape of the plutons. A similar situation is associated with undated tonalite, granodiorite and large granite plutons in the Heninga Lake area and north of Padlei (Fig. 3), where S2 foliation closely parallels the three dimensional form of the roofs of the plutons (Hanmer et al., 1998c). The timing of the regional D2 deformation is bracketed by pre-foliation 2681 Ma volcanic and ~2680 Ma sedimentary rocks, and the post-kinematic, 2666 Ma granite near Quartzite Lake (Fig. 3). However, at one locality (#7 in Fig. 3), a folded, 2691 +4/−3 Ma tonalite vein cross-cuts an already folded, amphibolite-facies foliation, and is itself cut by a post-kinematic, 2679 +3/−2 tonalite vein (see Davis et al., 2004). The metamorphic grade of the folded fabric suggests that it formed while heat was being advected by an early phase of tonalite-granodiorite plutonism. We attribute these pre-2691 Ma structures to a D1 event. In summary, it appears that the regional D2 event in the Kaminak Group occurred at ~2680 Ma and affected rocks as young as assemblage II and group 2 plutons. However, an older D1 deformation that occurred at, or just prior to, 2691 Ma, and could only have affected rocks of assemblage I and group 1 plutons, has only been identified locally (Table 1).

3. Tavani segment

Many lithological, plutonic and structural aspects of the Tavani segment of the Central Hearne supracrustal belt are very similar to those of the Kaminak segment (Fig. 3). Archean supracrustal rocks are continuous between the two areas, and the term "Kaminak Group" has been employed in both segments (Heywood, 1973; Ridler and Shilts, 1974a, 1974b). However, Park and Ralser (1992) proposed replacing the Kaminak Group in the Tavani segment by a "*Kasigialik Group*", divided into "*Atungag, Akliqnaktuk* and *Evitaruktuk formations*", plus an older "*Tagiulik formation*" that they interpreted to be allochthonous with respect to the Kasigialik Group.

According to Park and Ralser (1992), the Atungag formation represents a stratigraphically lowermost unit of gabbro and mafic pillow lavas with tuffaceous sediments. The Akliqnaktuk formation comprises mafic pillow lavas overlain by mafic volcanic arenites and pillowed flows ± hyaloclastite, overlain in turn by felsic volcanic and volcaniclastic rocks, all divided into 11 "*facies associations*". Inter-flow deposits in many of the facies associations comprise breccias with granitoid, chert and quartzite clasts, in addition to mafic pillow fragments, arenite and black slates. A thick (>1 km) package of felsic volcanic rocks, dated at 2700 ± 1 Ma, is contemporaneous with assemblage I of the Kaminak segment (Fig. 3, Table 1; Davis et al., 2004). Potentially, this succession also contains components equivalent to assemblage II of the Kaminak segment, but their identification will require further geochronology. The Evitaruktuk formation is a succession of slate-greywacke-conglomerate turbidites, overlain by arkoses with granitic clasts. Contacts between the three formations were interpreted to be conformable (Park and Ralser, 1992).

In the eastern part of the Tavani segment, Park and Ralser (1992) identified a structurally discordant, map-scale panel of their Tagiulik formation. This contains conglomerates, overlain by quartz-poor turbidites and turbidite-hosted banded iron formation that lithologically resemble the ~2680 Ma sedimentary rocks on the south side of the Kaminak segment. Detrital zircons that indicate a maximum depositional age of 2680 Ma for the formation (Fig. 3, Table 1; Davis et al., 2004) favour this correlation. Although Park and Ralser (1992) interpreted the basal part of the Tagiulik formation as a ductile thrust zone at the base of an exotic, allochthonous nappe, we investigated the proposed "shear zone" and found it to contain widespread, delicate, well preserved, sedimentary structures (Fig. 4I). Rather, we suggest that these sedimentary rocks stratigraphically overlie the volcanic succession.

Plutonic rocks in the Tavani segment are gabbroic to granitic in composition (Park and Ralser, 1992). The principal deformation structures are a regionally pervasive foliation associated with a dip-parallel extension lineation. Map-scale folds are rare, except in well layered turbidite units where layer-parallel foliation is deformed by upright, open to tight folds. Park and Ralser (1992) interpreted granites, dated at 2677 $\pm$ 2 Ma and 2666 $\pm$ 2 Ma, as synkinematic with respect to the foliation-forming and folding events, respectively, and reported an age of 2666 $\pm$ 9 Ma from a late-synkinematic porphyry (Table 1, Fig. 3). However, local cross-cutting relationships indicate that an older foliation developed, at least locally, during or prior to the intrusion of a 2686 $\pm$ 4 Ma tonalite pluton (Fig. 3; Davis and Peterson, 1998). Taken together, the preceding observations imply that Park and Ralser's Kasigialik Group is essentially equivalent to the Kaminak Group, intruded by late group 1 and group 3 granitic plutons (Table 1). Moreover, as in the Kaminak segment, the regional foliation is S2, formed at ~2680 Ma, and there is local evidence that an earlier S1 foliation-forming event occurred prior to 2686 Ma (Table 1). The map-scale folds of S2 represent a locally developed D3 event.

4. Henik segment

Neoarchean supracrustal rocks in the Henik segment of the Central Hearne supracrustal belt, referred to as the "*Henik Group*" (Eade, 1974; Aspler and Chiarenzelli, 1996), are divided into four informal units, "A1–A4", in ascending stratigraphic order (Aspler and Chiarenzelli, 1996; Aspler et al., 1999a). Using Aspler and Chiarenzelli's terminology: unit A1 comprises sandstone to mudstone fining-upward successions with local mafic volcanic layers; unit A2 is a sedimentary-volcanic assemblage containing sandstone to mudstone fining-upward successions, iron formation, felsic breccias and tuffs, mafic flows and tuffs, and rare conglomerate; unit A3 is a mafic volcanic-gabbro sill/dyke complex with a high proportion of flows to volcaniclastic rocks; and unit A4 consists exclusively of sandstones to mudstones. The basement upon which unit A1 was deposited remains undetermined. Extensive iron formation units that serve as marker horizons enabling correlation between stratigraphic sections of otherwise laterally discontinuous lithological units (Aspler and Chiarenzelli, 1996; Aspler et al., 1999a) appear to be correlative with similar ~2680 Ma rocks in the Kaminak segment. (Davidson, 1970b; Davis et al., 2004).

Map units in the Henik segment are remarkably similar to those of the Kaminak segment, albeit with a lower proportion of volcanic breccia, conglomerate and arenite (Aspler et al., 1999a). A single U-Pb zircon age of ~2695 Ma (Mortensen and Roscoe, unpublished data reported in Cousens et al. (2004)) from felsic volcanic rocks in the northeastern part of the Henik segment, interbedded with mafic volcanic rocks stratigraphically below the main iron formation horizon, suggests that volcanism was, in part, coeval with assemblage I in the rest of the supracrustal belt (Fig. 3, Table 1). Moreover, regional mapping indicates that the similarity between the Henik and Kaminak segments extends to the southwest as far as Sealhole Lake (Fig. 3; Aspler, 2000).

Archean structures in the Henik segment were inhomogeneously developed. For example, in the vicinity of Montgomery Lake (Fig. 3), supracrustal rocks were tilted, but not cleaved or folded during the Neoarchean (Aspler et al., 1992). In particular, this includes the mafic volcanic and siliciclastic rocks of units A3 and A4. In marked contrast, north of Montgomery Lake, Archean rocks carry a penetrative regional cleavage, cut by a late synkinematic granitic pluton dated at 2681 $\pm$ 1 Ma (Fig. 3, Table 1; Davis et al., 2004), that is therefore correlative with S2 in the Kaminak segment. Other granitoid plutons were emplaced syn- to post-kinematically with respect to S2, and Neoarchean map-scale fold closures appear to be absent throughout the Henik segment, despite the presence of local younging reversals. In brief, the timing and style of deformation are, therefore, in general agreement with relationships from the rest of the supracrustal belt (Table 1). However, available timing constraints require that the mafic volcanic and siliciclastic rocks of units A3 and A4 are younger than the ~2680 Ma sedimentary rocks in the rest of the supracrustal belt (see Fig. 3), and potentially younger than D2 (Davis et al., 2004).

5. Central Hearne supracrustal belt: synthesis

The geographical segments of the Central Hearne supracrustal belt present a common set of supracrustal lithological units formed over a relatively short time span (∼2710–2680 Ma, but principally after ∼2700 Ma; Fig. 3, Table 1). Mafic lavas, intermediate volcaniclastic rocks and isolated felsic volcanic centres were extruded and deposited in a submarine environment (assemblages I and II). The principal variation is an increased proportion of clastic sediments in the Henik segment, interpreted in terms of basin deepening toward the southwest (Aspler and Chiarenzelli, 1996). However, extensive iron formation and associated sedimentary rocks (∼2680 Ma) between the Henik Lakes and Hudson Bay (Davidson, 1970b) could represent the eastward continuation of the deeper water basin along the southern margin of the supracrustal belt (Fig. 2). If this is indeed the case, it implies that the axis of the depositional basin was oriented approximately east–west, and that lateral facies variation was north–south, at least during the later stages of basin development.

Conceivably, deep-water sandstone–mudstone beds at the top of the section in the Henik segment (unit A4; Fig. 3) could be equivalent to the extensive homogeneous semipelites north of Kaminak Lake (Fig. 2) that Davidson (1970a) had suggested were younger than the volcanic rocks of the Kaminak segment. However, the volcanic rocks associated with these sediments are the oldest dated rocks in the belt (Victory Lake; Fig. 2, Table 1), which would imply some form of tectonic intercalation.

5.1. Lateral discontinuity

Comparison of the map patterns from all three segments (Davidson, 1970a, 1970b; Bell, 1971; Park and Ralser, 1992; Aspler, 2000, Aspler, 1993; Aspler et al., 1992, 1994, 1997, 1999b, 2000b, 2000c; Aspler and Chiarenzelli, 1997b; Irwin, 1994, 1995, 1996, 1997; Relf, 1995; Peterson, 1997; Irwin et al., 1998; Hanmer et al., 1998c) indicates that the laterally discontinuous nature of most of the supracrustal map-units of assemblage I is a characteristic feature of the Central Hearne supracrustal belt as a whole. The excellent state of preservation of primary features (Fig. 4A and C–I; see also Park and Ralser, 1992; Aspler and Chiarenzelli, 1996) suggests that lateral discontinuity is not the product of intense, penetrative deformation. Alternatively, if lateral discontinuity were the result of discrete faulting, the map unit boundaries should outline kinematically linked fault networks, such as contractional, extensional or strike-slip duplexes (e.g. Boyer and Elliot, 1982; Swanson, 1989, 1990), for which there is no evidence. Accordingly, we infer that the lateral discontinuity of map units is essentially a primary feature of the supracrustal rocks, albeit modified by shortening, that reflects laterally-restricted, short-lived, depositional sub-environments. Supported by the geochronological data (Davis et al., 2004), we envisage an extensive, subaqueous, mafic to intermediate crust, with independent depositional sub-basins, that was dotted with local felsic volcano-plutonic centres, at least throughout the development of the assemblage I volcanic succession.

The plutonic and structural histories of the three segments are also broadly comparable. The tonalite-granodiorite and granite plutons are less voluminous toward the southwest (Fig. 3). Combined with the absence of Neoarchean foliation in parts of the Henik segment, this might suggest a transition toward shallower Neoarchean structural levels. In the Kaminak and Tavani segments, the S2 regional foliation clearly affects rocks of assemblage II and group 2 plutons, some of which are synkinematic, and is constrained to have formed at ∼2680 Ma. However, dated intrusive relationships demonstrate the presence of an earlier foliation, S1, otherwise indistinguishable in the field from S2, formed at, or prior to, 2691 Ma in the Kaminak segment and at, or prior to, 2686 Ma in the Tavani segment (Table 1). In either case, the geochronology tells us that S1 only affected assemblage I rocks and group 1 plutons. The absence of either a stratigraphic discordance between the two assemblages, or evidence for reworking of S1 by S2 is problematic and remains to be explained.

6. Discussion: tectonic setting

Early models for the construction of the Central Hearne supracrustal belt focused on the northeastern segments, but without the benefit of comprehensive geochemical or modern geochronological support. For the Kaminak segment, Davidson (1970b) proposed a single cycle stratigraphy with mafic volcanic flows at the base, overlain by intermediate to felsic volcanic and volcaniclastic rocks that formed localised, transient centres, all capped by clastic sediments and banded iron formation with minor chert. Ridler and Shilts (1974a, 1974b) extended Davidson's stratigraphy and identified four complete, and one incomplete, mafic–felsic volcanic/sedimentary cycles within the Kaminak and Tavani segments. However, these cyclic, "layer-cake" stratigraphic models imply extensive lateral continuity of map units that, with the exception of iron formations, is not present. In the Tavani segment, Park and Ralser (1992) were more explicit in interpreting the Atungag formation as oceanic basement, the Akliqnaktuk formation as an intra-oceanic arc, and the Evitaruktuk formation as turbidites derived from a similar arc source.

Integrating largely reconnaissance-level geological data from the Western Churchill Province as a whole, Aspler and Chiarenzelli (1996) suggested tectonic models that envisaged continental rifting and sea floor spreading (Rae domain), and development of either a back-arc basin or merging oceanic arcs (Hearne domain). However, the new geological, geochronological (Davis et al., 2004) and geochemical data (Sandeman et al., 2001, 2004a, 2004b; Cousens et al., 2004) caution against strict application of actualistic plate tectonic models for the Rae and Hearne domains in general, and the Central Hearne supracrustal belt in particular.

6.1. *Mature arc or oceanic plateau?*

Although modern plate tectonic models involving mature island arcs and associated marginal basins have been increasingly used to explain the evolution of Archean greenstone belts elsewhere, Hamilton (1988, 1998) has drawn attention to numerous inconsistencies, many of which apply in the present example, at least during the period ∼2710–2690 Ma (assemblage I). First, despite juvenile, oceanic arc-like geochemical signatures (Sandeman et al., 2004a, 2004b; Cousens et al., 2004), a laterally continuous volcanic edifice has not been identified, either within the Central Hearne supracrustal belt, or the Hearne domain as a whole (Hanmer and Relf, 2000; Relf and Hanmer, 2000). Second, geochronological data (Davis et al., 2000, 2004) indicate that ∼2710–2680 Ma supracrustal belts show no evidence of temporal polarity at the scale of the Hearne domain. Third, intermediate to felsic volcanic centres occur throughout the period of mafic volcanism (Table 1; Davis et al., 2004), suggesting that an "oceanic basement" upon which an intra-oceanic arc might have been built has not been found. Similarly, in the absence of evidence for intimate tectonic imbrication, intercalation of MORB-like and oceanic arc-like geochemical signatures on a scale of hundreds of metres to ∼1 km (Sandeman et al., 2004a; Cousens et al., 2004) is incompatible with a classical volcanic arc edifice built on oceanic crust. These observations are also incompatible with oceanic plateau models for oceanic crustal growth (Sandeman et al., 2001). Fourth, systematic mapping of the supracrustal belts of the Hearne domain (Park and Ralser, 1992; Tella et al., 1993, 1994, 2001; Tella, 1993, 1994, 1995; Tella and Schau, 1994; Aspler et al., 1997, 1998, 1999b, 1999c, 2000c; Peterson, 1997; Hanmer et al., 1998c; Irwin et al., 1998) has not identified lithological or structural units typical of accretionary wedges (cf. Hamilton, 1988; Polat and Kerrich, 1999). Fifth, unlike modern settings, penecontemporaneous juvenile crust with arc-like geochemical signatures is now known to extend across the Hearne domain (Sandeman et al., 2001), a minimum across-strike width of 225 km after deformation, an order of magnitude greater than for typical, modern, oceanic arcs. (e.g. Hamilton, 1988). Taken together, these observations suggest that classical arc and back-arc plate tectonic models, or the oceanic plateau paradigm, may not apply to the Central Hearne supracrustal belt. In addition to these domain-scale considerations, a tectonic model for the Central Hearne supracrustal belt should also account for the discontinuous nature of the supracrustal map units.

6.2. *Aborted "infant" arc?*

Insight into the kind of processes that could have operated during the construction of the Central Hearne supracrustal belt comes from the earliest (Eocene) phase of construction of the Izu–Marianas–Bonin and Tonga arc-trench systems of the Southwest Pacific Ocean (e.g. Stern and Bloomer, 1992; Taylor, 1992; Bloomer et al., 1995; Clift, 1995; Hawkins, 1995), and suggested analogues (Pearce et al., 1981,

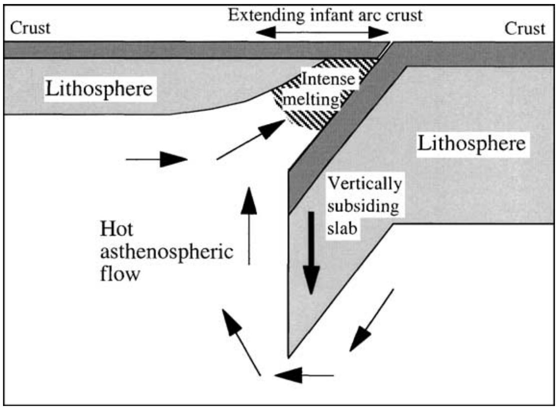


Fig. 5. Schematic illustration of the extensional, suprasubduction "infant arc" model. After Stern and Bloomer (1992) and Bloomer et al. (1995). See text for explanation.

1984). To briefly summarise, prior to formation of a localised, narrow, but laterally extensive island arc, juvenile crust developed in an extensional, suprasubduction environment, 300–450 km wide. This crust, formed above a steeply "subsiding" lower plate, was characterised by very rapid extension and high magmatic production rates, reflecting rapid trench retreat and decompression melting of upwelling asthenospheric mantle displaced by the falling oceanic slab (Fig. 5). In addition to arc-like chemical signatures, boninitic magmas are a characteristic feature of the Izu–Marianas–Bonin "pre-arc" crust. However, they are practically absent from the Tonga setting, where MORB compositions prevail. Unlike standard MORB, however, the pre-arc crust contains a high proportion of intermediate to felsic lavas and subvolcanic dioritic to granodioritic plutons. The new crust was divided by kilometre-spaced extensional faults into sub-basins floored by arc-like crust with basalt fill. Intermediate to felsic volcanoes (seamounts) developed on the sub-basin shoulders, and voluminous debris flows flowed down the flanks of the volcanic centres contributing to the sub-basin fill. It is important to emphasise here that, because of the rapid magmatism and high heat flow derived from the upwelling mantle, these mafic to felsic volcanic phases were produced penecontemporaneously, throughout the volcanic succession. Extension of the juvenile crust was also accommodated by emplacement of subvolcanic intermediate to felsic plutons, derived by melting of the lower parts of the volcanic pile. Another consequence of the rapid extension was that accretionary wedges did not develop. This phase of extensional, suprasubduction, juvenile crustal growth, with a duration of the order of ∼10 My, is referred to as an "infant" arc stage. In the Southwest Pacific Ocean, the "infant" arc stage ended when the subsiding lower plate reached the base of the asthenosphere and cut off the return flow of mantle material, after which extension ceased, convergent subduction was established, and suprasubduction magmatism was localised along a narrow island arc.

Without necessarily drawing a direct analogy with the "infant" arc model as it has been applied in the Southwest Pacific Ocean, processes attributed to it might be invoked to explain a number of primary features of the assemblage I (∼2710–2690 Ma) volcanic succession in the Central Hearne supracrustal belt. These include: (i) the location of the belt in a wide swath (225 km after deformation) of penecontemporaneous juvenile crust that extends across much of the Hearne domain; (ii) the close, primary intercalation of contemporaneous volcanic rocks of MORB-like and arc-like geochemical signatures throughout the belt, coupled with the disposition of volcanic deposits in discontinuous map units, consistent with fault-delimited, extensional sub-basins; (iii) the lack of a localised, laterally extensive volcanic arc edifice, despite the abundance of intermediate to felsic volcanic deposits; (iv) the development of isolated, independent, felsic volcanic centres throughout the magmatic history of the belt; and (v) the absence of accretionary wedges anywhere within the Hearne domain.

Vertical subsidence of the lower plate in an Archean thermal regime runs counter to the prevailing models of shallow subduction leading to tectonic underplating of continental crust and the accretion of mantle roots invoked for that time (e.g. Helmstaedt and Schulze, 1989; Abbott, 1991; Abbott et al., 1994; Cousens et al., 2001; Zegers and van Keken, 2001). However, other workers have suggested that extensional environments associated with high rates of voluminous magmatism can be generated by asymmetrical delamination and subsidence of sinkers of dense (eclogitised) mafic lower crust (Vlaar et al., 1994; Schott et al., 2000; Zegers and van Keken, 2001). We suggest that processes associated with an "infant arc" model modified by such a mechanism could provide a viable analogue for the Central Hearne supracrustal belt.

Finally, we note that the presence of abundant subvolcanic and/or penecontemporaneous intermediate to felsic group 2 plutons associated with the assemblage II volcanic succession (~2685–2680 Ma) may reflect a change in boundary conditions, termination of the extensional "infant arc" stage consistent with locally identified D1 deformation prior to 2685 Ma (Table 1), and a belated attempt to build some form of localised, arc-like edifice.

7. Conclusions

The geological, geochemical and geochronological characteristics of the Neoarchean Central Hearne supracrustal belt, in particular assemblage I, are not compatible with either classical or modified arc-subduction or oceanic plateau tectonic models. Rather, we suggest that the processes that operated during the early construction phase of the belt are analogous with those associated with an "infant arc" model, suitably modified to allow for slab subsidence under the high heat flow regime of the Neoarchean Earth. Locally identified deformation that separates the early (assemblage I) and later construction phases of the belt (assemblage II) may reflect a switch from an extensional, suprasubduction environment to a shortening regime associated with an attempt to initiate classical subduction and arc construction.

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