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**GEOLOGICAL SURVEY
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SCIENTIFIC AND TECHNICAL REPORTS

RAPPORTS SCIENTIFIQUES ET TECHNIQUES

AN UPPER CRETACEOUS AMMONITE AND INOCERAMIDS FROM THE HONNA FORMATION, QUEEN CHARLOTTE ISLANDS, BRITISH COLUMBIA

Project 800026

A.C. Riccardi
Institute of Sedimentary and Petroleum Geology

Riccardi, A.C., An Upper Cretaceous ammonite and inoceramids from the Honna Formation, Queen Charlotte Islands, British Columbia; in *Current Research, Part C, Paper 81-1C*, p. 1-8, 1981.

Abstract

The lower part of the Honna Formation contains the Turonian bivalve species *Mytiloides labiatus* (Schlotheim) s.l., and *Inoceramus* sp. cf. *I. multiformis* Pergament, whereas its uppermost beds have yielded the presumably Santonian ammonite *Plesiotexanites?* sp. indet. Therefore the formation appears to range from the early Turonian to the Santonian. The presence of *Mytiloides labiatus* s.l. in the upper part of the underlying Haida Formation suggests the absence of a prolonged hiatus between the two formations and confirms their partial interfingering.

Introduction

The Honna and Skidegate formations of the Queen Charlotte Islands have yielded only a few generically and specifically indeterminate ammonite and inoceramid fragments to date. Therefore, they were assigned only a general Late Cretaceous (i.e. post-early Turonian) age by Sutherland Brown (1968, p. 98). The post-early Turonian age of the Honna Formation was indicated by the presence of the "*Inoceramus*" *labiatus* fauna in the immediately underlying Haida Formation (Sutherland Brown, 1968, p. 92).

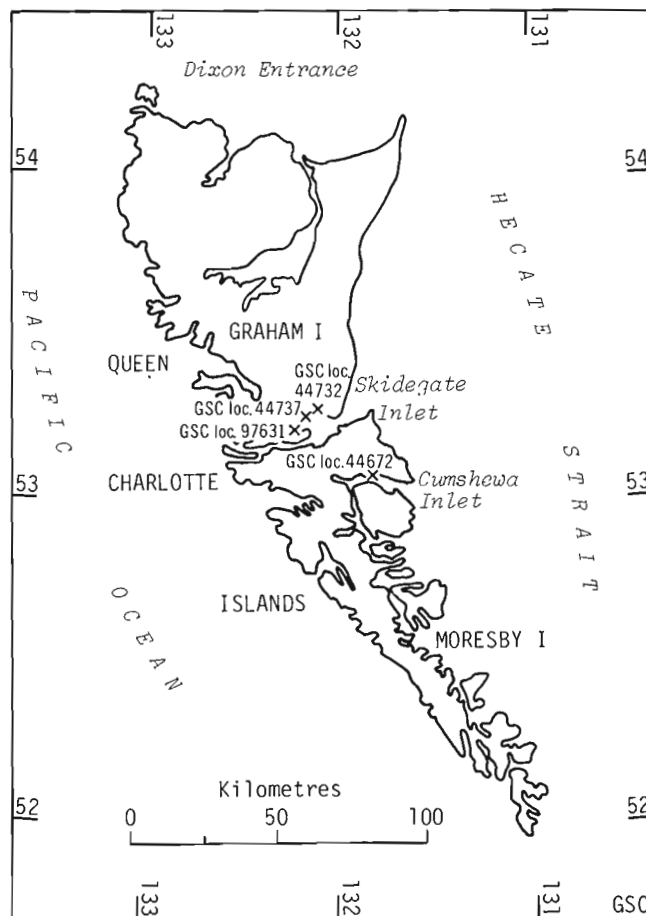


Figure 1.1. GSC localities mentioned in the text.

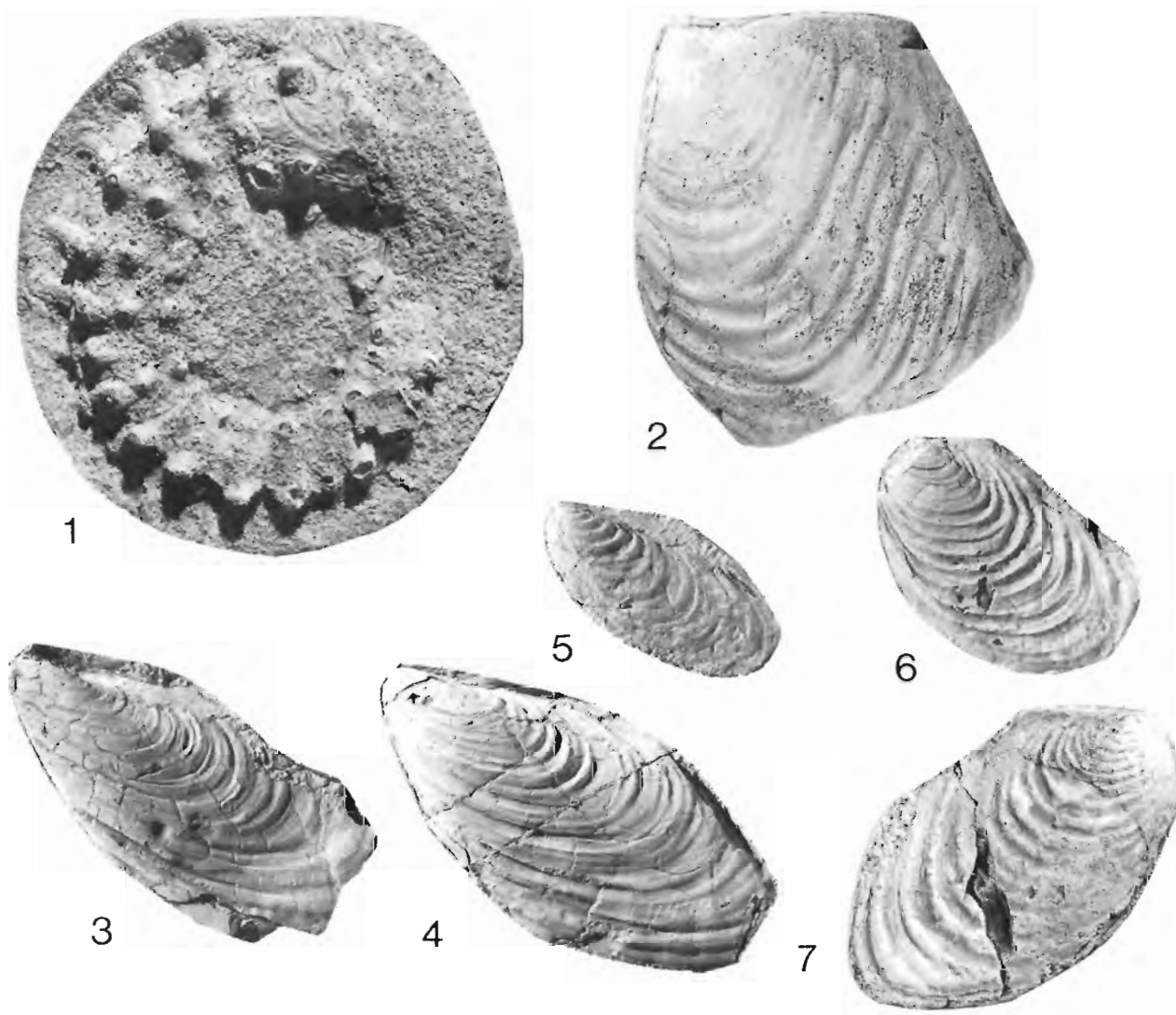


Plate 1.1

Figure 1. **Plesiotechanites?** sp. indet., GSC 67006, uppermost Honna Formation, GSC loc. 97631. Lateral view.

Figure 2. **Inoceramus** sp. cf. **I. multiformis** Pergament, GSC 67007, basal beds of Honna Formation, GSC loc. 44737. Left valve.

Figures 3-7. **Mytiloides labiatus** (Schlotheim) s.l.

3-6, left valves, GSC 67008 to 67011, upper Haida Formation, GSC loc. 44732.

7, right valve, GSC 67012, basal beds of Honna Formation, GSC loc. 44672.

An ammonite mold was recently found by C.J. Yorath in the uppermost beds of the Honna Formation and its rubber cast was submitted for identification to the writer. This cast proved to be generically identifiable as *Plesiotelexites*? and closely datable (probably Santonian). Its study has stimulated a reappraisal of all fossil material previously found in the Honna Formation by Dr. A. Sutherland Brown. An extensive preparation of inoceramids (only generically determined previously) from the lower part of the formation (see Sutherland Brown, 1968, p. 98) resulted in their assignment to the well-known early to mid-Turonian species *Mytiloides labiatus* and *Inoceramus* sp. cf. *I. multiformis*. To verify the somewhat unexpected stratigraphic range of these inoceramid species, the writer has also studied the inoceramid material previously found in the upper part of the type section of the underlying Haida Formation (Sutherland Brown, 1968, p. 88). This material, also identified as *M. labiatus*, is described and figured in this paper for comparative purposes.

Illustrated
Specimens

Fossil Localities

GSC 67008 to 67011	Mytiloides labiatus (Schlotheim) s.l., GSC loc. 44732. Type section of Haida Formation (3390' [1034 m]), Bearskin Bay, locality 29 in Sutherland Brown, 1968, p. 88 and Table IX (notice that on Table IX the GSC loc. is wrongly given as 44739 instead of 44732). A. Sutherland Brown coll. 1960.
GSC 67012	Mytiloides labiatus (Schlotheim) s.l., GSC loc. 44672. Basal beds of Honna Formation, Conglomerate Point, Cumshewa Inlet. A. Sutherland Brown coll. 1960.
GSC 67007	Inoceramus sp. cf. <i>I. multiformis</i> Pergament, GSC loc. 44737. Basal beds of Honna Formation, western Lina Island. A. Sutherland Brown coll. 1960.
GSC 67006	Plesiotelexites ? sp. indet., GSC loc. 97631. Uppermost Honna Formation, 53°12.0'N Lat., 132°10.4'W Long., C.J. Yorath coll. 1980.

PALEONTOLOGICAL DESCRIPTIONS

Class Bivalvia (Buonanni, 1681) Linné, 1758
ORDER PTERIOIDA Newell, 1965
Superfamily PTERIACEA Gray, 1847
Family INOCERAMIDAE Giebel, 1852
Genus **Mytiloides** Brongniart, 1822
Mytiloides labiatus (Schlotheim, 1813 [?]) s.l.
Plate 1.1, figures 3-7

- 1813. **Ostracites labiatus** Schlotheim, Leonhard's Taschenbuch für Mineralogie, v. 7, p. 93 (*fide* Woods, 1911, p. 281, incl. synonym list).
- 1935. **Inoceramus labiatus** var. **mytiloides** Mantell, Seitz, p. 435, Pl. 36, fig. 1-4; Pl. 37, fig. 4-5; text-fig. 2-3.
- 1935. **Inoceramus labiatus** var. **opalensis** forma **elongata** Seitz, p. 458; Pl. 38, fig. 4-6; Pl. 39, fig. 2-4; textfig. 14-15.
- 1935. **Inoceramus labiatus** var. **subhercynica** Seitz, p. 465, Pl. 40, fig. 1-5; text-fig. 17-18.
- 1968. **Inoceramus** sp. A, Sutherland Brown, p. 98 (loc. 29).
- 1968. **Inoceramus** sp., Sutherland Brown, p. 98 (GSC loc. 44672).
- 1977. **Inoceramus labiatus** (Schlotheim), Jeletzky, p. 119.

Material. 4 left (GSC 67008 to 67011), and 1 right (not figured) valves from GSC loc. 44732 (upper Haida Formation); 1 right valve (GSC 67012) from GSC loc. 44672 (basal Honna Formation).

Description. Shell of small to medium size, maximum height about 35 mm, length about 65 mm. Subequivalve, inequilateral, slightly to moderately biconvex. Maximum convexity dorsocentral; anterior and lateral umbonal slope moderately to very steep; shell gradually decreasing in convexity with growth, posterior part flattened.

Shell moderately to strongly prosocline, angle of inclination ranges from about 30° to about 50°, oblique, elongate to ovate in outline, with length greater than height. Beaks anterior, terminal or almost terminal, prosogyrous, slightly projected above hinge line. Posterior auricle small to moderate in size, subtriangular, flattened and separated from umbone by a break in postumbonal slope.

Hinge line straight, ranging from one third to one half of total shell length, an angle of about 90°. Anterior margin curved, passing into a broadly rounded venter; posterior margin nearly straight, forming an obtuse angle with the hinge line and a narrowly curved acute angle with the ventral margin. Juvenile shell similarly elongated or more rounded than adults.

Juvenile and adult ornamentation rather similar, consisting of low, rounded and subevenly to evenly spaced and regularly to subregularly developed rugae, gradually becoming more widely spaced, sometimes becoming weaker on older growth stages. The most unevenly spaced and irregular rugae are present in some of the most inclined specimens, where they can develop two or three riblets or crests in the area of maximum curvature.

Discussion. A systematic review of the *Mytiloides labiatus* lineage from the Americas and Western Europe is now being carried out by E.G. Kauffman (Kauffman, 1975, 1977a,b; 1978a, b; Kauffman in Kauffman and Powell, 1977; Kauffman in Wiedmann and Kauffman, 1978). This worker has already concluded that the "varieties" and "forms" recognized by Seitz (1935) within *M. labiatus* have distinct stratigraphic ranges and should be treated as separate species or subspecies.

Kauffman (in Kauffman and Powell, 1977) has thus recognized five different species within the *M. labiatus* lineage represented in the lower Turonian of northwestern Oklahoma. Furthermore, he (Kauffman, 1977a) has assigned to several different species the material from Japan originally ascribed to "*Inoceramus labiatus* (Schlotheim)" by Matsumoto and Noda (1975).

The morphological range of every one of these "species" based on large, well preserved and stratigraphically closely spaced populations, has yet to be adequately illustrated. Meanwhile, it could be misleading to assign isolated specimens to one or other of the closely related taxa recognized to date within the *M. labiatus* lineage. Therefore, a broad concept of that species has been followed here, all the more as the material from the Queen Charlotte Islands does not seem to fall within the restricted concept of *M. labiatus* (see Woods, 1911, Pl. L, fig. 5; Seitz, 1935, Pl. 38, fig. 1; Kauffman and Powell, 1977, Pl. 7, fig. 5; Jeletzky, 1970, Pl. 26, fig. 6; Kauffman, 1978b, Pl. 4, fig. 9).

The specimens here described and figured are instead similar to some of the taxa originally recognized by Seitz (1935) within *M. labiatus* s.l. For example, those figured on Plate 1.1, figures 3-5 (GSC loc. 44732) retain an elongate outline throughout ontogeny. This outline is similar to that of the *M. mytiloides* (Mantell) (see Seitz, 1935, fig. 2a-f, Pl. 37, fig. 5) and *M. submytiloides* (Seitz, 1935, Pl. 37, fig. 1-3). In these specimens, however, the junction between adult and juvenile ornament seems to be less distinct than in the first named species, while there is no posteroventrally trending sulcus such as occurs in the second species. A similar shape and ornament are also present in *M. duplicostatus* (Anderson, 1958, Pl. 17, fig. 3-4) from the lower to middle Turonian of California, as well as in the Alaskan material described by Jones and Gryc (1960, Pl. 20, fig. 5) as *I. labiatus*.

A more rounded valve (Plate 1.1, fig. 6), with more regular adult ornament, which resembles one of the specimens found in the basal part of the Honna Formation (Plate 1.1, fig. 7) also occurs in the sample from the Haida Formation. These specimens, however, seem to differ in inclination and variation of outline throughout the ontogeny. Thus, the upper Haida specimen (Plate 1.1, fig. 6) is slightly more inclined and has a similarly elongated outline throughout the ontogeny. However, the Honna specimen (Plate 1.1, fig. 7) is less inclined and exhibits a clear change from a rounded juvenile to an elongate mature outline. The first specimen resembles *M. subhercynicus* Seitz (1935, Pl. 40, fig. 3-4), whereas the second is close to *M. opalensis* (Böse) (see Seitz, 1935, Pl. 38, fig. 4-6).

The above mentioned forms of *Inoceramus labiatus* s.l. were recorded in the Lower Turonian, except for the *Mytiloides subhercynicus*, which is also present in the lower part of the Middle Turonian.

The specific nature and the early Turonian age of Sutherland Brown's (1968) material derived from the upper beds of the Haida Formation were previously pointed out by Jeletzky (1977, p. 119).

Genus *Inoceramus* J. Sowerby, 1814

Inoceramus sp. cf. *I. multiformis* Pergament 1971

Plate 1.1, figure 2

1968. *Inoceramus* sp. Sutherland Brown, p. 98 (GSC loc. 44737).

1971. cf. *Inoceramus multiformis multiformis* Pergament, p. 61, Pl. IX, fig. 2, 3; Pl. X, fig. 2, 3; Pl. XI, fig. 1, 2; Pl. XII, fig. 1-5; Pl. XXXIII, fig. 3.

Material. One left valve, GSC 67007 from GSC loc. 44737. Basal Honna Formation.

Description. Shell of medium size, height about 62 mm, length about 53.5 mm. Inequilateral, moderately convex. Maximum convexity near the anterior margin, in coincidence with the line of maximum inflexion of the concentric ornament; anterior slope steep, posterior gentler. Shell evenly curved along the growth axis.

Shell suberect, angle of inclination about 70°, ovate-subrectangular in outline, slightly higher than long. Beak anteriorly terminal, narrowly rounded, prosogyrous, slightly projected above hinge line. Posterior auricle moderate in size, subtriangular, continuous with postumbonal slope.

Hinge line relatively long, straight, amounting to about two thirds of the total shell length, forming nearly a right angle with the anterior margin. Anterodorsal margin almost straight, becoming curved to join the rounded ventral margin, which passes into the broadly curved posterior margin. The latter forms an obtuse angle with the hinge line.

Juvenile and adult ornament similar, consisting of about 22 coarse, widely spaced, and rather regular concentric rugae, which are stronger and wider in the central part of the valve. The rugae have an asymmetric curvature, with the point of maximum inflexion close to the anterior margin with which they merge at very low angles. On the posterior part of the valve the rugae are weakly inclined towards the margin but they bend forward near the hinge line, which they reach forming nearly a right angle.

Discussion. The specimen here described is close to the material from the Coniacian of the Pacific Coast of the USSR that Pergament (1971, Pl. IX, fig. 3; Pl. X, fig. 2-3; Pl. XII, fig. 1-5; Pl. XXXIII, fig. 3) has included in *Inoceramus multiformis multiformis* Pergament. The Canadian specimen, however, seems to have a smaller inflation, less prominent umbo and more regular concentric rugae.

A related species seems to be *Inoceramus waltersdorfensis* Andert (1934, p. 112, Pl. 4, fig. 2, 3, 4-7) from the highest Turonian and lower Coniacian of Euramerica. However, as pointed out by Pergament (1971, p. 64), the inflation of Andert's species seems to be larger and the curvature of its concentric rugae is more asymmetric. Nevertheless, the two are connected by transitional specimens (cf. Tröger, 1967, Pl. 12, fig. 1, 3; Pergament, 1971, Pl. 12, fig. 1, 4; Kauffman, 1978b, Pl. 1, fig. 9; 1978c, Pl. 2, fig. 2, 10).

Inoceramus multiformis also bears some resemblance to *Inoceramus lusatie* Andert, from the late Turonian of Europe (see Andert, 1934, fig. 14a-b; Tröger, 1967, Pl. 8, fig. 3b), although the latter seems to have a less asymmetric outline with a projected anterior margin (see Andert, 1934, Pl. 7, fig. 2; Kauffman, 1978b, Pl. 2, fig. 5), and a finer and more irregular ornament.

Kauffman (1977a, p. 178) has considered *I. multiformis* to be a junior subjective synonym of *I. (I.) teshioensis* Nagao and Matsumoto (1939, Pl. 24, fig. 6, 7, 9; Pl. 26, fig. 7; Noda, 1975, Pl. 35, fig. 2-7) from the late Turonian and (?) lowest Coniacian of Japan. However, as already pointed out by Pergament (1971, p. 64), his species seems to differ in the extended wing and outline of the concentric ornament. The same differences appear to be present in the European material figured by Tröger (1967, Pl. 9, fig. 1-9; Pl. 10, fig. 3), and Heine (1929, fig. 13), but not in that of Andert (1934, Pl. 4, fig. 10), which Kauffman (1977a, p. 178) included in *I. (I.) teshioensis* Nagao and Matsumoto.

The material available to the author is insufficient to take a definite stand on this issue. Furthermore, a change of specific names would not substantially change the age of the Canadian material. Therefore, Pergament's name is used here tentatively.

The type material of *I. multiformis* Pergament comes from the Coniacian of the Pacific coast of the USSR. The same or related species seems to occur in the Upper Turonian and Lower Coniacian of Euramerica (Kauffman, 1977a, p. 178).

The specimen described here was found at an isolated locality. However, it could indicate a slightly younger age for the basal part of the Honna Formation than the above described specimens from GSC loc. 44672 included in *M. labiatus* s.l.

SUBCLASS AMMONOIDEA Zittel, 1884
Superfamily ACANTHOCERATACEA de Grossouvre, 1894
Family COLLIGNONICERATIDAE Wright and Wright, 1951
Subfamily Texanitinae Collignon, 1948
Genus *Plesiotexanites* Matsumoto, 1970
Plesiotexanites? sp. indet.
Plate 1.1, figure 1

Material. Rubber cast of a whorl fragment. GSC 67006, from GSC loc. 97631. Uppermost Honna Formation.

Description. The specimen is about 63.4 mm in diameter and consists of one very evolute whorl ($U/D=0.46$). The line of overlap with the preceding (not preserved) whorl apparently lies immediately outside of a row of tubercles which is located in a ventro-lateral position. The whorl section appears to be subquadrate and perhaps as wide as high, with the broadest part in the middle of the flanks. The umbilical wall is low and somewhat inclined.

Ornament consists primarily of 3 to 4 rows of nodes, which are superposed on low, flat and very weak ribs. The ribs are almost straight on the lower part of the flanks and acquire a slight forward projection on the ventro-lateral shoulder. The interspaces are two times wider than the ribs. There are about 12 ribs on the last half whorl, and the entire whorl seems to have approximately 20 ribs.

The innermost whorls are not preserved. The earliest part of the preserved whorl with the diameter of 45 mm exhibits 3 rows of nodes. Although the venter is not visible these rows are interpreted respectively as the umbilical, ventro-lateral and ventral (umbilical, marginal and external of Collignon's 1948 nomenclature). The umbilical tubercles are located on the umbilical shoulder and are rounded and relatively small. The ventro-lateral (or marginal) nodes are more prominent and separated by a concave and wide depression from the ventral ones, which are almost spinose. At a diameter of about 45 mm, lateral tubercles begin to differentiate just below the marginal ones. Farther adorally, the lateral tubercles increase in size and shift into a more median position on the

flanks. Finally, at the greatest preserved diameter the four rows of nodes are nearly equidistant and almost equal in size. Up to 50 mm in diameter there are some occasional secondary ribs intercalated on the upper part of the flank, which also bear ventro-lateral (marginal) and ventral (external) nodes.

The suture line is not visible.

Dimensions (in mm)	Diameter		Umbilical Width (% of Diameter)		Height of Whorl (% of Diameter)	
	63.4		29	(0.46)	c. 21.2	(0.33)
	47.4		23.8	(0.50)	13.6	(0.29)
	43.7		20.6	(0.47)	13.3	(0.30)

Discussion. All described features are those of the Acanthocerataceae, within which the closest affinities are with the Texanitinae. The systematics of these ammonites have been dealt with by Collignon (1948), Matsumoto (1955, 1970), and Klinger and Kennedy (1980). These workers have mainly relied on the number of tubercle rows and the order in which they develop throughout the ontogeny for the recognition of texanitid genera and species.

The presence in our specimen of 3 rows of nodes, followed by the development of a 4th (lateral) row are features characteristic of **Protexanites** (**Anatexanites**) Matsumoto. However, all species included in the subgenus **Anatexanites**, i.e. **P. (A.) fukazawai** (Yabe and Shimizu, 1925, Pl. 30, fig. 1; Pl. 31, fig. 1-2; Pl. 33, fig. 1-2), **P. (A.) nomii** (Yabe and Shimizu, 1925, Pl. 32, fig. 1-3) and **P. (A.) reymonti** Matsumoto (1970; see Reymont, 1955, Pl. 23, fig. 3, text-fig. 46a), have a more prominent ribbing combined with relatively less conspicuous tubercles.

Paratexanites Collignon (1948, see Klinger and Kennedy, 1980, p. 13-19) has four rows of tubercles, umbilical, submarginal, marginal and external, in the adult stage. However, it does not exhibit lateral tubercles and its submarginal and marginal tubercles are approximated, sharing a single base, in the early growth stages.

The Canadian specimen could correspond to the inner whorls of a large species which would acquire a pentatuberculate stage in the outer whorls. If that were the case it would have to be included in **Plesiotexanites** Matsumoto. Such assignment is supported by the circumstance that some species, e.g. **Plesiotexanites thompsoni** (Jones, 1966) from the Early Santonian of California, and **P. yezoensis** Matsumoto (1970, p. 99, fig. 79-83) from the Santonian of Japan and South Africa exhibit an intermediate growth stage similar to that found in **Protexanites** (**Anatexanites**) due to the delayed doubling of the ventrolateral tubercles compared with the appearance of the lateral tubercles.

P. thompsoni and **P. yezoensis**, however, have more prominent and numerous ribs than our specimen, and their umbilical and lateral tubercles are relatively smaller while the ventrolateral (marginal) become more prominent. **P. matsumotoi** has relatively stronger and spinose ventrolateral tubercles combined with better defined ribs.

The specimen from the Honna Formation has a general resemblance to the inner whorls of other Santonian species of **Plesiotexanites**, such as **P. stangeri** (Baily) and **P. collignoniforme** Klinger and Kennedy (see Young, 1963, Pl. 42, fig. 3-4; Pl. 43, fig. 2-4; Pl. 71, fig. 1-4; Collignon, 1966a, Pl. 484, fig. 1958; Kennedy et al., 1973, Pl. 5, fig. 2; Klinger and Kennedy, 1980, fig. 68, 70), although in those species the ventrolateral tubercles split into two rows (marginal and submarginal) before the lateral tubercles begin to appear.

In the evolute whorls and strong tubercles the Canadian specimen exhibits some superficial resemblance to some **Texanites** Spath, such as **T. americanus** (Lasswitz, 1904, Pl. 8, fig. 1) from the lower to middle Santonian of the Gulf Coast, **T. texanus** (Roemer, 1852, Pl. 3, fig. 1a-c; Collignon, 1966b, Pl. 33, fig. 1; Klinger and Kennedy, 1980, fig. 123-124) from the Santonian of North America and Africa, **T. rarecostatus** Collignon (1966a, Pl. 487, fig. 1965; Klinger and Kennedy, 1980, fig. 121, 122) from the Santonian of South Africa and Madagascar. However, **Texanites** is characterized by the presence of five rows of tubercles from a relatively early stage (see also Haas, 1942; Collignon, 1966a; Klinger and Kennedy, 1980).

Considering that the venter of the specimen from the Honna Formation is unknown, the existence of yet another row of ventral tubercles cannot be completely ruled out. In such a case the rows that are here regarded as marginal and external should instead be interpreted as submarginal and marginal. This conclusion would only confirm the placement of our specimen into **Plesiotexanites**, unless the number of tubercles in the external row would be larger than in the other two rows, as happens in **Menabites** and **Bevahites** of Collignon (1948).

However, **Menabites** has a trituberculate stage persisting almost to maturity, and the pentatuberculate stage follows almost immediately. Similarly, in **Bevahites** a pentatuberculate stage is attained early in the ontogeny. Furthermore, this last genus is characterized by the proximity of submarginal and marginal tubercles, and due to bifurcations of ribs it has many more external than internal tubercles (see Collignon, 1948; Klinger and Kennedy, 1980).

A tubercle development similar to that described herein is also found in "**Menabites**" **walnutensis** Young (1963, p. 109, Pl. 58, fig. 1, 4) from the (?) lower Campanian of Texas, which also agrees in size, evolution and preponderance of tubercles over ribs. This species, however, seems to have fewer

ribs in the last whorls, i.e. 8 versus 12 on the last half whorl, while the external tubercles are slightly clavate and more numerous than the ventrolateral (marginal), i.e. 13 versus 9 in the last half whorl. Despite the increase in the number of external nodes this species has inner whorls with tubercle development of **Plesioteanaxites**.

The Texanitidae are known from the lower Coniacian to the middle Campanian, being most widely represented in the Santonian (see Matsumoto and Haraguchi, 1978; Klinger and Kennedy, 1980). **Plesioteanaxites** is known from the Santonian to the lower Campanian, while **Protexanites** (**Anatexanites**) seems to be restricted to the middle Santonian.

Texanitids are relatively rare in the Upper Cretaceous of the Pacific coast of North and South America (see Benavides Caceres, 1956), while they are abundant in the Gulf Coast, Madagascar, South and Eastern Africa, Europe and Japan. From the Pacific Coast of North America the following species have been described: **Submortonicer** **chicoense** (Trask) from the Campanian Chico Formation of California (see Matsumoto, 1959) and Ganges Formation, Nanaimo Group of British Columbia (Ward, 1976); **Plesioteanaxites** cf. **P. kawasakii** (Kawada) from the Santonian of California (Matsumoto, 1959) and **Plesioteanaxites thompsoni** (Jones) from the Lower Santonian Funks Formation of California (Jones, 1966; Matsumoto and Haraguchi, 1978).

Ages of the Faunas and the Age Limits of the Honna Formation

The affinities of **Plesioteanaxites?** sp. indet. suggest a Santonian (s.l.) age for the uppermost beds of the Honna Formation, although the known time range of the Texanitinae, i.e. early Coniacian-middle Campanian leaves an outside possibility of these beds being either slightly older or younger.

The inoceramids found in the basal part of the Honna Formation indicate a Turonian age. Inoceramids of the early-middle Turonian **Mytiloides labiatus** lineage described here occur about 117 m below the top of the underlying Haida Formation (Sutherland Brown, 1968, p. 88) and at the base of the Honna Formation (Sutherland Brown, 1968, p. 98). These data indicate a partial interfingering of the Haida and Honna formations, as was already suggested by Sutherland Brown (1968, p. 84, 93, 98). They appear to contradict the existence of a prolonged hiatus between these formations recently suggested by Jeletzky (1977, p. 121, fig. 4).

All above data suggest an early Turonian-Santonian age for the Honna Formation, unless a paraconformity occurs somewhere between the known fossiliferous levels.

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Abstract

The pre-Triassic rocks of Grand Manan consist of intensely deformed sedimentary and volcanic rocks, a small area of less deformed volcanic and intrusive rocks, and an isolated undeformed granite. Five generations of cleavage and four generations of folding can be recognized in the intensely deformed sedimentary and volcanic rocks. S_1 foliation is subparallel to primary layering and F_1 folds are lacking. S_2 , S_3 , and S_5 cleavages trend northwest. F_4 folds associated with westward-dipping S_4 cleavage verge eastward. Lithological correlation with mainland rocks suggests that the sedimentary and volcanic rocks are Cambro-Ordovician, whereas the volcanic-intrusive rocks and the granite are probably Silurian and Late Silurian respectively. Structural correlation with the mainland suggests that the polyphase deformation is either pre-Late Silurian or post-Carboniferous. Posttectonic metamorphism associated with the granite appears to preclude the post-Carboniferous age of the deformation.

Introduction

The western part of the island of Grand Manan exhibits a thick Triassic succession of flat-lying tholeiitic lavas and sills overlying red siltstones and sandstones. The pre-Triassic rocks, exposed on the eastern side of the island and on smaller islands offshore, consist of intensely deformed sedimentary and volcanic rocks; less deformed volcanic and intrusive rocks outcrop north of a fault at North Head. Just south of the mapped area (Fig. 2.1), undeformed granite is exposed on Three Islands. The ages of the rocks are uncertain. Obscure fossils have been recorded near North Head (Bailey, 1917). Granite cobbles in a conglomerate horizon interbedded with the volcanic rocks on Ross Island (Alcock, 1948a) have yielded K-Ar ages of 600 and 650 Ma (Leech et al., 1963; recalculated using constants suggested by Steiger and Jäger, 1977); the conglomerates are deformed, and the age of the cobbles is probably greater than the isotopic dates.

In the absence of diagnostic fossils or a comprehensive program of age dating, the rocks and the deformation can be dated by lithological and structural correlation with rocks of known age on the adjacent mainland. Alcock (1948a) correlated the sedimentary and volcanic rocks with the Green Head and Coldbrook groups respectively of the Saint John area, both of which are Precambrian. Re-examination of the rocks of Grand Manan casts doubt on Alcock's correlations. In this paper, we outline the lithological succession and describe the deformation of the sedimentary and volcanic rocks. The age and deformation of the pre-Triassic rocks relative to other sequences in southern New Brunswick and southeastern Maine are then examined.

Lithological Succession

The oldest rocks consist of interbedded black pelite and quartzite exposed in the vicinity of The Thoroughfare and northern Ross Island (Fig. 2.2a). The quartzites on Ross Island are continuous through Gull Islet and several smaller islands and reefs with quartzites on Nantucket Island and near Woodwards Cove. The abundance and bed thickness of quartzite increase toward volcanic rocks which surround and unconformably succeed the older sedimentary assemblage. On Nantucket Island and the east side of Whitehead Harbour, volcaniclastic rocks contain quartzite pebbles identical to the underlying quartzite. Although erosion of the quartzites occurred during deposition of the younger rocks, the unconformity between the two assemblages does not appear to be appreciably angular.

The volcanic assemblage consists of volcanic flows, pillowed lavas, hyaloclastites and volcaniclastic sandstones and siltstones. The clastic sedimentary rocks include conglomerate, sandstone and siltstone derived from a nonvolcanic provenance. This assemblage comprises all of Alcock's (1948a) Coldbrook Group south of North Head and much of his Green Head Group rocks.

On Ross, Cheyney and White Head islands, the succession consists mainly of pillowed lavas, lava flows, hyaloclastites and minor volcaniclastic sediments (Fig. 2.2a). Farther to the northwest between Grand Harbour and Red Head the rocks are mainly volcaniclastic sediments with

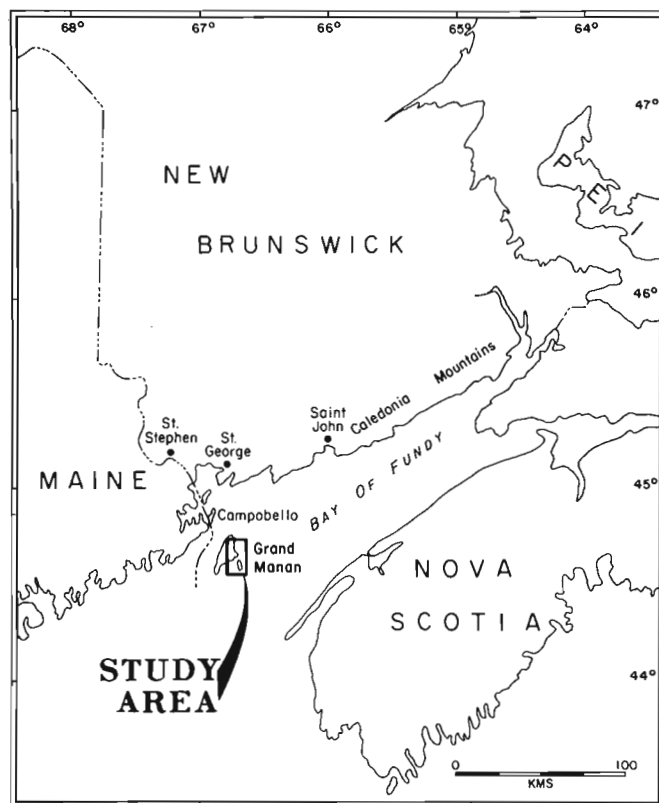


Figure 2.1. Map showing area investigated on east side of Grand Manan and localities referred to in the text.

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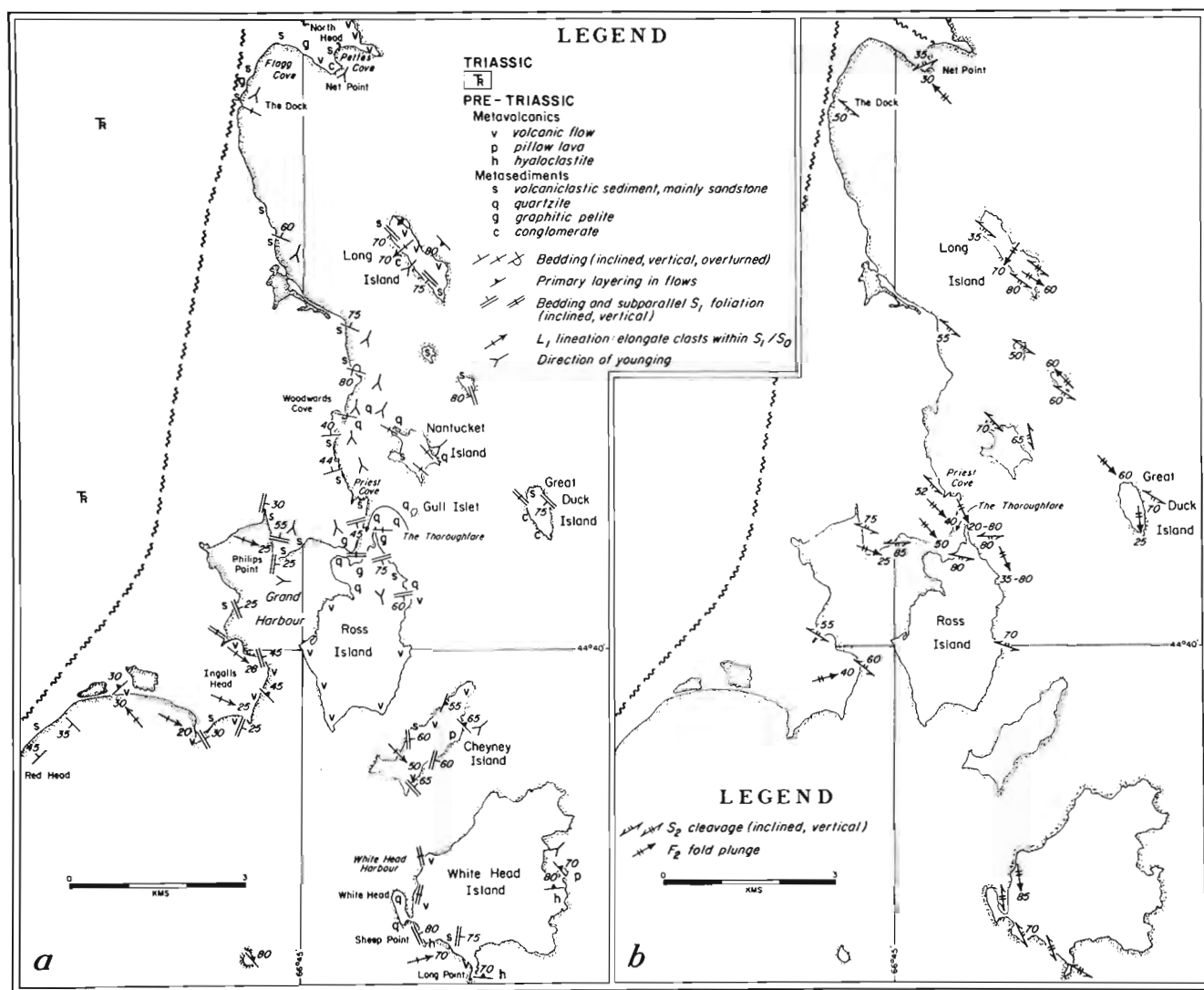
minor volcanic rocks, while between Priest Cove and The Dock volcaniclastic sandstones with minor interbedded siltstones and pelites form a northward younging sequence approximately 4000 m thick. The volcaniclastic sediments are considered to be stratigraphic equivalents of the predominantly volcanic succession to the south on Ross, Cheyney and White Head islands. Predominantly pelitic sediments, conglomerates and volcanic rocks, striking northwest in a belt which includes Great Duck Island, Long Island, and the coast between The Dock and Net Point, possibly represent the youngest part of the succession.

The conglomerates form lenticular or channelized deposits which contain clasts of (a) quartzite and granite (Alcock, 1948a), deformed granite and gneiss (Ross Island), (b) quartzite, argillite and limestone (south of Flagg Cove, Leavitt, 1963), (c) argillite and quartzite (The Dock), and (d) various intraformational lithologies (Great Duck and Long islands, Net Point). The gneissic clasts resemble the Brookville gneiss at the core of the Precambrian succession at Saint John (Currie et al., 1981). Some of the quartzite and

argillite fragments derive from the older sedimentary assemblage. Presumably the instability indicated by the conglomerates reflects the commencement and continuation of volcanism.

The volcanic and intrusive rocks at North Head differ lithologically from the rocks to the south, and closely resemble the Silurian volcanic-intrusive suite in the St. George area of southwestern New Brunswick (Donohoe, 1978). Potter et al. (1968) likewise considered these rocks to be Silurian, but they also assigned a Silurian age to the volcanic rocks farther south.

The undeformed granite on Three Islands is mineralogically and texturally similar to the early Acadian granites of New Brunswick. A Late Silurian K-Ar age of 419 Ma (Leech et al., 1963, recalculated) from petrographically similar granite on East Wolf Island (Alcock, 1948b), 20 km north of Grand Manan, strongly suggests that the Three Islands granite is part of the Acadian plutonic suite (Fyffe et al., 1981).



- bedding and primary layering, composite S_1/S_0 foliation (subparallel S_1 foliation and bedding), and L_1 lineation; the direction of younging of beds and primary layering, and the lithology of sedimentary and volcanic rocks, are also shown.
- S_2 cleavage and F_2 folds.

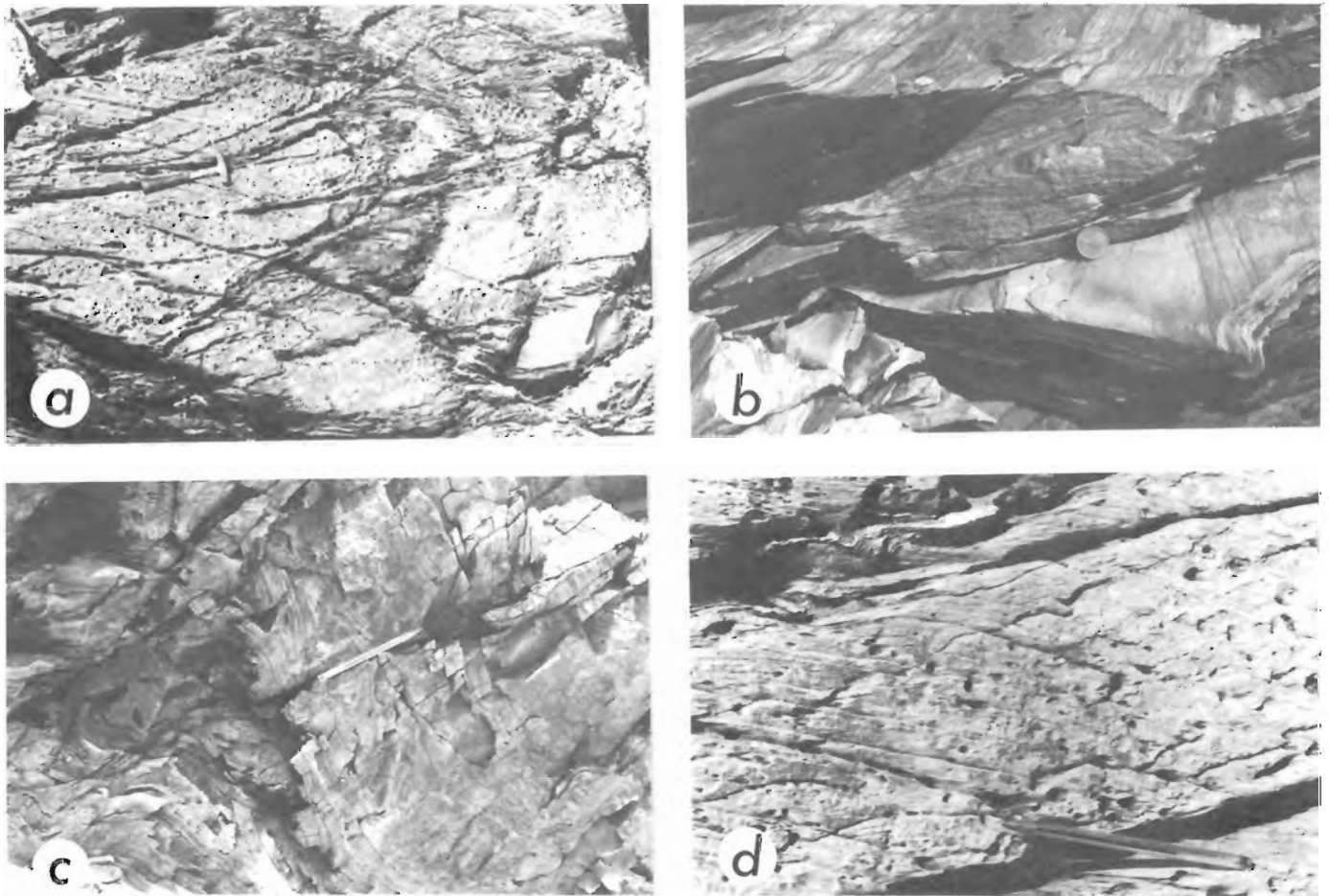
Figure 2.2. Maps showing the orientation and distribution of D_1 and D_2 minor structures in pre-Triassic rocks of Grand Manan.

Deformation of the Sedimentary and Volcanic Rocks

The dip and strike of bedding in the sedimentary rocks and of primary layering in the volcanic rocks show considerable variation (Fig. 2.2a). The S_1 foliation is subparallel to bedding and primary layering, and no major F_1 folds have been recognized. Changes in the dip and strike of bedding and subparallel S_1 foliation are attributed to local folding during D_2 to D_5 deformations, but may be due in part to post- D_1 /pre- D_2 folding. No major F_2 folds have been identified, and a relatively constant regional strike of S_2 cleavage (Fig. 2.2b) suggests that major F_3 , F_4 , and F_5 folds are also absent.

The polyphase sequence of minor structures is most completely developed in the predominantly pelitic sedimentary rocks exposed at the following localities; Net Point and Flagg Cove, west side of Long Island, Great Duck Island, The Thoroughfare and northern Ross Island, Philips

Point, and White Head Island between Sheep Point and Long Point. Five phases of deformation have been established on the basis of overprinting relationships, and of characteristic style and orientation of minor structures formed during each phase. The D_1 deformation is represented by S_1 foliation subparallel to bedding. Widespread D_4 structures and locally developed D_5 structures can be correlated from outcrop to outcrop on the basis of characteristic style and orientation. Structures which deform the S_1 foliation and are deformed by D_4 and D_5 structures have been designated as either D_2 or D_3 on the basis of overprinting relationships, but the correlation of D_2 and D_3 structures from outcrop to outcrop is often difficult due to local variation in the original style of D_2 and D_3 structures and to deformation and disorientation by later structures; the structures ascribed to D_2 and D_3 deformations may represent more than two generations.



- a. East-west cross section of F_2 fold, S_2 cleavage and S_3 cleavage at Net Point. S_2 cleavage (parallel to hammer handle) dips gently northwestward in a reclined F_2 fold plunging north-northwest at 20° . F_2 fold deforms interbedded psammitic and pelitic sediments and subparallel S_1 foliation. S_3 cleavage inclined north-northeast at 45° shows apparent dip to the east (right).
- b. Up-plunge view of reclined tight F_2 fold plunging southeastward at 40° in tuffaceous sediments. Fold deforms bedding and subparallel S_1 foliation. S_2 cleavage partings are discernible in the hinge and upper limb of the F_2 fold. Northwest side of The Thoroughfare.
- c. Asymmetrical tight F_2 fold (near pen) plunges $125^\circ/30^\circ$ in the limb of an F_3 fold in tuffaceous sediments. Spaced S_3 cleavage (parallel to pen) dips southwestward at 45° , cutting through the F_2 fold. Northwest side of The Thoroughfare.
- d. Closer view of S_3 cleavage (parallel to pen) immediately east of the hammer-head in Fig. 2.3a.

Figure 2.3. D_2 and D_3 minor structures.

D₁ Deformation

The S₁ foliation is defined by an alignment of fine grained, mainly sericitic micaceous minerals and elongate quartz grains which constitutes a penetrative fabric subparallel to bedding in most of the sedimentary rocks. In the volcanoclastic sandstones along the coast between Priest Cove and The Dock, the S₁ fabric appears to be absent. At several localities in the volcanic rocks, spaced platy cleavage partings subparallel to the primary layering are interpreted as S₁ foliation. Inequant volcanoclastic fragments oriented parallel to the S₁ foliation appear flattened within S₁ and are in places markedly elongate, forming a lineation (L₁) which trends predominantly northwest (Fig. 2.2a) within the composite S₁/S₀ surface. F₁ minor folds have not been observed. The S₁ foliation is interpreted as a deformational fabric formed during low grade metamorphism.

D₂ Deformation

The S₂ cleavage varies from a crenulation cleavage to a locally developed foliation penetrative at grain scale. The crenulation cleavage is defined by microfolds of the composite S₁/S₀ foliation, and by films of dark, very fine grained irresolvable minerals which form closely spaced S₂ cleavage partings along the limbs of the microfolds. The S₂ foliation is defined by an alignment of micaceous minerals and elongate quartz grains which overprints and largely obliterates the microfolded S₁ fabric. In thin sections of dark graphitic pelite with interbedded pale silty laminae at northern Ross Island and The Thoroughfare, S₂ crenulation cleavage in the pelitic layers changes to S₂ foliation in the silty laminae. The S₂ cleavage strikes predominantly between west-northwest and north-northwest, and dips steeply toward the northeast or southwest (Fig. 2.2b). The orientation of the S₂ cleavage varies locally due to later folding. At Net Point, S₂ cleavage dips gently to moderately northwestward associated with reclined F₂ folds which appear recumbent in cross section (Fig. 2.3a). F₂ folds are tight (Fig. 2.3a, b, c), asymmetrical, and mostly plunge steeply (Fig. 2.2b). Between Priest Cove and The Dock, F₂ folds are lacking and S₂ cleavage is rare.

D₃ Deformation

Minor structures ascribed to D₃ deformation are largely restricted to The Thoroughfare and Flagg Cove. The S₃ cleavage is a crenulation cleavage in which earlier planar structures are microfolded between closely spaced films. It is locally transitional to foliation penetrative at grain scale. The S₃ cleavage strikes northwest and is mainly subvertical (Fig. 2.4). F₃ folds are upright, open to tight, and symmetrical or slightly asymmetrical, and they deform F₂ folds in the vicinity of The Thoroughfare (Fig. 2.3c), at the south side of Flagg Cove, and at Philips Point. The F₃ folds mostly plunge gently, but are locally subvertical where rotated by F₄ folds. Northwest striking, subvertical crenulation cleavage that cuts S₂ cleavage and is in turn cut by S₄ cleavage on Long, Cheyney and White Head islands is designated S₃ although associated F₃ folds are lacking. At Net Point, S₃ crenulation dips relatively gently northward and is superimposed on reclined F₂ folds and S₂ cleavage (Fig. 2.3a, d).

D₄ Deformation

The S₄ cleavage is defined by closely to coarsely spaced (1 to 30 mm) partings between which earlier planar structures are crenulated (Fig. 2.5a). It often occurs in discrete zones (Fig. 2.5b), and penetrates pelitic and psammitic sediments, and locally volcanic rocks. Intense S₄ cleavage is present in thick bedded volcanoclastic sandstones at The Dock and in massive volcanic rocks on the east side of Long Island.

The S₄ cleavage dips gently to moderately westward (Fig. 2.5c) and is associated with open to tight asymmetrical F₄ folds that persistently verge eastward. The regular orientation of S₄ cleavage suggests that the variation in F₄ fold plunge (Fig. 2.5c) is largely due to pre-D₄ variation in dip and strike of the earlier planar structures.

D₅ Deformation

The S₅ cleavage is defined by coarsely spaced (5 to 50 mm) partings between which earlier planar structures are slightly to moderately crenulated. It is present in only a few localities, and is subvertical with a west to northwest strike (Fig. 2.5c). At Priest Cove, S₅ cleavage occurs in volcanoclastic sandstones. It is present in sedimentary and volcanic rocks on Long Island and White Head Island associated with gentle upright F₅ folds in which the S₄ cleavage is rotated into gentle northeastward dips. Local variations in the strike of S₄ cleavage throughout the area may be due to broad F₅ folding.

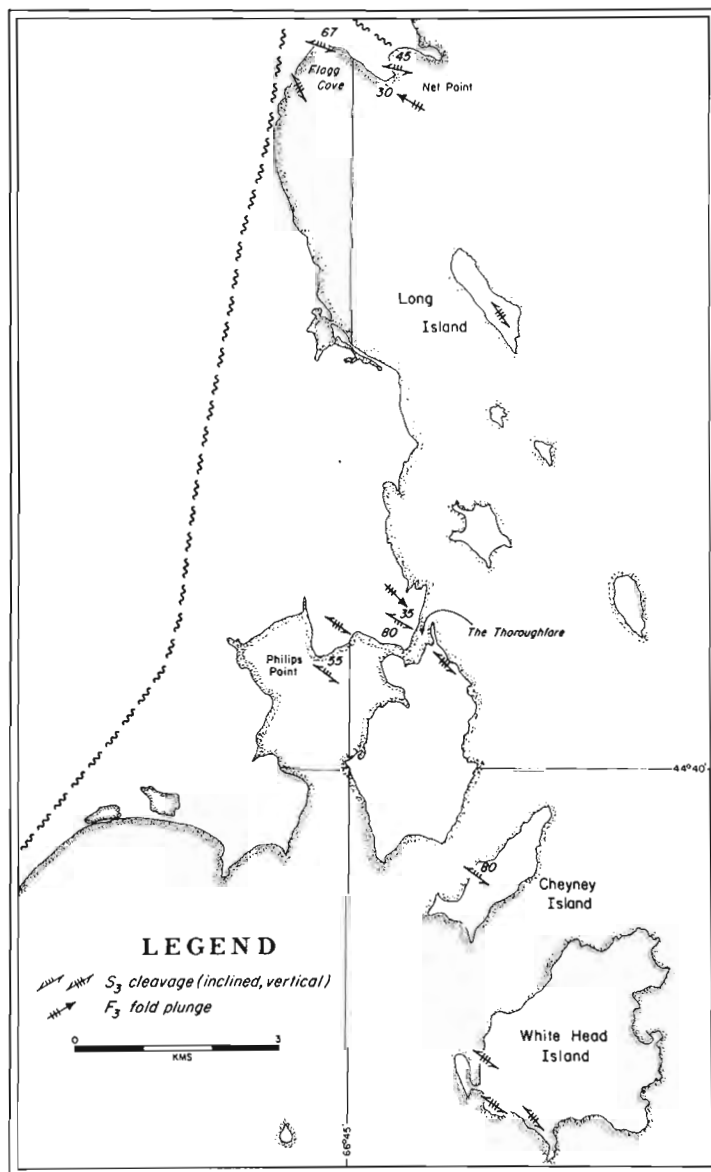
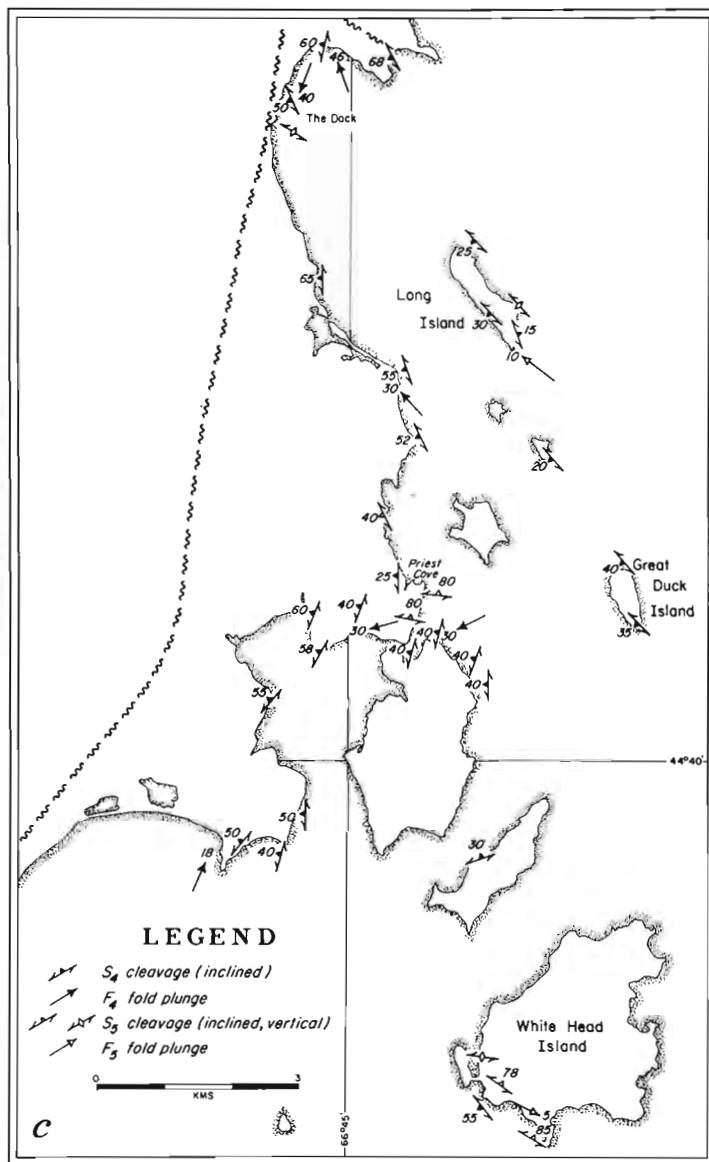


Figure 2.4. Orientation and distribution of S₃ cleavage and F₃ folds.

Figure 2.5. D_4 and D_5 minor structures.

- Coarsely spaced S_4 cleavage (parallel to hammer handle) dips west-northwest at 40° in pelitic sediments, deforming S_1 foliation dipping steeply southward (left), 1 km northeast of Philips Point.
- Closely spaced S_4 cleavage strikes 018° and dips 45° west (right) in pelitic sediments, and forms a discrete zone which cuts through S_2 cleavage (parallel to hammer handle) dipping southward at 70° . East shore at northern tip of Ross Island.
- Orientation and distribution of S_4 cleavage and F_4 folds, and S_5 cleavage and F_5 folds.



Metamorphism

Chloritoid crystals 0.1-0.5 mm in length are locally abundant in pelitic sediments at the south end of Long Island and Great Duck Island, and in graphitic sediments at The Thoroughfare and northern Ross Island. In the latter localities the chloritoid crystals overprint S_2 cleavage films. At the northern tip of Ross Island chloritoid crystals overprint F_3 microfolds and S_3 cleavage films. On Long Island and Great Duck Island chloritoid laths and rosettes overprint F_4 microfolds and S_4 cleavage films. Crystallization of chloritoid evidently succeeded D_4 deformation. Time relations between chloritoid and D_5 deformation have not been observed. The rosettes of chloritoid suggest that the mineral crystallized under static conditions, which may have succeeded D_5 deformation. Hornfelsic metamorphism of the rocks appears to increase southward toward the granite exposed on Three Islands.

Correlatives on the Mainland

Correlation of the pre-Triassic rocks of Grand Manan with rocks on the adjacent mainland must take into account the lithological succession, the structural history and style, and the metamorphic and intrusive history. Based on these criteria, correlation with the Green Head and Coldbrook groups (Alcock, 1948a) is improbable. The Grand Manan succession comprises a lower quartzite and graphitic pelite assemblage, overlain disconformably or with angular unconformity by an assemblage containing predominantly basic flows, hyaloclastites and volcanoclastics. Both parts of the succession exhibit five phases of deformation. By contrast the type Green Head Group localities around Saint John are dominated by marble (Alcock, 1938), while the Coldbrook Group consists almost entirely of acid volcanics with only minor amounts of sedimentary material. There is a major unconformity between the two successions, representing metamorphism and diapirism followed by erosion

(Currie et al., 1981). This unconformity has also been identified from differences in deformation of the rocks (Alcock, 1938; Rast et al., 1976a; Wardle, 1978).

The sedimentary and volcanic succession is most closely correlative to Cambro-Ordovician rocks on the mainland. The Cambrian(?) and Ordovician Cookson Formation of southwestern New Brunswick and southeastern Maine comprises massive quartzite grading upward into graphitic pelite and intercalated quartzite, succeeded by thick bedded well-graded sandstone and graphitic pelite, gradationally overlain by argillaceous sandstone and pelite; pillowed and massive basalt with black slate and chert interbeds, tuffaceous basalt, and sparse calcareous sandstone form the youngest rocks of the succession (Ruitenberg and Ludman, 1978; Ludman, 1981). Preliminary chemical analyses of pillowed lavas from the Cookson Formation 3 km southeast of St. Stephen resemble those of the pillowed sequence on Grand Manan.

Examining mainland correlatives of the polyphase deformation on Grand Manan suggests two possibilities. The Cookson Formation exhibits a steep S_1 foliation trending predominantly northeast, and bedding and foliation are parallel even where folding is complex (Amos, 1963). Ruitenberg (1968) recorded well developed S_1 slaty cleavage, mainly subparallel to bedding, but F_1 folds occur in places mainly in quartzite layers. Asymmetrical F_2 folds verge southeastward accompanied by gently inclined S_2 crenulation cleavage (Ruitenberg, 1967, 1968). The polyphase deformation has been attributed to the Lower Devonian Acadian orogeny (Ruitenberg et al., 1977). Chlorite grade metamorphism accompanied S_1 deformation and biotite grade metamorphism accompanied D_2 deformation so that S_2 cleavage is locally defined by a penetrative biotite fabric (Ruitenberg, 1967). The persistent parallelism of S_1 foliation and bedding can be compared with that on Grand Manan, but the style and orientation of the polyphase deformation do not obviously match those on Grand Manan.

Another possible correlative of the deformation of the Grand Manan rocks can be found in the polyphase structures formed during the Appalachian-Variscan orogeny in the Carboniferous Mispec Group and older rocks west and east of Saint John (Rast and Grant, 1973a, b; Ruitenberg et al., 1973). Grand Manan lies about 15 km southeast behind the probable southwestward continuation of the Variscan Front in the Bay of Fundy (Rast and Stringer, 1974; Rast et al., 1976b). S_1 foliation dips moderately to steeply southeast associated with northwestward verging folds west of Saint John (Rast and Grant, 1973a), but S_1 is generally parallel to bedding in the Caledonia Mountains to the northeast, where F_1 folds are rare and stretching lineations within S_1 with a slightly predominant northwest trend are common (Ruitenberg et al., 1979, fig. 4-11). S_2 crenulation cleavage dips northwestward, associated with southeastward verging folds, and a locally developed subvertical S_3 crenulation cleavage trends north to northwest. Metamorphism is mostly in lower greenschist facies, and posttectonic chloritoid occurs in Mispec rocks at Chance Harbour, 60 km northeast of Grand Manan (Rast et al., 1978). The Mispec Group ranges up to Westphalian B or C (Rast et al., 1978; McCutcheon in Ruitenberg et al., 1979), indicating a late Westphalian or younger age for the deformation.

The style and orientation of the polyphase deformation of the Mispec Group appear to fit the deformation on Grand Manan more closely than that in the Cookson Formation. However, major reverse faults present in the Mispec Group have no apparent counterpart on Grand Manan, and appreciable parts of the stratigraphic succession in nappes associated with the Variscan Front are overturned or upside down, whereas the degree of overturning is less on Grand Manan.

Predicting that a Late Silurian age for the Three Islands granite will be demonstrated by radiometric measurements, we tentatively assign a Late Silurian date to the hornfelsic metamorphism and a pre-Late Silurian age to the deformation of the pre-Triassic rocks.

Acknowledgments

Based on a field trip to Grand Manan, Dr. R.J. Wardle suggested that the pre-Triassic rocks cannot be correlated with the Green Head and Coldbrook Group rocks of Saint John. We thank Dr. Paul Williams for discussion of the microstructures and Dr. Ken Currie for a critical review of the manuscript. The fieldwork was supported (P.S.) by Energy, Mines and Resources (EMR) Research Grant G.S. 33-71 and (G.E.P.) by EMR Research Agreement 59-4-80.

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LATE HOLOCENE SOLIFLUCTION RATES AND RADIOCARBON SOIL AGES, CENTRAL CANADIAN ARCTIC

Project 750071

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Abstract

Twelve radiocarbon dates on humus buried beneath four solifluction lobes indicate long-term average solifluction rates of 2.2, 3.3, >3.5, and 1.3 mm/year. The lobes are composed of till, sand, bouldery gravel, and cobble gravel on slopes of 2° to 25°. Availability of moisture seems to be more important than slope angle or texture in determining solifluction rates. The rates of advance of two lobes have varied through time, though not synchronously. Radiocarbon ages of soils at time of burial range from 285 to 953 years and are important components of the radiocarbon ages of the paleosols. Twelve other radiocarbon-dated lobes from other parts of the world experienced similar rates of movement.

Introduction

For the purpose of this report, "solifluction" is defined as the process or set of processes that gives rise to small landforms commonly referred to as solifluction lobes, gelifluction lobes, or turf-banked lobes (or terraces, steps, or stripes). The main processes probably are gelifluction – the downslope flow of saturated soil due to gravitational force – and frost creep – the heave of soil perpendicular to a slope by growth of segregated ice followed by vertical settling of the soil upon melting of the ice (Benedict, 1976). Considerable attention has been paid to contemporary rates of solifluction; Benedict (1976) summarized measured rates from 102 experimental sites in polar, subpolar, and alpine localities around the world. Far less attention has been paid to calculating long-term average rates of solifluction even though this can be done easily and it would allow evaluation of the measured contemporary rates. A further benefit of measuring long-term rates of solifluction is that changes of rates through time can be established and these can be examined for relationship to paleoclimatic records.

As solifluction lobes advance across vegetated surfaces they override organic soils and prevent further accumulation of organic material. Radiocarbon dates on overridden humus layers beneath four solifluction lobes in the central Canadian Arctic are used here to calculate long-term average rates of solifluction, to establish a chronology of movement, and to estimate radiocarbon ages of soils at time of burial. Because rate of solifluction is a function of several properties and conditions, the regional climate as well as the setting, composition, and structure of each lobe is described.

Climate

The four solifluction lobes discussed here occur at two sites at low elevations (29 and 150 m a.s.l.) near the community of Spence Bay (Fig. 3.1). Spence Bay has a mean annual temperature of -15.4°C; a mean annual temperature range of 41.7°C; and mean June, July, and August temperatures of 0.8°, 7.1°, and 6.5°C, respectively. The station receives 153.4 mm of precipitation per year with about half of that falling as rain (Atmospheric Environment Service, 1975).

Description of Lobes

Lobe 1 is one of many turf-banked lobes ornamenting the leading edge of a solifluction sheet (or terrace) on the well drained southwest-facing slope of a drumlin. The lobe is about 5 m long by 3 m wide, the riser is 30 to 40 cm high, and

the tread forms a 13° slope angle. The riser and the ground in front of it are completely vegetated but the central part of the tread is bare (Fig. 3.2). The soliflucted material consists of till, nonsorted material with about half the matrix falling in the sand fraction and with a slight modal peak in the medium sand range (Fig. 3.3, 3.4). The material has a low liquid limit (13 to 16 per cent) and a narrow plasticity range (2 to 4 per cent). It overlies a 10 to 20 cm-thick humus layer that is continuous with the humus layer of the soil in front of the lobe. This buried soil had developed on marine beach gravel, which was formed by wave reworking of the till.

Lobe 2 resembles lobe 1 in that it is one of many turf-banked lobes on the leading edge of an extensive solifluction sheet (Fig. 3.5) with a southwest exposure. It is also similar

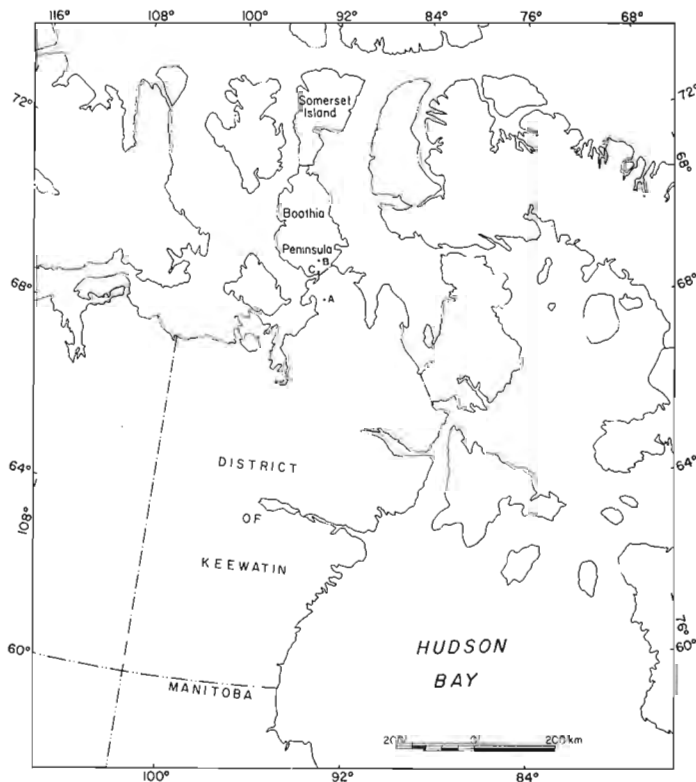


Figure 3.1. Location map showing the study area and sites discussed in the text; A = lobe 1, B = lobes 2, 3, and 4, C = Spence Bay.

in size and in vegetation cover (vegetated riser, bare tread axis). It differs from lobe 1 in that the site is less well drained, the tread forms a much lower slope angle – only 2°, and the soliflucted material is less heterogeneous (Fig. 3.4, 3.6). The soliflucted material of lobe 2 consists of glaciofluvial sand with a strong modal peak in the medium sand range. Some mud was added to the sand during marine submergence. The solifluction sheet originates at marine limit, which indicates that the fine grained marine material that was added to the sand could have been critical in making the sand susceptible to solifluction. The solifluction lobe overlies a 10 to 20 cm-thick humus layer that is continuous with the surface humus in front of the lobe. The buried soil had developed on glaciofluvial gravel (Fig. 3.6).

Lobe 3 is one of many turf-covered lobes, similar in size to the lobes described above, but occurring on the lower slope of a steep (25°) end moraine ridge composed, in this vicinity, of bouldery gravel. The site supports particularly lush vegetation because water is supplied to it nearly throughout the summer by melting of a semi-permanent snowbank along the end moraine crest (Fig. 3.7). The matrix also has sufficient fine grained, amorphous organic material disseminated from the covering sod to give it a brownish colour. The soliflucted material overlies a continuous, 10 to 30 cm-thick humus layer which in turn overlies bouldery gravel (Fig. 3.8).

Lobe 4 is a lobate extension of one of many turf-banked steps on the 25° southeast slope of the same end moraine as lobes 2 and 3. The steps occur on very well drained spurs between minor gullies (Fig. 3.9). The step and lobe risers are conspicuous because they are colonized by dense *Cassiope* shrub mats, whereas the treads have a discontinuous and much more fragile grass and lichen cover with odd clumps of *Cassiope*. The soliflucted material consists of cobble gravel with the clasts being only barely matrix supported. The matrix is about 80 per cent sand with a strong modal peak in the very coarse sand fraction (Fig. 3.4). The soliflucted gravel is only 40 cm thick (Fig. 3.10), much thinner than the soliflucted layers of the three lobes described above. It overlies a continuous humus layer 10 cm thick, which in turn overlies gravel that is identical to the soliflucted material.

Solifluction Rates and Soil Ages at Time of Burial

Twelve radiocarbon ages were determined for samples of the humus layers beneath the solifluction lobes (Fig. 3.3, 3.6, 3.8, and 3.10, Table 3.1). Where possible, only the top 2 to 5 cm of the humus layers were sampled in order to minimize the time span represented by the material. The samples were wet sieved in the laboratory to concentrate the organic matter and to divide it into size classes. Most samples contained plentiful organic matter in the 0.063 to 1.0 mm size range but had little coarser material. Hence, the age determinations were done on the finer of the size classes.

The radiocarbon age of a humus sample is the sum of time elapsed since burial of the soil and the radiocarbon age of the soil at time of burial¹. The latter is caused by the fact that the modern or surface soil contains dead as well as living organic matter. The radiocarbon ages of the modern soils in front of the lobes were not determined by radiocarbon dating, but they are estimated below by extrapolating the rates of advance of the lobe fronts as determined from the radiocarbon dates on the buried soils.

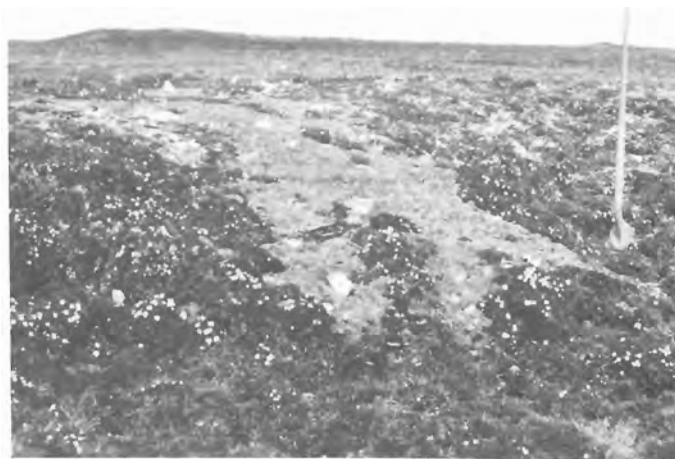


Figure 3.2. Lobe 1, a turf-banked solifluction lobe in till; note unvegetated (light-toned) tread. GSC 203509-J

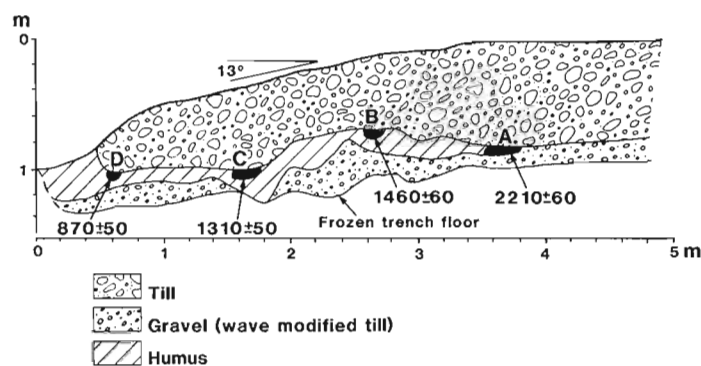


Figure 3.3. Section along axis of lobe 1 showing soliflucted till over humus over gravel, radiocarbon dates, and points (A, B, C, D) referred to in Table 3.2.

Solifluction Rates

Calculation of the rates of advance of the lobes from the radiocarbon dates involves the assumption that the radiocarbon age at the time of burial of all samples of each buried soil is the same. The average rate of advance of a solifluction lobe between any two sample points is, therefore, the distance between the two sample points divided by the age difference. The results are summarized for each lobe in Table 3.2. Except for lobe 3, all radiocarbon dates are in the proper sequence, progressively younger towards the fronts of the lobes (Fig. 3.3, 3.6, 3.8, and 3.10). All that can be stated for lobe 3 is it advanced 2 m in less than 570 years, hence at a rate of more than 3.5 mm/year. The average rates of advance of lobes 1, 2, and 4 range from 1.3 to 12.5 mm/year, but the overall average rates range only from 1.3 to 3.3 mm/year.

Soil Ages at Time of Burial

Because lobe 3 advanced across the innermost sample at some unknown time less than 570 years ago, the radiocarbon age of the sample at time of burial was something less than 570 years. Furthermore, because the outermost sample, which is 710 years old, also has been buried for less than 570 years, its radiocarbon age at time of burial must have been more than 140 years (710-570 years).

¹ This is often referred to as the "apparent age" of a soil. Theoretically, for equilibrium soils, where the rate of production of organic matter equals the rate of decomposition of organic matter, the "apparent age" is the "mean residence time" of organics in the soil.

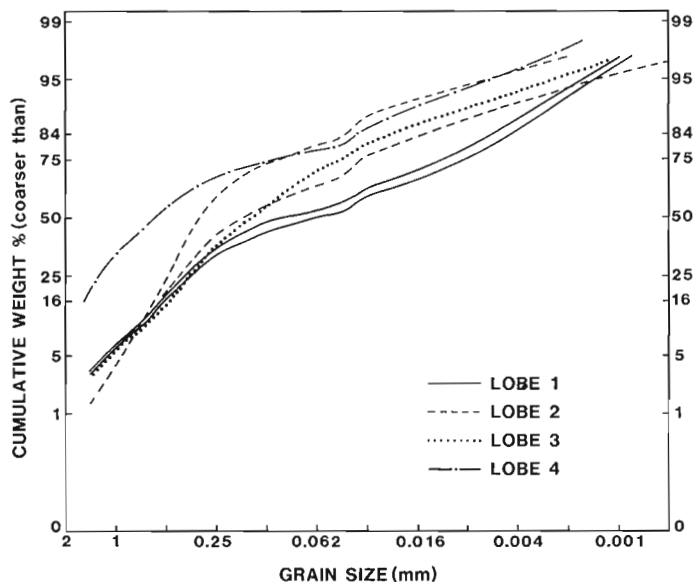


Figure 3.4. Cumulative grain size curves for the <2 mm fraction (matrix) of material from the four solifluction lobes.



Figure 3.5. Large solifluction sheet of muddy sand on a gentle slope at the base of a 60 m-high end moraine (background); the person is standing at lobe 2. GSC 203674-I

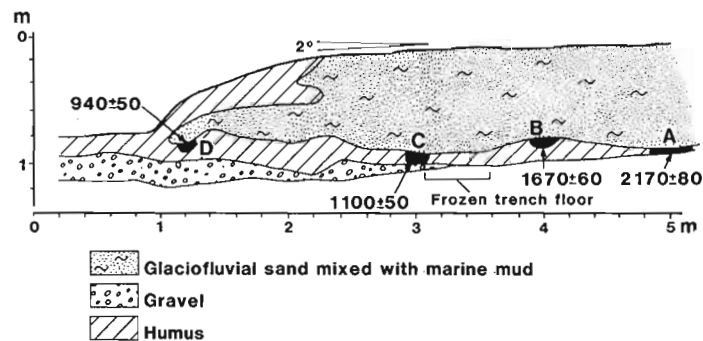


Figure 3.6. Section along axis of lobe 2 showing soliflucted muddy sand over humus over gravel, radiocarbon dates, and points (A, B, C, D) referred to in Table 3.2.



Figure 3.7. Turf-covered solifluction lobes on mid and lower slope of 60 m-high end moraine. Late-lasting snowbanks near top of moraine provide moisture throughout the summer; the person is standing at lobe 3. GSC 203674-J

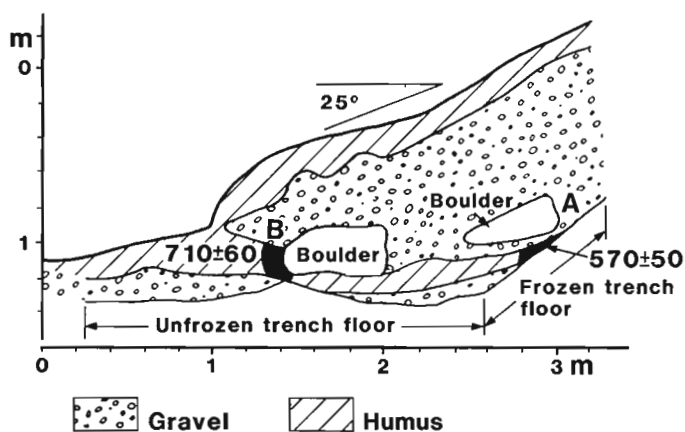


Figure 3.8. Section along axis of lobe 3 showing soliflucted bouldery gravel over humus over gravel, radiocarbon dates, and point A referred to in Table 3.2.



Figure 3.9. Parallel, lobate, *Cassiope*-banked solifluction steps on well drained spurs between gullies on the slope of a 60 m-high end moraine, here composed of cobble gravel; site of lobe 4. GSC 203535-U

Table 3.1

Radiocarbon dates on four solifluction lobes

Lobe	Sample	Date (years B.P.)	Lab no.	Distance from lobe front (cm)
1	A	870 ± 50	GSC-3100	20
1	B	1310 ± 50	GSC-3131	120
1	C	1460 ± 60	GSC-3168	220
1	D	2210 ± 60	GSC-3017	320
2	A	940 ± 50	GSC-3069	20
2	B	1100 ± 50	GSC-3144	200
2	C	1670 ± 60	GSC-3189	300
2	D	2170 ± 80	GSC-2994	400
3	A	710 ± 60	GSC-3080	40
3	B	570 ± 50	GSC-3011	200
4	A	410 ± 50	GSC-3085	20
4	B	1360 ± 50	GSC-3021	150

Ages are quoted in radiocarbon years before present (B.P.), where "present" is taken to be 1950, and are based on a 2σ standard error.

All samples were treated with NaOH, HCl, and distilled water. GSC-2994 was counted for 2 days in the 2 L counter; all others were counted for 2 days in the 5 L counter.

Table 3.2

Long-term average rates of solifluction based on ¹⁴C dates

Lobe 1	A+B	1000 mm in	750 years =	1.3 mm/year
	B+C	1000 mm in	150 years =	6.7 mm/year
	C+D	1000 mm in	440 years =	2.3 mm/year
	A+D*	3000 mm in	1340 years =	2.2 mm/year
Lobe 2	A+B	1000 mm in	500 years =	2.0 mm/year
	B+C	1000 mm in	570 years =	1.8 mm/year
	C+D	2000 mm in	160 years =	12.5 mm/year
	A+D*	4000 mm in	1230 years =	3.3 mm/year
Lobe 3	A+Front	2000 mm in	<570 years =	> 3.5 mm/year
Lobe 4	A+B	1200 mm in	950 years =	1.3 mm/year
(see Fig. 3.3, 3.6, 3.8, and 3.10 for points A, B, C, D)				
*overall average rates for lobes with more than 2 dates.				

The radiocarbon ages at time of burial of the foremost samples can be estimated for the other three lobes if we assume that the rate of advance of each lobe front between the foremost sample and its present position was equal to the rate of advance during the time interval immediately preceding. For example, the front of lobe 1 (Fig. 3.3) advanced from point C to point D at 2.3 mm/year (Table 3.2); If it continued to advance at the same rate to its present position, 20 cm beyond point D, then point D must have been passed 87 years ago, or 58 years B.P. on the radiocarbon time scale¹. If the true time of burial of the soil at point D was 58 radiocarbon years B.P., then the radiocarbon age at time of burial of the sample from point D was 812 years (870-58 years). The radiocarbon ages at time of burial of other samples (Table 3.3) range from 285 years to 953 years.

Changes of Rates Through Time

In calculating the rates of solifluction between dated sample points, the radiocarbon ages at time of burial of all samples from each lobe were assumed to be the same. If this assumption is retained, the real chronology of movement can be established for three of the lobes by subtracting the appropriate radiocarbon ages at time of burial from the radiocarbon ages of the samples. For lobe 4, only an overall average rate of movement can be determined because only two samples were dated. Four samples were dated from each of lobes 1 and 2, and the history of their movements is plotted in Figure 3.11. Lobe 1 greatly increased its rate of advance in the interval 650 to 500 years B.P., and lobe 2 increased its rate of advance even more sharply in the interval 250 B.P. to A.D. 1979.

Discussion

Rates and Mechanisms

All four lobes are at sites with similar present climates (all on southeast-facing slopes at low elevation) and all have experienced the same changes of climate over the past 1500 years. They differ widely from each other, however, in composition, slope angle, drainage conditions, and form, all of which are factors that might be expected to influence the rate of solifluction considerably. Despite this, with the possible exception of lobe 3, the overall average rates of movement of the lobes are closely similar (1.3 to 3.3 mm/year) and the variations in average rates through time for lobes 1 and 2 far exceed the differences in overall average rates between the lobes. Again, excepting lobe 3, slope angle does not appear to exert a dominating influence on solifluction rate. Lobe 2, with a slope angle of only 2°, had both the highest overall average rate (3.3 mm/year) and by far the highest rate for any time interval (12.5 mm/year), whereas lobe 4, with a 25° slope angle, had the lowest rate (1.3 mm/year). Perhaps the most important condition in this case is soil moisture availability; lobe 2 has much more moisture available than lobe 4 (the driest site of all) and lobe 3, which possibly had the highest rate of movement, occupies the wettest site as well as having a steep slope.

The textural composition of a soil should exert a strong influence on the rate of solifluction. Soils rich in fine grained material are highly susceptible to the growth of segregated ice, which both causes frost creep and upon thawing, lowers the resistance of the soil to shear stress. Fine grained soils also have much greater ability to retain moisture and hence are susceptible to flow by plastic deformation or liquefaction. For these reasons, till should be much more prone to solifluction than sand or gravel and this seems to be borne out by the impression that solifluction forms apparently are more common on till than on sand and gravel². In this regard lobe 1 should be more susceptible to solifluction than lobes 2 and 3, but both these lobes experienced faster movements than lobe 1. Again, the explanation probably lies in the fact that lobes 2 and 3 occupy sites that have a better supply of moisture. Lobe 4, however, which is composed of gravel with a coarse sand matrix, occupies a site that is dry throughout most of the thaw season, but still its rate of movement was not much less than that of lobe 1. These gravel lobes and steps that occupy dry sites on steep slopes (Fig. 3.9) are unlikely to "flow". Normally, because of their texture and high permeability, they would not be expected to host segregated ground ice in the active layer during the winter, a condition that seems necessary to promote frost creep. Perhaps, however, the tough, dense mat of *Cassiope* that covers the risers provides sufficient impedance to the flow of groundwater during the fall freeze-back that enough moisture is retained behind the risers to form segregated ice, and hence promote frost heave.

¹Year zero in the radiocarbon time scale is A.D. 1950; field measurements were made in 1979.

²This could be a misimpression, however, caused by the fact that till is much more common than sand and gravel, and sand and gravel tend to form stable level surfaces. Solifluction forms are certainly not uncommon on vegetated slopes on sand and gravel.

This mechanism would ensure their continued advance one they had developed; the mechanism of their initial formation, however, is less easy to visualize. Perhaps the trigger was local changes in frost table configuration and permeability brought on by development of ice wedges, which are ubiquitous in gravels in this region, and subsequent vegetation zonation along ice-wedge troughs.

Comparison with Other Studies

I am aware of only 12 other cases where radiocarbon dates have been used to calculate long-term average rates of solifluction. Benedict (1976) summarized those available up to that date and since then two other studies have been published. Table 3.4 summarizes the results of these studies and shows that the average rates, in all cases, are a fraction of a centimetre per year, mostly 1 to 2 mm/year. Thus, they are similar to the rates calculated above. This uniformity of rates is surprising in that the various study sites are located in vastly different environments around the world and the soliflucting materials are widely different in texture. Perhaps it is simply fortuitous. Because we have a sample of only 16 from a world population of many millions of lobes, it is impossible to generalize at present.

Ages of Soils at Time of Burial

The radiocarbon ages of pockets of humus at time of burial can be estimated reliably from the radiocarbon dates and solifluction rates. These ages vary significantly from site to site (Table 3.3). This variation shows that radiocarbon ages of surface soils should not be assumed to be constant across wide regions of the Arctic. Because radiocarbon ages of even thin soils can be 950 years or more, the timing of geological events in the Holocene that are based on radiocarbon dates on buried humus and not corrected for radiocarbon age at time of burial could be seriously in error. This problem could be eliminated by dating macrofossils (wood) from the upper boundary of the buried soil because these should have radiocarbon ages at time of burial of or approaching zero. In most parts of the Arctic, however, this is not possible because of the scarcity of large woody plants. Therefore, it might be useful if some attention were paid to establishing the variability of radiocarbon ages of surface soils and determining the causes of such variability.

Changes of Rates Through Time

Benedict (1966) found that the rate of movement of a stone-banked terrace in the Colorado Front Range fluctuated widely through time, and he related the changes to the Neoglacial moraine record in nearby cirques, thus implying that the changes of solifluction rates reflected changes of climate. The rates of advance of lobes 1 and 2 in this study also have fluctuated through time (Fig. 3.11). The timing of the changes, however, does not coincide, and that lack of

Table 3.3

Radiocarbon ages at time of burial of humus samples

Lobe	^{14}C age of sample	Date of burial	Apparent age (years)
1	879 B.P.	58 B.P.	813
2	940 B.P.	13 years after 1950	953
3	710 B.P.	<570 B.P.	>140, <570
4	410 B.P.	125 B.P.	285

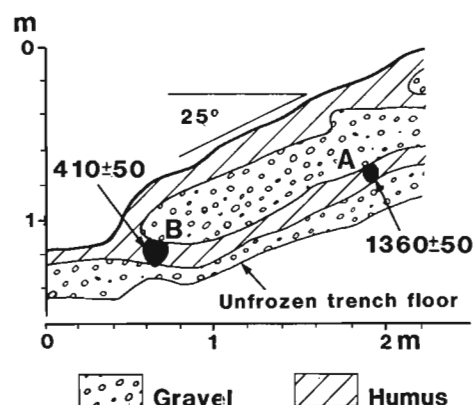


Figure 3.10. Section along axis of lobe 4 showing soliflucted cobble gravel over humus over cobble gravel, radiocarbon dates, and points (A, B) referred to in Table 3.2.

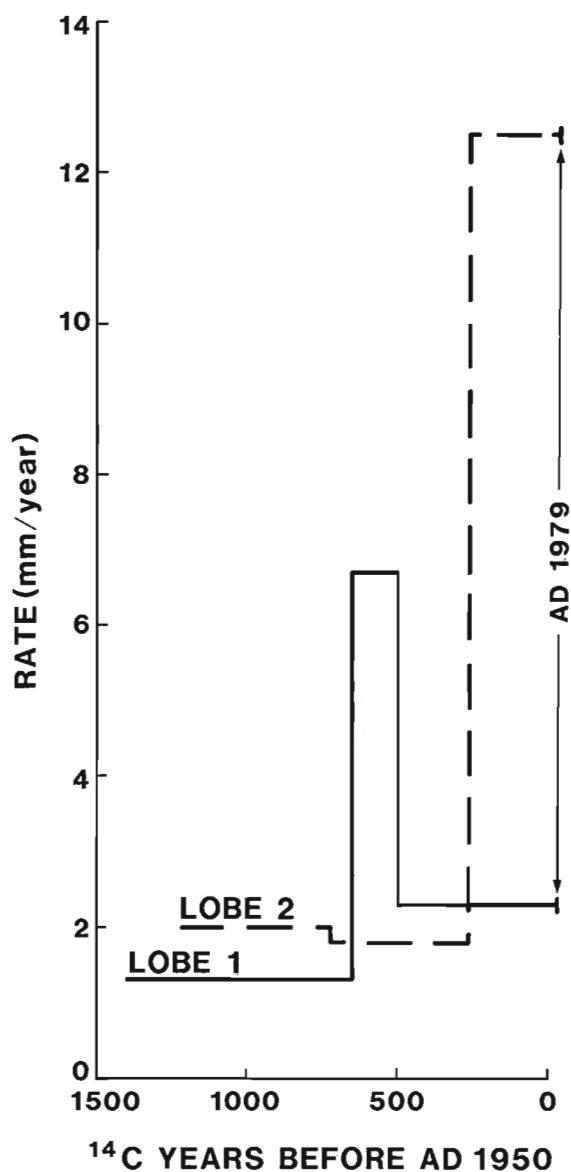


Figure 3.11. Fluctuations in the rate of advance of lobes 1 and 2 over the last 1400 radiocarbon years (chronology based on radiocarbon dates corrected for radiocarbon ages of soil upon burial).

Table 3.4
Radiocarbon-dated solifluction lobes from other areas

Location	Reference	Oldest ¹⁴ C Date	Average rate of movement (mm/year)
Colorado	Benedict 1966, 1970	2470 ± 110	3.2
Colorado	Benedict, 1970	2340 ± 130	1.8
Australia	Costin et al., 1967	2980 ± 180	1.6
Greenland	Everett, 1967	1695 ± 140	1.2
Greenland	Everett, 1967	935 ± 120	1.8
Scotland	White and Mottershead, 1973	5145 ± 135	>1.7
Alaska	Hamilton in Benedict, 1976	2075 ± 130	1.1
Alaska	Hamilton in Benedict, 1976	2670 ± 180	0.8
Norway	Worsley and Harris, 1974	2480 ± 90	1.2
Norway	Worsley and Harris, 1974	2550 ± 80	1.1
Norway	Ellis, 1979	4750 ± 130	2.1
Yukon	Alexander and Price, 1980	(28 dates)	7-10

coincidence could indicate that the changes are not responses to a common cause such as a change of regional climate. The fluctuations of rates could be reflecting site-specific changes of a condition such as ease of drainage of the active layer brought on by local aggradation or degradation of the permafrost table, gullying or slumping of adjacent soil, or any of several other possible causes. This is not to say that climatic changes will not engender changes in solifluction rates, but a much larger sample size will be necessary to determine if such has happened in any region. On the other hand, the apparent uniformity of long-term average rates of solifluction in many parts of the world (Table 3.4) could be used to argue that climate is not a dominant determinant of solifluction rate.

Summary

Radiocarbon dates on humus buried by four turf-banked solifluction lobes that have developed on till, sand, bouldery gravel, and cobble gravel and on slopes ranging from 2° to 25° show that:

1. The overall average rates of solifluction were similar (1.3 mm/year, 2.2 mm/year, 3.3 mm/year, and >3.5 mm/year).
2. The rates of movement of two of the lobes have varied over the past 1400 years (from 1.3 mm/year to 6.7 mm/year for lobe 1; from 1.8 mm/year to 12.5 mm/year for lobe 2), but the variations were not synchronous.
3. Radiocarbon ages of humus samples from the top 5 cm of soil at time of burial range from 285 to 953 years.

Long-term average rates of movement calculated here are similar to rates calculated for 12 other lobes at widely scattered arctic and alpine sites around the world. This apparent global uniformity of long-term average rates and the lack of synchrony of fluctuation in rates could be used to argue that climate is not the major determinant of solifluction rates.

Acknowledgments

Logistical support for field work in 1979 was provided by polar Continental Shelf Project of Energy, Mines and Resources, Canada. Field assistance was rendered by Aavo Taal and Robert Hélie. Comments on the typescript by Dr. R.J. Fulton and J.A. Heginbottom were most useful.

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Project 800024

L.A. Dredge
Terrain Sciences Division*Dredge, L.A., The leaching of carbonates in discontinuous permafrost, Boreal Manitoba; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 23-25, 1981.***Abstract**

In boreal areas, peaty vegetation promotes the dissolution of calcareous parent materials. In nonpermafrost terrain carbonate removal has progressed to a depth of 2.5 to 3 m; half of the original carbonates have been leached from the upper 0.5 m of the soil column, where dissolution is most intense. In permafrost terrain leaching is less intense and is restricted to the upper 0.8 m of the mineral soil. Mudboils have relatively high carbonate contents because of upwelling of unleached material from the subsoil.

Soils in permafrost may have low buffering capacities despite their high carbonate contents because meteoric water does not circulate below the active layer.

Because of present and past leaching, carbonate contents may not be good indicators of till provenance.

Introduction

This study arose from a regional Quaternary geology mapping project in northern Manitoba. During the course of routine laboratory analysis it became apparent that the carbonate soils characteristic of the region had been leached to a considerable extent. This report presents data relating to the depth and intensity of leaching of calcareous soils in a boreal environment, compares results from sites with and without permafrost conditions, and suggests some broad implications stemming from the results. Because of the homogeneous nature of parent materials, vegetation, and topography, the results shown here are thought to be applicable over a broad area.

Acknowledgments

F.M. Nixon assisted with the collection of samples used in this report. The paper was critically read by P.A. Egginton and R.J. Fulton.

Regional Conditions Influencing Carbonate Weathering in Northern Manitoba

Leaching of calcareous soils occurs when calcium and magnesium carbonates are dissolved by percolating meteoric water charged with carbon dioxide. The process begins at the surface and gradually progresses deeper. The thickness of the leached zone and its character depend on the hydrological and biochemical environment, the thermal regime, the texture and original carbonate content of the substrate, and the time available for the leaching process to occur.

The study area is in Manitoba between 57° and 59°N and between 92° and 96°W, near the west coast of Hudson Bay, above the area covered by postglacial marine deposits. The land rises gently from 120 to 250 m, with a regional slope of 2 m/km; surface drainage is generally poor. There is little variation in the terrain apart from Churchill River valley, which is deeply incised into till and underlying bedrock. The region is vegetated by spruce forest, but most of the area consists of forested bog, with peat 1 to 4 m thick, interrupted by fens and shallow thermokarst pools. The peaty vegetation creates a strongly acid environment, with the pH of the pools ranging from 3 to 5.

Annual precipitation is 40 cm; mean annual air temperature is -7.3°C, but temperatures reach 12°C in July. The area lies within the zone of discontinuous permafrost.

Where peat is thick, the depth of thaw does not exceed 60 cm; the mineral substrate remains frozen all year but ground temperatures do not fall below -5°C (Brown, 1978; Dredge, 1979). Where the peat cover is thin (i.e., less than 1 m), permafrost is rarely encountered. Seasonal frost sorting and cryoturbation of materials in unfrozen terrain (i.e., without permafrost) is manifested by mudboils.

The parent material is calcareous till and related glaciolacustrine sediment, which were first exposed to subaerial conditions about 7700 to 8000 years ago. The parent material and leached soil are uniformly fine textured, consisting of about 35% sand, 43% silt, and 22% clay (Table 4.1). In unweathered material the carbonate content (c) is homogeneous areally and makes up between 32% and 38% of the total soil weight (Table 4.1).

Procedures

Near-surface and subsurface samples used to construct the leaching profile were collected at 28 sites over a broad area (10 000 km²).

The sites chosen have geological, vegetational, topographical, and drainage characteristics that are similar from site to site and that are characteristic of the region. Because sampling was conducted for mapping purposes, most samples were obtained from a variety of depths, but at four sites soil samples were taken at regular intervals throughout vertical profiles (Table 4.1). Carbonate contents in Table 4.1 for widespread spot samples from different depths fit consistently with the leaching pattern determined from sites where samples were taken at a number of depths. This suggests that the detailed profiles are representative of a broad area. Figure 4.1 is a generalized composite of the results and shows the typical relationship between carbonate content and depth for the region.

Near-surface samples in areas of unfrozen terrain were taken from hand-dug soil pits. A Hoffer coring probe and Stihl power corer were used to obtain near-surface samples in frozen terrain and all subsurface samples to a depth of 3 m. Samples from greater depths (3-20 m) were obtained by excavating in from riverbanks.

Carbonate contents of the <2 mm fraction were determined by digesting the samples in hydrochloric acid. Minor amounts of other minerals may also have been dissolved.

Results and Discussion

Figure 4.1 presents generalized curves (derived from data in Table 4.1), which show how carbonate content changes with depth in permafrost and nonpermafrost terrain.

Degree and Depth of Leaching in a Boreal Forest Environment (Without Permafrost)

The profile (Fig. 4.1) shows that carbonates have been removed from the soil to a depth of 2.5 to 3 m. The greatest leaching has occurred near the surface, and about half of the carbonates originally present have been leached from the upper 0.5 m. Below about 3 m carbonate values approach 35 per cent (the amount of carbonate in unweathered parent material) and are relatively constant to depths of at least 20 m.

Soil and climatic conditions in this area favour relatively extensive leaching. The soil parent material is carbonate rich, and its silty texture enables a large surface area to come into contact with meteoric water and groundwater despite its relatively low permeability. Because low soil temperatures (5°C in summer) promote solution of carbonates, solubility may be 50 per cent greater at the temperatures prevailing in this boreal environment than in warm temperate climates (Arkley, 1963). The acidity of the environment also promotes dissolution of the carbonate fraction. The ubiquitous bogs and fens, which are the

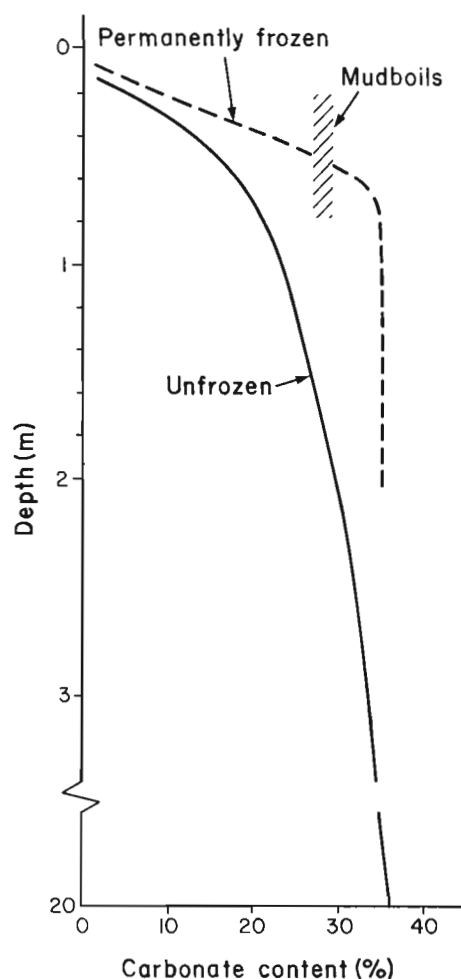


Figure 4.1. Carbonate depletion in frozen and unfrozen terrain. The curve was constructed using the data shown in Table 4.1.

principal groundwater reservoirs, have pH values between 3 and 5, and consequently a greater rate of carbonate solution might be expected here than in areas where the groundwater is neutral (Arkley, 1963). Depth of leaching in this region, therefore, may be expected to exceed that in warmer climates, providing sufficient time has elapsed for the weathering process to reach equilibrium conditions.

Data from dated sites below the limit of postglacial marine submergence in the nearby Hudson Bay Lowland indicate that the leaching profile shown here could have developed in about 5500 years. Since this area has been exposed to soil development for 7700 to 8000 years, Figure 4.1 may represent an equilibrium profile for the northern boreal forest environment.

Table 4.1
Carbonate contents and grain size distribution
for leached soil samples

	Site No.	Depth (cm)	c% ¹	% sand	% silt	% clay
Permafrost terrain						
	1A	20	9	31	40	29
	1B	60	34	35	43	22
	1C	120	34	32	44	24
	2A	20	10	34	43	23
	2B	50	24	36	42	22
	2C	80	34	35	45	20
	3	30	15	37	44	19
	4	15	10	35	41	24
	5	15	6	36	42	22
	6	30	24	38	43	19
	7	15	9	28	40	32
	8	200	35	36	44	20
	9	250	32	32	46	22
Mudboils						
	10	30	26	37	44	19
	11	48	28	30	41	29
	12	50	26	35	42	23
	13	70	25	33	45	22
Nonpermafrost terrain						
	14A	50	17	29	45	26
	14B	75	20	33	40	27
	14C	100	33	36	44	20
	14D	200	30	36	40	24
	15A	20	5	32	42	26
	15B	50	15	38	47	15
	15C	500	32	32	41	27
	16	40	11	36	43	21
	17	20	8	35	41	24
	18	20	5	33	43	24
	19	25	6	37	42	21
	20	20	4	33	40	27
	21	40	11	34	46	20
	22	90	20	32	46	22
	23	120	24	35	41	24
	24	200	29	34	40	26
	25	218	33	34	42	24
	26	250	36	38	45	17
	27	400	36	35	43	22
	28	900	35			

¹c is carbonate content

On the basis of 96 samples, the carbonate content of the parent material is 35% (standard deviation 3.2).

In the northern environment, the depth of leaching may be more closely controlled by depth limitations in the flow of the meteoric component of groundwater than by factors related to chemical equilibrium. A zone of carbonate accumulation, typically present in the lower part of leaching profiles at sites from warmer and drier climates, is not present in the profile shown here. Its absence suggests that groundwater may be leaving the soil system (and entering rivers) before it is saturated with bicarbonate ions.

Effect of Permafrost

A substantial difference exists in both depth and amount of leaching between substrates with permafrost and those without. In soils under perennially frozen peat, leaching is limited to the upper 60 cm of mineral substrate, and the amount of carbonate removal at a given depth is much less than in unfrozen conditions. The difference is attributed to restricted groundwater movement in permafrost terrain. Otherwise, these areas with extremely low pH due to peat cover should be extensively leached. The fact that any leaching was observed in permafrost terrain suggests that some circulation of unfrozen water occurs through soil capillaries in this relatively warm permafrost environment (soil temperatures reach -2°C in summer), but it is also possible that depletion of carbonates occurred prior to the onset of permafrost.

Effect of Frost Churning

Near-surface and surface samples taken from mudboils in unfrozen silts have carbonate values similar to those in frozen terrain. The higher percentage of carbonate in mudboils than at other unfrozen sites suggests the incorporation and mixing of material from depths of 2 m or more. Another hypothesis is that mudboil centres act as wicks or as evaporation centres (Egginton, 1981); if groundwater is drawn to and evaporated from these wicks, an unleached component of the salt might remain.

On Carbonate Soils as Buffers of Acid Rain

It has been previously assumed that soils of the Hudson Bay Lowland, because of their high carbonate contents, are capable of buffering natural or man-induced acidity of meteoric water. Results from this study suggest that unfrozen soils in the region can neutralize the effects of

acidification of surface waters and thus have a high buffering capacity. Where permafrost is present, however, circulation is restricted mainly to the active layer, and carbonates in the frozen subsoil do not act as a buffering material. In areas of discontinuous permafrost, therefore, it should not be assumed that calcareous soils have a high buffering capacity on the sole basis of high carbonate contents in samples from upwelling mudboils or from frozen subsoils.

On Carbonates as Till Provenance Indicators

Results from this study show that a substantial amount of leaching has occurred in unfrozen soils in northern Manitoba; for this reason care must be exercised in using carbonate contents for till provenance studies. Spurious differences in carbonate values can arise from (a) sampling at inconsistent depths and (b) sampling from frozen and unfrozen terrain. Furthermore, the overburden over much of northeastern Manitoba consists of multiple till sheets separated by thin non-till units. A substantial part of the overburden may have been leached to some degree at various times in the past. Carbonate contents obtained from partially depleted sites will not reflect the regional carbonate contents of parent materials.

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SURFICIAL GEOLOGY, LOUGHEED ISLAND, NORTHWEST ARCTIC ARCHIPELAGO

Project 760010

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Hodgson, D.A., *Surficial geology, Loughheed Island, northwest Arctic Archipelago; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 27-34, 1981.*

Abstract

The surficial materials on Loughheed Island and adjacent Edmund Walker Island are mainly weathered soft shale and sandstone at high elevations, and a complex of weathered rock, and glaciomarine, marine, deltaic and fluvial sediments over the remaining area. Glacial till, and associated deformation of the soft bedrock, are exposed only in section. The direction of the deforming force, the source of erratics, and alignment of drumlinoid ridges and a possible esker, indicate ice movement from the southeast; the age of the responsible glacial event is not known and may well be pre-late Quaternary. Rapid deposition of glaciomarine, and subsequently marine, sediment was in progress at $10\,500 \pm 130$ years ago, while emergence of at least 90 m has occurred since $10\,240 \pm 280$ years ago. The cause of uplift and the origin of the glaciomarine sediments is suggested to be break-up of an ice sheet to the south, covering at least Bathurst Island 10 500 years ago, and producing floating ice, or even an ice shelf which extended to Loughheed Island; however, latitudinal compressive tectonic forces might also be responsible for uplift.

Introduction

Glacial and glaciomarine deposits found on Loughheed Island have not been reported from adjacent areas of surrounding, larger Queen Elizabeth Islands. This information, derived from a survey of surficial materials, plus confirmation of at least 90 m emergence since the late Quaternary (GSC-356, Table 5.1), introduces data from the northwestern islands into the debate on extent of late Quaternary glacial ice in the Queen Elizabeth Islands. Blake's (1970) concept of the Innuitian Ice Sheet covering much of the islands has been challenged by England (1976) and Boulton (1979), who favour a restricted ice cover. In addition to its significance in regional late Quaternary history, the island is surrounded by a number of offshore gas fields and, if development proceeds, knowledge of surficial materials will be required for engineering and environment studies.

Raised beaches and marine shells were noted by the first recorded visitors to the Findlay Group, which comprises Loughheed and the three smaller islands to the southeast (Stefansson, 1921, p. 544). Fyles (1965) observed landforms of probable glacial origin and measured raised marine deposits, which were plotted on the Glacial Map of Canada (Prest et al., 1968). For this study, Loughheed Island was visited between mid-July and early August, 1979, when, with S.A. Edlund who studied vegetation (Edlund, 1980), traverses were made by Honda A.T.C. and on foot.

Acknowledgments

F.M. Nixon shared the field work and provided advice at a later date, and L.W. Sobczak (Earth Physics Branch) provided ideas on crustal movement mechanisms. The comprehensive logistical support of Polar Continental Shelf Project is appreciated. This manuscript was critically read by L.A. Dredge and W. Blake, Jr.

Physiography

Loughheed Island (1300 km²) is elongated north-northwest; the north half rises to a ridge at 130 m a.s.l., whereas the south half comprises low rolling hills rarely rising to 100 m elevation and an extensive lowland in the west. Fluvial dissection locally sharpens relief in all parts of the island. The three islands to the southeast are low, and the largest, Edmund Walker Island, peaks at 135 m a.s.l. The Findlay Group rises from waters at least 150 m deep, except

to the southwest where a depth of about 50 m is maintained towards Melville and Cameron islands (Fig. 5.1). Previous studies summarized by Pelletier (1966) suggest that inter-island channels and basins of the Queen Elizabeth Islands are products of a glacially modified fluvial system. However as tectonism is now known to have formed Parry Channel (Kerr, 1980), concurrent rifting might have shaped Loughheed and the other northwest Queen Elizabeth Islands.

Structural Geology

Thick units of soft sandstone alternate with shale in the shallow Loughheed Syncline, the axis of which coincides with the longer axis of the island. Bedrock is described by Balkwill et al. (1977, 1979), and units are outlined on Figure 5.2. Surface characteristics of rock are described under surficial materials in this report.

Drainage and Permafrost

The flow of surface water ceases soon after snowmelt in all but the highest order channels or where perennial snowbanks survive in drainage incisions. Surface water is otherwise restricted to several lakes dammed by coalescing prograding deltas, and to the few areas where thaw-enlarged frost fissures occur.

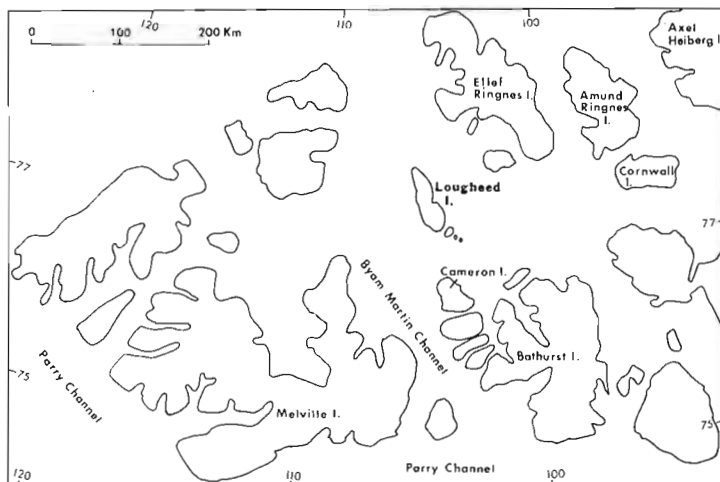


Figure 5.1. Location map.

Table 5.1
Radiocarbon dates, Loughheed and Edmund Walker islands

Laboratory no. (Field no.)	Radiocarbon age (years B.P.) corrected (uncorrected)	Material dated	Location	Sample elevation (m)	Geological environment	Significance	Collector	References
IKGSC)-24	8200 ± 180	Marine pelecypods	NE coast Loughheed I., 77°33'N, 105°W	ca. 23-30	Clay in shallow gully	As higher shells uncommon on this coast, sample may not be contaminated by them (see text).	V. Stefansson. Coll. during Canadian Arctic Expedition, 1916; subm. 1959 by J.G. Fyles	Craig and Fyles, 1960, p. 11. Walton et al., 1961, p. 53
GSC-320 (FG-64-113 at site 7/27C)	10 100 ± 150	Hiatella arctica	17 km W of C. Rondon, Loughheed I., 77°20'N, 105°07'W	ca. 60	Surface of linear ridge (see text); on glaciomarine sediments?	Maximum age of 60 m water plane on central Loughheed I.	J.G. Fyles, 1964	Lowdon et al., 1967
GSC-356 (FG-64-46d at site 7/19D)	10 240 ± 280	Hiatella arctica	E. Edmund Walker I., 77°09'N, 104°02'W	ca. 90	Surface of shale ridge	Maximum age of 90 m water plane on Edmund Walker I.; highest definite late Quaternary shells	J.G. Fyles, 1964	As above
GSC-2928	10 400 ± 260 (10 400 ± 260)	Hiatella arctica 15.7g	30 km SSE of C. Ahnighito, Loughheed I., 77°30'N, 105°36'W	43.5	Near top of 3.5 m marine pelite and glaciomarine sediment exposed in nivation hollow	Indicates rapid rate of sedimentation for 3 m of marine pelite	D.A. Hodgson, 1979	This report
GSC-2986	10 500 ± 130	Hiatella arctica 26g	as above	40.5	In stony marine pelite (glaciomarine) sea and deposition over rock, beneath 3 m marine pelite sediment	Minimum age of late Quaternary of glaciomarine sediment	D.A. Hodgson, 1979	This report

Permafrost thickness is >300 m at Pat Bay well site (Fig. 5.2), 17 m a.s.l. and 2 km from the modern shoreline (Judge et al., 1980). This site emerged from marine waters after 8000 years B.P. (IGSC-24, Table 5.1; Fig. 5.3), but probably before 5000 years ago. Active layer thickness varies from 10 to 100 cm depending on material and on vegetation type and cover.

No data exist on ground ice. No surface manifestation of ice occurs except possibly as frost fissures, though these may not contain ice wedges owing to the paucity of moisture.

Climate – Wind

Prevailing and strongest winds are from the northwest or north-northwest over much of the northwest Queen Elizabeth Islands (Maxwell, 1981), hence eolian deposits have this alignment. There is a remarkable parallelism on Loughheed Island among bedrock structure trends (and thus topography), glacial ice flow of unknown age, and dominant wind direction, which causes confusion in the air photo interpretation of landforms.

Surficial Deposits

Surface Rock

Material directly derived from underlying bedrock is exposed over 25 per cent of the island (Fig. 5.2). Periglacial weathering, possibly aided by glacial deformation, has reduced the poorly cemented bedrock to constituent particles, dominantly medium to fine sand and silt size, down to a depth of several metres. Probably all surface rock, which includes the highest land, has been washed by marine waters during late Pleistocene and early Holocene high sea levels; hence scattered pockets of marine sediments occur on this unit. A discontinuous lag of clasts related to underlying rock contains scattered erratics of southern provenance (Fyles, 1965).

In vertical section, the contact between bedrock and overlying sediments ranges from abrupt to gradational owing to intermixing by glacial or fluvial processes. Unconsolidated silty shale beds, and to a lesser extent sandstones, are easily confused with Quaternary marine or deltaic sediments, unless Cretaceous macrofauna are present.

The three dominantly sandstone formations described by Balkwill et al. (1977, 1979) are treated as one unit here, whereas the two shale formations, which have dissimilar weathering characteristics, are each considered separately.

Isachsen, Hassel, and Eureka Sound Formations – Sandstone. These formations are composed of weakly cemented carbonaceous sandstone which readily weathers to buff or light grey sand. Scattered resistant cemented beds reinforce areas of highly dissected scarpland. The Hassel Formation, which is the most widely distributed sandstone, contains shale beds.

Christopher Formation – Shale. As on the Ringnes islands, the Christopher Formation underlies rounded hills which form the highest land even though the weathered rock can be highly unstable in the geotechnical sense. Balkwill et al. (1977) described this shale as more silty than the Kanguk Formation; however the weathered rock is more plastic and prone to failures, particularly active layer detachment slides and flows. Stream beds, however, are commonly more stable owing to an armouring of mudstone concretions derived from this formation. As this is the best vegetated and most water-retentive unit, it is the unit most likely to contain substantial ground ice bodies.

Kanguk Formation – Shale. The low lying, acidic, and barren silty terrain is quite unlike the Christopher Formation (cf. Edlund, 1980). Thin sandstone beds reinforce low scarps. Wide shallow channels filled with silty sand are characteristic of terrain dominated by large-scale fluvial processes rather than by mass movement and rills as on Christopher Formation shale.

Quaternary Deposits

Much of Loughheed Island, other than land at high elevations, is covered by Quaternary deposits. With the exception of fluvial deposits and thick marine and deltaic sediments, it proved difficult to identify on air photos materials with different origins. Hence one third of the materials map (Fig. 5.2) is composed of a complex of glacial, glaciomarine, marine, and deltaic sediments, together with areas of rock where drainage is incised, or weathered rock under a discontinuous marine veneer especially near the central west coast.

Glacial Deposits. Till outcrops only in sections found in the southeast and west of the island (Fig. 5.2), and the extent of subcrop is unknown. Carbonate and crystalline erratics, together with striated sandstone clasts, scattered on the surface at all elevations, are either remnants of till or of glaciomarine deposits. A diabase block 4 m across was found 125 m a.s.l. at the northern height of land. Till is best exposed over an area of Kanguk Formation shale and minor sandstone in the extreme southeast of Loughheed Island at about 30 m a.s.l. Here, in up to 1 m of a calcareous, in places pinkish-hued, diamicton the dominant clasts are sandstone but include Paleozoic limestones. The till overlies rock in which the upper few centimetres to metres has commonly been deformed, and in some cases convoluted or even overthrust and sheared, by a compressive force from a southerly direction. Rafts of unconsolidated bedrock up to 50 cm thick and 3 m long are incorporated in the till. Rock exposed in a river cut near the west coast has been sheared to a depth of 3 m by a force from the southeast (Fig. 5.4). Glacial deposits are overlain by a stony marine pelite, in places in conformable contact, elsewhere unconformable. Underlying rock is everywhere too friable or soft to preserve striae.

Glaciofluvial (?) Deposits. An esker-like ridge 1.5 km long, 10 m high, and 25 m a.s.l. at the southeast extremity of Loughheed Island is oriented north-northwest. The surface is composed of sandstone and a few carbonate clasts up to 1.5 m diameter, some striated; the subsurface is dominantly sand or sandy silt. The concentration of surface boulders is unique on the island, and even if reworked by beach processes, deposition by a glacial process is the most plausible origin.

Glaciomarine (?) Deposits. A dark grey or black stony marine pelite up to 3 m thick, and unstructured to faintly stratified, is widespread in the west centre, centre, and southeast of Loughheed Island. The sediment overlies rock or till and commonly grades upwards into stonefree marine pelite. Stones, composing up to 10 per cent of the volume, are chiefly sandstone and less commonly carbonate, and are commonly striated. Contact with underlying rock or till may be abrupt (Fig. 5.5) and in places obviously disconformable. Clasts of poorly consolidated material resembling the underlying rock occur in the pelite. Nevertheless, as fragile and paired marine pelecypods occur at all levels, it is considered that most clasts are dropstones, and not reworked from subjacent deposits. Drifting glacial ice or an ice shelf are likely sources. At a number of localities thin (1–25 cm) lenses or beds of finely stratified sand underlie stony pelite in a conformable manner (Fig. 5.6).

Figure 5.2 Surficial materials and geology, Lougheed Island.

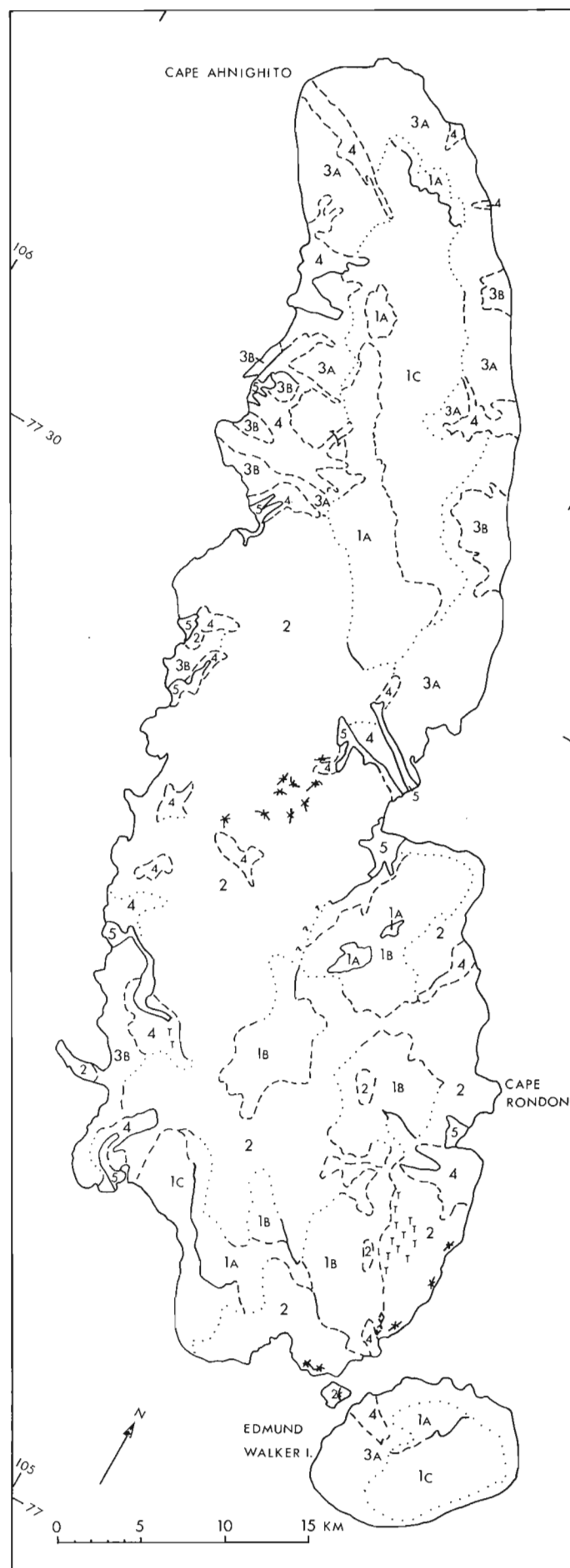
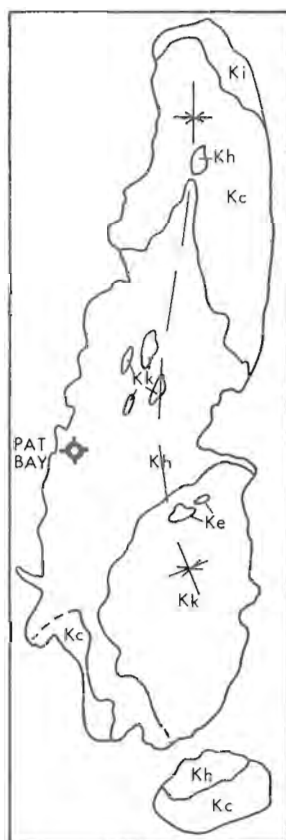
LEGEND

- 5 FLUVIAL DEPOSITS: sand and silt, in active channel zones >200 m wide
- 4 DELTAIC DEPOSITS: sandy topset beds, planar, generally overlying marine pelite.
- 3A MARINE PELITE: <2 m thick
- 3B MARINE PELITE AND DELTAIC DEPOSITS: >2 m thick
- 2 QUATERNARY UNDIFFERENTIATED: surficial rock; till (minor outcrop); glaciomarine (stony pelite); marine (pelite or silty and sandy littoral deposits); active fluvial channels <200 m wide.
- 1 SURFICIAL ROCK
 - A: sandstone (Isachsen, Hassel, and Eureka Sound formations); friable, generally disintegrated to sand.
 - B: shale (Kanguk Formation), silt, clay, shale fragments; low plasticity.
 - C: shale (Christopher Formation), silt, clay; moderate plasticity.

- Geological boundary (defined, approximate, assumed/transitional)
- TTT Known till subcrop
- <>< Esker-like ridge
- ✱ Raised beach flight (subdued morphology)

GEOLOGY (see inset below)

- Ke: Eureka Sound Formation: sandstone
- Kk: Kanguk Formation: shale
- Kh: Hassel Formation: sandstone
- Kc: Christopher Formation: shale
- Ki: Isachsen Formation: sandstone



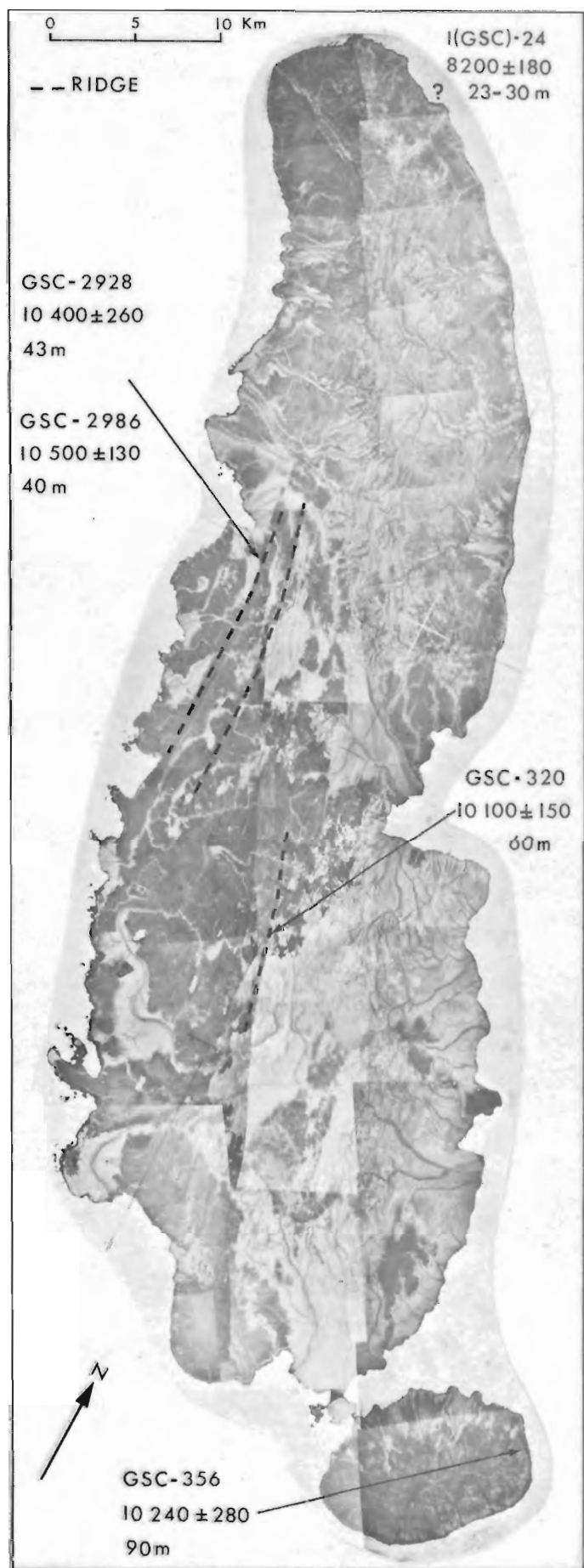


Figure 5.3. Photomosaic of Loughheed Island showing locations of radiocarbon-dated material. (RE No. 11934 EMR).

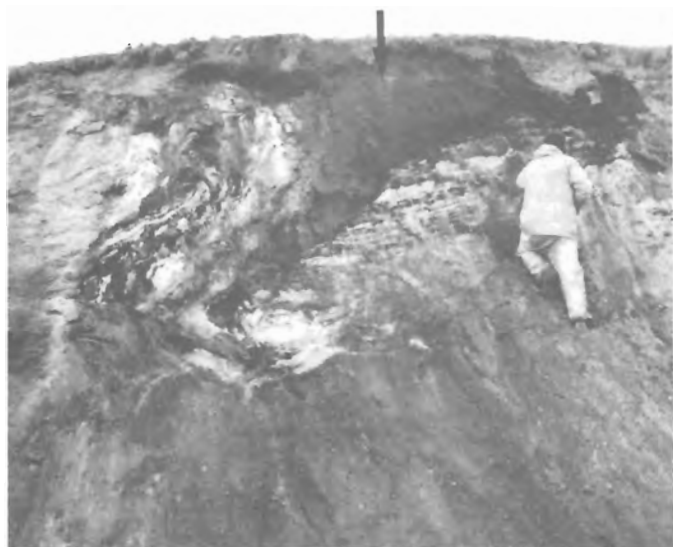


Figure 5.4. Glacially(?) deformed sandstone and shale overlain by marine pelite (base of marine sediments shown by arrow); north of Pat Bay well site, Loughheed Island (see Fig. 5.2, geology inset). GSC 203754-D



Figure 5.5. Large nivation hollow, northwest Loughheed Island, showing marine pelite (black exposures) over Hassel Formation sandstone. The section parallels a drumlinoid ridge, which forms the horizon. Marine pelecypods from the glaciomarine and marine sediments in the section (arrow) dated 10 400 ± 260 years old (GSC-2928) at top and 10 500 ± 130 years (GSC-2986) at bottom. GSC 203754-C

The thickest sections were observed in stream cuts and nivation hollows on the flanks of three linear ridges, shown in Figure 5.3. Each ridge is at least 15 km long, about 250 m wide, and up to 30 m high (Fig. 5.7) though most of the relief is developed in rock.

Marine Deposits. Inundation by a late Quaternary sea to at least 90 m a.s.l. (see marine shells, GSC 356, 10 240 ± 280 years old, at 90 m on Edmund Walker Island, Table 5.1) left few deposits at high elevation, but below 50 m



Figure 5.6. Typical contact (arrow) between sandstone, stratified sand, and glaciomarine sediment, grading up to marine sediment shown in the centre of Figure 5.5. GSC 203754-E



Figure 5.7. North end of westernmost drumlinoid ridges. Nivation hollow (Fig. 5.5, 5.6) and alignment of ridges shown by arrows. GSC 203754-B

there is a discontinuous cover. The thickest accumulations (Fig. 5.2) are adjacent to large rivers and are probably bottomset deposits of prograding deltas. Much of the sediment is silt size, even over Hassel Formation sandstone; conversely, sandy littoral sediments, probably derived from fluvial and till deposits, cover coastal areas underlain by the Kanguk Formation in the southeast.

Few beach forms exist, even at the modern shoreline, owing to lack of coarse material and to vigorous ice push which occurs in all but the most protected waters. Modern coastal geology is described by Woodward-Clyde Consultants (1981).

Fluvial and Deltaic Deposits. Most large rivers are prograding, while upstream a wide channel is cutting into the former delta surface. This surface is rarely terraced, for lateral fluctuations of a river mouth are rare on a coast where a near permanent sea ice cover moderates normal coastal processes. A similar fluvial regime in the adjacent Ringnes islands is described in more detail by Hodgson (in press).

Bedloads are dominantly fine sand or silt, except where till, glaciomarine materials, or coarse competent rock are intersected by channels.

Eolian Deposits. The singularity of wind direction aided by low precipitation results in erosion of unvegetated areas. This is well illustrated in Figure 5.3 where sand and silt from wide active river channels have been deposited up to 5 km to the south-southeast (light-toned patches), inhibiting vegetation growth (dark-toned areas), especially in the otherwise well vegetated west-central area. Erosion and deposition take place in thin sheets at any one time, thus no distinctive landforms occur.

Granular Material Sources

Gravel and sand are scarce in Quaternary deposits on Loughheed Island. Gravel covers, but does not appear to core, the esker-like landform in the southeast (Fig. 5.2), and till deposits are thin, bouldery, and generally only outcrop in section. The discontinuous lag gravel on inactive fluvial surfaces and over glaciomarine sediments could be concentrated by dragging. Sandy fluvial deposits are common, but silt content may be high. Bedrock is the best source for aggregate: sand is available as friable quartzose sandstone, and both sandstone and shale units include indurated sandstone beds suitable for riprap.

Late Quaternary Glacial History

Direct Evidence of Glacial Ice

The till described above is exposed at a number of localities in the southern half of Loughheed Island. Sandstone, the dominant clast lithology, outcrops widely on this and surrounding islands, including much of the land area immediately to the south (eastern Melville and northwestern Bathurst islands). Carbonate rocks, which are less commonly present in till, only outcrop in abundance more than 150 km away, particularly to the southeast in central and eastern Bathurst Island. This, and the orientation of glacial deformation, indicate Bathurst Island as the most likely source of the erratics.

No evidence exists of a carapace ice cover on Loughheed Island, and in particular, no meltwater channels are present. Such evidence of late Quaternary age is hardly expected, however, as little or none of the island projected above the late Quaternary sea at its maximum extent (when deglaciation most likely would occur).

Blake (1964) shows drift and till patches on Bathurst Island and on Helena Island – 100 km from southern Loughheed – otherwise the closest clearly identified till deposits are over 200 km distant: till of unknown age on northeast Amund Ringnes Island and on Cornwall Island (Hodgson, in press), and a moraine ca. 10 000 years old on southern Melville Island (Fyles, 1967). Although the till on Loughheed Island is overlain by glaciomarine sediments dated 10 000 to 10 500 years old (though not at a location directly over till), it cannot be inferred that the till is only slightly older as the glaciomarine sediment disconformably overlies till in places and generally overlies rock – not till.

Clasts in glaciomarine sediments are of similar lithologies to the till, except for the few diabase and granitic rocks assumed to originate south of Parry Channel. A minimum age for this sediment has been established from three collections of marine shells made in marine pelite over drumlinoid ridges. From the surface of the most easterly ridge (Fig. 5.3) at 60 m, *Hiattella arctica* valves are $10\,100 \pm 150$ years old (GSC-320, Table 5.1). From a massive nivation hollow in the most westerly ridge (Fig. 5.5, 5.7), *Hiattella arctica* valves were collected near the top and bottom of 3.5 m of pelite, stony near the base, and overlying a rock surface at 40 m a.s.l. (Fig. 5.6). The basal shells (from a rare concentration at this level) are $10\,500 \pm 130$ years old (GSC-2986). Shells are absent in mid-section but common in sandy partings near the top (banks of shells are exposed in the vicinity). *Hiattella arctica* valves from 20 cm below the ground surface and 3 m above GSC-2986 are $10\,400 \pm 260$ years old (GSC-2928), indicating rapid accumulation of the fine grained sediment.

A terrestrial or littoral origin on unglaciated land seems unlikely for the lenses of stratified sand; the necessary subsequent or concurrent transgression should result in poorly or unsorted deposits. Sorted deposits might be expected from submarine currents, possibly at the base of an ice shelf attached to, and fed by, an ice cap or sheet.

The most equivocal landforms are the drumlinoid ridges described above. A structural origin must be considered as they have a similar trend to the low-amplitude, regularly spaced folds described by Balkwill et al. (1977); however ridges both coincide with and slightly diverge from anticlinal and synclinal axes shown by Balkwill et al. (1979). The ridges are sharp-crested whereas the bedrock dips gently, and even appears to be homoclinal beneath one ridge. If they are glacial (or even ice shelf) bedforms, no age can be assigned.

Other equivocal evidence of glaciation comes from Marlowe's (1968) study of bottom sediments within a 50 to 100 km radius of Loughheed Island; to the west of the island he found an upper oxidized unstructured stratum, 6 to 54 cm thick, overlying more structured reduced material in which cyclical laminates resembling varves occurred. The lower layer was equated with restricted water circulation and possible glacial sedimentation at a time of lower sea level. The upper stratum is explained by more open circulation brought about by a rise in sea level. As no datable material was recovered from cores, as no means of comparison exists between rates of sedimentation of Marlowe's cores and the marine pelite on Loughheed Island (note the >3 m deposition on the drumlinoid ridge), and as a record of a rise in sea level has not been found, no correlation of cores with land exposures has been possible.

Indirect Evidence of Glacial Ice

The most plausible explanation to a glacial geologist for the 90 m of emergence which has occurred during the last 10 000 years is disintegration of an overlying or nearby ice sheet, resulting in elastic and viscous crustal responses and removal of gravitational influence on the sea (quantitative values for an arctic area (Somerset Island) are discussed by Dyke (1979)). However as Boulton (1979) questioned the value of postulating an ice sheet on the basis of 'proxy' data, other mechanisms for this uplift suggested by a geophysicist (L.W. Sobczak, personal communication, 1981), will be given. Unfortunately the form of emergence is not known, since the profusion of shells at high elevations and paucity at low elevations makes contamination of younger collections likely unless they are found in situ. Only one sample younger than 10 000 years B.P. has so far been dated: shells from 23 to 30 m a.s.l. are 8200 ± 180 years old (GSC-24).

High seismic activity which occurs below Byam Martin Channel (Basham et al., 1977; Forsyth et al., 1979), latitudinal compressive forces and longitudinal tensional forces which are predicted from four large earthquakes (Hasegawa, 1977), northeast trending mafic dikes which occur within the region (Reford, 1967), and a system of northeast trending faults and northwest trending folds and uplifts which are proposed (Sobczak, in press) may explain an uplift in the short term (10 000 years) of 90 m. In the long term (millions of years), however, general subsidence of many kilometres has been noted (Sweeney, 1977) and possibly up to 17 km (Sobczak, in press). Glacial unloading may have triggered the latitudinal forces in the last 10 000 years and thus have resulted in local uplift occurring subparallel to the north trending Boothia Uplift (Kerr, 1977). Other possible mechanisms for local uplift include diapiric evaporite intrusions, gas hydrate migration, and massive ice growth induced by temperature changes or world-wide sea level changes.

Conclusions

Glaciomarine deposits on Loughheed Island have a minimum age of $10\,500 \pm 130$ years B.P. (GSC-2986). Since this time, some 90 m of emergence has taken place. This uplift can be explained by glacio-isostatic mechanisms following removal of an ice sheet in close proximity at ca. 10 000 years ago or as a result of latitudinal compressive tectonic forces. The nearest late Quaternary ice margin proved by direct evidence is that of the Laurentide Ice Sheet 250 km to the south on southern Melville Island (Fyles, 1967). It is probable that at this time an ice sheet expanded to the sea on Axel Heiberg Island, 250 km east of Loughheed Island, and on western Melville Island, 200 km southwest. None of these is likely to induce much emergence. Blake (1964, Fig. 1) showed by striae, streamlined landforms, and meltwater channels, an undated ice flow radiating from the centre of Bathurst Island. An ice sheet covering Bathurst Island, and contiguous with ice extending to Ellesmere Island (Hodgson, 1978), or even more widespread, might depress Loughheed Island and generate floating ice or an ice shelf extending to or grounding on the island.

The age of till deposits on Loughheed Island has not been established; they may be the result of a late Quaternary glaciation, or may be older.

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Abstract

Redbed and associated coastal marine gypsum-bearing, sabkha sequences occur within the lower member of the Neohelikian Society Cliffs Formation of eastern Borden (Rift) Basin. Intricate folds and numerous steep, small-scale, faults occur in these strata on northeastern Bylot Island. These structures probably developed prior to emplacement of Hadrynian Franklin diabase dykes, and Phanerozoic sedimentation in the region. The folds developed partly in response to movement along nearby major fault zones, but the style of folding was controlled predominantly by the volume of contained evaporites. Although the preserved evaporite mineral is gypsum, data suggest that percolating water removed original halite.

Introduction

In this paper, we describe and interpret folding and faulting in the evaporitic Society Cliffs strata of northern Bylot Island, and compare these structures with those in some younger evaporitic sequences, as well as in other Bylot Supergroup strata.

Gypsum was first recognized in eastern Borden (Rift) Basin by members of Geological Survey of Canada during "Operation Bylot" in 1968 (Jackson and Davidson, 1975; Jackson et al., 1975, 1978b). Gypsum was identified subsequently near Arctic Bay and also in southeastern Borden Basin by Gildsetzer (1973, 1974) and Olson (1977).

Spectacular castellated cliffs formed along the inlets and fiord-like sounds of mountainous northern Baffin and Bylot Islands display up to 6100 m of varicoloured Neohelikian (late middle Proterozoic) strata. These strata, defined as the Bylot Supergroup (Jackson and Iannelli, in press), nonconformably overlie an Archean-Aphebian high grade metamorphic complex. The supergroup is intruded by Hadrynian Franklin diabase dykes, and is overlain unconformably by Paleozoic and Cretaceous-Eocene strata.

The general Neohelikian stratigraphic succession was established, and previous exploratory work documented, by Lemon and Blackadar (1963), and by Blackadar (1956, 1970). Recent studies include those by Blackadar (1968a-d), Fahrig et al. (in press), Galley (1978), Geldsetzer (1973, 1974), Graf (1974), Iannelli (1979), Jackson (1969), Jackson and Davidson (1975), Jackson and Iannelli (in press), Jackson and Morgan (1978), Jackson et al. (1975, 1978a, b, 1980), Olson (1977), and Sangster (1981). Paleomagnetism of the Franklin dykes has been studied by Fahrig et al. (1971) and Fahrig and Schwarz (1973).

Bylot Supergroup

The Bylot Supergroup is one of several successions that contain basic extrusive and/or intrusive rocks correlated with the Mackenzie Igneous Event (e.g. Fahrig et al., in press), and which were deposited in penecontemporaneous, possibly interconnected, basins now partially preserved along the northeast margin of the Canadian Shield. Jackson and Iannelli (in press) interpreted these basins as having formed during a 1200-1250 Ma opening of the Proto-Arctic Ocean – the Poseidon Ocean.

Contemporaneous faulting was an important tectonic element during deposition of the Bylot Supergroup in the Borden Basin, which developed as a second order structure

within the North Baffin Rift Zone (Jackson et al., 1975). Paleocurrents and facies changes suggest a major landmass to the east or southeast. Several grabens and horsts, third order structures, developed during sedimentation and were modified by subsequent faulting in Hadrynian-Cenozoic time (Jackson et al., 1975; Jackson and Iannelli, in press). The most extensive third order structure, the Milne Inlet Trough, which extends for 1200 km, is interpreted as an aulacogen.

The Bylot Supergroup comprises three groups of strata in the Borden Basin, whose contact relationships range from gradational to abruptly conformable and disconformable with one another (Jackson and Iannelli, in press):

1. The lower part of the basal Echaluk Group is chiefly fluvial and shallow marine quartz arenites that contain one to two plateau basalt sequences (e.g. Galley, 1978). The upper part of the group is chiefly marine, pyritiferous, shales that grade laterally into marine influenced delta fan complexes of conglomeratic sandstones.
2. The Society Cliffs and overlying Victor Bay formations are abruptly to gradationally conformable with one another, and constitute the Uluksan Group, a carbonate platform deposit. These strata are chiefly shelf dolostones with lesser limestones, shales and sandstones. Redbed and coastal sabkha sequences occur in the eastern part of the basin. The distribution, in northeastern Bylot Island, of Society Cliffs and Victor Bay strata (Uluksan Group), as well as the underlying Echaluk Group, is shown in Figure 6.1.
3. The lower part of the uppermost Nunatsiak Group is mostly alluvial and intertidal shales, siltstones and sandstones. Conglomeratic alluvial fan complexes, a biohermal dolostone carbonate platform, and turbidite sequences predominate in various areas, whereas limestones predominate in the southeastern part of the Borden Basin. The upper part of the Nunatsiak Group is mostly marine shelf quartz-rich sandstones.

Society Cliffs Formation

This formation consists mostly of dolostones, interbedded with lesser amounts of various terrigenous clastics. Throughout Borden Basin, an abrupt upward decrease in the terrigenous clastic content of the Society Cliffs Formation allows it to be divided into two members: a lower member of abundant terrigenous clastics interbedded with the dolostones, and an upper member containing very little terrigenous clastics (Jackson and Iannelli, in press). Most of the gypsum occurs in the eastern part of Borden Basin.

¹Precambrian Geology Division

²Economic Geology Division

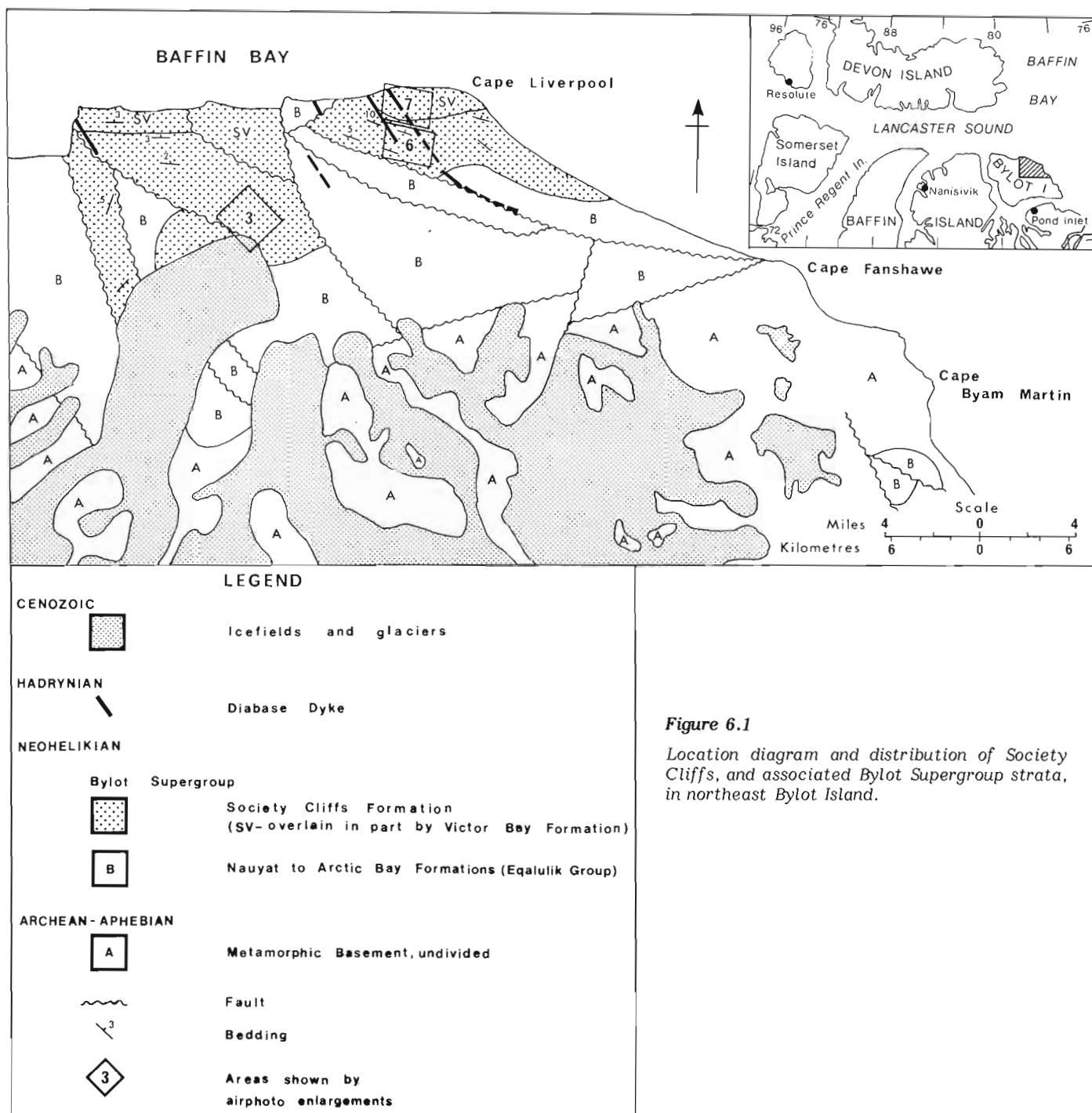


Figure 6.1

Location diagram and distribution of Society Cliffs, and associated Bylot Supergroup strata, in northeast Bylot Island.

The Society Cliffs Formation thickens from 260 m west of Nanisivik to 856 m at Milne Inlet south of western Bylot Island. Measurements and rough estimates suggest that thicknesses may range from 450-600 m on Bylot Island.

The Society Cliffs Formation of northeastern Bylot Island is typical of the formation elsewhere throughout eastern Borden Basin. However, unlike elsewhere, it is commonly poorly exposed and weathered to a mud. Where the sedimentary contacts with adjacent formations are exposed, they appear conformable and sharp to gradational. The contact between the two Society Cliffs members is gradational.

Description

Most of the formation comprises several types of laterally and vertically interfingering grey to buff and greyish brown, finely crystalline, dolostones that weather buff to locally brown, orange, and reddish brown. The dolostones commonly emit a petroliferous odour when freshly broken and contain blebs of black bituminous material. The dolostones include thin bedded to massive types with cryptalgal laminates, plus other stromatolitic dolostones, and dolostone breccias and conglomerates. Minor limestone occurs locally. Many dolostones contain shaly to sand terrigenous material. A large variety of sedimentary structures are present and include local halite casts, and soft

sediment deformation features such as slump folds. Breccias are abundant and widespread. Most breccias are related to karstification, syndepositional erosion and solution of evaporites. Dolomitization and locally faulting also may account for some of the breccia developments. The Society Cliffs commonly contains many vugs, some of which contain gypsum. Minor varicoloured replacement chert is common, and stylolites occur locally.

The terrigenous clastics of northeastern Bylot Island are chiefly red, green, and grey to black, locally calcareous shales and siltstones that occur in units up to 2 m thick, but which are locally 12 m. Terrigenous sandstones occur locally, and one arkosic conglomerate unit has dolomitic arkose and siliciclastic dolostone clasts up to 0.4 m in diameter. The terrigenous clastics are interlayered with dolostone units of similar to greater thicknesses and with laminated to massive gypsum beds up to 1 m thick. Gypsum-rich beds constitute 5-20 per cent of the strata locally.

Most strata of the lower member are arranged in shallowing-up and fining-up cycles (terrigenous clastic up into dolostone) which are also common locally in the upper member. The stratigraphic level of the boundary between the two members is gradational, and varies from one region to another because the development of gypsum and associated redbeds is lenticular. Thus the lower member comprises the whole formation on western Bylot Island, and constitutes much of the formation south and west of Maud Bight and Cape Liverpool. However, the lower member is relatively thin south of the east end of Maud Bight and west of Cape Hay (Fig. 6.1, 6.2).

Interpretation

The gypsiferous redbed cycles have been interpreted as lenticular coastal sabkha sequences developed chiefly in embayed areas within grabens (Jackson and Iannelli, in press). They may also have developed in foreland basins, and/or interbar zones as described by Borchert and Muir (1964). Major marine evaporite deposits are considered to be confined mostly to two tectonic settings (e.g. Davies, 1976):

- a) relatively stable intracratonic or interior basins on a single plate.
- b) linear rift troughs or basins of continental rift valley systems or of continental-oceanic margins, both resulting from divergent lithospheric plate movements.

Society Cliffs evaporites are clearly related to a continental-oceanic margin rift environment. In addition to the data presented by Jackson and Iannelli (in press), the ubiquitous presence of bitumen traces also supports this conclusion (Borchert and Muir, 1964). In addition, the bituminous nodular dolostone in southwestern Borden Basin (e.g. Arctic Bay area west of Nanisivik) may be equivalent, environmentally, to the typical pre-saline bituminous marl or dolostone of deep (subtidal) basins free of currents, where circulation has begun to be restricted (Borchert and Muir, 1964), rather than an intertidal deposit (Geldsetzer, 1974; Jackson and Iannelli, in press).

The lower member of the Society Cliffs Formation represents a variety of depositional environments ranging from alluvial plain to subtidal and intertidal (Jackson and Iannelli, in press). Most of the upper member represents deposits of shallow subtidal to intertidal environments.

Siliciclastic dolostones and conglomeratic strata in the upper part of the Society Cliffs Formation suggest some rejuvenation of source areas, possibly by faulting, along the Cape Hay Fault Zone and mild uplift of the Byam Martin High. This high, however, probably was not significantly uplifted until near the close of deposition of the overlying Victor Bay Formation.

Folds

Interesting fold patterns are displayed by gypsiferous Society Cliffs strata of northeastern Bylot Island (Fig. 6.1, 6.2). These structures, which are rare in the Bylot Supergroup, are excellently portrayed on several airphotos (e.g. Fig. 6.2, 6.3, 6.6, 6.7) and are best exposed north of glacier D91 (Fig. 6.3-6.5) and southwest of Cape Liverpool (Fig. 6.6, 6.7). These folds are associated with numerous steep, small displacement faults that commonly trend both subparallel to the northwest-directed Hadrynian Franklin diabase dykes and north-northeast at 75° to this northwest trend. The apparent separation varies, but is primarily right-lateral for the northwest faults and left-lateral for the north-northeast faults (see Fig. 6.3, 6.6, 6.7).

Folds southwest of Cape Liverpool (Fig. 6.1) are gentle, open and plunge very shallowly in various directions, but primarily to the northwest and northeast. The exposed fold pattern is enhanced on the airphotos by the gentle relief of the area; small displacement faults are common. The folds north of glacier D91 (Fig. 6.3, 6.4, 6.5) are relatively tight and plunge gently northwest. The dip of bedding on the flanks of the folds is up to 60°. Small displacement faults are relatively uncommon, possibly due to the nature of the outcrop and the overburden.

Slightly larger, gentle open folds are developed in Ekalulik Group strata adjacent to the Central Borden Fault Zone along the southern margin of the Borden Basin (Jackson and Iannelli, in press). Folds in Bylot Supergroup strata occur locally in several other localities adjacent to fault zones.

The present folding in the northeast Bylot Island Society Cliffs strata may be the result of several periods of deformation. Extensive evidence for syndepositional faulting throughout Borden Basin (Jackson and Iannelli, in press) suggests that some deformation may have occurred during or shortly after deposition.

Folding in the northeast Bylot Society Cliffs strata is probably related, at least in part, to movement chiefly along the Cape Hay Fault Zone that separates the north Bylot coastal plain from the Byam Martin Mountains to the south (Fig. 6.2) and possibly along the Northern Baffin Fault that trends east just north of Baffin and Bylot islands (e.g. Kerr, 1980). Evidence elsewhere in Borden Basin indicates that much, if not most, of the Bylot Supergroup folding and some faulting preceded deposition of early Paleozoic strata, which are less folded and faulted than the Bylot Supergroup, but more so than the Cretaceous-Eocene strata on northern Baffin and Bylot islands. In all these strata the folding occurs primarily adjacent to, and is probably related to, faults.

Hadrynian Franklin dykes cut across some of the Society Cliffs structure (Fig. 6.6, 6.7), but locally may be slightly deformed themselves. Also, some Franklin dykes were emplaced along fault zones. If these were subsequently reactivated, resulting deformation may have been more extensive on one side of the dyke than the other. Probably, however, most of the Society Cliffs deformation occurred some time between deposition of the formation and emplacement of the 750 Ma old Franklin dykes, possibly during the later closing of the Poseidon Ocean, that opened at 1200-1250 Ma.

Evidence for Evaporites

Fold patterns within the Society Cliffs strata in northeast Bylot Island are better developed, more intricate, and tighter than elsewhere in the Bylot Supergroup throughout Borden Basin. These folds (Fig. 6.4, 6.5) and associated faults bear a marked similarity to the structural pattern present in evaporitic sequences elsewhere, such as those in

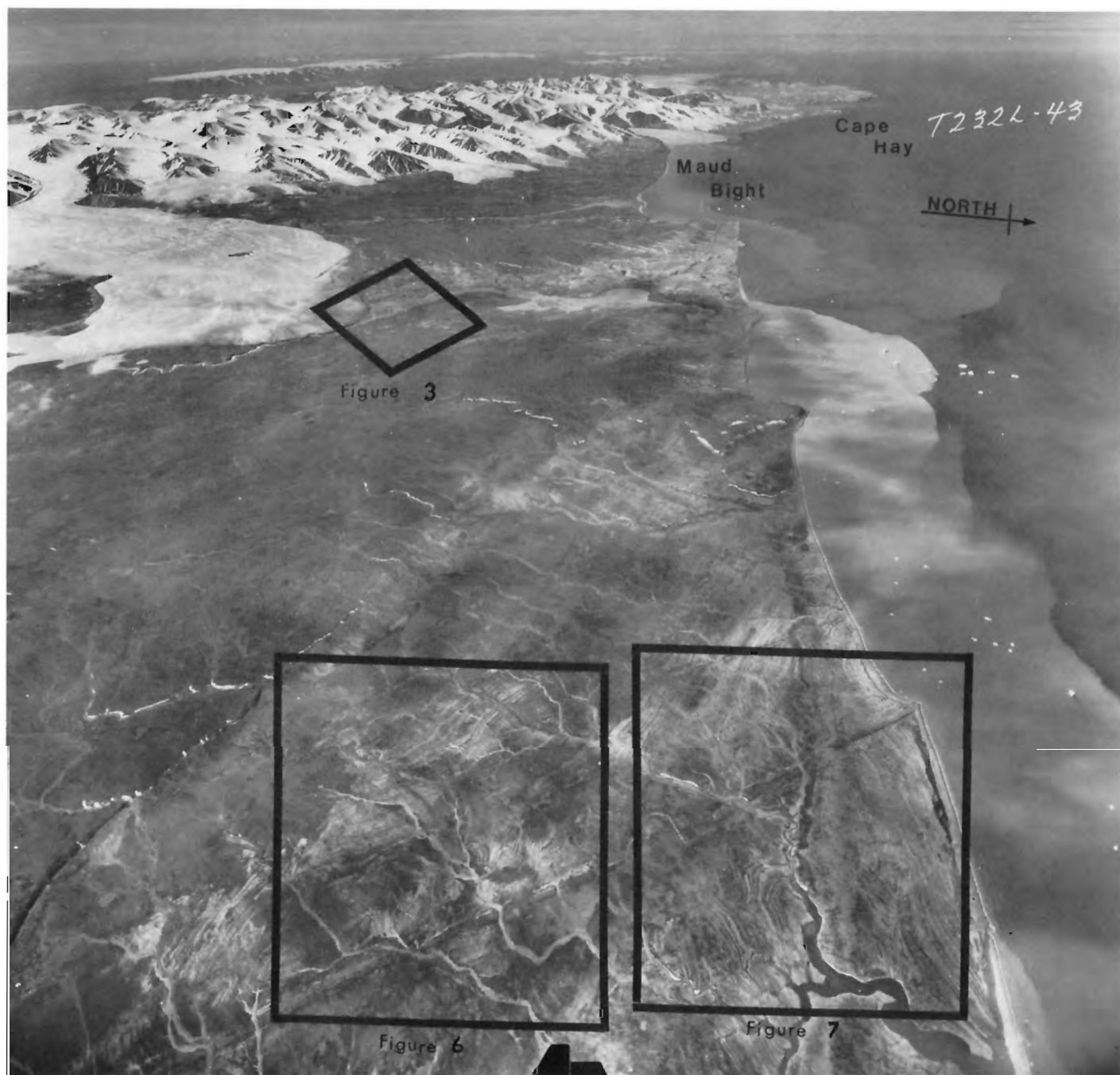


Figure 6.2. View west of north-central Bylot Island coastal lowlands, showing folded gypsiferous beds of the Neohelikian Society Cliffs Formation intruded by Hadrynian Franklin dykes. Areas shown in Figures 6.3, 6.6 and 6.7 are outlined. The Byam Martin Mountains and Borden Peninsula are in the background. Oblique airphoto T232L-43

the Mississippian of New Brunswick (e.g. Howie and Cumming, 1963) and Nova Scotia (e.g. Evans, 1965). Similar structures in evaporitic sequences elsewhere have been related to salt or gypsum tectonics (e.g. Evans, 1967). On a smaller scale some minor structures in Mississippian gypsum beds are similar to the larger pattern seen on northeast Bylot Island (Fig. 6.8). On northeast Bylot Island the fold structures, and small-scale faults, although related to regional fault activity, may also have been controlled in their formation by evaporitic sequences which may originally have contained salt bodies as well as gypsum.

Diapiric structures occur in most Phanerozoic salt deposits around the present Atlantic Ocean (Rona, 1976). In northeast Bylot Island, salt, if present, may have been too thin to form diapirs and was removed in solution, or diapirs may have formed but were removed when strata overlying the Society Cliffs formation were eroded.

The regional abundance of karst-related breccias and solution breccia, the close spatial relationship between redbed and/or sabkha sequences and extensive varicoloured replacement chert in eastern Borden Basin, and possibly the ubiquitous traces of bitumen (Borchert and Muir, 1964)

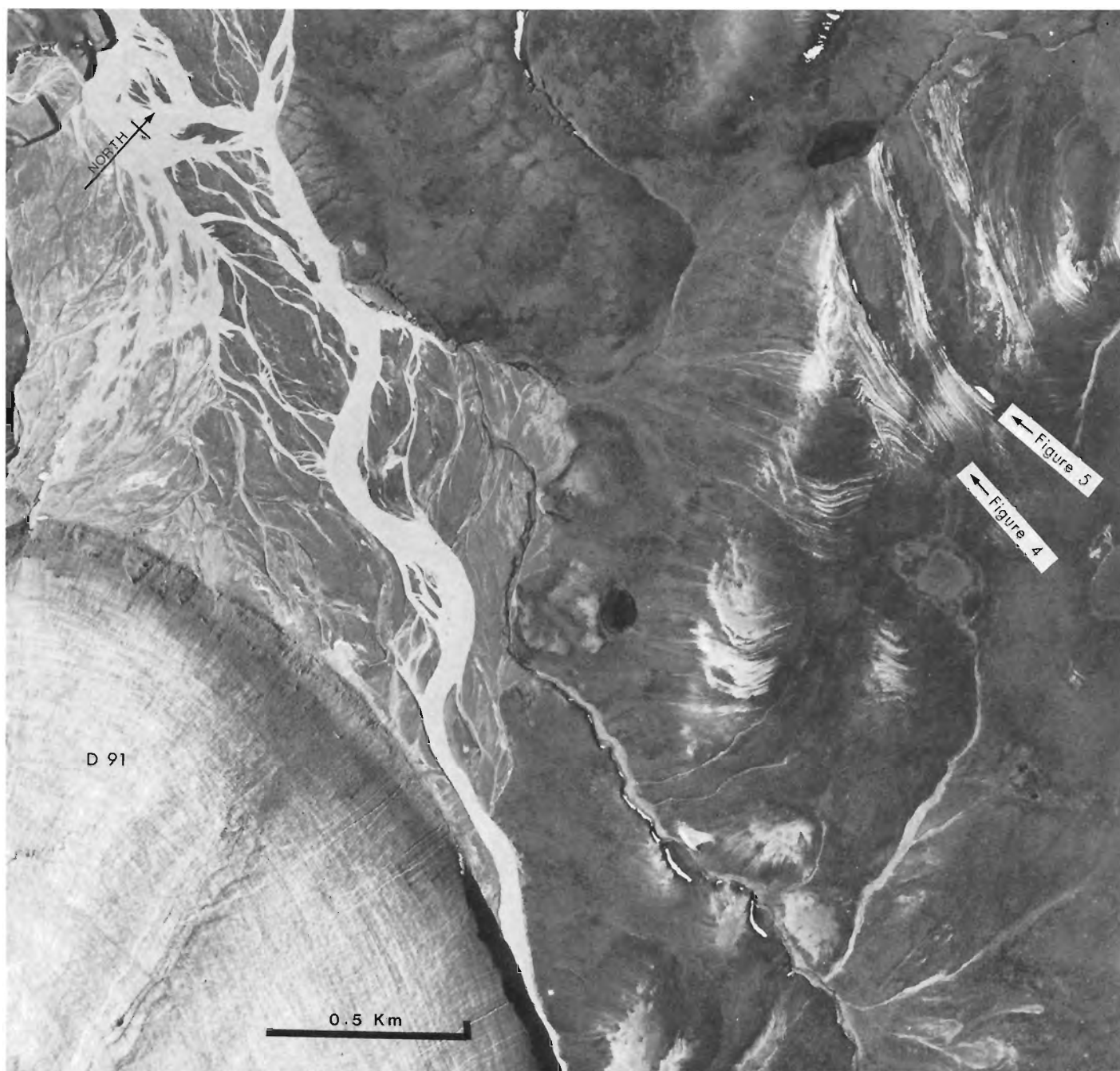


Figure 6.3. *Folds in gypsiferous Society Cliffs Formation at 73°25'N, 78°36'W, 11 km southeast of Maud Bight. The orientations of views in Figures 6.4 and 6.5 are indicated. The melting margin of piedmont glacier D91, with flow lines perpendicular to the margin, is on the lower left. Enlargement from vertical airphoto A-16214-56*

all attest to an abundance of percolating waters that could have dissolved and removed salt. The ubiquitous non-tectonic carbonate breccias, such as those in the Society Cliffs, are an indication that salt may originally have been present in the formation (e.g. B.V. Sanford, personal communication). The common occurrence of vugs, some containing gypsum, may also be an indication of removal of evaporites. The Society Cliffs sabkha sequences, together with the occurrence locally of fluorite in the overlying Victor Bay Formation, are indicators of arid climate during deposition of the Uluksan Group.

Economic Resources

Several mineral deposits in the region, with a wide range of geologic age, have aroused interest. Lead-zinc and iron deposits of economic and near economic size occur in northern Baffin Island. In addition, oil, coal, copper, uranium and other minerals have attracted prospecting activity.

Some of the highest grade iron deposits in the world were discovered in 1962 by Murray Watts in the late Archean Mary River Group at Mary River, 250 km south of northern Bylot Island, where several ages of iron ore may be present



Figure 6.4. Folds in Society Cliffs Formation accentuated by white weathering gypsiferous beds. View to west from helicopter at altitude 200 metres at 73°35'N, 78°36'W (see Fig. 6.3). GSC photo 202821-I by L.M. Cumming

(Gross, 1966; Jackson, 1966). Pods of hematite in the Society Cliffs Formation near Nanisivik may represent oxidized iron sulphides and closely resemble some of the younger ore at Mary River.

Mississippi Valley-type, silver-bearing, lead-zinc mineralization occurs chiefly in the Society Cliffs Formation within the Nanisivik area of the Milne Inlet Trough (Geldsetzer, 1974; Graf, 1974; Olson, 1977). Nanisivik Mines has been shipping concentrates of about 150 000 tonnes per year since 1977. The Nanisivik mine, next to coal mining on Spitsbergen Islands, is at present the most northerly mine in the world. The Nanisivik mineralization was discovered originally by a prospector, A. English, who accompanied Bernie (1911) on his 1910-11 trip to the region. The area was subsequently mapped by Blackadar (1956, 1970).

Major lead-zinc mineralization in Paleozoic strata on Little Cornwallis and Cornwallis islands to the northwest, and traces of lead-zinc mineralization in late Archean lower Mary River Group strata to the southeast on Baffin Island are both probably within the North Baffin Rift Zone (Jackson and Iannelli, in press). Traces of copper mineralization are associated chiefly with the Neohelikian plateau basalts but at one locality, copper occurs in red gypsiferous Society Cliffs shale (Jackson et al., 1975).

Beds of gypsum are most abundant in Bylot Island. Society Cliffs strata, are known to be as much as 2.5 m thick, and predominate with minor interbedded shale in units up to 8.5 m thick. The gypsum occurrences on Bylot Island are adjacent to a major Arctic shipping route for ore carriers travelling to and from Nanisivik. Massive gypsum (alabaster) might provide stone for carving by the Inuit who sometimes use the massive, altered, banded carbonate of the Society Cliffs Formation which occurs at baked contacts along Franklin dykes.



Figure 6.5. Hinge zones of northwest-trending folds in gypsiferous Society Cliffs beds immediately north of 73°35'N 78°36'W (see Fig. 6.3). View to west from helicopter at altitude 150 metres. Northeast margin of uplifted block of Archean-Aphebian crystalline rocks of central Bylot Island (Byam Martin Mountains) in background. GSC 202821-H by L.M. Cumming

Within the Bylot Supergroup, specks of bituminous materials and petroliferous odours are particularly common within Society Cliffs dolostones. A limestone sample from the Nunatsiak Group in southeast Borden Basin indicated a faint trace of oil when examined with ultraviolet light (Jackson et al., 1975). Age of the bituminous material has not yet been determined.

Coal beds and lenses occur within the Cretaceous-Eocene strata on Bylot and northern Baffin islands, and a small amount of coal was mined locally near Pond Inlet (Weeks, 1924). Much interest has recently been shown in the oil potential of strata believed to be of Cretaceous-Eocene age in Lancaster Sound north of Bylot Island and in Baffin Bay.

Acknowledgments

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Figure 6.6. Shallow-plunging folds in gypsiferous Society Cliffs Formation southwest of Cape Liverpool. View to the west with northwest-trending Franklin dykes (D) in foreground and background. Enlargement from oblique airphoto T232L-43 (Fig. 6.2)



Figure 6.7. View to the west of shallow-plunging folds and minor faults in gypsiferous Society Cliffs strata west of Cape Liverpool. Minor faults are both parallel to and at high angles to Franklin dykes (D). Enlargement from oblique airphoto T232L-43 (Fig. 6.2)



Figure 6.8.

Polished surfaced on a sample of intraformationally folded banded gypsum show, on a small scale, structures similar to those of the Society Cliffs Formation in north-central Bylot Island (compare Fig. 6.4). Sample (GSC no. 8577) collected by A.T. McKinnon from Mississippian strata at Hillsborough, New Brunswick; scale in inches. GSC photo 112361-C by L.M. Cumming

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EQUILIBRIUM BETWEEN U AND eU (^{214}Bi) IN SURFACE ROCKS OF CANADA

Project 760045

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Charbonneau, B.W., Ford, K.L., and Cameron, G.W., *Equilibrium between U and eU (^{214}Bi) in surface rocks of Canada; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 45-50, 1981.*

Abstract

Viewed on a regional basis average U concentrations obtained by delayed neutron activation and by gamma ray spectrometry measurement based on ^{214}Bi (eU) for surface bedrock samples are approximately equal (in equilibrium) even though individual samples may show considerable difference (disequilibrium). Arithmetic means of U are 6.9 ± 0.3 ppm and eU 6.3 ± 0.3 ppm for a total of five hundred samples from five different parts of Canada. Thus whether lithogeochemical or gamma ray spectrometric programs are carried out results should be comparable on a regional basis.

The world literature indicates that U concentrations measured at surface in many instances are representative of U concentrations at depth because most of the U is fairly tightly bound in crystalline rocks. However, some lithologies with a high percentage of labile U may have lost from surface exposures a substantial proportion of their original U endowment even though U and eU may remain in approximate balance. For example a two-mica granite in Maritime Canada was found to be depleted nearly 40% in U in the near surface 6 cm layer as compared to the underlying 6 cm layer even though U and eU (^{214}Bi) concentrations were in near balance in each layer.

Introduction

Both U and Th decay by a series of steps each characterized by emission of alpha or beta particles and in some cases gamma rays. In the process a chain of radioactive daughter elements is formed until eventually stable nuclides are reached: ^{206}Pb for the ^{238}U series and ^{208}Pb for the ^{232}Th series. Figure 7.1 is a simplified illustration of the ^{238}U decay series (Lang et al. 1962).

The U and Th concentrations of rocks can be measured in the field or laboratory by gamma ray spectrometry using a sodium iodide crystal detector to record the intensity of gamma radiation of specified energies that specific daughters emit (Hurley, 1956).

Commonly in the U decay scheme gamma rays of energy 1.76 MeV emitted by ^{214}Bi are measured, and in the Th decay scheme gamma rays of 2.62 MeV emitted by ^{208}Tl are counted.

The calculated equivalent U and equivalent Th concentrations are valid only if these elements are in radioactive equilibrium with the daughter products being measured. Equilibrium exists in a decay chain when all the daughters decay at the same rate as they are produced from the parent nuclide. Thus each daughter product in equilibrium will be present in constant proportion to its parent isotope. The gain or loss of any of the radionuclides by various geological processes will result in disequilibrium (Rosholt, 1959).

Equilibrium is generally assumed in the Th decay scheme because the short half lives of the daughters does not afford time for physical separation (Levinson and Coetzee, 1978).

In the ^{238}U series certain nuclides (Fig. 7.1) have specific chemical and physical characteristics which make them particularly susceptible to separation from the decay chain: in particular ^{238}U , ^{234}U , ^{226}Ra , ^{222}Rn , (Levinson and Coetzee, 1978). The result may be either U excess or deficiency relative to ^{214}Bi .

Magnitude of the Phenomenon of Disequilibrium

Investigators in many parts of the world have reported disequilibrium in the U decay scheme especially in near surface materials. Rocks, soils, lake and stream sediments,

vegetation, and water have been shown to be in disequilibrium in various publications, (Dyck and Boyle, 1980; Levinson and Coetzee, 1978; Titayeva and Veksler, 1978; Szoghy and Kish, 1978; Ostrihansky, 1976; Rosholt, 1959). The combined effect of much of this literature creates an impression that in surface rocks one would not expect balance between the concentration of elemental U and the calculated equivalent U concentration based on ^{214}Bi . However, in the experience of the authors, in Canada, such is not the case. Usually when U and equivalent U concentrations for a representative number of samples are averaged, the results are closely comparable. The scope of this paper is limited to a discussion of some results that illustrate this relationship between U and eU determinations in surface bedrock samples in Canada, and implications derived from these results. The purpose of the paper is to shift emphasis from a concept of widespread disequilibrium to one of equilibrium between average U and eU in surface rocks.

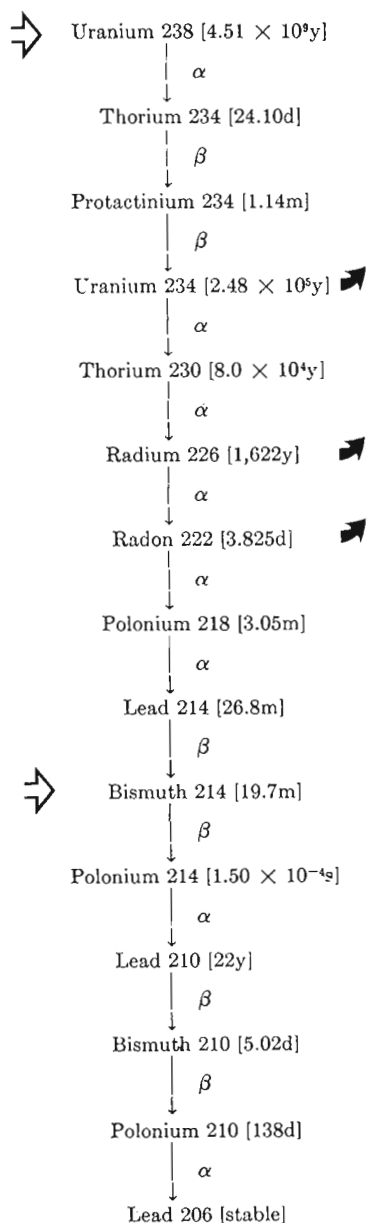
However, even if U and eU determinations are relatively close at the surface of outcrops this does not preclude the possibility that U has been depleted from the surface with equilibrium or balance between U and eU (^{214}Bi) being, maintained. This could arise because either the depletion process has been slow or daughter products such as Ra have been removed along with the labile U. (See Fig. 7.1). An example of this situation will be discussed later in this paper. Removal of U from surface rocks does not result in disequilibrium if the balance is maintained between U and eU. It is important to differentiate between the two concepts:

1. Equilibrium or balance in the surface rocks between U and eU (^{214}Bi).
2. Depletion of U from surface rocks even though eU may be in balance with U.

Field and Laboratory Investigations

Sampling

The sample material discussed relates to five different areas in Canada (Fig. 7.2): The Maritimes; Elliot Lake, Ontario; Prince George area, British Columbia; Hearne Lake, NWT; Fury and Hecla area, NWT. The samples relating to four of the areas were taken as geological hand specimens



(Principal members only; isotopes constituting less than 0.2 per cent of the decay products are omitted)

Figure 7.1. Simplified disintegration series of ^{238}U .

broken from the surface of outcrops during the course of ground follow-up investigations of airborne radiometric survey patterns. These samples were collected for subsequent petrological study, chemical analysis, laboratory gamma ray spectrometric K, eU, eTh analysis and delayed neutron activation analysis for U. More information about the sample material can be found for the Maritimes – K.L. Ford (in preparation); Elliot Lake – Charbonneau et al. (1976; British Columbia – K.L. Ford (Personal communication); Fury and Hecla area – Chandler et al. (1980). The Hearne Lake samples were taken during regional geological mapping by Henderson (1976). The study areas comprise rocks of Archaean age (Elliot Lake, Hearne Lake), Hudsonian age (Fury and Hecla) and Paleozoic age (Maritimes and British Columbia). In total five hundred samples are included in this study of which 415 are "crystalline granitoid" rocks petrologically ranging from diorite to granite

pegmatite. The remaining samples include paragneiss, volcanics and sediments. Additional sampling included a shallow drilling program carried out in the New Ross area, Nova Scotia, (Fig. 7.1) to investigate possible U leaching in the near surface. Twenty-eight 12 cm-deep holes were drilled at seven different sites. U and eU as well as Th and eTh values were averaged for the top and bottom 6 cm of each hole for each site.

Laboratory Methodology

Details of the delayed neutron procedure for determining U and Th concentrations have been given by Millard (1976). Equivalent U analysis based on measurement of ^{214}Bi gamma rays has been described by Bunker and Bush (1966, 1967). Results for delayed neutron activation and gamma ray spectrometry are estimated to be precise and accurate to approximately 5%. Over the past number of years many samples have been analyzed for U at commercial laboratories by delayed neutron activation in order to compare these values (U) with concentrations determined by laboratory gamma ray spectrometry.

U vs. eU Concentrations

Table 7.1 summarizes the results of the U and eU measurements. The average U concentrations quoted are arithmetic means (AM) $\pm$ standard error of the mean (SEM), and the geometric mean (GM). Also included is the correlation coefficient (R) for the U and eU data for the total data set, and for the individual areas. The same results are plotted in Figure 7.3. The error bars are the standard error of the mean. The U and eU data are equal (within the standard error of the means). In particular no apparent systematic depletion of elemental U with respect to equivalent U can be observed. The close overall balance between elemental U and equivalent U is most simply shown by the average U of 6.85 ppm and eU of 6.25 ppm for all 500 samples in Table 7.1. The data were sorted by granitoid and "other" lithology but since no apparent difference in the comparison between U and eU relating to these groupings was noticeable this subdivision was not included in Table 7.1.

Scatter Diagram U vs. eU

The results shown in Table 7.1 and Figure 7.3 are averages and they do not depict graphically the scatter which is present within the data. Figure 7.4 is a scatter which is present within the data. Figure 7.4 is a scatter diagram showing the comparative determinations of U and eU for all 500 samples. The best fit normal line has a slope of 0.98, a correlation coefficient of $R = 0.94$, and a Y intercept of 0.76 ppm U. The main reason for the intercept above is thought to be that eU values reported are about 1/2 ppm low due to a systematic laboratory error which will not be discussed further at this time.

Some of the scatter evident on Figure 7.4 is explainable by sample inhomogeneity. The laboratory gamma ray spectrometric analysis is based on approximately 400 grams of sample whereas the delayed neutron analysis utilizes a sample size of only about one gram. The inclusion of an extra grain or two of radioactive mineral could significantly change the U concentration by delayed neutron analysis. It has been found that variations of $\pm 20\%$ can easily occur when duplicate samples are submitted for delayed neutron activation analysis for U. However, it is felt that since there is no bias to this error only the spread on the U vs. eU plot is increased and no systematic bias is introduced. Much of the spread evident in Figure 7.4 is undoubtedly due to small scale local disequilibrium as described by Levinson and Coetzee (1978) and Ostrihansky (1976).

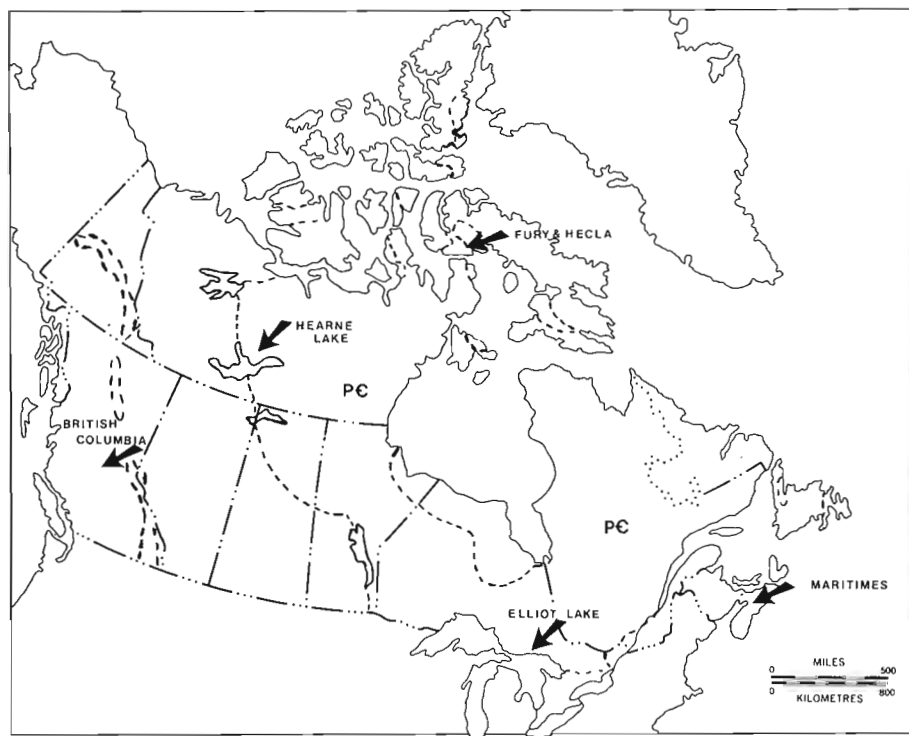


Figure 7.2. Index map showing "study" areas. New Ross area Nova Scotia is located near Maritimes arrowhead.

Table 7.1
Results of U and eU measurements for five study areas and for "total samples"

Area	No.	U (DNC)			eU (^{214}Bi)		R
		AM	SEM	GM	SEM	GM	
Total	500	6.85	0.30	3.85	6.25	0.30	0.94
British Columbia	028	11.15	1.49	6.88	10.63	1.77	0.87
Maritimes	074	11.09	1.00	7.50	9.84	0.97	0.95
Fury and Hecla	050	9.07	0.99	7.24	7.63	1.01	0.98
Elliot Lake	229	5.82	0.40	4.10	5.42	0.42	0.93
Hearne Lake	119	4.26	0.33	2.91	3.99	0.38	0.90

Shallow Drilling

Table 7.2 shows the averaged results from the shallow drilling experiment conducted near New Ross, Nova Scotia (Fig. 7.1) in two-mica peraluminous granite (K.L. Ford in preparation). Table 7.2 summarizes the results of twenty eight shallow drill cores from seven sites. Average Th values (7 ppm) are consistent for the upper 6 cm and lower 6 cm of the core. However, U values decrease near surface to 9 ppm from 14 ppm in the lower 6 cm layer (delayed neutron activation). U/Th ratios are 2.0 in the lower layer and decrease to 1.3 in the upper layer. The average eU is 13 ppm for the lower layer and 10 ppm for the upper.

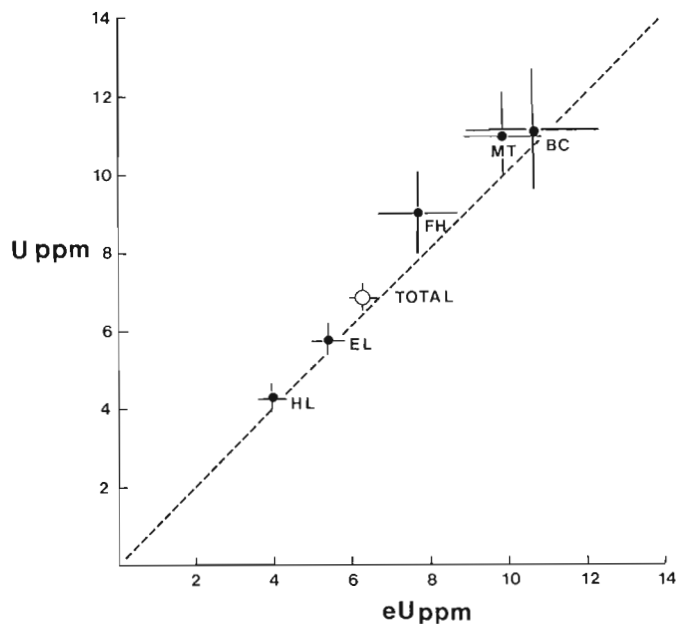
Richardson (1964) noted a decrease and more variable U/Th ratio in near surface rocks in the Conway granite of New Hampshire. Nash (1979) found that U in two-mica granite with high U and U/Th ratio is typically very labile (fine grained uraninite?) and this could explain the results described for the New Ross area. However, it should be noted that even though there is evidently a sharp decrease in U in the near surface 6 cm layer as compared to the underlying 6 cm layer (9 ppm vs. 14 ppm), comparison between U and eU in the two layers is quite close. Comparative analyses show averages of 9 ppm U vs. 10 ppm eU in the upper 6 cm and 14 ppm U vs. 13 ppm eU in the lower 6 cm of the cores. While these average values reflect overall equilibrium, individual sites can show considerable local disequilibrium (Ford, in preparation). Extreme changes in the state of equilibrium over distances of a few centimetres have been reported by many authors, in particular, Ostrihansky (1976) and Nash (1979).

Discussion

Thus an average U and eU are typically in fairly close balance at the surface of outcrops. This is not surprising in light of the fact that typically crystalline rocks have most of their U tied up in relatively resistant radioactive accessory minerals (Nash, 1979). Szalay and Samsoni (1969) showed that only about 20% of U was leachable by 0.2N HCl from a suite of European granites. Further U in the leach attains within a few hours a concentration in the 20-60 ppb range which represents an equilibrium between rock and liquid. They state: "it is obvious from these observations that the leaching of uranium from crushed rock is not a simple dissolution process. Reabsorption occurs in the course of which a large part of the dissolved U is reabsorbed on the mineral surface. The solid and the aqueous phases are in 'dynamic sorption equilibrium'".

Thus only a relatively small percentage of the U in a typical rock would be labile under surface weathering conditions and furthermore a general equilibrium could be set up in surface water of given chemical composition between liquid and solid rock. This would limit the dissolution of U. Nash (1979) stated furthermore that U mobilized by weathering often migrates to nearby grain boundaries where it is complexed with iron oxide. Thus the U would not really escape from the volume of the rock.

A satisfactory explanation for the relatively sharp depletion which occurs in some granites in the upper few centimetres even though U and eU may be in near balance must include the leaching of ^{234}U (Cherdynstev, 1971), as well as removal of ^{226}Ra (see Fig. 7.1).



HL - Hearne Lake
EL - Elliot Lake

FH - Fury and Hecla
MT - Maritimes
BC - British Columbia

Figure 7.3. Average U ppm vs. eU ppm for surface samples from the five study areas.

Levinson and Coetzee (1978) indicate that Ra can be mobile in the surface environment. "The solubility of RaSO_4 is particularly low being 2×10^{-5} g/l (~20 ppb) at 25°C. It is this very low solubility of the sulphate that is often given as a reason why radium is (or should be) immobile. However, because of the extremely low abundance of radium in water, even in water considered to have anomalously high radium contents, the above limiting solubility is rarely (if ever) attained, and radium may remain dissolved in natural waters of any composition. In other words, there is no 'solubility barrier' to the mobility of radium" (Levinson and Coetzee, 1978).

Regarding the question of whether the U determinations on surface material are representative of unweathered material at depth, numerous examples of U concentrations remaining constant from surface to depth have been published: for example, Wollenberg and Smith (1968), from the Nevada batholith; Tilling and Gottfried (1969), from the Boulder batholith, Montana; Stuckless et al. (1977), from the Granite Mountains, Wyoming. Even though there is no apparent surface related U disequilibrium in the Granite Mountains, there has apparently been extensive U loss amounting to greater than 50% of the original concentration. This conclusion has been based on lead isotope work (Stuckless et al. 1977). As well at depth, U and eU levels are actually in close balance in the Granite Mountains (Stuckless and Ferreira, 1976).

The fact that certain rocks may be somewhat depleted in U in the near surface cannot be questioned. Barbier (1968) described such a situation in granites of the central massif of France. Rogers et al. (1965) state that U has been lost from the surface rocks of the Conway granite, New Hampshire.

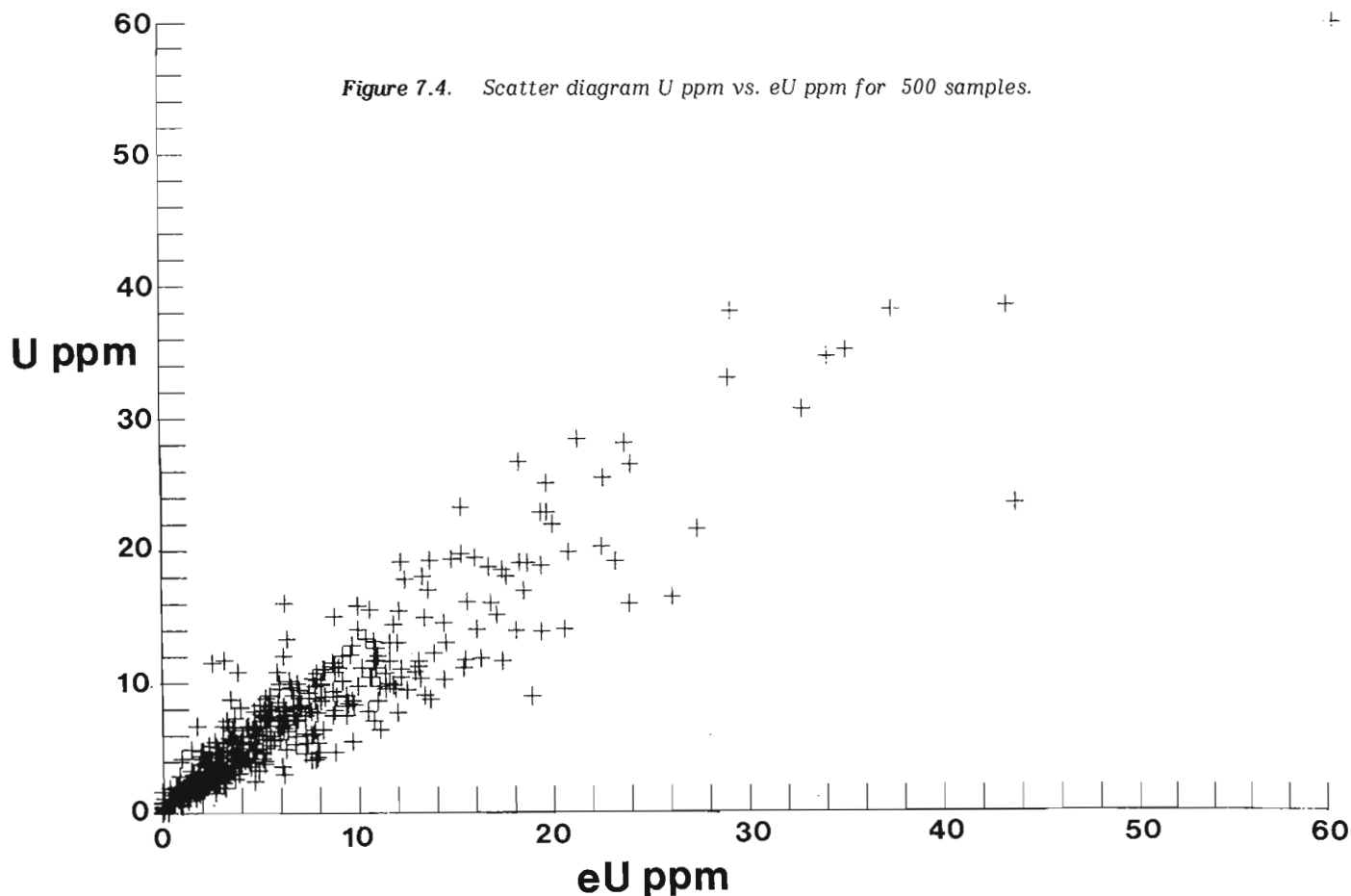


Figure 7.4. Scatter diagram U ppm vs. eU ppm for 500 samples.

Table 7.2

Average U, eU, Th, eTh for seven shallow drill cores from the New Ross area South Mountain batholith, Nova Scotia

	U ppm	eU ppm	Th ppm	eTh ppm	U/Th	eU/eTh
Upper 6 cm	9	10	7	6	1.3	1.7
Lower 6 cm	14	13	7	6	2.0	2.2

Whether or not all of this loss is a weathering phenomenon or is related to release of pressure in the near surface, that is, dilatancy (Goldrich and Mudrey, 1972) is debatable (Stuckless et al. 1977). Nash (1979) states that two-mica granites from the states of Washington and Idaho are susceptible to U loss because a high percentage of the U in the rocks is in a "labile form".

Generally deep drilling must be undertaken to profile the U concentration with depth in order to document removal of U from the near surface because as we have shown in this paper (Table 7.2) even when U has been extensively removed U and eU may be in close balance in outcrops. The extent to which U has been removed from the surface rocks will be governed mainly by the amount of the U in the rock which is labile and the extent to which U once liberated can be removed from the rock mass. If a substantial proportion of the U is labile but only liberated, for example, to grain boundaries and complexed with iron oxides, then U will not be depleted in the rock. This same argument applies to Ra.

Conclusions

Viewed on a regional basis average U concentrations obtained by delayed neutron activation and by gamma ray spectrometry measurement based on ^{214}Bi (eU) for surface bedrock samples are approximately equal even though individual samples may show considerable difference. Thus whether lithogeochemical or gamma ray spectrometric programs are carried out results should be comparable on a regional basis. The literature indicates that U concentration measured at surface in many instances are representative of U concentrations at depth because most of the U is fairly tightly bound in crystalline rocks. However, some lithologies with a high percentage of labile U may have lost from surface exposures a substantial proportion of their original U endowment.

Acknowledgments

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BOREHOLES NEAR OTTAWA FOR THE DEVELOPMENT AND TESTING OF BOREHOLE LOGGING EQUIPMENT – A PRELIMINARY REPORT

Project 740085

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Bernius, G.R., Boreholes near Ottawa for the development and testing of borehole logging equipment – A preliminary report; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 51-53, 1981.

Abstract

Three vertical boreholes 121 m deep and one 300 m deep have been drilled recently near Ottawa to test borehole logging equipment being developed by the Geological Survey of Canada. The holes will also be available for experiments by other research groups. The holes are NQ size and have been drilled 10, 20, and 70 m apart so that hole-to-hole electromagnetic and seismic studies can be made. Very high core recovery allows for detailed petrological and chemical examinations. The top 65 m of Upper Cambrian Nepean sandstone is separated from the underlying Precambrian mixed granite and gneiss by an unconformity marked by a 17 m thick weathered alteration zone. Numerous faults and fractures were recognized in the core.

Introduction

Four boreholes have been drilled recently near the existing gamma ray spectrometry calibration facilities at the Canada Centre for Mineral and Energy Technology (CANMET) complex in Bells Corners west of Ottawa (Fig. 8.1). They were drilled under contract to EMR by Longyear Canada Inc. in March 1981 as part of an ongoing program by the Resource Geophysics and Geochemistry Division of the Geological Survey of Canada to construct calibration and test facilities for the development of geophysical instruments (Killeen and Conaway, Killeen, 1978). The Geological Survey is presently involved directly or indirectly in the designing and building of several types of borehole logging probes and computerized logging equipment which will measure temperature, density, electrical, nuclear, geochemical, magnetic and acoustic properties.

The holes are drilled vertically and are spaced horizontally along a line at intervals of 10, 20, and 70 m to conduct experiments in the use of hole-to-hole electromagnetic

and seismic equipment. The site (Fig. 8.2) was chosen for easy access and also to be located far from the Canadian Explosives Research Laboratory which is on the northwest side of the CANMET complex. (Although there should be no problem, experiments involving electromagnetic transmitting equipment must receive prior approval.)

Core (47.6 mm diameter) recovery was very high – 99 per cent in most cases – which will allow detailed petrological and chemical studies to be made. Table 8.1 lists the holes and their main statistics.

Brief Geological Logs of the Drill Core

Borehole BC-81-1, which is located to the northeast of the other holes (Fig. 8.3), was drilled to a depth of 300.5 m. The first 4.9 m is overburden which was cased with NW steel casing. This is followed by 2-3 m of broken core and, to a depth of about 11-12 m, the rock is a rusty brown to grey dolomitic shale which contains pockets of pink calcite and is indicative of the Oxford Formation (Wilson, 1946). For a thickness of another 8-9 m to a depth of about 20 m the grey sandstone and sandy shale suggests the March Formation. The Nepean Formation, a relatively pure quartz sandstone of Late Cambrian age, lies below the above mentioned formations and overlies the Precambrian basement complex of granite and gneiss. Generally the Nepean is a fairly uniform light grey to white quartz sandstone containing cross-and graded-bedding in places as well as thin layers of darker minerals, vertical fossilized worm holes, patches of dark red hematite and a few thin (2-3 cm) layers of pure quartz conglomerate near the bottom of the section. The distinction between the Oxford, March, and Nepean formations is not clear due to the gradational nature of the contacts.

An unconformity occurs at a depth of 65.5 m followed by a 17 m thick zone of altered and weathered granite which consists mostly of chlorite and clay minerals or friable pinkish red feldspar. The alteration gradually becomes less noticeable at depth.

From the unconformity to the bottom of the hole, the rock is a mixture of pinkish, medium grained granitic (frequently syenitic) rock and a dark green hornblende-biotite gneiss. There are often small sections of gneiss within the granitic sections (indicated in Fig. 8.3) as well as small pods of granite within the gneissic sections. Contacts between the two rock types are usually gradational over a distance of 1 cm or less. Some of the granite appears to be a result of

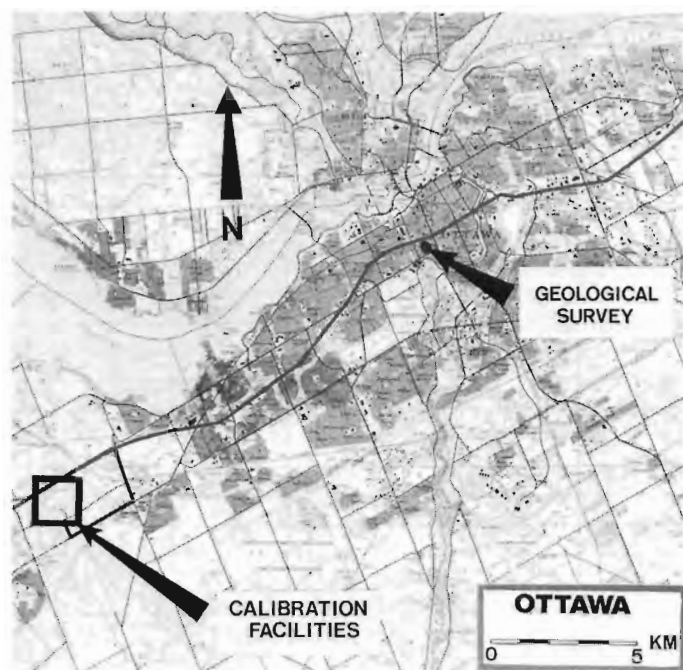


Figure 8.1. Map showing location of Ottawa borehole calibration facilities.

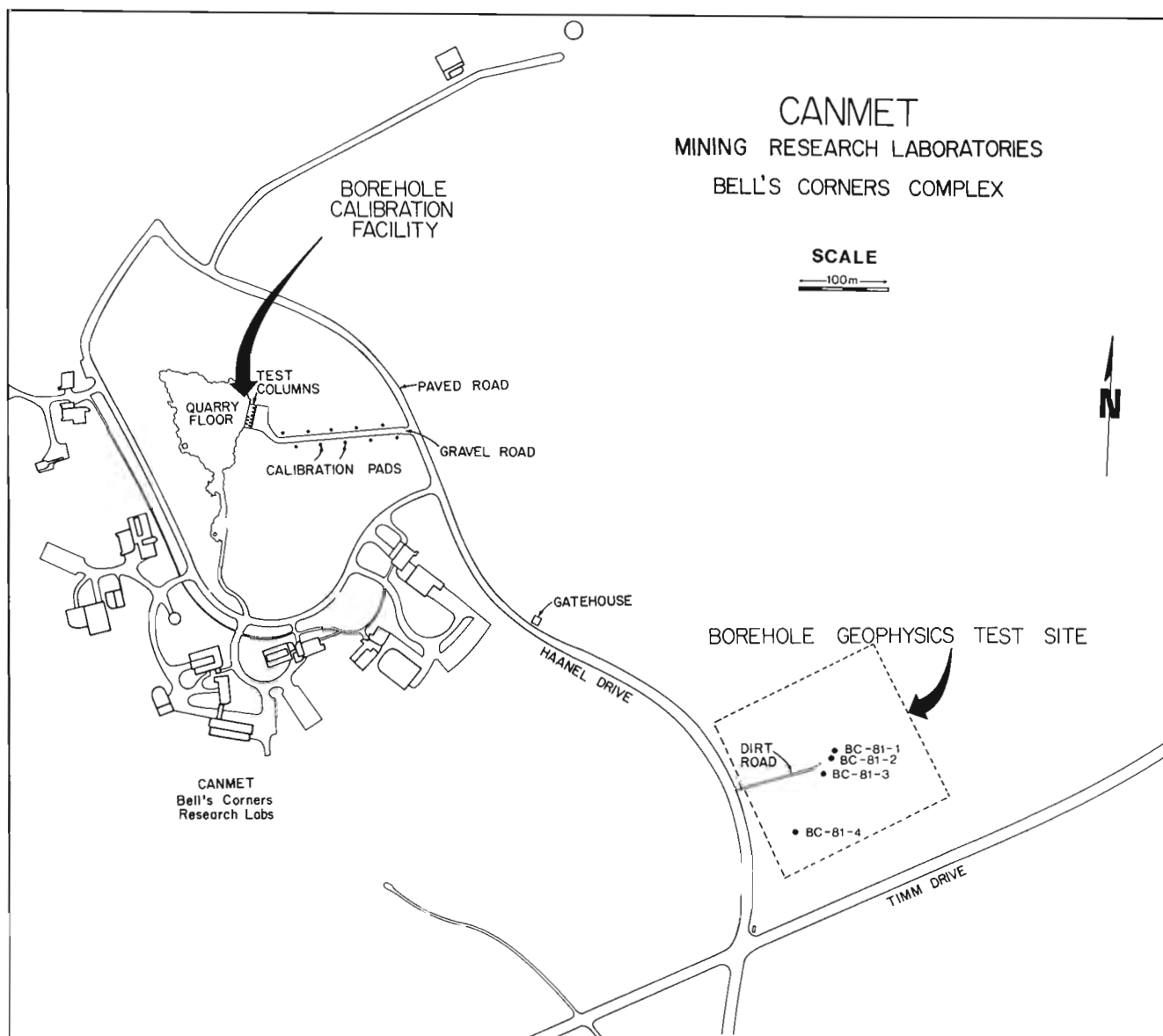


Figure 8.2. Detailed map showing location of test boreholes and existing calibration facilities at CANMET complex near Bells Corners.

Table 8.1
Statistics on test boreholes near Bells Corners

Hole No.	Depth Drilled	Diameter	Casing	Depth to Unconformity	Thickness of Alteration Zone
BC-81-1	300.5m	75.7mm	4.9m	65.5m	17m
BC-81-2	121m	75.7mm	4.9m	64.9m	17m
BC-81-3	121.2m	75.7mm	2.1m	64.6m	17m
BC-81-4	121m	75.7mm	1.5m	61.4m	17m

partial melting of the gneiss. The granite-gneiss contacts usually dip from 45° to nearly vertical; the direction of the dip is not clear.

Several small fractures and faults are visible in the core. Fractures are usually less than 1 cm in thickness and are filled with calcite. Faults frequently appear as broken core with the rock showing a considerable degree of alteration. Fault gouge 2 cm thick was noted at 104.9 m.

BC-81-2 is drilled 10 m southwest of BC-81-1 to a depth of 121 m. No difference in rock types was noticed between this hole and BC-81-1 except that the unconformity occurs at 64.9 m. There is a possible fault zone from 72.7 m to 75.1 m. Steel casing was used for the first 4.9 m.

BC-81-3 is located exactly 20 m from BC-81-2 and 30 m from BC-81-1. It is also 121 m in depth and the unconformity occurs at 64.6 m. An altered zone with fault gouge was noted from 81.4 to 81.5 m. Steel casing was used for 2.1 m.

BC-81-4 is placed 70 m from BC-81-4 and 100 m from BC-81-1. Depth is 121 m and casing is 1.5 m. A possible fault occurs at 107.3 m. The unconformity is at 61.4 m which indicates at least a partial dip to the northeast. Further drilling to the northwest or southeast would be necessary to clarify this.

A considerable amount of pyrite occurs in the gneiss. Usually it is disseminated but in places, particularly near fault or fracture zones, it appears massive. It was also seen to a limited degree in the granite and to a slight extent in the sandstone.

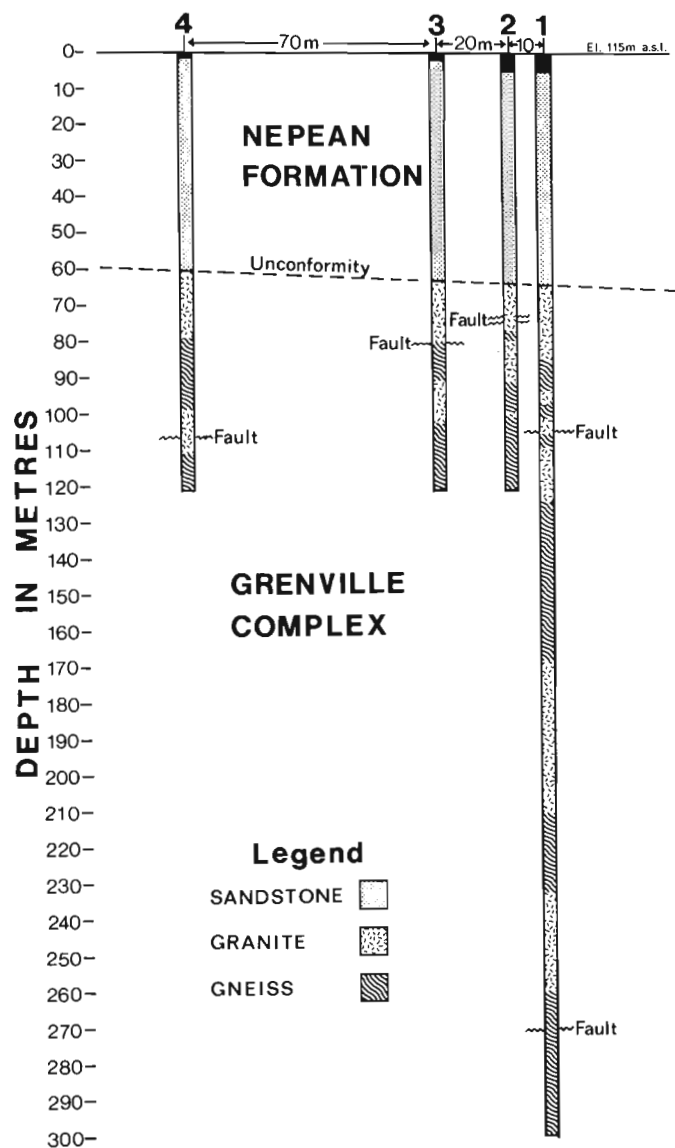


Figure 8.3. Generalized profile of test boreholes. Diameter of holes is exaggerated while vertical and horizontal scale are equal.

Conclusions

Four boreholes have been drilled near Ottawa for testing of logging equipment developed by the Geological Survey and for use by other research groups. The holes were designed to accommodate most of the existing logging equipment and anticipated new geophysical equipment which will become available. The cores from the holes will provide close geologic control which can be directly related to physical data acquired by logging. Alterations or additions to this array of boreholes may be made in the future, and any suggestions or recommendations regarding such changes are invited from potential users of this facility.

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GEOCHEMICAL INVESTIGATION OF SEDIMENT AND PORE WATER SAMPLES FROM THE NORTHEAST PACIFIC OCEAN, OFF THE COAST OF CALIFORNIA

Project 790019

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Cranston, R.E., Buckley, D.E., Fitzgerald, R.A., and Winters, G.V., *Geochemical investigation of sediment and pore water samples from the northeast Pacific Ocean, off the coast of California; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 55-61, 1981.*

Abstract

A 100 km square study area of the seafloor off Cape Mendocino, California, currently under investigation as a low level nuclear waste disposal area, has been studied for a number of geological and chemical parameters. This report contains an initial summary of geochemical results from sediment and pore water samples collected at 4 locations in the study area. The pelagic/hemipelagic sediments are soft, unconsolidated clays with the upper 20 to 30 cm consisting of brown, oxidized muds, rich in manganese oxides. Localized layers of manganese oxide at depth in the cores may represent interglacial periods. Traces of dissolved sulphide were detected in isolated zones, with the shallowest occurrence at 1 m. Horizontal variations in the data suggest that nearshore samples are more reducing than those collected farther from the continental margin.

Introduction

For a number of years Canada has been a member of the International Seabed Working Group which fosters co-operative research and information exchange on the environmental assessment of the seabed as an area for potential disposal of nuclear waste materials. As a result of our continuing research in the Northwest Atlantic on the geochemical and sedimentological nature of deep sea environment, a team of Canadian scientists was invited to join with Oregon State University to carry out similar studies

in the Northeast Pacific on board their research vessel **R/V Wecoma**. This invitation provided us with an opportunity to compare geochemical processes and ion retention capacities for sediments from various deep ocean environments.

The environmental assessment of the seafloor off Cape Mendocino, California has been underway for a number of years (Heath et al., 1979; Karlin et al., 1979; Heath, 1980, 1981a). The assessment is part of a study to determine whether the area located south of the Mendocino Fracture Zone and west of the Delgada Fan (Fig. 9.1a, b) is potentially suitable as a disposal area for low level nuclear waste (Heath et al., 1979; Carter, 1980). The 100 km square study area was selected because it has water depths greater than 4 km, is within 200 nautical miles of the coast, appears to be a quiescent depositional zone and contains fine grained pelagic and/or hemipelagic sediments. This is an initial progress report of the geochemical investigation of sediment and pore water samples collected in March 1981 (Heath, 1981b) on board the **R/V Wecoma** from Oregon State University.

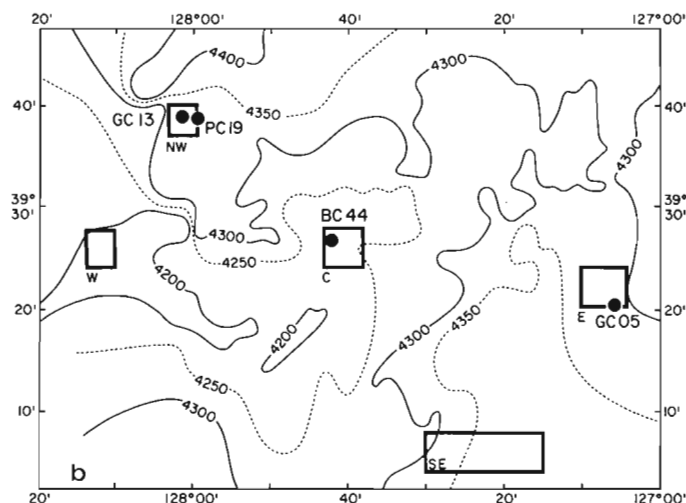
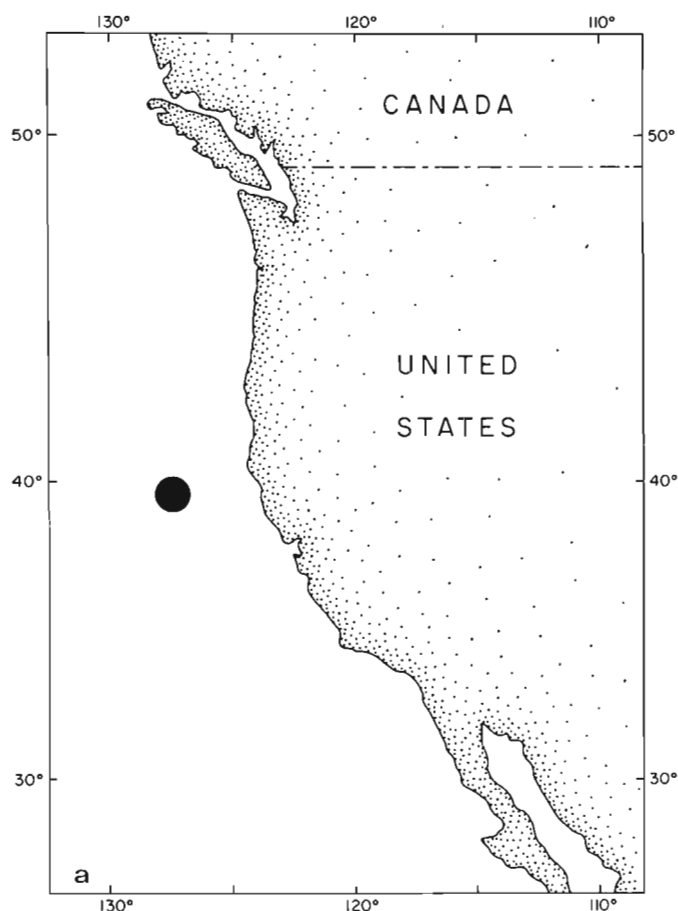


Figure 9.1a. Location of study area in the Northeast Pacific Ocean.

Figure 9.1b. Detailed study area, showing sampling locations.

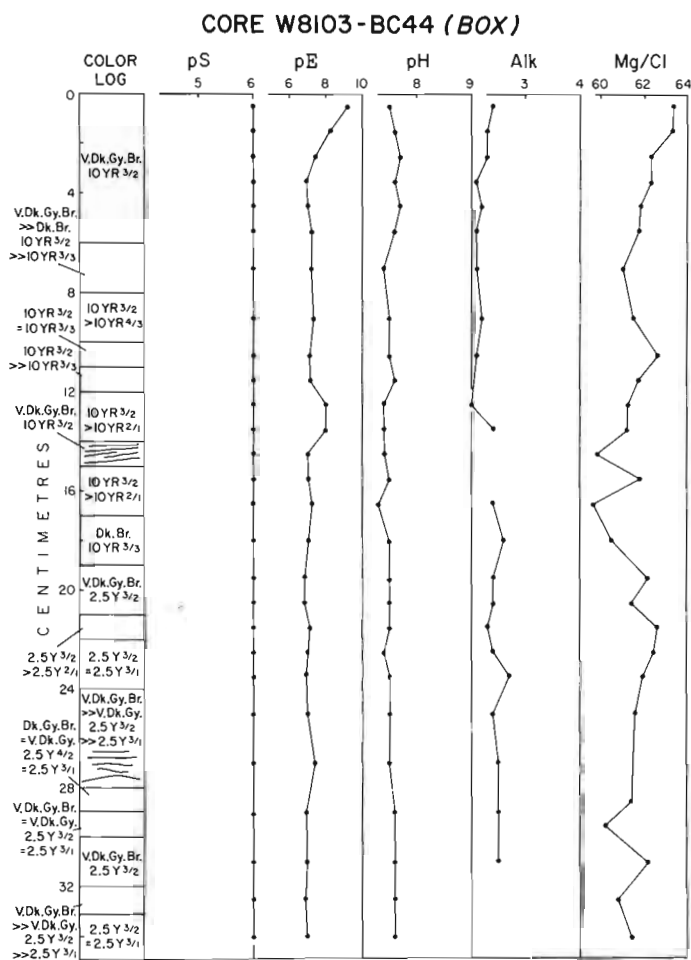


Figure 9.2. Core log and geochemical data for core BC-44.

Acknowledgments

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Methods

Core Sample Collection

Three types of coring devices were used to obtain sediment samples. A 0.5 m box corer recovered undisturbed sediment samples to a depth of 35 cm. A 10 cm diameter PVC gravity corer was used to obtain sections of undisturbed sediment up to 5 m in length. Cores of up to 16.6 m long were obtained with a double piston Benthos corer. The core was retained in a 6.5 cm diameter plastic liner inside a steel barrel.

Shipboard Laboratory Operations

A 7 m refrigerated shipping container was used for subsampling and storing the core samples. The cooled lab ($4 \pm 2^\circ\text{C}$) contained two environmental chambers that were continuously flooded with nitrogen gas. Sections of cores were extruded in the nitrogen-purged glove box immediately following collection. Core descriptions and colour logging were done prior to selecting subsample intervals for geochemical analyses. Portions of each selected interval were stored in separate vials for pH, pE and pS^{2-} analyses. The remainder of the interval (100 to 200 cm^3) was squeezed under nitrogen pressure to extract pore water.

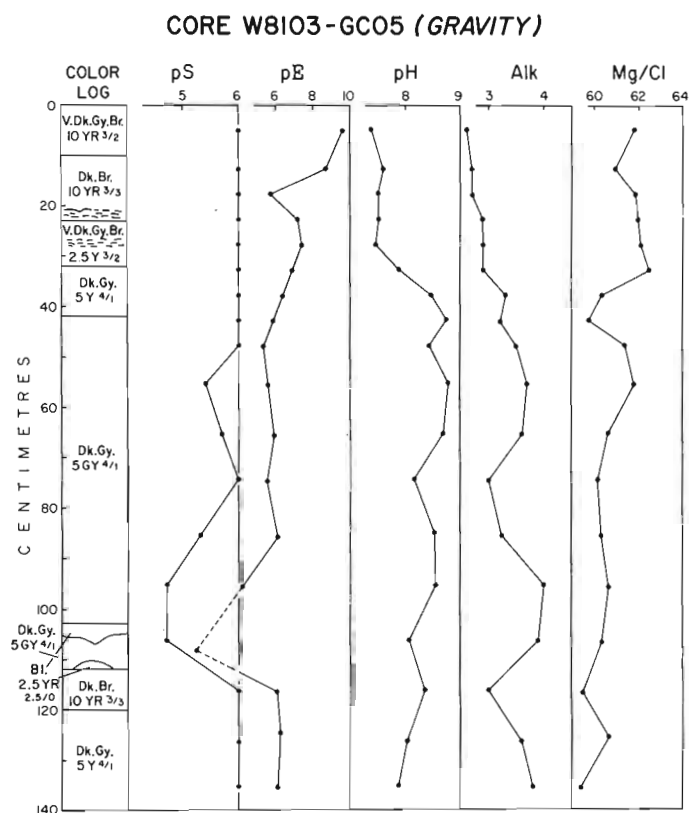


Figure 9.3. Core log and geochemical data for core GC-05.

The squeezers were constructed of cast acrylic fitted with a screwcap and a base plate which contained a 47 mm membrane filter holder. A manifold holding 12 squeezers was connected to a liquid nitrogen tank (PGS-45) through a double stage regulator (CGA-580) set at 3 atmospheres. Ten to 50 mL of filtered pore water was collected depending on the amount of sediment, the type of sediment and the water content.

A 50 g portion of sediment was collected from each interval and stored in a capped vial for pH and pE measurements. A combination pH electrode was standardized with Palitsch buffer at pH 8.2 (Whitfield, 1969). Careful placement of the electrode in the subsample was important in order to exclude gas pockets. Stable readings were routinely reached within two minutes after inserting the electrode. Reproducibility was ± 0.05 pH units. Redox potential (pE) was measured in the same vial as pH using a combination platinum electrode standardized in Zobell solution (Zobell, 1964). The electrode was carefully inserted to eliminate gas pockets. Voltages were recorded at 10 s intervals over 90 s. Redox potential was calculated from E_H voltages relative to the standard hydrogen electrode (Orion Research, 1980). Both pH and pE analyses were carried out under nitrogen gas at 4°C .

Total dissolved sulphide as H_2S , HS^- and S^{2-} , was determined on 5 mL portions of wet sediment which had been immediately spiked with antioxidant buffer (Frant et al., 1970) and homogenized with a Vortex mixer. The buffer raised the pH to 13.2, converting H_2S and HS^- to S^{2-} during a 1 hour waiting period. A silver sulphide specific ion electrode was placed in the buffered mud and allowed to equilibrate for 30 s. Drift-free voltage responses were compared to a calibration curve to calculate sulphide concentrations (Orion Research, 1978). Detection limit for the method is $1 \mu\text{M}$ ($\text{pS}^{2-} = 6.0$), with a reproducibility of $\pm 0.2 \text{ pS}^{2-}$ units.

Table 9.1
Mean concentrations for geochemical parameters
for 4 cores from the Northeast Pacific

Parameter	x	GC-13 sd	n	x	PC-19 sd	n	x	BC-44 sd	n	x	GC-05 sd	n
depth (cm)	190	130	48	1110	302	72	14	10	29	57	40	18
pH (sed)	7.8	0.4	48	8.3	0.2	72	7.5	0.1	29	8.1	0.5	18
pE	6.9	0.8	48	5.3	0.6	72	7.3	0.5	29	6.1	1.6	18
pS ²⁻	5.9	0.2	18	5.5	0.5	72	6.0	0.0	29	5.8	0.4	18
pH (water)	7.6	0.1	36	7.9	0.1	70	7.9	0.1	25	7.9	0.1	18
alkalinity (meq/L)	2.7	0.2	36	3.7	0.4	69	2.3	0.3	25	3.3	0.5	18
sulphate (mM)	29.0	0.4	36	28.9	1.2	69	29.3	0.6	25	29.9	1.5	18
Na/Cl	552	5	36	547	6	70	554	7	28	544	10	18
K/Cl	20.9	0.3	36	20.8	0.7	70	21.1	0.4	28	22.5	0.6	18
Ca/Cl	22.5	0.6	36	23.1	0.4	70	22.4	0.2	28	21.6	0.2	18
Mg/Cl	62.2	1.0	36	61.7	1.2	70	61.7	0.9	28	60.9	1.0	18
Water (%wet)	53.9	2.7	48	50.5	2.9	68	57.0	6.1	29	52.8	5.5	18
nitrate (μM)	50.6	33.7	36	39.3	13.9	63	68.5	17.0	20	27.8	10.0	18
nitrate (μM)	0.73	0.29	36	0.96	0.32	43	1.32	0.59	20	1.12	0.88	18
phosphate (μM)	8.15	3.21	36	15.3	2.9	63	9.55	3.75	20	14.1	5.8	18
silicate (μM)	281	59	36	200	58	65	367	62	20	452	29	17
org. C (%)	0.52	0.07	36	0.48	0.08	73	0.74	0.08	29	0.75	0.08	18
total C (%)	0.62	0.19	36	0.96	0.88	73	0.78	0.08	29	0.90	0.17	18
Mn(Pw, μM)	63	75	36	152	41	70	34	37	23	79	39	18
Mn(r.lch,ppm)	4300	3600	36	1700	3200	73	14 100	7000	29	2800	3500	18
Fe(r.lch,ppm)	2200	1000	36	1600	700	73	4000	300	29	2300	800	18
Mn(a.lch,ppm)	350	270	36	970	1400	73	980	430	29	350	240	18
Fe(a.lch,ppm)	550	140	36	720	160	73	850	130	29	790	100	18
Cu(a.lch,ppm)	28	23	36	39	46	73	48	8	29	36	12	18
Zn(a.lch,ppm)	10	6	36	6.6	3.9	73	28	7	29	15	10	18
NOTE: All data are measured either in pore water or in sediments. Leach data are shown as (r.lch,ppm) for reducing leach in ppm, dry weight, while (a.lch,ppm) represents acid leach data in ppm, dry weight.												

A 2 mL portion of filtered pore water was immediately analyzed for pH, total alkalinity and dissolved sulphate. A combination pH microelectrode was used to determine pore water pH to ± 0.05 pH units. Total alkalinity was determined by titrating with 0.008 M HCl in 0.6 M NaCl, using an auto-titrator and a programmable circuit closure. The alkalinity equivalence point was determined from a Gran plot (Edmond, 1970). The reproducibility was ± 0.02 meq/L.

The alkalinity sample was then titrated with Pb²⁺, causing PbSO₄(s) to form. Excess Pb²⁺ was measured with a specific ion electrode (Goertzen and Oster, 1972; Orion Research, 1977). The precipitate was collected on 0.4 μm Nuclepore filters and later analyzed for Pb and Ca by flame atomic absorption spectroscopy to confirm the amount of Pb precipitated and to ensure that there was no interference due to CaSO₄(s) forming. The sulphate equivalence point was determined by regression, with a reproducibility of ± 0.2 mM.

Land-based Analyses

Pore water samples were frozen and stored for nutrient analyses by autoanalyzer at Oregon State University. Subsamples for dissolved trace metals, organic and total carbon and major cations were stored in sealed containers at natural pH and 4°C and returned to our lab for analyses. Sediment subsamples were stored for subsequent analyses of water content, total and organic carbon, particle size, mineralogy, and leachable and total metal. Metal analyses for the water and sediment are being done by atomic absorption spectroscopy. Dissolved organic and total carbon will be determined with a liquid carbon analyzer. Total and organic carbon in sediments have been determined with a combustion analyzer. Particle size will be determined by Coulter counter and mineralogy by X-ray diffraction.

Table 9.2
Mean concentrations for geochemical parameters
for upper 35 cm of 3 cores

Parameter	GC-13			BC-44			GC-05		
	x	sd	n	x	sd	n	x	sd	n
pH(sed)	7.2	0.2	12	7.5	0.1	29	7.6	0.2	6
pE	7.6	0.7	12	7.3	0.5	29	7.6	1.4	6
pS ²⁻	-	-	-	6.0	0.0	29	6.0	0.0	6
pH (water)	7.7	0.0	12	7.9	0.1	25	7.8	0.0	6
alkalinity (meq/L)	2.5	0.1	12	2.3	0.3	25	2.8	0.1	6
sulphate (mM)	29.0	0.5	12	29.3	0.6	25	29.4	0.6	6
Na/Cl	554	3	12	554	7	28	553	5	6
K/Cl	21.3	0.2	12	21.1	0.4	28	21.8	0.2	6
Ca/Cl	22.1	0.2	12	22.4	0.2	28	21.4	0.7	6
Mg/Cl	62.3	0.4	12	61.7	0.9	28	62.1	0.7	6
water(%wet wt)	55	2	12	57	6	29	55	4	6
nitrate (μM)	76.8	7.9	12	68.6	17.0	20	34.8	13.3	6
nitrate (μM)	0.7	0.4	12	1.3	0.6	20	1.3	1.1	6
phosphate (μM)	4.7	0.6	12	9.6	3.8	20	8.7	2.1	6
silicate (μM)	227	37	12	367	61	20	466	10	6
org. C(%)	0.55	0.05	12	0.74	0.08	29	0.70	0.11	6
total C(%)	0.57	0.06	12	0.78	0.08	29	0.87	0.10	6
Mn(pw, μM)	<5	-	12	35	36	23	32	29	6
Mn(r.lch,ppm)	8200	2000	12	14 100	7000	29	6700	3400	6
Fe(r.lch,ppm)	3400	300	12	4000	300	29	3400	300	6
Mn(a.lch,ppm)	140	50	12	980	430	29	610	40	6
Fe(a.lch,ppm)	700	70	12	850	130	29	790	160	6
Cu(a.lch,ppm)	25	9	12	48	8	29	46	8	6
Zn(a.lch,ppm)	16	5	12	28	7	29	25	12	6
NOTE: All data are measured either in pore water or in sediments. Leach data are shown as (r.lch,ppm) for reducing leach in ppm, dry weight, while (a.lch,ppm) represents acid leach data in ppm, dry weight.									

Results

Detailed geological and geochemical studies were done on 4 cores from the study area. They were split and logged visually to take note of colour variations (Munsell soil colour codes), obvious changes in texture, and occurrences of internal structures. A summary of the core logs and some geochemical parameters is presented in Figures 9.2 to 9.5. A summary of chemical data is presented in Tables 9.1 and 9.2. Due to space limitations, the detailed profiles and data tables containing all analytical results to date are not included in this initial report.

Discussion

The box corer provided an undisturbed water-sediment interface sample (BC-44, Fig. 9.2). The upper 6 cm consisted of fine, very dark grey-brown clay. This layer was oxidizing, allowing nitrification to occur which produced a nitrate maximum at 3 cm. Below 6 cm, the clay or silty clay was streaked with very dark grey sediments, identified as a MnO₂ precipitation zone. A dark zone at 15 cm was also enriched in leachable Mn. A continual decrease in nitrate below 10 cm is indicative of anaerobic respiration resulting in

denitrification. Phosphate concentrations in the pore waters increased steadily from the sediment-water interface, a result of phosphorous release from organic matter decomposition.

The gravity core taken from the eastern margin of the study area (GC-05, Fig. 9.3) had similar sediments to those of the box core in the upper 21 cm. These sediments were overlying clay or silty clay, mottled with very dark grey sediments to 32 cm. Leachable metal concentrations showed anomalously high values for Fe and Mn, indicating metal oxide-oxyhydroxide precipitation. Nitrate concentrations decreased and dissolved phosphate and manganese values increased to 32 cm, indicating anaerobic respiration with denitrification. At 32 cm, the colour abruptly changed to dark grey, with no mottling to 103 cm. The pore water pH, alkalinity and nitrate increased at 32 cm and coincided with a 2 to 5 fold decrease in leachable Fe and Mn, suggesting a zone capable of reducing Fe(III) and Mn(IV). Two small lenses of black grainy minerals were found at 103 and 112 cm. Dissolved sulphide concentrations reached a maximum at 90 to 112 cm. Minima in dissolved Ni, Cd and Cu profiles at 90 to 112 cm also suggest that these metals may be forming sulphide minerals or may be being scavenged from solution. Sediments below 112 cm appeared to be similar to those found in the upper part of the core.

CORE W8103-GC13 (GRAVITY)

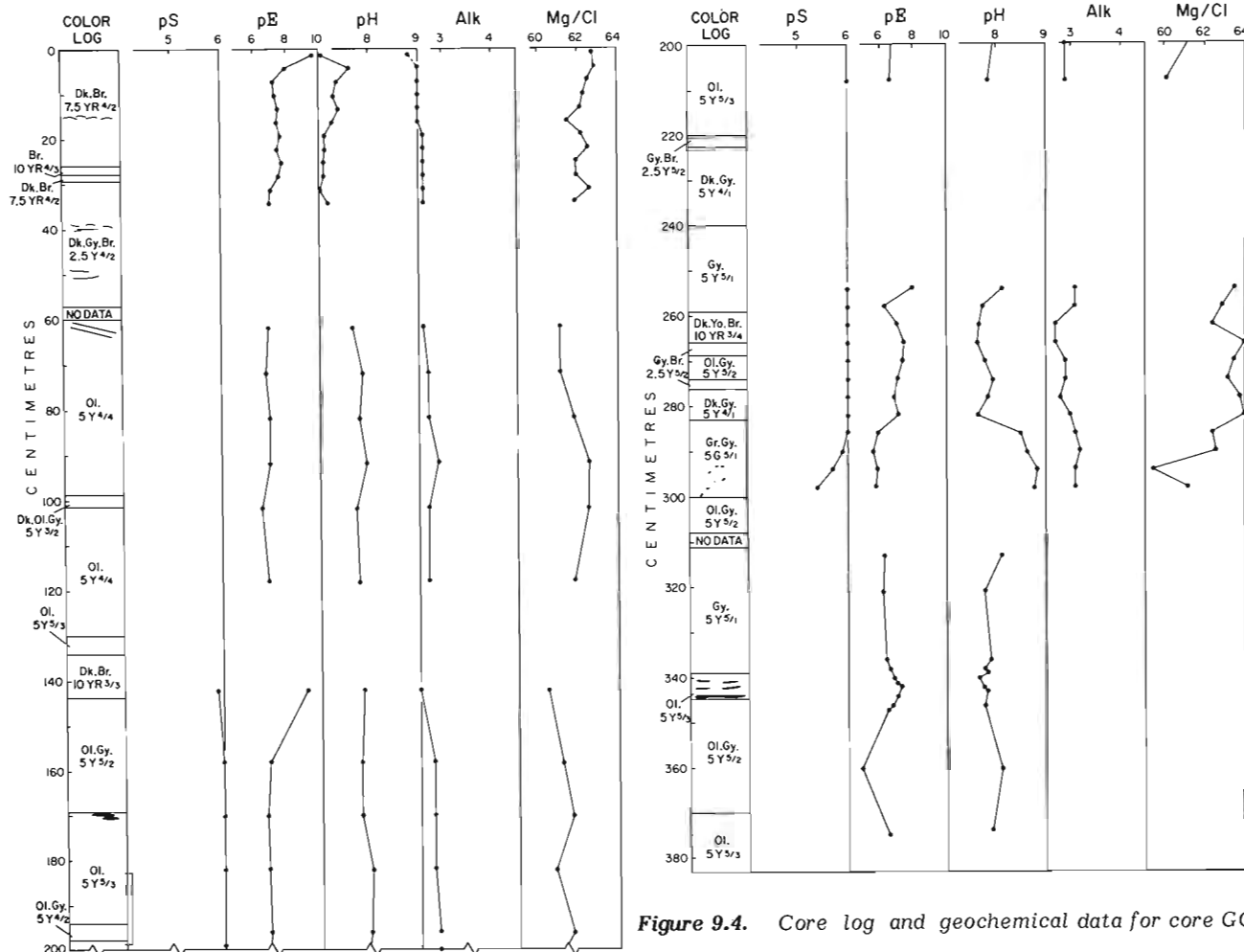


Figure 9.4. Core log and geochemical data for core GC-13.

A 383 cm gravity core taken from the northwest area contained watery dark brown clayey sediment in the upper 26 cm with a few darker bands and streaks at 26 to 29 cm, indicating some bioturbation (GC-13, Fig. 9.4). An abrupt transition to olive coloured sediments occurred at 60 cm where dissolved Mn was first detected. The Mn values increased to 120 cm, then dramatically decreased at 140 cm. A small zone of dark brown clayey sediment was found at 134 to 140 cm, coinciding with Fe and Mn enrichment in the solid phase and a marked removal of phosphate from the pore water, indicative of phosphate complexes forming with Fe and Mn. Olive, grey or dark grey sediments persisted to 259 cm while dissolved Mn steadily increased in this zone. A very sharp colour change to dark yellow-brown occurred at 259 cm, where leachable Fe, Mn, Cu and Zn concentrations increased. Phosphate was also removed from the pore water. It appears that Fe and Mn precipitation removed a number of cations and anions from solution due to coprecipitation and/or adsorption. These samples will be studied in detail with a scanning electron microscope and an energy dispersive X-ray analyzer at a later date. Some indication of bioturbation was seen in the grey-brown sediments from 266 to 274 cm. The remainder of the core consisted of greenish grey, olive grey or grey clayey sediment. At 284 cm, disseminated black spots were visible, and a drop in pE coincided with the occurrence of dissolved sulphide and a marked increase in pH, dissolved in Mn and phosphate.

Piston core 19 was collected from the northwest section of the study area, within 4 km of GC-13. The lower half of the core, from 680 to 1660 cm was logged and subsampled for geochemical study (Fig. 9.5). The entire core log depicted the sediment to be grey or greenish grey, with an exception at 1599 to 1634 cm. A series of alternating bands of dark brown, brown and dark grey-brown clays and silty clays were found overlying olive grey sediments. These brownish sediments are unique because they resemble material seen only in the upper sections of GC-13. Geochemical data for pE, pH, alkalinity, dissolved Mn, silicate and leachable metal also resemble those found in the upper metre of the sediment column. Preliminary palynological analyses indicate that the dark brown sediment in the upper metre is Late Holocene in age. The lowest section of the piston core intersects a glacial-interglacial sequence with the greenish grey sediment at 1588 cm being representative of a fossil-poor glacial sediment while the dark brown sediment immediately below is fossil-rich and represents an interglacial period (Mudie, personal communication). Further paleontological studies will be done to determine the origin of the sediments.

Dissolved sulphide was detectable below 900 cm, with a concentration maximum occurring at 1010 cm where black spots were observed. Dissolved silicate, pE and water content values were lower in the piston core than those found in the shallower cores, while pH, alkalinity, phosphate and dissolved Mn were noticeably higher in the piston core (Table 9.1).

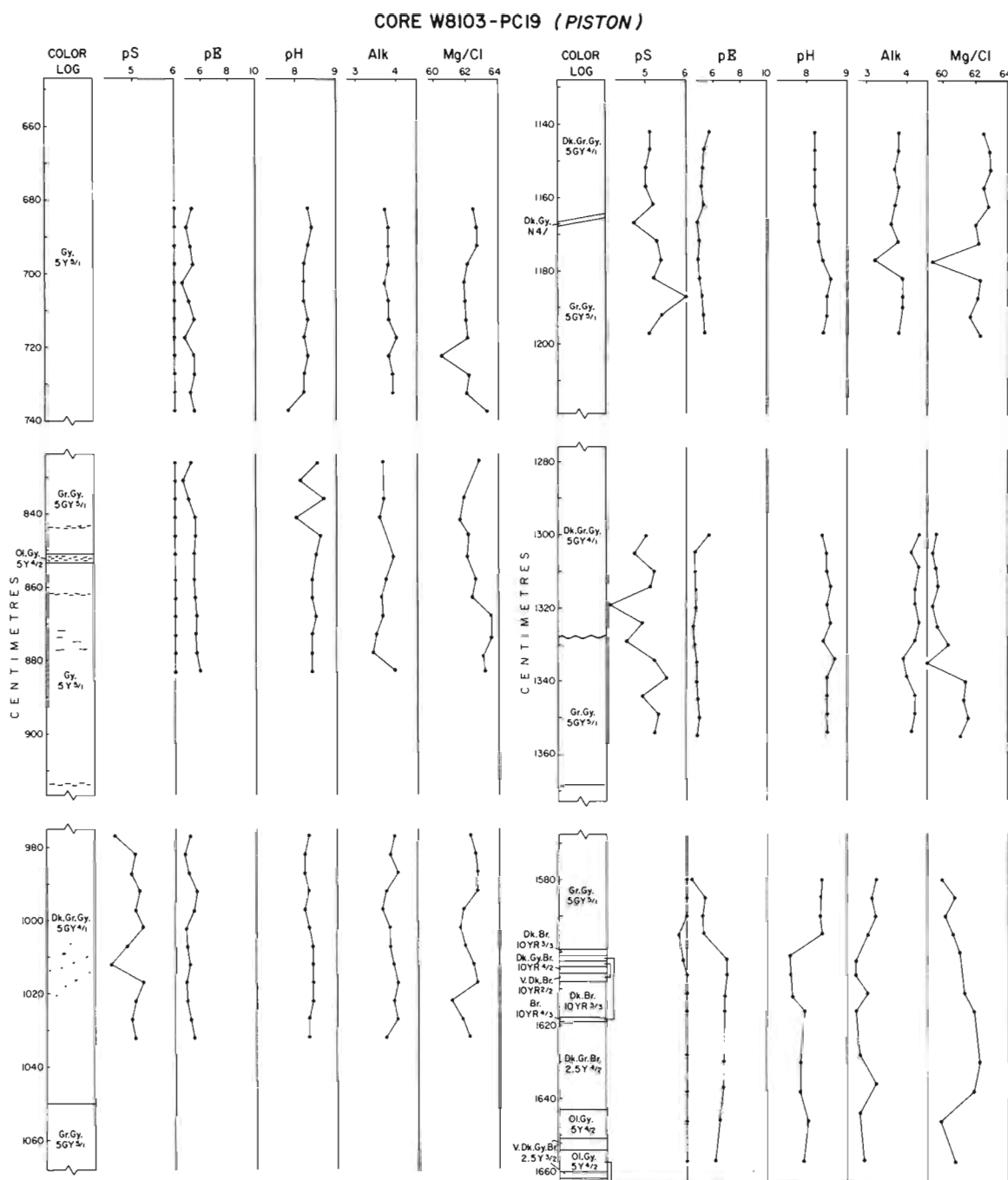


Figure 9.5. Core log and geochemical data for core PC-19.

Mean concentrations for geochemical parameters measured in the upper 35 cm of the sediment column have been compiled to compare the surface sediments on a regional basis (Table 9.2). A consistent difference appears between GC-13 and the other two cores. There appears to be less organic and total carbon at station 13, causing a less reducing environment to be present. Dissolved Mn, silicate, phosphate, pH and nitrite are lower at GC-13 than the other areas, while nitrate is marginally higher. The marked decrease in dissolved silicate at GC-13 suggests that less biogenic silica is available for regeneration.

Summary

1. The upper 20 to 30 cm of the sediment column consists of dark brown clay overlying grey and/or olive coloured material. These upper sediments are generally enriched in $\text{MnO}_2(\text{s})$, while the deeper zones contain some localized oxidation layers enriched in leachable Mn. A number of sulphide-rich zones also appear, occurring at 110 cm for GC-05, at 290 cm for GC-13 and at 1010 cm for PC-19. Removal of trace metals from solution appears to occur both in the $\text{MnO}_2(\text{s})$ and the sulphide-rich zones, while marked phosphate precipitation occurs only in the Mn enriched layers.

2. Large horizontal variations are apparent for surface sediments in the 100 km square study area. Geochemical data suggest that the central (BC-44) and eastern (GC-05) areas are more reducing than the northwestern area (GC-13). The distance from shore and coastal production may account for the westward decrease in organic matter and biogenic silica reaching the sediments.

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SCIENTIFIC AND TECHNICAL NOTES

NOTES SCIENTIFIQUES ET TECHNIQUES

ANORTHOSITE ERRATICS OF PROBABLE LAURENTIAN ORIGIN IN THE VICINITY OF BUFFALO, NEW YORK

Project 780017

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Introduction

During May 16-18, 1980 I participated on the field trip of Northeast Friends of the Pleistocene: "Late Wisconsin stratigraphy in the Upper Cattaraugus basin, New York" (Lafleur et al., 1980). The guidebook used the expression "bright" for the light coloured, dominantly crystalline erratic pebbles and boulders that are in sharp contrast with the "drab" or dark green to dark grey and black local bedrock. Although most of the bright rocks were granites, granite gneisses, and K-feldspar porphyries which are common over a broad region of the Laurentian Highlands, few were distinctive. Many such rocks outcrop in the Frontenac Axis and in the Adirondacks to the north and east of the Buffalo area. However, at three localities visited during the field trip, I identified erratics of anorthositic rocks that do appear to be distinctive and most probably have a restricted source area northwest of Montreal, Quebec.

Observations

At Lord Hill (Stop no. 1, May 17; Lafleur et al., 1980, p. 22) several well rounded boulders of anorthosite were seen, ranging from 30 to 60 cm in diameter. These rocks were dominantly light grey, but large, reddish purple to maroon crystals (1 to 2.5 cm) of feldspar, probably plagioclase, occurred throughout the rock. This particular facies of anorthosite is not known in outcrop to the author, but the colour of the phenocrysts is similar to that of other rocks described below. Such colours are not known in the anorthosites of the Adirondacks (J. Craft, personal communication, 1980).

At Stop no. 5 (May 18; Lafleur et al., 1980, p. 27) in a flashflood gravel bar in the bed of Connoisarauley Creek, I discovered another distinctive anorthosite cobble. This clast was elliptical, well rounded, some 25 cm in long dimension, and had been broken in half. The feldspars of the freshly broken surface were a light bluish purple, and the weathered surface of the rock was lavender grading to light grey.

At Stop no. 3 (May 18, Jackson Hill Road; Lafleur et al., 1980, p. 26) an erratic block of deep reddish purple to maroon anorthosite, more than 3 m diameter, was partly exposed in the stream bed on the east side of the road. This one block and the smaller cobble from Connoisarauley Creek resemble outcrops and erratics identified in Canada by Gadd (1980a).

Tentative correlation of these two is made, and possibly also the rocks with reddish purple to maroon phenocrysts, with a distinctive maroon to purple rock that occurs in the westernmost massif of anorthosite in Grenville rocks of the Gatineau River map area (Emslie, 1975; Baer et al., 1978). The most accessible outcrop of this rock is on the main street of the Sainte-Adèle, Quebec.

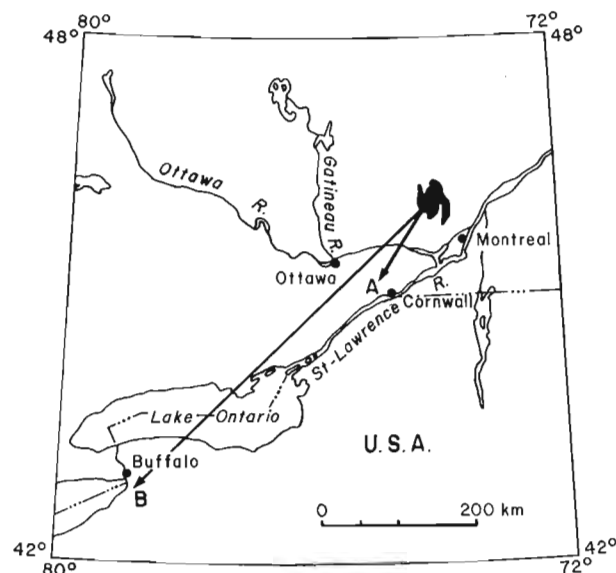


Figure 1. Location map showing anorthosite body at Sainte-Adèle, Quebec (from *Principal Mineral Areas of Canada*, 1980), and the most direct path of transport for maroon anorthosite erratics in the Cornwall, Ontario (A), and Buffalo, New York (B) areas.

Comments

Gadd (1980a) showed southwesterly transport of large erratics of purple (maroon) anorthosite from the apparent source at or near Sainte-Adèle, Quebec, to the vicinity of Cornwall, Ontario (A of Fig. 1). Gadd (1980b) also showed that a major ice stream (Ottawa-Lake Ontario lobe) flowed from the Laurentian Highlands north of Ottawa and Montreal into and through Lake Ontario basin south of Oak Ridges moraine. The report of erratics found in the Buffalo, New York area supports and extends these previous observations.

Although maroon or purplish anorthosite facies exist in Ontario west of Ottawa River valley (S. Lumbers, personal communication, 1981), there is no known means of glacial transport of those materials to the east end of Lake Erie. It is assumed here, therefore, that these erratics in the Buffalo area most probably are from source rocks in the vicinity of Sainte-Adèle and most probably were transported by the late glacial Ottawa-Lake Ontario ice lobe or ice stream.

The erratics near Buffalo have colour and textural characteristics in common with the Cornwall area erratics and the outcrops near Sainte-Adèle. Samples of these Canadian erratics and the outcrop have been provided to Robert G. Lafleur¹ and Parker Calkin², for petrographic comparison with erratics from the Buffalo region.

Purple to maroon anorthosite erratics of large size have been transported about 560 km from their probable source. Although evidence is not sufficient, it appears that erratics in the Cornwall area are concentrated in an interlobate zone (Gadd, 1980a) and thus possibly mark the marginal limit of an ice lobe or ice stream. Presence in numbers in the Buffalo and Cornwall areas and sparsity in or absence from large areas up-ice from these places suggests a zonal concentration of erratics. Such a concentration would be normal in the terminal zone of an ice lobe where large, resistant materials, possibly carried in part supraglacially or englacially, would

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accumulate. Their presence, therefore, may mark the terminus of a major ice lobe, probably the Ottawa-Lake Ontario lobe, partly defined earlier (Gadd, 1980b).

This is supporting evidence that a late Wisconsin lobe of Laurentide ice flowed southwesterly from the vicinity of Montreal via Ottawa and/or St. Lawrence valleys towards the southwest end of Lake Ontario near Buffalo. Further study of the distribution of distinctively maroon to purple anorthosite erratics may help define an ice lobe related to that flow pattern.

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PRE-LAST-GLACIAL ORGANIC REMAINS IN OTTAWA VALLEY

Project 780017

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Recent syntheses of the stratigraphy for the Wisconsin Stage in the Ottawa Valley portion of Upper St. Lawrence Lowland present no sedimentary record of pre-Late Wisconsin events: "The Upper St. Lawrence Lowland which lies between the above two areas has not preserved any deposits that can be identified with certainty as of Early or Middle Wisconsin age" (Dreimanis, 1977, p. 42). Moreover, recent borings made through the Quaternary sediments of Ottawa valley during the stratigraphic drilling program of 1975 and 1976 failed to reveal any pre-Late Wisconsin deposits: "The borings in Ottawa Valley reveal, on the contrary, that a single, thin till sheet discontinuously overlies bedrock. Since deposition of this till, the Ottawa Valley basin has received the apparently continuous suite of nonglacial sediments..." (Gadd, 1977, p. 379). Recent work by Gadd on organic sediments underlying deposits of the last glaciation is the first observation of a unique occurrence of such material in St. Lawrence Lowland west of Lac Saint-Pierre, Quebec (cf. *St. Pierre Sediments* in Gadd, 1971).

A working face in a sand pit near Pointe-Fortune, Quebec (NTS 31G/9; UTM grid location 478431), exposes about 3 m of oxidized till overlying some 6 m of oxidized, stratified, and crossbedded sand, with some silt bands (Fig. 1). The section was first noted by Grant and described by Gwyn and Thibault (1975) as follows: "...three metres (10 feet) of Fort Covington Till overlie crossbedded glacial-fluvial sands". Richard later reported that "No fossils were found in either the till or in the underlying sand, but the sand is believed to be subaqueous outwash deposited in a marine environment" (Richard, 1978a, p. 116; 1980). Because no geological event older than the Late Wisconsin glaciation had been recognized in the Quaternary deposits of Ottawa valley up until that time (Gwyn and Thibault, 1975; Dreimanis, 1977; Gadd, 1977), it was believed that the sand and the overlying till represented younger events that had occurred during the Champlain Sea episode 12 800 to 9800 years ago (Richard, 1975b, 1978b).

During a field excursion in 1978 led by Richard, in a section of the sand pit Gadd and R.J. Fulton noted disturbed masses of dark grey sand containing disseminated organic matter including flattened twigs, but the material was not seen in place; excavation below the floor of the pit was prevented by high water levels. A grab sample of the organic-rich sand submitted to J.V. Matthews, Jr. provided "interesting" plant and animal macrofossils that suggested that the material might be "relatively old"; the biotic affinities appeared to be freshwater and terrestrial. The dated wood sample contains coniferous and deciduous species, but because of compression and distortion, identification of the wood was not possible (unpublished GSC Wood Identification Report 79-31 by L.D. FarleyGill).

M. Réjean Bélanger, the owner-operator of the pit, noted that the organic-rich sand had been excavated from below the water table in an operation to bury unwanted large boulders; he also reported large clam shells in the same materials. During Gadd's 1979 field season, M. Bélanger



Figure 1. Crossbedded fluvial sand overlain by coarse bouldery sandy till at Bélanger sand pit, Pointe-Fortune, Quebec. The excavation from which wood and other organic remains were collected was made in the floor of the pit in the foreground of Figure 1a. Figure 1b shows thick sets of crossbeds in the fluvial sand.

excavated, voluntarily and free of charge, an undisturbed portion of the pit approximately 1 x 3 m at the surface and about 2.5 m deep; but because of the degree of saturation of the material, Gadd had only a brief opportunity to examine and collect materials from below the water table before the pit collapsed. The sand was dark grey to black, uniformly fine grained and silty, vaguely stratified, and apparently undisturbed. It had a fetid odour and disseminated organic matter could be seen throughout. Two samples were collected: one a bulk sample of sand, silt and organic matter, the other a small collection of bits of wood. The wood was compressed and worn. Some of the larger pieces of wood, weighing 9.8 g dry, yielded a radiocarbon date of >42 000 years (GSC-2932) obtained from a 3-day count in the 5 L counter at 1 atmosphere.

Discussion and Conclusions

Dating of such transport-worn bits of wood, can only give a maximum age for the enclosing sediment. The true age of the wood could be much greater – perhaps 54 000 years (the working limit of the counter at 4 atmospheres) or more; however, sufficient material had not been available for counting at high pressure. Hopefully more wood may be obtained, preferably from materials in situ, thereby providing a better opportunity to obtain a finite date and eliminate some areas of speculation.

The relatively massive structure of the wood-bearing sand indicates rapid sedimentation in an environment of deposition such as the broad lake-like expansion of the present Ottawa River. The absence of pebbles argues against the "glacial-fluvial" origin suggested by Gwyn and Thibault (1975).

The deposit lies on the southern slope of a wide part of Ottawa River valley in which marine sediments of Champlain Sea are extensively exposed. The section reveals neither glaciomarine nor marine facies. It is presumed, therefore, that the fluvial system inferred from the sand and the subsequent glaciation that deposited the till both occurred before the Champlain Sea episode. The nonfinite age rules out a correlation with the young readvance into Champlain Sea proposed by Richard (1975a, b). This sand at Pointe-Fortune which predates the last glaciation is, therefore, more probably related to the >75 000 year-old St. Pierre Sediments, and the overlying till to the Gentilly glaciation of the St. Lawrence Lowland (Gadd, 1971).

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DRAINED LAKE EXPERIMENT FOR INVESTIGATION OF GROWTH OF PERMAFROST AT ILLISARVIK, NORTHWEST TERRITORIES – INITIAL GEOPHYSICAL RESULTS

Project 730006

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Introduction

A multidisciplinary study of the growth of permafrost and its properties under natural field conditions was proposed in 1973 by J. Ross Mackay, University of British Columbia. The natural field conditions would be created by the drainage of some northern lakes which are on the verge of self-drainage. An ad-hoc working group of interested personnel was formed, with J.R. Mackay as chairman, to co-ordinate activities of government and university researchers.

A lake was selected for draining on Richards Island, Mackenzie Delta area, Northwest Territories, approximately 60 km west of Tuktoyaktuk (see Fig. 1). The lake dimensions prior to draining were 600 m by 350 m with a maximum water depth of 4.7 m. The name Illisarvik was chosen for the lake – an approximation of an Inuvialuktun word meaning "a place of learning".

This report deals with a portion of the pre-drainage program in which the Geological Survey of Canada seismic section was involved.

Our field program began in April 1978, with the following objectives:

1. To establish a survey grid around the lake.
2. To obtain lakebottom bathymetry on a reconnaissance scale.
3. To install thermistor cables beneath the lakebottom for thermal studies carried out by the Earth Physics Branch of Energy, Mines and Resources.
4. To obtain geological information, and samples of sub lakebottom materials including estimates to the top of ice-bonded permafrost.
5. To obtain geophysical logs in ice-bonded and non-ice-bonded sediments as baseline data for other (seismic) geophysical surveys.
6. To map the shape of the "talik" zone beneath the lake by marine seismic refraction techniques.

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Geol. Surv. Can., Paper 81-1C.*

The Survey Grid

A metric grid was established on the lake in April 1978 (Fig. 2). The north-south base line is approximately 045° true north and is roughly positioned along the long axis of the lake. Wooden markers at 50 m intervals were established circumjacent to the lake; in late summer when the active layer was deepest, these markers were re-established.

Lake Bottom Bathymetry

Through-ice soundings were made using a sounding pole at 50 m intervals on the grid in the deeper portions of the lake and at closer intervals near the shoreline. Since the bottom sediments were soft, the measurement accuracy was only 5 cm. Sounding data from a through-ice resistivity survey (Scott, in press) were also incorporated in the bathymetry. The compilation is shown in Figure 3; contour intervals are in metres using the ice surface as datum.

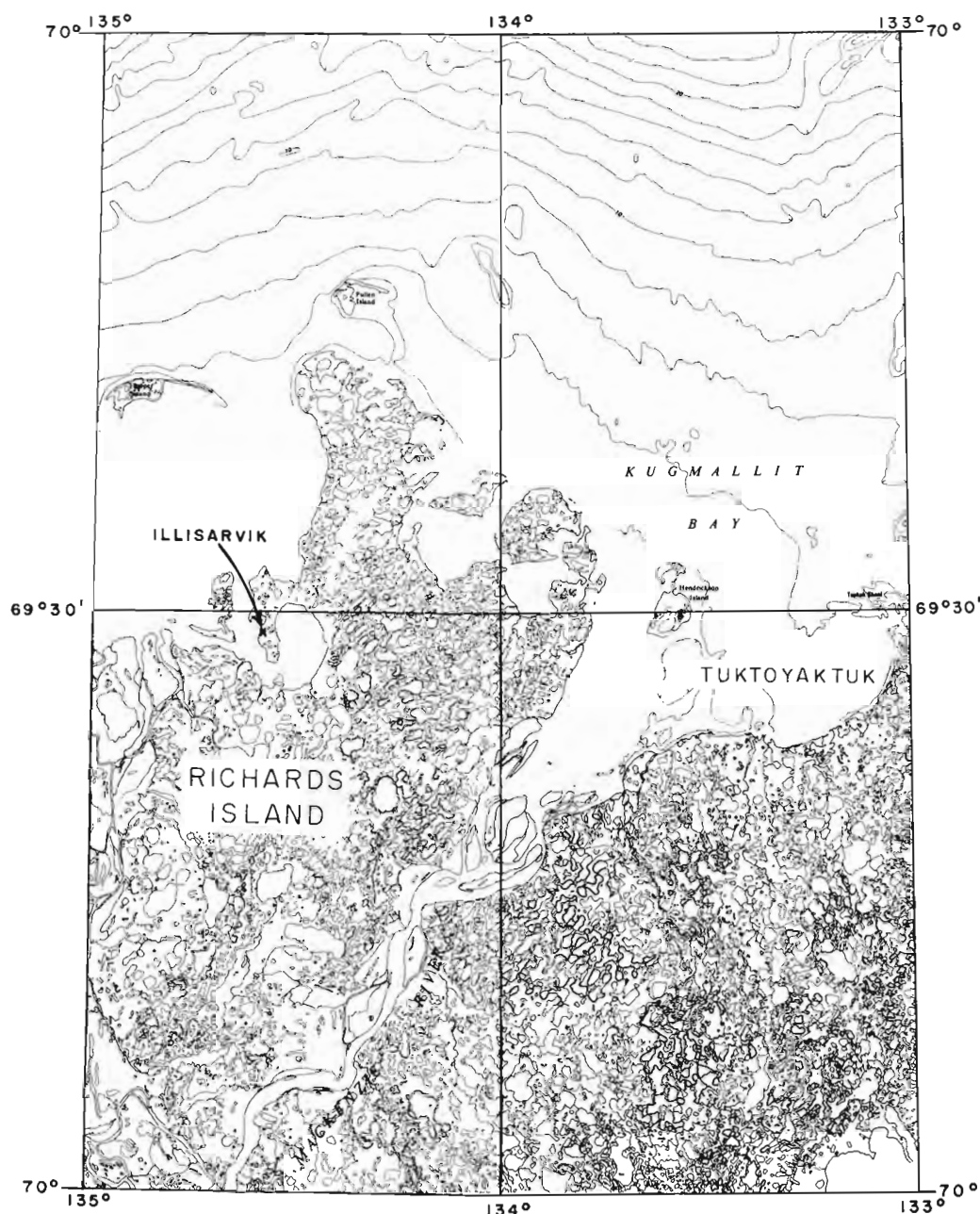


Figure 1. Location of the Illisarvik drained lake site.

During the summer months most of the lines were re-surveyed with a marine fathometer and we found that these measurements were within 10 cm of those obtained from the on-ice measurements. Further, the marine work indicated that no major bathymetric features had been missed using the 50 m on-ice grid.

Most of the lakebottom lies at a depth below 2 m S; there are also two minor depressions (below 2.5 m) in the northern quadrants. Using a 10 m x 10 m grid increment, the volume of water is estimated to be 306 000 m³.

Drill Hole Locations

Figure 2 shows the locations of all holes drilled by all participants during the 1978 field season. An additional hole not shown on the map was drilled at approximately 100 N and 400 W and is designated as the "Cabin Hole".

During April, holes 78-1 to 78-10 were drilled on the ice by the GSC drilling crew. Additional holes 78-12, 78-13, 78-14, 78-15 and the "Cabin Hole" were drilled during the summer months by GSC and Earth Physics Branch crews.

The Hydraulic Jet Drilling Technique

The jet drilling technique used for the program has been described in detail by MacAulay et al. (1977). Normally, the 1" steel drill pipe is left in position at the completion of drilling and thermistor cables are installed in it. We attempted to install 2" ABS casing in the offshore holes using the following techniques:

1. installing the 2" casing around the 1" drill pipe after completion of drilling. The drill-pipe is then retrieved.
2. pulling the 1" drill-pipe and installing the 2" casing in an open hole.
3. installing 2" casing during the drilling operation.
4. jetting with the 2" casing using a 6 m leader of 1" steel drill pipe.

None of the above techniques were particularly successful because of the thick layer of loose non-ice bonded sediments covering the lake bottom; however, we were able to case hole 78-7 using technique no. 1.

For the lakeshore holes drilled during the summer which were spudded in ice-bounded sediments technique no. 1 was used successfully.

To obtain washed samples were installed 4" ABS casing in the upper section of the hole with the casing top close to ice surface. Samples were then caught with a dipper in the drill hole. This method was only partially successful; and the drilling proceeded, the seal of the 4" casing to the lake bottom was lost and return flow was not achieved.

Drilling rate was used to indicate the top of ice-bonded sediments at depth in the holes. Depths to ice-bonding could be determined to an accuracy of 0.25 m.

Generalized Geology

Figure 4a shows the geological logs obtained along the north-south base line. The lithologies were established from a combination of sample returns and drilling penetration characteristics. Incorporated in this interpretation is the data on the top of ice-bonding obtained from holes 78-16 to 78-21 drilled by J.R. Mackay in August 1978. Figure 4b shows the geological logs obtained along the east-west base line.

The lakebottom is underlain by an organic-rich clay layer to depths as great as 10 m below ice level. Interspersed locally in this layer are lenses of gravel. Beneath the organic layer the section is uniformly fine- to medium-grained sand with some local silty zones (hole 78-4).

The top of ice-bonding is identified by a change in drilling rate suggesting a rapid transition to ice-bonding near the 0° isotherm. Figures 5 and 6 (after Judge et al., in press) show the N-S and E-W temperature isothermal sections constructed from temperature observations in the drillholes. Comparing the indicated top of ice-bonding with the temperature observations, the boundary observed from the drill rate observations correlates within 1 m of the position of the 0° isotherm in each of the holes. Hence, much of the interstitial water is transformed to ice close to 0°C as might be expected in coarse grained materials.

There is little indication of massive ice lensing from the drilling characteristics. Some minor lensing occurs in holes 78-4, 78-7, 78-8 and 78-10.

In holes 78-1 and 78-10 voids were encountered and gas was observed escaping from the hole.

Geophysical Logging Results

Seismic uphole logs were run on holes 78-7, 78-13, 78-14, 78-15 and "Cabin Hole". Figure 7 shows the logging results from the "Cabin Hole" drilled on the hilltop west of the lake. The hole was drilled to a depth of 17 m with the waterjet technique. Beneath the icy organic layer, silt was encountered to a depth of 7 m where a transition to fine- to

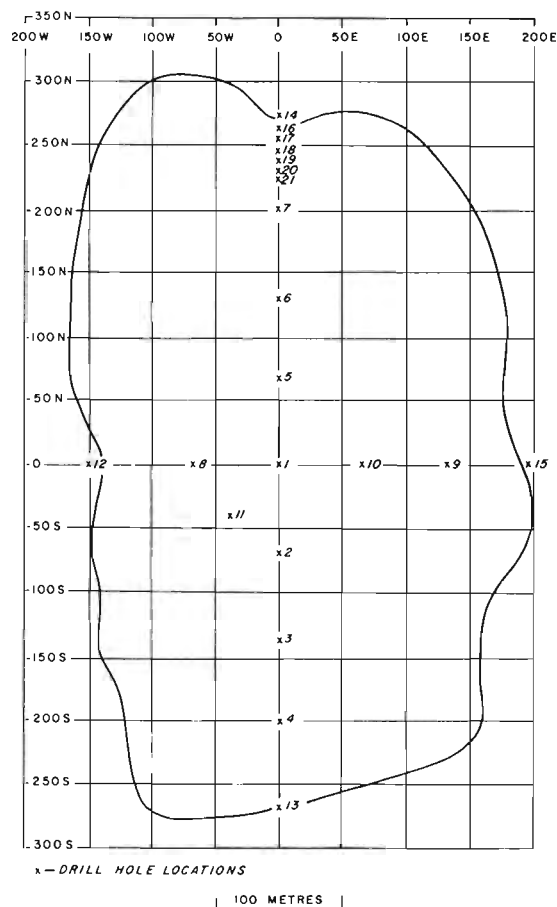


Figure 2. Location of survey grid and 1978 drillholes with respect to the lakeshore.

medium-sand was encountered. Immediately after drilling a caliper log was run which showed a change in hole size associated with the silt-sand boundary. The gamma-gamma density log reflects the low densities associated with the icy organic layer. A change in count rate is also observed in the range of 6.5 – 7.0 m depth which probably results from the change in hole size since the tool was suspended freely in a 2" PVC casing and not clamped to the borehole wall.

Uphole seismic logging was run using a single broad-band hydrophone suspended in a water-filled 2" PVC casing. The seismic source used was a 16 lb hammer struck against a steel plate embedded in the icy surface layer at the top of the borehole. A velocity of 3300 m/s was observed for the silt and 4000 m/s for the sand. Both velocities are typical for the respective materials in ice-saturated conditions for temperature less than -2°C.

Figure 8 shows the logs taken from hole 78-7 immediately after hole completion and casing. Organics and clays were encountered to a depth of 7 m below ice surface; between 7 and 8 m depth, a stiff clay layer was encountered; the remainder of the hole contained fine sand which was non-ice-bonded to a depth of 14 m. The gamma-gamma log shows the effect of organics in the upper section, immediately below the stiff clay layer a low density "kick" was observed which probably denotes the presence of gas. The lower section of the gamma-gamma log shows little variation with the exception of peak (low density) at 15.5 m. This anomaly was present only in the log run within two hours of hole completion; on re-running the log 24 hours later this anomaly had disappeared although all other variations in the curve were repeatable. We suggest that this anomaly resulted from excess ice in the vicinity of the hole at this depth which had subsequently melted from the effects of the warm drilling fluid. Calibration of gamma count rate with density had not been done for a fluid-filled freely suspended tool, hence observed count rates can only be used as a qualitative guide to density variations.

The uphole seismic run in hole 78-7 was done with a suspended broadband hydrophone using 1/4 lb geogel dynamite charge shot at lakebottom as the source. A velocity of 1750 m/s was observed for non-ice-bonded sand and clay and 2540 m/s for ice-bonded sand below the top of permafrost. The velocity observed for ice-bonded sand is considerably lower than that observed in the "Cabin Hole" and is attributed to marginal ice-bonding resulting from warm permafrost temperatures.

Figure 9 shows the seismic uphole logs shot on holes 78-13, 78-14 and 78-15 drilled at the edge of the lake in predominantly medium grained sands. A single broad-band hydrophone was suspended in 2" ABS casing and a 1/4 lb geogel shot placed within 10 m of the hole was used as a source. The velocities observed in all holes are relatively low are compared to ice-bonded sands observed in the "Cabin Hole" yet somewhat higher than that observed in the 78-7 hole in the lake. This may reflect the temperature variation from warm permafrost beneath the lake to the much colder permafrost away from the lake. Figure 10 shows a plot of observed seismic velocities versus observed temperatures from all holes where both parameters were measured. The

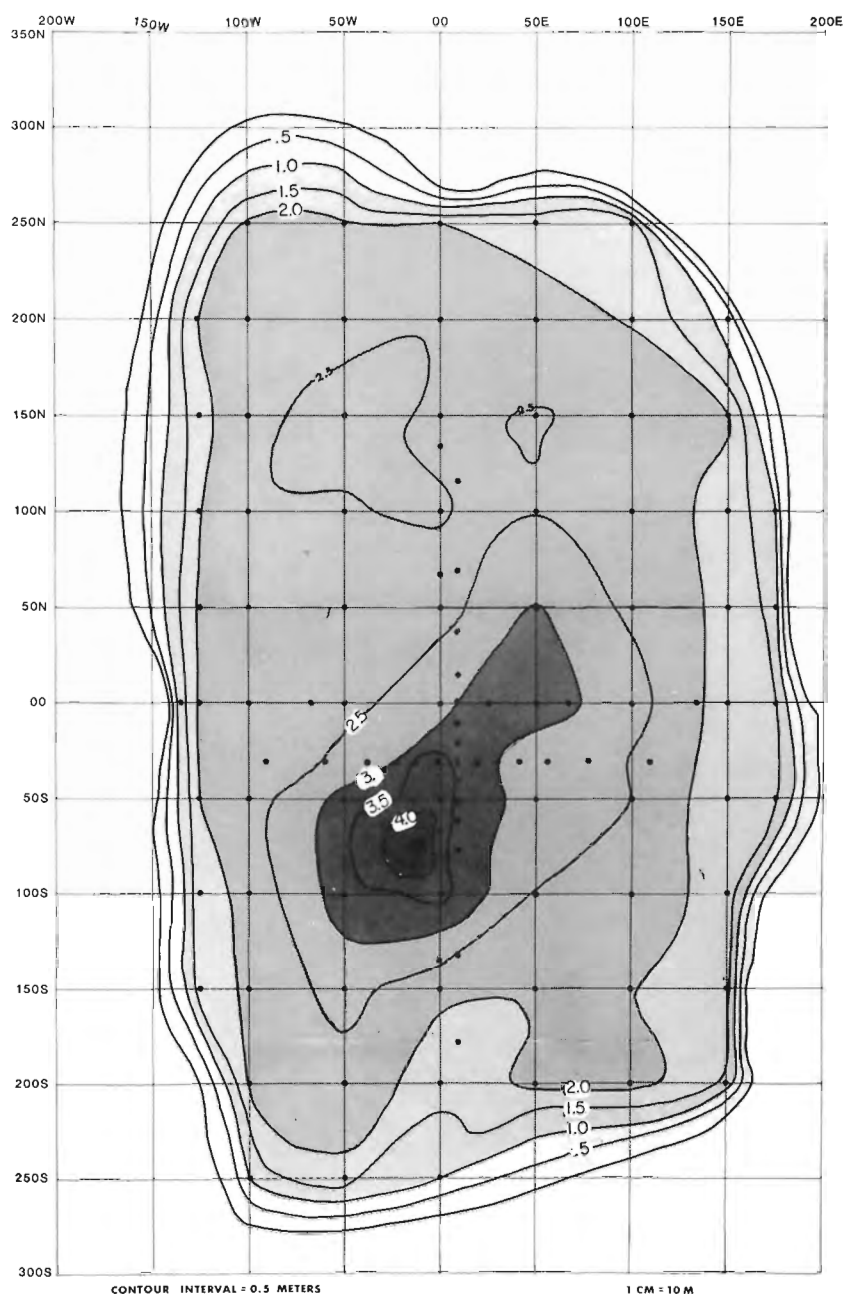


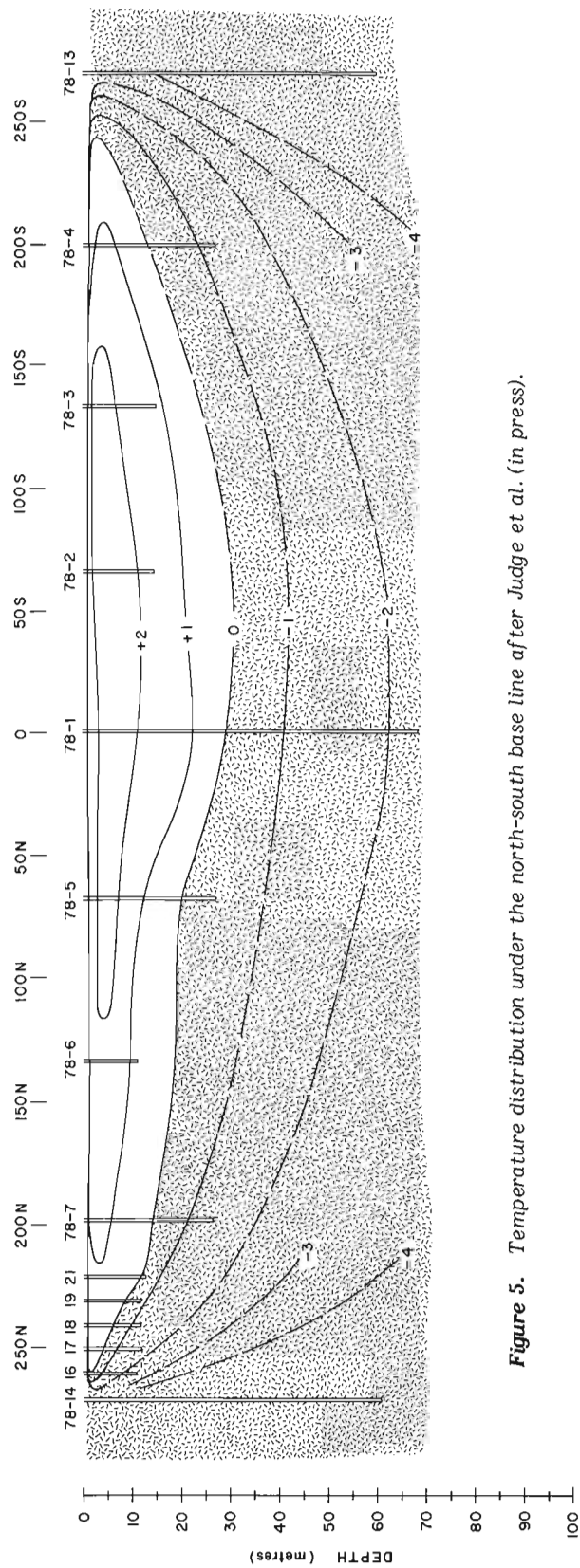
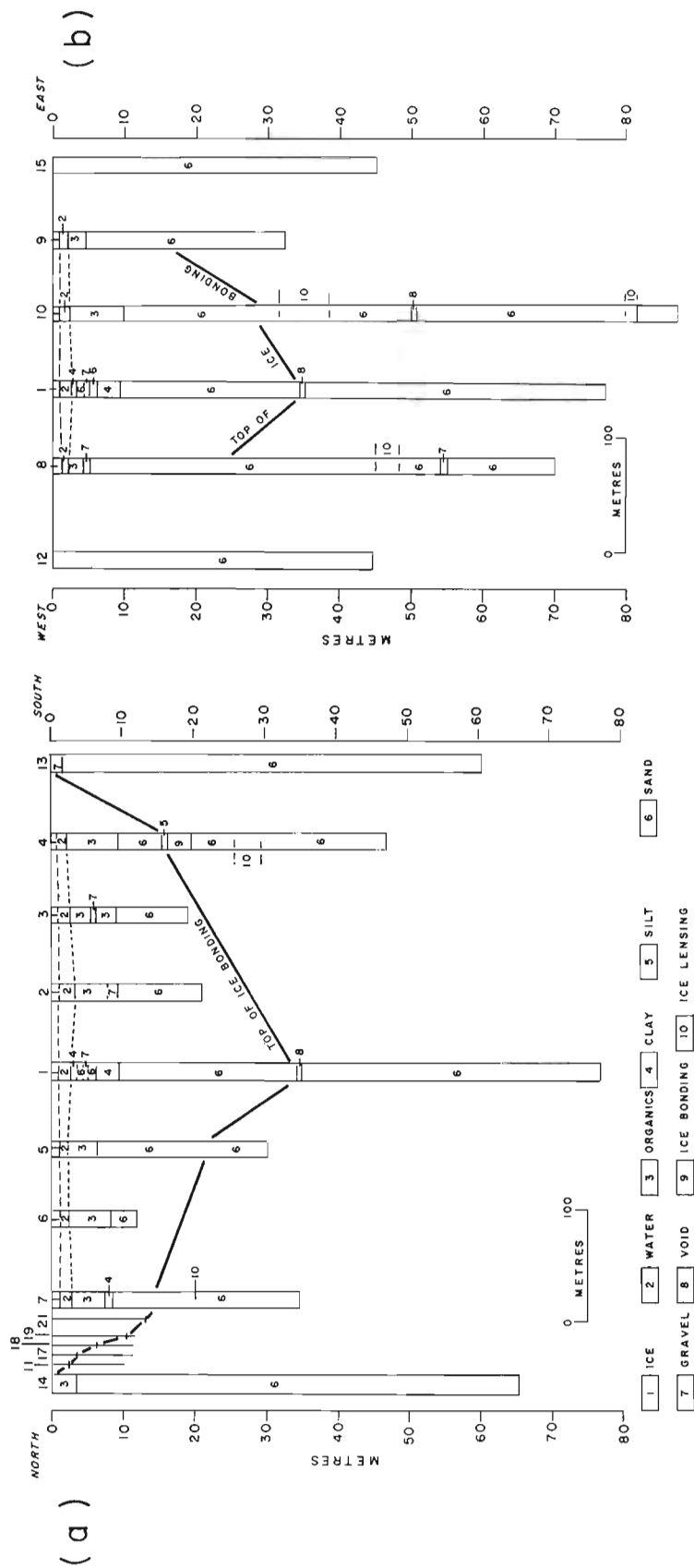
Figure 3. Illisarvik bathymetry established by through-ice soundings.

values plot between the curves for sand and clay as given by Kurfurst (1976) and higher than the loam curve by Aptikaev (1964). Hence, assuming that the observed values come from the same sedimentary unit we suggest that the interpreted sand section may contain a large silt fraction.

Marine Seismic Refraction Survey

Survey Procedure

The refraction program was carried out in July 1978, from a small boat using a 12 hydrophone marine cable with 15 m spacings between "phones" in-line with a dynamite source detonated at various intervals off the end of the array. Both the source and hydrophone array were placed on the lakebottom.



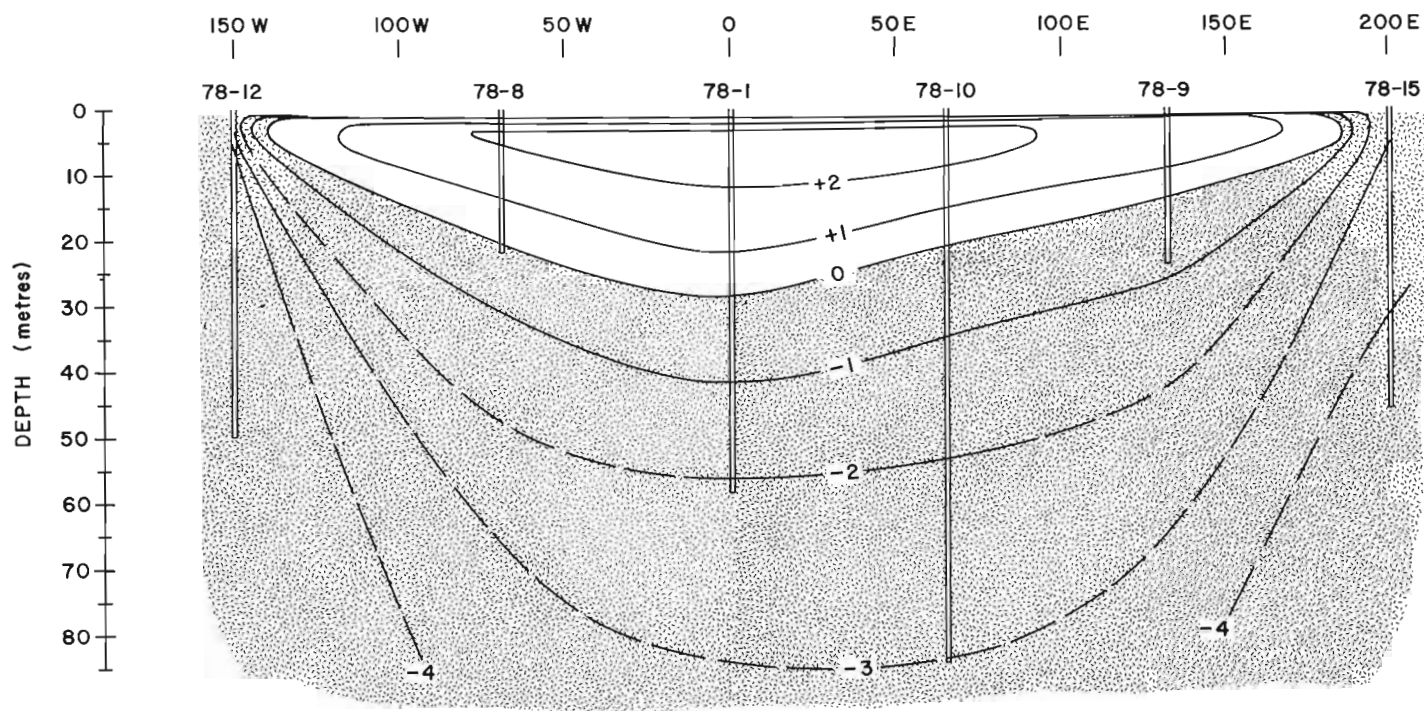


Figure 6. Temperature distribution under the east-west base line after Judge et al. (in press).

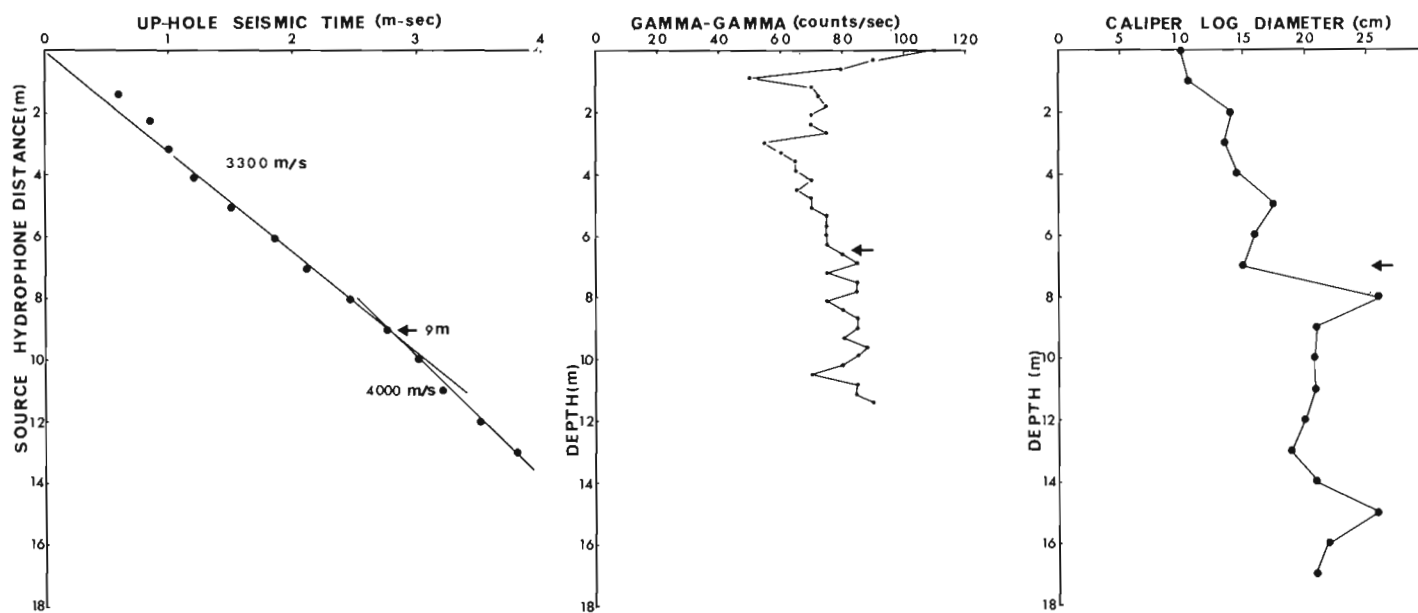


Figure 7. Geophysical logs from the "Cabin Hole" at grid reference 100 N and 400 W.

ILLISARVIK LAKE DH NO. 7

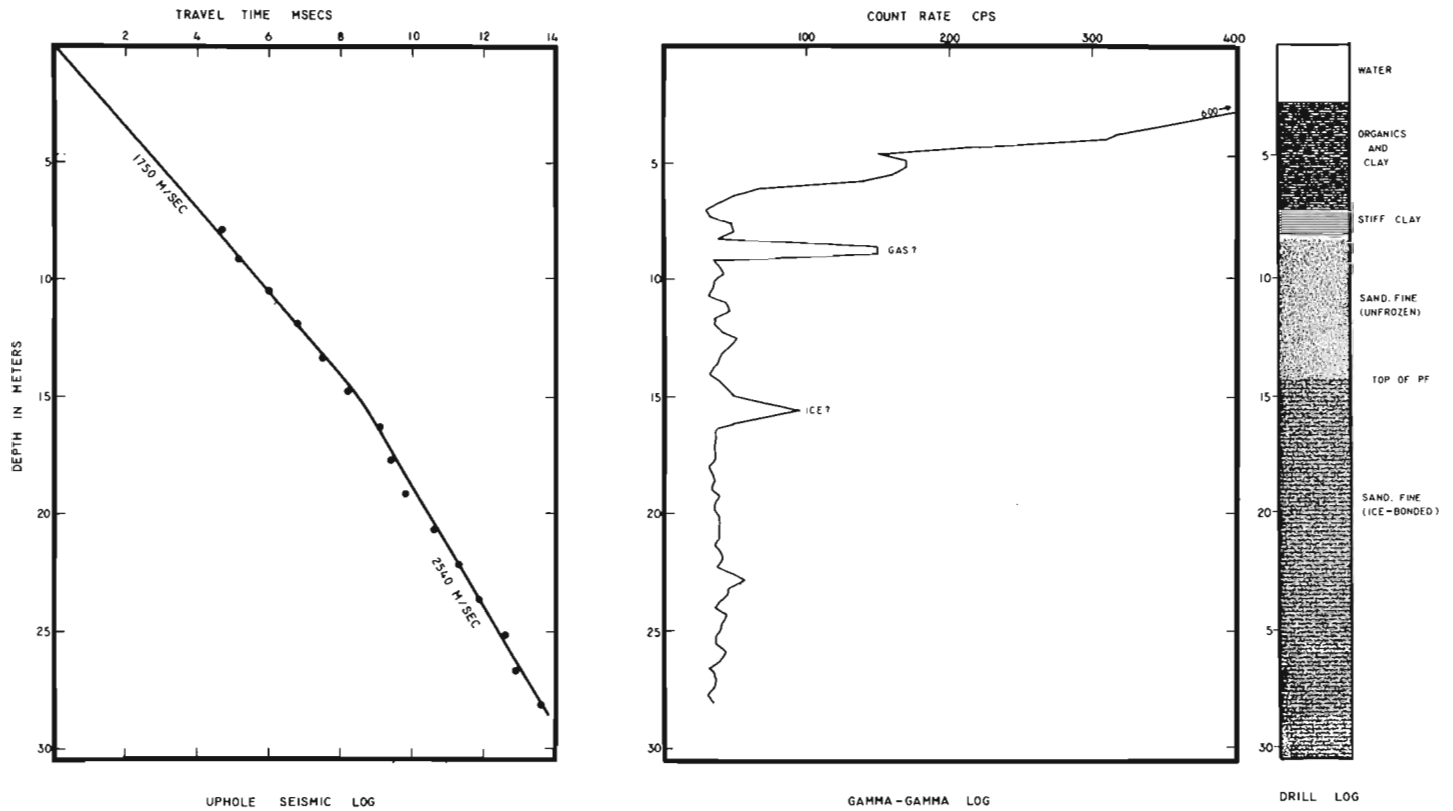


Figure 8. Generalized geology, gamma-gamma and seismic up-hole logs for DH-78-7.

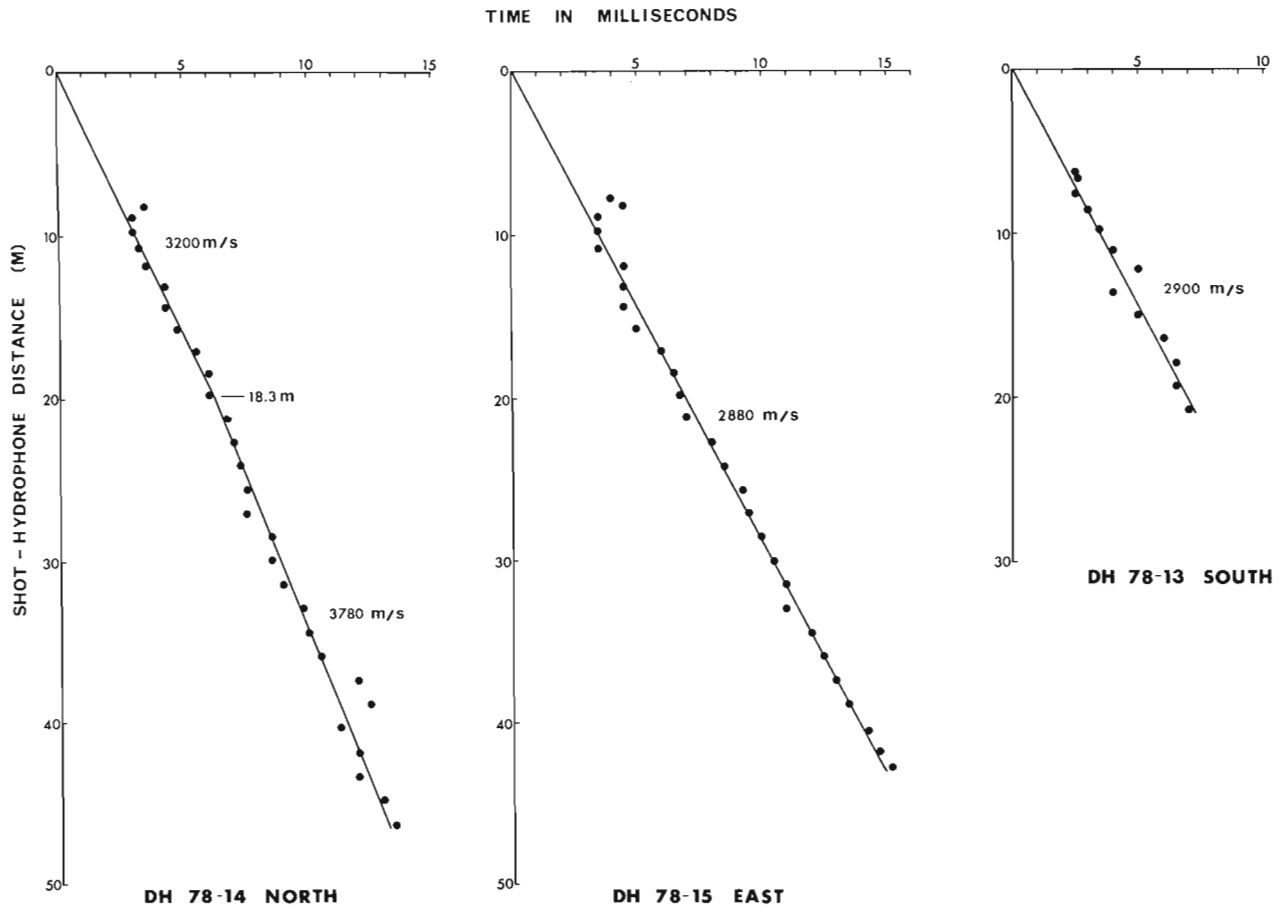


Figure 9. Seismic uphold logs for DH-78-13, 14, 15. Depth to seismic interface is shown in DH-78-14.

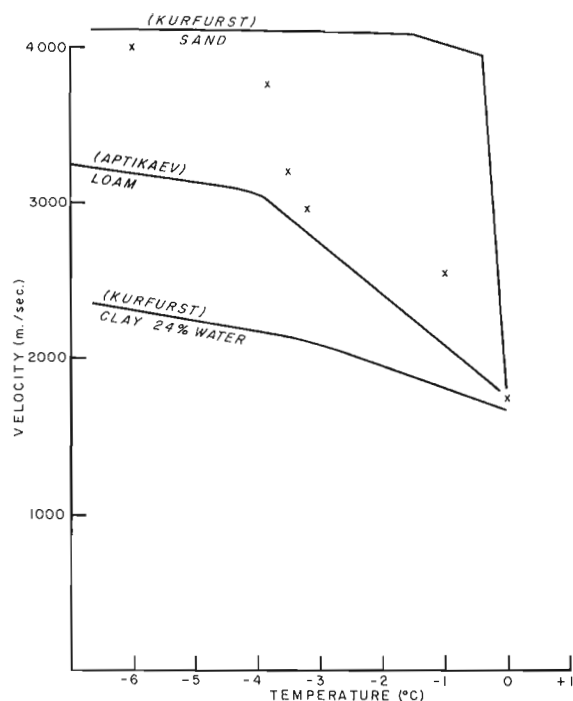


Figure 10. Observed borehole velocity-temperature data compared to type curves of Kurfurst (1976) and Aptikaev (1964).

The cable was laid along N-S and E-W lines of the established grid at 50 m line spacings. At each end of the line, a shot was placed close to the shoreline such that effectively the shot was placed on the top of the permafrost table hence no delay time in travel of the down-going wave through the unfrozen zone could be attributed to the shot (see Fig. 11). Observed delay times for each of these arrays would result entirely from travel-time in the unfrozen layer beneath the hydrophones. All array positions were shot from both ends; that is, for a fixed line of hydrophones on the lakebottom, shots were detonated in line off each end of the array. This was done to obtain reverse measurements of the permafrost refraction velocity to obtain velocity estimates in the presence of dip or structure on the permafrost refractor.

Considerable uncertainty as to the exact position of the shot and hydrophone array resulted since only approximate shot positions could be estimated from the shoreline grid. The exact shot positions could only be determined after lake draining when shot hole depressions on the lakebottom were accurately mapped. Shot to hydrophone distances were measured by identifying the water wave arrival time on the seismic records and by using a water velocity of 1480 m/s.

Seismic wave arrivals at the hydrophones were recorded with an SEI RS-4 12-channel recorder with a timing accuracy of 1 ms. Charge sizes varied between 1 and 5 pounds of high-velocity dynamite depending upon seismic coupling with the lakebottom and water depth.

The Data

Data quality throughout the survey was not good compared with other refraction surveys done in adjacent areas of the Mackenzie Delta and Kugmallit Bay area of the Beaufort Sea (Hunter, 1973). The water arrival was the most prominent event of the seismogram. Refractions through the unfrozen lakebottom sediment and from the top of the ice-bonded permafrost zone were often low in amplitude and in

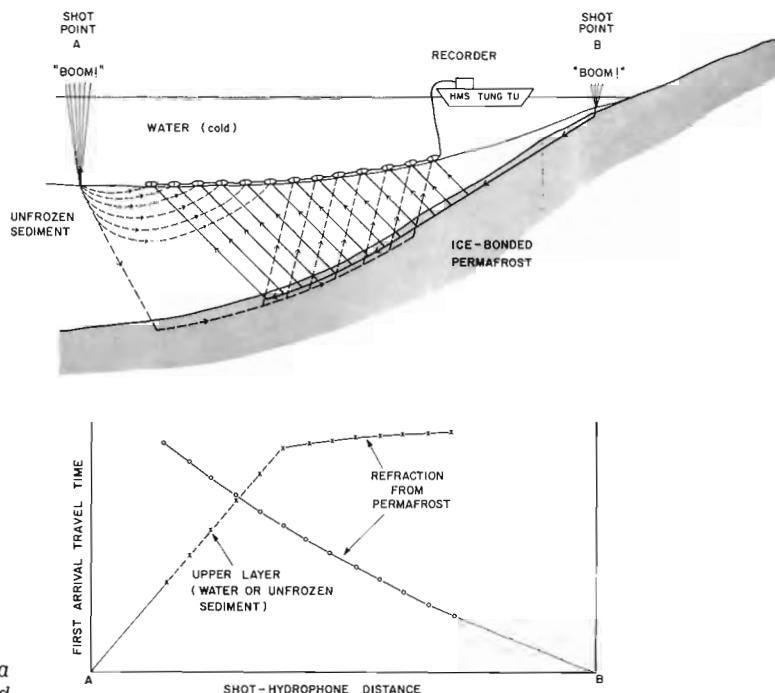


Figure 11. Marine seismic hydrophone array configuration showing refracted wave travel paths and time-distance plots.

frequency. It was apparent that the loose organic sediment covering the bottom was an unusually high absorber of seismic energy. Approximately 75 per cent of the seismic data were subsequently used in the interpretation, hence some areas of the lake bottom are not adequately covered.

The Interpretation

Forward the reverse profile data were used to compute the velocity of the permafrost refractor beneath each hydrophone array using a method given by Overton (1967). Delay times under each hydrophone resulting from travel through the unfrozen layer were computed. The delay time (T) is a function of the thickness of the unfrozen zone Z , the velocity of the unfrozen zone V_0 and the velocity of the ice-bonded permafrost refractor V_1 and is given by:

$$T = \frac{Z V_0 V_1}{\sqrt{V_1^2 - V_0^2}} \quad (1)$$

Velocities observed on the seismograms, which were associated with the upper layer, were consistently 1480 m/s (identical to that in water). These velocities may not represent the upper layer since the water wave arrival may have masked lower velocity events from the unfrozen layer. To obtain a delay-time depth relationship for the unfrozen zone, delay times measured in the vicinity of drillholes encountering the permafrost layer were plotted against depth to top of permafrost. The data are shown in Figure 12. Although scatter exists, the mean of the data groups around a line with a slope of 1.0.

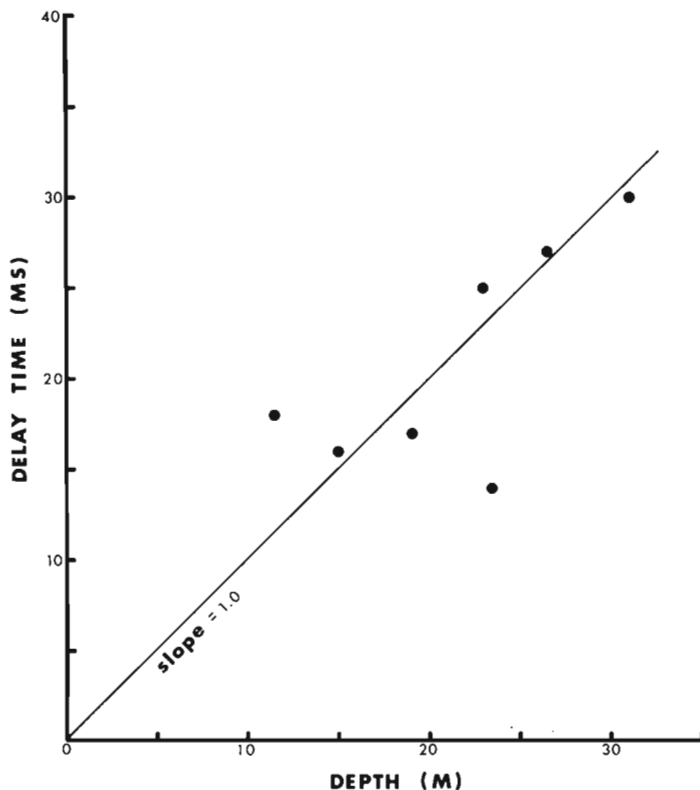


Figure 12. Measured time delays from marine seismic refraction interpretation plotted against depth to top of ice-bonded permafrost at seven drillholes.

The Results

The delay time sections for each of the grid lines shot were digitized at 25 m intervals where data existed and were contoured to produce a map of delay time shown in Figure 13. Average refractor velocities along segments of each line are also indicated. For the data obtained over the drillholes shown in Figure 12, unfrozen layer velocities were computed from equation 1. Velocities ranged from 610 m/s to 1770 m/s with the majority of velocities less than 1480 m/s (velocity in water).

Discussion of Results

1. From the correlation with borehole results the unfrozen layer velocities are unusually low. Most unconsolidated materials which are water-saturated give velocities in excess of that of water (1480 m/s). A gas concentration in marine sediments of less than 10% is known to result in a substantial decrease in sediment velocity (Domenico, 1976). Since numerous gas seeps were observed in the lake during the survey it is reasonable to assume that the low velocities computed for the unfrozen layer result from minor concentrations of gas.
2. The velocity of the ice-bonded permafrost refractor varies considerably under the lakebottom. Lower velocities (2400 – 2900 m/s) were observed in the central part of the lake whereas higher velocities occur around the rim of the lake. This variation may result from either:
 - a) Variation of grain size with similar permafrost temperatures. Fine grain materials display lower velocities than coarse grain materials because of the lower proportion of water in the frozen state.
 - b) Similar grain size materials with variation in permafrost temperatures. For Fine- to medium-grained water-saturated sand below 0°C more interstitial ice is found at -2°C than at -0.5°C.

The latter interpretation is preferable. From drilling results, a similar fine sand was identified in all holes. Because of larger negative temperature gradients occurring near the shore of the lake, (Judge et al., in press), the refracted seismic energy probably travelled in the highest velocity portion of the permafrost refractor with the shot-hydrophone array geometry used in the survey. In the centre of the lake where negative temperature gradients are lower, the refracted seismic wave travelled in relatively high temperature permafrost with lower ice-content.

3. Due to the probable variation of unfrozen layer velocity as indicated from borehole comparisons, the layer thickness estimates are given in the form of delay times in Figure 13. From the depth-delay time curve given in Figure 12 it is possible to obtain approximate estimates of layer thickness by multiplying the delay times by the slope of the curve (1.0). Since it was not possible to obtain accurate velocity measurements on the unfrozen layer throughout the grid, the delay times shown can be used only as an approximate guide to the shape of the "talik" zone.

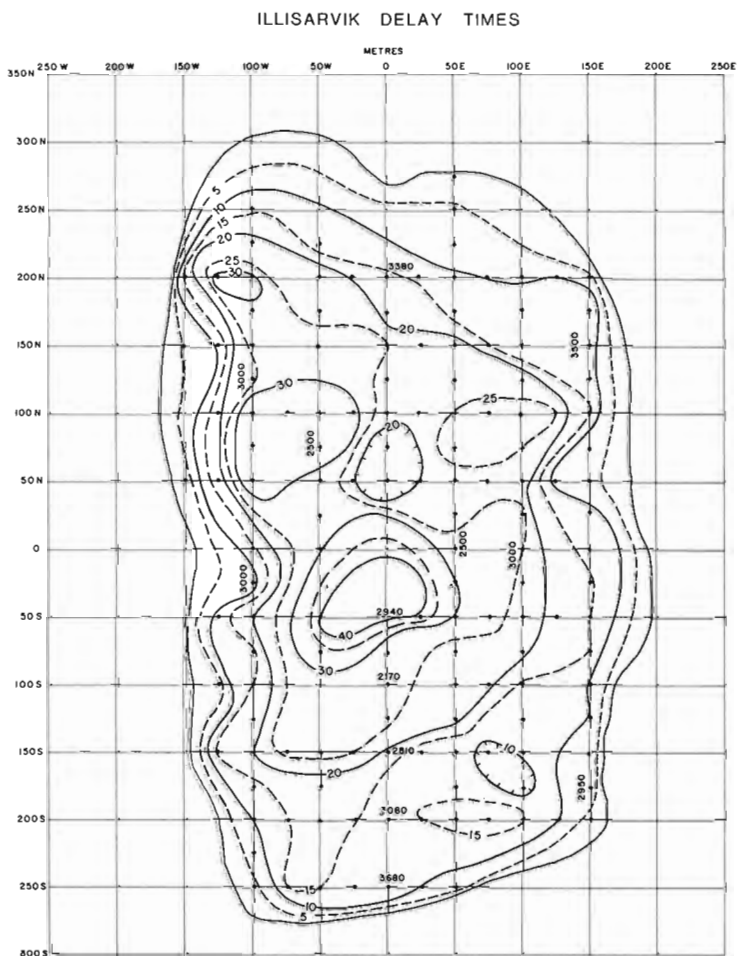


Figure 13. A contoured map of delay-times based on data intervals of 25 m. Delay times correspond approximately in depth in metres to the top of ice-bonded permafrost.

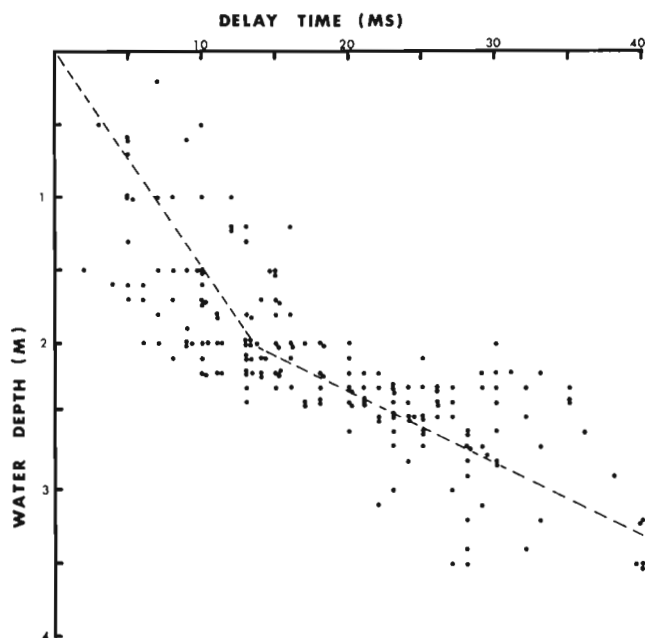


Figure 14. Delay times vs water depth showing a break in travel at approximately 2 m water depth.

4. The major depression in the talik zone is centred at 25 W, 25 S with other depressions centred at 75 W, 75 N and 74E, 75 N. The top of the ice-bonded permafrost table drops abruptly close to the shoreline and the contours mirror the water-depth contours closely (Fig. 3).

To determine if a relationship existed between water depth and depth to top of the ice-bonded permafrost layer, delay times were plotted against water depth for the 25 m grid spacing and are shown in Figure 14. A trend towards larger delay times with increased water depth is observed. A change in the trend at a water depth near 2 m is evident and is slightly greater than the expected maximum thickness of winter ice. Hence, the depth to the permafrost layer in the shoreline area is probably controlled by the combined effects of water to lakebottom and the colder mean annual onshore soil temperatures.

An alternative method of showing the insulating effect of the lake water (hence depression of the permafrost layer) is to plot estimated volume of water against estimated volume of unfrozen sediment for each 50 m grid square (Fig. 15). There is a possible break in slope occurring at approximately 5000 m³ of water which defines 2 m average water depth for the 50 m square.

Summary

1. A survey grid and lake bathymetry were established for Illisarvik Lake.
2. Hydraulic jet drilling and geophysical logging of holes in the lake and circumjacent to it has:
 - a) indicated the occurrence of a thaw bulb or "talik" beneath the lake,
 - b) established that a thin layer (<10 m) of organics and clay forms the lakebottom material throughout the area and is underlain by fine- to medium-grained sand, and
 - c) indicated permafrost beneath the lakebottom is ice-bonded, however, the occurrence of ice lensing is probably low.

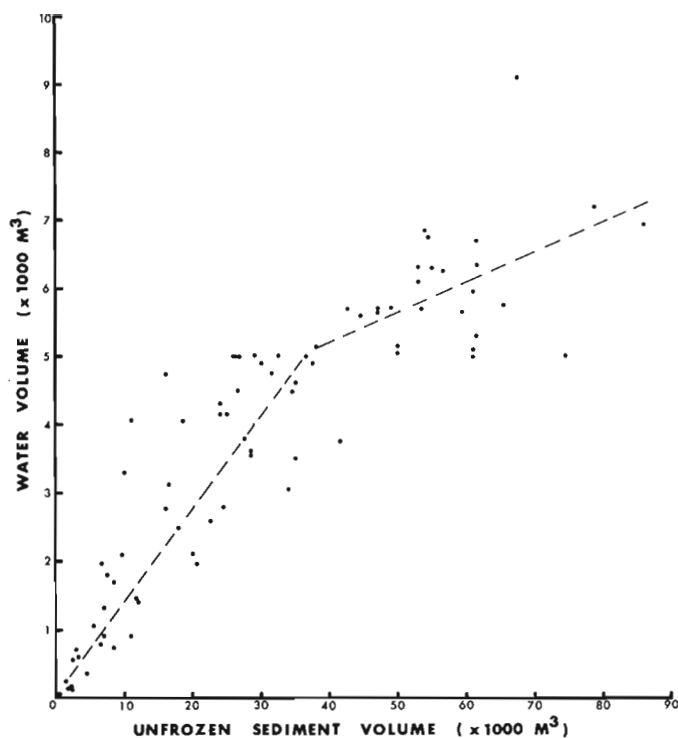


Figure 15. Volume of water vs volume of unfrozen sediment per 50 x 50 m grid.

3. Marine seismic refraction surveying in the lake has indicated the shape of the subbottom talik zones. The subbottom materials probably contain small quantities of gas resulting in only approximate seismic estimates of depth to the top of ice-bonded permafrost. Seismic velocities of permafrost beneath the lake decrease from shoreline to centre of the lake, probably in response to decreased ice in sediments at warmer permafrost temperatures away from the shoreline.

Since the lake was drained (Aug. 13, 1978) a residual pond has developed in the bathymetrically low area of the lakebottom (Mackay, in press) beneath which is the deepest portion of the talik zone. This area of the lakebottom may develop substantial ground heave possibly resulting in pingo formation (Mackay, 1978) and will be the subject of continued observations for several years.

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THE GEOLOGY OF THE CROWDIS MOUNTAIN VOLCANICS, SOUTHERN CAPE BRETON HIGHLANDS

EMR Research Agreement 51-4-80

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Introduction

The Crowdis Mountain volcanics occur at the southern end of the Cape Breton Highlands northwest of Baddeck (Fig. 1). The area is located on map sheets 11K/2 and 7, between 46°08' and 46°23'N, and 61°45' and 61°57'W; it is bounded on the west by the Cabot Trail. Excellent access is provided by Nova Scotia Forest Industries roads. Outcrop is restricted to the steep stream valleys that dissect the Highlands, and to cleared areas bordering woods roads.

Previous workers described this part of the Highlands as underlain by low to high grade metavolcanic and metasedimentary rocks. The metasediments were tentatively correlated with the George River Group and the metavolcanics with the Fourchu Group of southeastern and central Cape Breton by Kelley (1967); Milligan (1970) included the entire area in the George River series. Keppie (1979) considered the area to comprise late Precambrian volcanics and George River Group equivalents, based largely on Milligan's work.

In the summer of 1979, reconnaissance mapping examined the transition from lower grade to higher grade rocks in the Middle River area and attempted to determine some definite basis for correlation between the Cape Breton Highlands and southeastern Cape Breton.

Four broad units were recognized on the basis of the first summer's work: a southern and eastern volcano-sedimentary sequence, a western and central zone of schists and amphibolites, a northern area of granitic gneisses, and various granitic plutons. In the summer of 1980, mapping of the volcano-sedimentary sequence, referred to here informally as the Crowdis Mountain volcanics, was undertaken at a scale of 1:15 000. This paper discusses the field relations and possible age of these volcanics; preliminary results suggest that the sequence may be early Paleozoic rather than Precambrian.

Lithology

The Crowdis Mountain volcanics consist of mafic to felsic volcanic rocks, lithic pyroclastics, and fine grained sediments, cut by diorite and microgranite sheets (Fig. 2). In the south, at Hunters Mountain and MacMillan Mountain, they are overlain by buff sandstones and conglomerates probably of Early Carboniferous (Horton) age. On the west and southeast, the volcanics appear to be in fault contact with brick red Horton sandstones and conglomerates. On the east, they are faulted against foliated monzodiorite, and to the northwest they grade into mylonite which is bounded on the north and west by granitic rocks.

The southern end of the volcanic complex (MacMillan Mountain, MacDonald Brook) consists of pink to purple-brown rhyolites, locally spherulitic, interlayered with coarse grey-green arkosic sandstones containing abundant volcanic fragments, green to maroon shales, and conglomeratic layers containing abundant maroon shale fragments, jasper fragments and rhyolite clasts. The sandstones are immature, containing doubly terminated quartz crystals, shale chips, and euhedral potassium feldspars. Minor amounts of lithic pyroclastics containing fragments of maroon shale and grey-green or buff ash and siltstone also occur in the southern area. Rare, isolated outcrops of mafic amygdaloidal

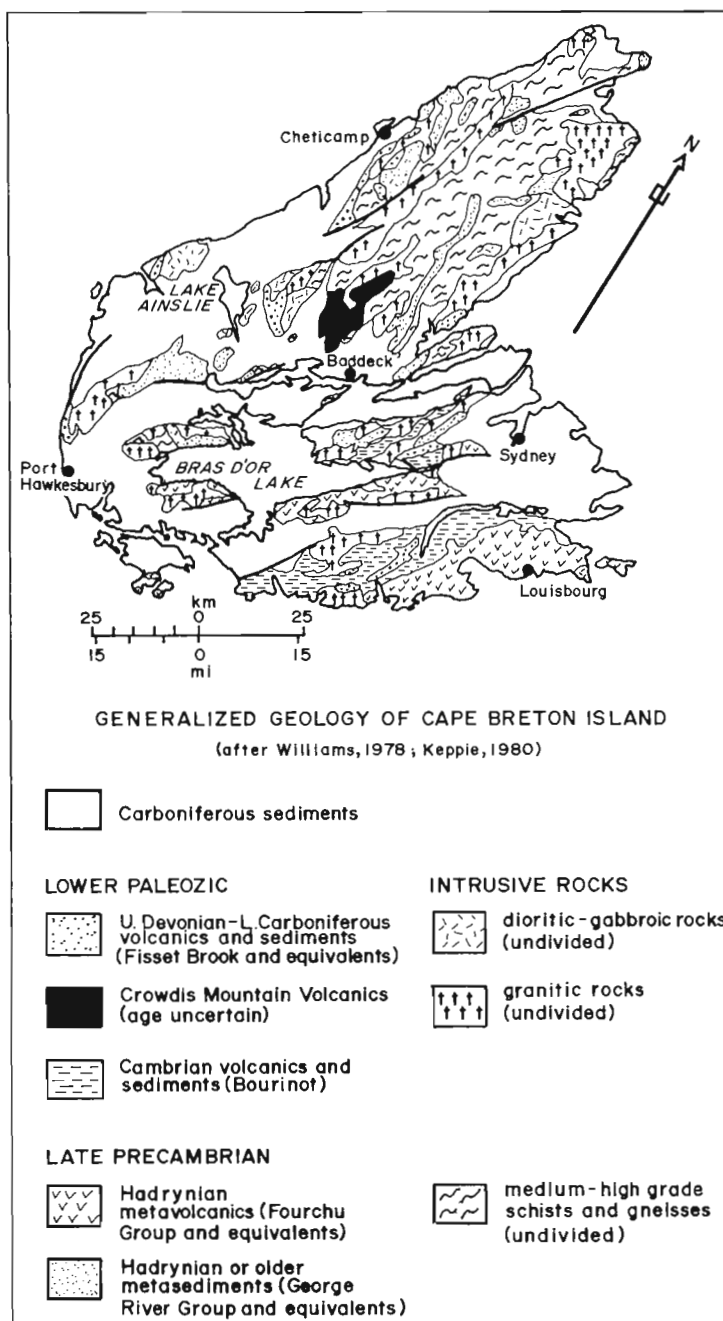


Figure 1. General geology, Cape Breton Island (after Williams, 1978; Keppie, 1980).

volcanics are present. The rocks are undeformed except for a slight phyllitic sheen and irregular cleavage in the shales and fracturing in all units.

Farther north, mafic and intermediate volcanics, lithic pyroclastics, and siltstones are more common, and the rocks are cut by green diorite and pink microgranite sheets, possibly subvolcanic sills (Leonard McLeod Brook). These sheets are locally mutually intrusive, although more commonly the microgranite intrudes the diorite, and they underlie much of the central part of the area (Big Barren). Possible hybrids, consisting of leucodiorite with large alkali feldspar crystals, are common in some areas. North of Big Barren, lithic pyroclastics become still more abundant; locally they contain cobble-sized rhyolite and siltstone

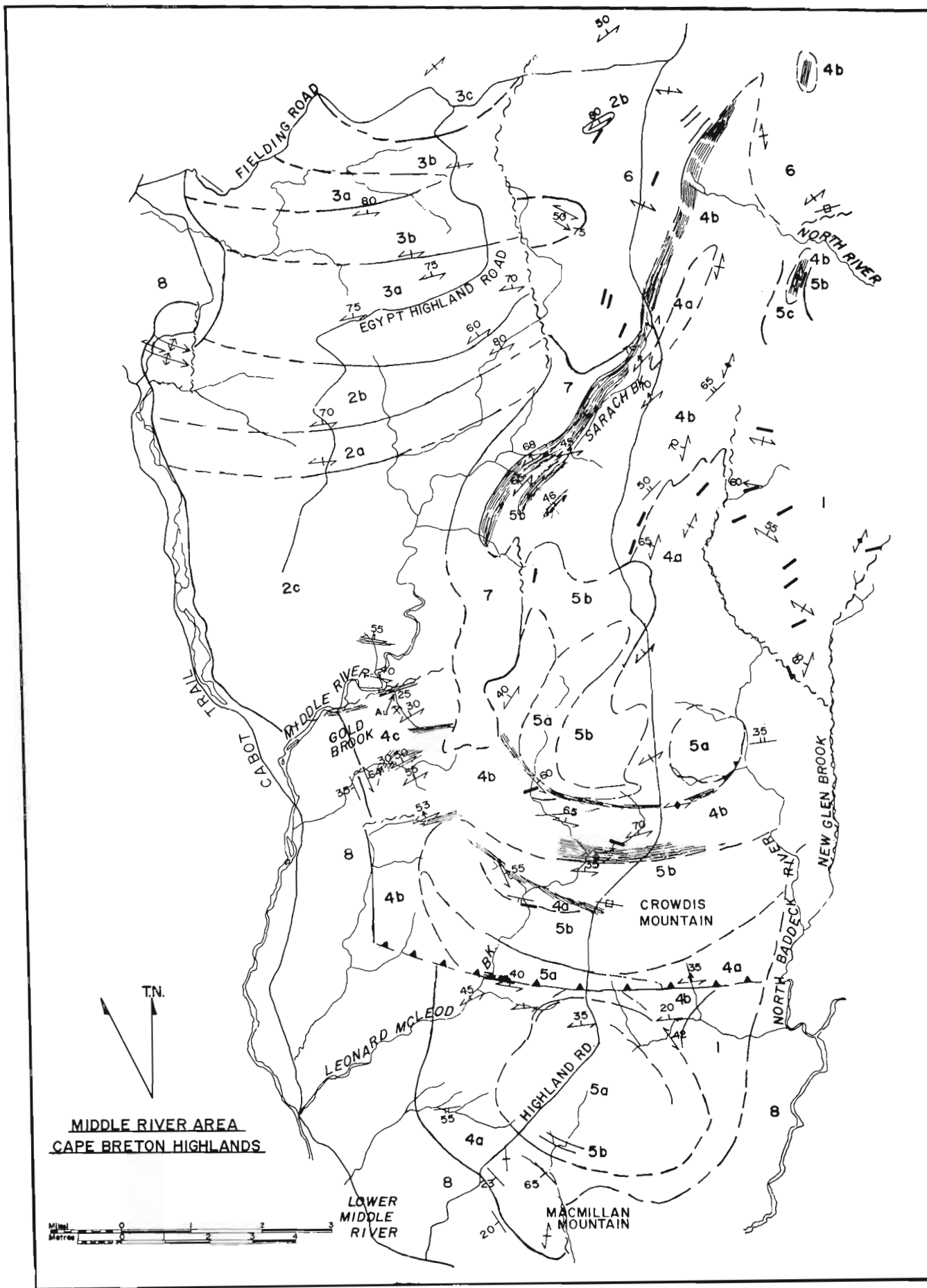


Figure 2. Geology, Middle River area, Cape Breton Highlands.

LEGEND FOR FIGURE 2

- 8 Undivided Carboniferous conglomerate, sandstone, shale (probable Horton equivalents)

INTRUSIVE ROCKS

- 7 Medium grained equigranular syenogranite, undeformed
6 Medium- to coarse-grained biotite ± hornblende monzogranite to granodiorite, locally weakly foliated

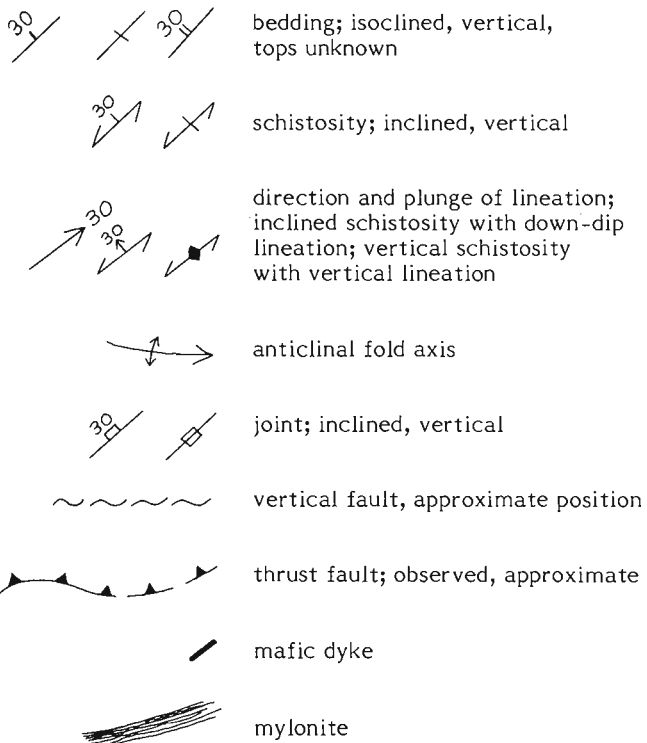
CROWDIS MOUNTAIN VOLCANICS

- 5 Fine- to medium-grained intrusive rocks, possibly subvolcanic sills.
5a - diorite;
5b - microgranite
4 Volcanic rocks and related pyroclastics and sediments.
4a - mafic and felsic flows, interbedded crystal and lithic tuffs, and minor sediments;
4b - mainly lithic pyroclastics, minor shales and siltstones;
4c - mainly fine grained sediments with minor conglomerate and pyroclastic horizons, also phyllite and schist

MIDDLE RIVER METAMORPHIC COMPLEX

- 3 Deformed granitic rocks.
3a - fine grained biotite granite schist;
3b - granitic augen gneiss;
3c - foliated syenogranite, in part gradational into 3b
2 Metasedimentary rocks.
2a - gneissic metasediments and amphibolite;
2b - pelitic and semi-pelitic schists, amphibolite common;
2c - pelitic and semi-pelitic schists, amphibolite rare
1 Foliated hornblende - biotite monzodiorite

KEY



fragments and are interlayered with blue-grey siltstones and green layered mafic tuffs. Just west of the central North Baddeck River, a zone of distinctive foliated white rhyolite containing green and red alteration spots is interlayered with pink and grey banded rhyolite and intermediate flows. Just east of the North Baddeck River, at the southern end of a logging road, foliated monzodiorite passes into pebble conglomerate, volcanogenic sandstone, and an intermediate dyke or flow. There is some suggestion of faulting, possibly modifying an unconformity, but the exposures are too poor to accurately define the nature of the contact.

Along the west side of the area (Garry Brook, McLean Brook), the Crowdis Mountain volcanics consist mainly of a fine grained pink felsic schist, possibly mylonitic felsic volcanics. Locally, deformed felsic pyroclastic horizons are preserved. This zone of deformed pyroclastics probably continues as far north as the deformed conglomerate and phyllites at Second Gold Brook described by Delahay (1979), where deformed felsic pyroclastics have been observed near the south end of the brook.

Structure

The intensity and style of deformation within the Crowdis Mountain volcanics are highly variable. At the southern end of the area, the volcanics dip and face at shallow angles to the south and southwest. However, elsewhere in the sequence vertical dips are common, and parts of the volcanics may face northwest although facing criteria are rare. Strikes vary from roughly east-west in the south to markedly northeast-southwest in the north. The rocks range from essentially undeformed, as at MacMillan Mountain, to mylonitized, as at Sarach Brook. However, there does not appear to be a simple strain gradient from south to north. Rather, there are several east-west trending fault zones, some marked by narrow mylonite belts. These faults apparently do not disrupt the sequence greatly; similar lithologies occur on either side. The fault zones dip moderately north and invariably show down-dip lineations, suggesting steep thrusting or shallow reverse faulting directed from the north towards the south. This interpretation provides some support for the suggestions of Milligan (1970) and Currie (1977) that parts of the Cape Breton Highlands are thrust over the surrounding Lower Carboniferous sediments.

The most prominent structure in the area is the northeast-trending Sarach Brook mylonite zone. This is developed from siltstones and lithic pyroclastics that grade into it with increasing development of penetrative foliation, isoclinal folding, transposition of sedimentary layering and deformation of lithic fragments. The resulting rock is a very fine grained, well foliated, commonly banded, grey to pink protomylonite to mylonite which has a well developed vertical lineation defined mainly by ash fragments. This indicates vertical movement on the zone, but the sense of movement was not determined in the field. The Sarach Brook mylonite zone separates the Crowdis Mountain volcanics from the metasedimentary schists, amphibolites, and rare gneisses that occur north of Middle River. It is cut off by granites to the north and east, and has not been traced to the southeast.

Metamorphism

Metamorphic grade in the Crowdis Mountain volcanics is generally low - the greenschist facies (chlorite and biotite zones) predominates. At this grade, felsic volcanics contain sericite and hematite, and are easily identified except where deformed. Mafic volcanics and sills locally contain secondary hornblende and epidote as well as chlorite and may be transitional between the greenschist and the amphibolite

facies. In the Gold Brook area, schistose metasediments contain syntectonic garnet and chloritoid, and late chlorite and biotite porphyroblasts associated with a crenulation cleavage. Garnet occurs locally in the Sarach Brook mylonite zone. Overall, the metamorphic grade is highest in the northwest, where the volcanics are intruded by granites and faulted against higher grade metasedimentary rocks.

Intrusive Rocks

Several types of intrusive rocks can be distinguished in the area. In the north, foliated syenogranite grades into augen gneiss in a belt that is concordant with the high grade schists north of Middle River. To the east, the volcanics are faulted against foliated biotite-hornblende monzodiorite. Both the syenogranite and monzodiorite probably predate the volcanics although no cross-cutting relationships have been observed.

Along Highland Road, the volcanics and mylonite are intruded by a black and white monzogranite to granodiorite with some pink potassium feldspar phenocrysts, that ranges from mildly foliated to undeformed. A homogeneous potassium-feldspar-rich granite bounds the Sarach Brook mylonite on the northwest. It probably intrudes the mylonite, although the field relations are not clear and the mylonite contains zones of what appears to be deformed granite. Its peculiar elongated shape may be fault-controlled.

The volcanic rocks are intruded by a set of sheet-like microgranites and diorites. They trend roughly east-west, that is, they are broadly concordant with the structural and lithological trends. They may be subvolcanic sills, since their compositions and metamorphic grade are similar to the volcanic rocks; however, the appearance of rare microgranite sills cutting the Sarach Brook mylonite zone suggests that these intrusions could be younger than the volcanic sequence. They predate the east-west trending faults.

A few mafic dykes cut the volcanics and metamorphic rocks in the north; these trend broadly northeast and some contain titanite, indicating a mildly alkaline bulk composition.

Age

No fossils or radiometric ages are available from the Crowdis Mountain volcanics; one aim of the study was to collect samples for dating. Therefore, their age can only be guessed at by comparison with sequences of known age elsewhere in Cape Breton. Obviously the rocks are pre-Early Carboniferous, as they are unconformably overlain by Horton-type sediments. They have been assigned to the Precambrian by previous workers, apparently on the assumption that all nonfossiliferous rocks in the Cape Breton Highlands are Precambrian. Milligan (1970) suggested that they were part of the George River series, a metasedimentary sequence of probable Hadrynian age. However, the metavolcanics that occur locally with the George River group (e.g. in the Creignish Hills) do not resemble the Crowdis Mountain volcanics in lithology, structure or degree of metamorphism. The Hadrynian Fourchu Group of eastern Cape Breton, with which the rocks were correlated by Kelley (1967), consists of calc-alkaline pyroclastic rocks including crystal tuff and ignimbrite with some mafic to felsic flows (Keppie et al., 1980). These are generally strongly foliated and locally isoclinally folded, and do not resemble the Crowdis Mountain volcanics in either lithology or structural style.

These volcanics, therefore, do not appear similar to known Precambrian volcanics elsewhere in Cape Breton. They may, of course, represent part of a distinct Precambrian sequence; however, their relative freshness and lack of deformation in an area that marks the convergence of

three Appalachian zones (Gander, Avalon and Meguma) and that must have undergone major lower Paleozoic and/or Precambrian deformation makes it difficult to believe that the sequence is, in fact, Precambrian.

At the other extreme, neither do these volcanics closely resemble the upper Devonian Fisset Brook Formation in its type area near Cheticamp. These rocks are mainly flows, have a much higher proportion of mafic and intermediate material, and are interlayered with brick red micaceous sandstones and siltstones, that are unlike any sediments associated with the Crowdis Mountain volcanics. However, the volcano-sedimentary sequence in the Lake Ainslie area, which contains upper Devonian/lower Carboniferous plant fossils and has been correlated with the Fisset Brook Formation (Kelley and MacKasey, 1965), bears at least a superficial lithological resemblance to the undeformed volcanics at MacMillan Mountain (Blanchard, personal communication, 1980).

The Crowdis Mountain volcanics are therefore probably Early Paleozoic, and at least in the south may be late Devonian. Problems remain in determining the equivalence of the undeformed rocks at MacMillan Mountain to the variably deformed rocks farther north. No break was observed in the field, and they have been included together, but the diorite-microgranite complex may obscure the contacts. The age of the sequence will thus remain uncertain until a reliable fossil or radiometric date is obtained.

Discussion

The work so far has established that the Crowdis Mountain volcanics are a sequence of mafic to felsic flows and pyroclastics cut by diorite and microgranite sills, possibly of middle Paleozoic age, that grade into mylonite in the northwest and are cut by numerous east-west faults. Numerous problems remain both with the field relations and with the interpretation of the data. Mapping problems include distinguishing among mylonite, rhyolite, deformed rhyolite and microgranite, and finely banded siliceous siltstone, determining the stratigraphy of the sequence, working out the relation of the volcanics to the higher grade rocks north of Middle River, and determining the relation of the microgranite and diorite sills to the volcanics and each other. Other problems include the local and regional significance of the deformation, the petrology of the sequence, and the age and correlation of the volcanics as discussed above.

The work being done in this area has implications for the geology of the Cape Breton Highlands in general. It is clear that the volcanics are not part of the George River group as this term is generally used in Cape Breton (albeit that the term is poorly defined) (cf. Milligan, 1970), although George River lithologies may occur in the metasedimentary sequence north of Middle River. The sequence does not resemble Precambrian volcanics of eastern Cape Breton and is probably Paleozoic. If the rocks at the southern end of the Highlands, closest to the known Precambrian units of Cape Breton, do not fit the Precambrian pattern, just how good is the widely accepted interpretation of the rest of the Cape Breton Highlands as Precambrian? Certainly, if a few Cambrian granite ages (Clarke et al., 1980) and the general "Avalonian" nature of the terrane are accepted (e.g. Williams, 1978), the Highlands should be underlain largely by late Precambrian rocks. However, large regions of the Highlands remain virtually unmapped, radiometric data are sparse, and metamorphic grade and degree of deformation cannot be considered sufficient criteria for determining age. If the volcanics are, in fact, late Devonian, the extent of post-Acadian volcanism in Cape Breton must be re-evaluated. This study suggests that much more work is needed in northwest Cape Breton before its role in Appalachian evolution can be unravelled.

Acknowledgments

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Addendum

Preliminary radiometric dates obtained at Memorial University in May 1981 indicate an age of 384 ± 10 Ma for rhyolites at MacMillan Mountain (Unit 4a), 414 ± 40 Ma for the monzogranite in the Highland Road - Fielding Road junction area (Unit 6), and 363 ± 38 Ma for the diorites on Leonard McLeod Brook (Unit 5a).

THE X-RAY CRYSTALLOGRAPHY OF TUNGSTITE

Project 680023

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Introduction

Tungstite, ideally $\text{WO}_3 \cdot \text{H}_2\text{O}$, is a rare secondary tungsten oxide mineral found in Canada at the Kootenay Belle mine, Salmo, British Columbia; Marlow, Beauce County, Quebec; Emerald, Inverness County, Nova Scotia and 2 miles west of Moose River, Halifax County, Nova Scotia. Interest in this particular mineral species arose from reading an excellent review article by T.G. Sahama on secondary tungsten minerals (Sahama, 1981). Regarding tungstite, Sahama stated, "As a mineral species tungstite is still inadequately described and needs to be studied with modern laboratory techniques.". The most recent comprehensive study (Kerr and Young, 1944) failed to uncover tungstite crystals large enough for crystallographic study. A search of the National Mineral Collection housed within the Geological Survey of Canada located several tungstite specimens from Canada and elsewhere. However, only one, GSC No. 16549 labelled as "Kootenay Belle mine, Salmo, B.C." had tungstite masses sufficiently coarse enough for detailed mineralogical examination.

Table 1

X-Ray powder diffraction data for tungstite

I est.	dÅ meas.	dÅ calc.	hkl
80	5.36	5.35	020
10	4.60	4.62	011
20	3.73	3.74	120
* 100	3.463	3.464	111
* 30	2.924	2.927	031
* 30	2.677	2.676	040
* 40	2.616	2.619	200
		2.560	002
50	2.556	2.555	131
* 20	2.383	2.383	140
* 20	2.351	2.352	220
* 30	2.312	2.309	022
* 10	2.113	2.113	122
* 20	1.975	1.975	051
* 15	1.951	1.952	231
5	1.867	1.872	240
		1.850	042
40	1.851	1.848	151
* 40	1.831	1.831	202
3	1.783	1.784	060
5	1.746	1.744	142
* 45	1.731	1.732	222
		1.689	160
10	1.689	1.685	013
* 40	1.634	1.633	311
* 30	1.603	1.604	113
* 10	1.576	1.577	251
* 15	1.512	1.511	242
* 10	1.500	1.500	331
15	1.478	1.477	133
		1.465	071
10	1.465	1.464	062
		1.462	340
* 20	1.410	1.410	162
- 114.6 mm. Debye-Scherrer powder camera, Cu radiation Ni filter ($\lambda\text{CuK}\alpha = 1.54178\text{\AA}$)			
- film no. 64066, sample from Kootenay Belle Mine, Salmo, B.C. (no. 16549)			
- indexed with $a = 5.238$, $b = 10.704$, $c = 5.120\text{\AA}$			
* powder lines used for unit cell refinement			

A complete mineralogical re-examination of tungstite, including microprobe analysis and new optical determinations is currently underway and the results will be reported elsewhere. However, the powder and single crystal X-ray diffraction studies have been completed and the new data are reported herein.

Experimental

Numerous yellow platy tungstite fragments were extracted from specimen no. 16549 and examined by the precession orientation method. Of these, only two crystals, platy {010} and approximately 0.02 mm in longest dimension, were isolated which displayed no evidence of twinning or multiple crystallites. The planes $h0l \rightarrow h4l$, $Ok\bar{l} \rightarrow 2k\bar{l}$, $hk0 \rightarrow hk2$ and 101^*Ab^* were collected using both crystals in different orientations on precession single crystal cameras and employing Zr filtered Mo radiation. Exposures of 100 hours or more were necessary for detection of weakly diffracting nodes.

Results

Tungstite is orthorhombic with measured unit cell parameters (zero level precession films) $a = 5.27$, $b = 10.8$ and $c = 5.15\text{\AA}$. Systematic absences of the type (1) hkl no conditions, (2) $Ok\bar{l}$ no conditions, (3) $h0l$ with $l = 2n$ and (4) $hk0$ with $k = 2n$, preclude that the space group choices are $Pmcb$ (55) and $P2cb$ (32).

Both zero and upper level precession films display a strong pseudo-A-centred orthorhombic lattice with pseudo-systematic absences of (1) hkl with $k + l = 2n$, (2) $Ok\bar{l}$ with $k + l = 2n$, (3) $h0l$ with $h = 2n$ and $l = 2n$ and (4) $hk0$ with $k = 2n$. Only seven weak to very weak reflections (041, 023, 221, 112, 261, 102, and 104) were detected which violate the conditions necessary for an A-centred cell. A complete crystal structure determination of tungstite is planned for the near future to confirm these results.

A fully indexed powder pattern, in reasonable agreement with PDF 18-1418, is presented in Table 1. Note that none of the reflections which characterize the P-lattice were sufficiently intense to be represented on the 114.6 mm Debye-Scherrer powder film. The refined unit cell parameters, based on 18 lines between 3.463 $\text{\AA}$ and 1.410 $\text{\AA}$ for which unambiguous indexing from precession single crystal films was possible, are:

$$\begin{aligned} a &= 5.238 (2) \text{\AA} \\ b &= 10.704 (4) \text{\AA} \\ c &= 5.120 (2) \text{\AA} \end{aligned}$$

and calculated cell volume of $287.07 \text{\AA}^3$.[†]

Assuming an ideal formula of $\text{WO}_3 \cdot \text{H}_2\text{O}$, the calculated density with $Z = 4$ is 5.78g/cm^3 , in reasonable agreement with the measured result of 5.517 for tungstite from the same locality (Walker, 1908). The Kootenay Belle tungstite is intimately associated with ferritungstite ($D_m = 5.02$, Sahama, 1981) and the low measured result clearly suggests that a combination of tungstite and admixed ferritungstite was used for the original specific gravity determination.

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[†] A preliminary crystal structure determination gave unit cell parameters $a = 5.249(4)$, $b = 10.711(9)$ and $c = 5.133(4)\text{\AA}$ (J.T. Szymanski, personal communication, 1981)

DISCUSSIONS AND COMMUNICATIONS

DISCUSSIONS ET COMMUNICATIONS

A PROVISIONAL STANDARD FOR CORRELATING THE PRECAMBRIAN ROCKS OF THE CANADIAN SHIELD: DISCUSSION

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Introduction

It should be noted that the GSC report by Douglas (1980a) for correlating Precambrian rocks of the Canadian Shield constitutes Part 2 of "Proposals for time classification and correlation of Precambrian rocks and events in Canada and adjacent areas of the Canadian Shield". C.H. Stockwell (in press) has, as noted in the foreword to Dr. Douglas' paper, prepared a comprehensive report including a revision and expansion of his time scale that will form Part 1 of these proposals. Since both of these authors, particularly Dr. Stockwell, have spent years studying the problems of Precambrian time scales, classification, and correlation, the views and ideas of each deserve most careful consideration. The writer suggests that before there is acceptance or rejection of any innovations concerning these matters, the proposals of both authors should be compared and studied thoroughly, together with any pertinent discussion. The following discussion is intended to focus attention on issues arising from Part 2.

As stated by Douglas (1980a, p. 2), the purpose of his paper is to "propose additions to Stockwell's standard for Precambrian time and to present the author's reasons for doing so"; the objectives are to arrive at useful subera time units in order to correlate rocks as closely as possible, arrange them in temporal order, and depict them suitably on 1:1 000 000 compilation maps of the Canadian Shield. In pursuing these objectives, Douglas has evolved the standard presented in his Figure 4. It differs from those of Stockwell (1973; in press) in the number and duration of time subdivisions, in nomenclature, and to a restricted extent, in method. The net result is an elaborate standard, but one that contains a number of controversial aspects. It should be mentioned that a few important points of discussion, including some of his specific correlations (Fig. 5 to 7), still remained at the time of Dr. Douglas' untimely death, and the paper as published lacks his final version. The writer is aware that Dr. Douglas had informally agreed to reconsider a number of the issues raised in this discussion.

Pages 1 to 5 inclusive of his paper are largely philosophical, and had already been published elsewhere (Douglas, 1980b); they include the same basic views as those of Stockwell and others favouring a truly geohistorical approach to Precambrian correlation and classification as opposed to "boundaries of convenience", and need no further comment here. Douglas' detailed discussion of standards and his own in particular, however, lead to several comments.

Time-stratigraphic Units

Douglas (1980a, p. 6) correctly pointed out, as has the writer in discussing Stockwell's forthcoming report, that the time scale of the latter is not a time-stratigraphic scale comparable to those of the Phanerozoic. According to Douglas, however, two units of his own standard can be construed as time-stratigraphic (chronostratigraphic) units

comparable to conventional ones, i.e. rock sequences whose beginning and end can be specifically defined, timed and correlated. These, as discussed later, are rocks of his "Huronian Subera", divided into early and late intervals, and his Middle Neohelikian division. It appears from this and from his remarks on paleomagnetism (p. 4) that he believed a chronostratigraphic scale for the Precambrian to be desirable and eventually feasible, although he discounted the use of fossils and the direct isotopic dating of sedimentary rocks (p. 3-4); Douglas based his view on the use of paleomagnetic characteristics in stratigraphic sequences. This writer believes that a chronostratigraphic scale is unrealistic for the rocks of the Canadian Shield, as for many other Precambrian regions. Given the ability to classify and correlate rocks directly against a time scale by means of isotopic dating and/or geological relations, there appears to be no particular need for a chronostratigraphic scale, but in any case metamorphism is a great barrier to the paleomagnetic approach, just as it commonly is to determination of primary ages. Although "magneto-stratigraphic" intervals are similar to conventional time-stratigraphic units provided the data from the stratigraphic successions are complete and unequivocal, there appears to be little possibility that enough well preserved sequences exist in the Canadian Shield to allow the construction of a scale for any more than a restricted portion of Precambrian time. Correlation of sequences over long distances using paleomagnetic data alone does not seem practicable. Apparent polar paths, of course, are completely dependent on isotopic age determinations.

Standards for Precambrian Time

In summarizing the Stockwell model, Douglas (1980a, p. 6) concluded rather critically with the statement that "certain" rocks must be classified into numerically-bounded categories by means of their isotopic age alone (i.e. those cases where there are insufficient geological relationships available for guidance), presumably an undesirable procedure in his view. This would be true whether the boundaries are arbitrary or rock-referenced, either to a type stratigraphic section or otherwise. The unstated alternative is, of course, to classify or correlate them much less rigorously or even not at all. Obviously, indiscriminate use of isotopic dates is to be avoided, but keeping in mind the restricted scope of Precambrian stratigraphic study to date (essentially confined to lithostratigraphy), the incomplete and dispersed supracrustal record of Precambrian time, and the effects of metamorphism, commonly repeated, over vast areas, isotopic age determination, whether of primary deposition or of metamorphism, must be recognized as the extremely valuable tool it is. Only through isotopic dating has there been the possibility of forming any sort of sophisticated or comprehensive time scale for the Precambrian, and without it Precambrian stratigraphic classification in Canada would scarcely have advanced in the last 30 years. Therefore, the use of the technique, either alone or with other aids, is not to be downgraded. Even with this advantage and the use of available geological relationships, Precambrian classification or correlation under either the Douglas or the Stockwell scheme, which appear complex to some, is still very rudimentary compared with Phanerozoic correlation, the Precambrian time divisions ranging up to 500 million years in length. Classification by any method into one of these large compartments is clearly justified and necessary.

Table 1
Modified standard for the Canadian Shield

EON	ERA	SUBERA	PERIOD	OROGENIC EVENT	NON-OROGENIC EVENT	POLARITY	AGE - Ma
P R O T E R O Z O I C	HADRYNIAN	LATE		AVALONIAN	CORONATION	R	700-
		EARLY					800- 900-
	HELIKIAN	NEOHELIKIAN	LATE	GRENVILLIAN			1000-
					DULUTH	N	1100-
					LOGAN	R	1200-
		EARLY		ELZEVRIRIAN			
					COPPERMINE	N	1300-
					SEAL	R	
		PALEO- HELIKIAN	LATE		NAIN ELSONIAN	M	1400-
					HARP	M	1500-
				KILLARNEAN	CROTEAU	R	
		EARLY					1600- 1700-
	APHEBIAN	LATE		HUDSONIAN	DUBAWNT	M	1800-
		MIDDLE		MORANIAN			1900- 2000-
					NIPISSING	N	2100-
		EARLY		BLEZARDIAN			2200- 2300- 2400-
							2500-
				KENORAN			2600-
A R C H E A N	1						2700-
	2			LAURENTIAN			2800- 2900-
	3			WANIPIGOWAN			3000- 3100- 3200- 3300-
	4			UIVAKIAN			3400- 3500-

The writer's comments on the Douglas standard are:

1. Archean

(a) From geological and isotopic evidence in Superior Province, the boundary at circa 2700 Ma appears to be assuming increasing importance, therefore it should have at least equal status with the other Archean boundaries. Such is not the case in the Douglas standard. Unlike either Douglas or Stockwell, the writer favours four subdivisions of equal status (eras) in the Archean as shown in Table 1; these should be named as soon as possible to avoid difficulties with informal names (early, middle, late, etc.).

(b) Proposed usage of the time terms "Keewatian" and "Timiscamian" for Late Archean time subdivisions (Douglas, 1980a, Fig. 4) is inadvisable because it would lead to confusion. These terms derive from "Keewatin" and "Timiskaming", lithological or lithostratigraphic names of long standing and still current use that are applied to specific rock groups of type areas which, unfortunately, are distant from each other. The Keewatin rocks of the type area (Lake of the Woods, Ontario) have an unknown maximum age, so that even their classification as "Keewatian" would not be justified. The time interval so designated is in any case greater than that of the deposition of the Keewatin sequence (which is also unknown) so that there is no direct time-stratigraphic connotation with respect to that sequence as one might expect from the name. Objections to "Timiscamian" are similar.

2. Early Aphebian (the "Huronian" of Douglas)

This interval begins at the close of the Kenoran Orogeny and terminates at the estimated close of intrusion of the Nipissing Diabase of Southern Province. This terminal boundary (2100 Ma) is close to that (2140 Ma) of Stockwell (in press) using the estimated close of the Blezardian Orogeny of the same region, set essentially by the intrusion of the Creighton and Murray granites, so it is of no particular advantage in classifying rocks. Adoption of Douglas' usage and nomenclature, however, would result in a time span for his "Huronian Subera" greater than that involved in the deposition of the Huronian Supergroup, which is also confusing, notwithstanding the suggested change to "Huron" Supergroup, a name that is preoccupied. It would include intrusion of post-Kenoran dykes, post-Kenoran uplift and erosion, deposition of the supergroup, early post-Huronian folding and metamorphism, intrusion of Creighton and Murray granites, and intrusion of Nipissing Diabase. Douglas' attempt to change the meaning of the lithostratigraphic term "Huronian" is particularly objectionable in view of its clear definition by Robertson et al. in 1969, specifically to clarify and simplify its usage and eliminate its previous misuse as a time-stratigraphic term (Huronian "series" or Huronian "system"). Douglas' argument for an intrinsic time connotation for "Huronian" is not convincing. The absolute time span of Huronian Supergroup deposition is very imprecisely known and there are no adequate stratigraphic criteria specifically defining its beginning or end; by modern criteria and definitions, "Huronian" has never been a valid chronostratigraphic term since the rocks were first designated a series in the middle of the last century, when virtually all supracrustal sequences, Phanerozoic or Precambrian, were indiscriminately called "groups", "series", or "systems". Although Douglas' desire for a name for the early Aphebian subera is understandable since he subdivides this interval into early and late parts, it is strongly recommended that as a time term, "Huronian", along with "Keewatian", and "Timiscamian", not be adopted.

Further, subdivision of this early Aphebian "Huronian" interval (see Douglas, 1980a, Fig. 4), based on an unconfirmed isotopic age of deposition of the Gowganda Formation of the Huronian Supergroup, is unnecessary and unjustified geologically and isotopically. Douglas has accepted, at least provisionally (p. 14), a single Rb-Sr isochron age from clastic (greywacke and argillite) beds of that formation as the older limit of his "Late Huronian". The fact that this number (2240 Ma) falls within the known time bracket (2500-2115 Ma) for the Huronian Supergroup does not make it sufficiently definitive in this writer's view for use as a boundary in a time standard. Stratigraphically, there is no appreciable discordance at this level in the Huronian succession, and the reversed magnetism mentioned by Douglas and shown on the standard is equivocal and in any event is not necessarily a diagnostic paleomagnetic "fix" in the Huronian succession useful for correlation with other sequences, strata other than the Gowganda greywacke and argillite not having been tested. There is really no adequate geological, paleomagnetic, or isotopic basis as yet for subdividing the early (i.e. pre-Nipissing Diabase) Aphebian. That there is no present need for the proposed boundary is evident from his Figure 7, where the assignment of early Aphebian sequences with respect to it is uncertain in every case except for the Huronian Supergroup itself.

3. Middle and Late Aphebian

Here the discussion of Southern Province and related comment and classification (p. 13-14, Fig. 4, Fig. 7 columns 3 and 4) call for some remarks:

(a) In the standard (Fig. 4), Douglas has placed the Penokean Orogeny at 1700 Ma, making it post-Hudsonian as he re-defines the latter (see below), and creating a serious source of confusion. Evidence from the type area south of Lake Superior indicates that the Penokean is early- or pre-Hudsonian, and relationships in the Sudbury area of Ontario indicate that it is a discrete pre-Hudsonian event that preceded deposition of the Whitewater Group, intrusion of the Sudbury Irruptive and their subsequent deformation. Thus, the post-Whitewater metamorphism and other effects attributed to the Penokean in the eastern part of Southern Province (p. 13) are better said to be Hudsonian. The age cited by Douglas for this metamorphism (1684 Ma) is anomalously young for either event and probably should not be accepted at face value; as pointed out by Stockwell (in press), this also applies to young ages from the micropegmatite of the Sudbury Irruptive. Following invited comment on his manuscript, Douglas indicated his intention to revise his position of the Penokean downward (personal communication, 1979). Its removal from the 1700 Ma point on the standard presumably would have necessitated deletion of his "Early Paleohelikian".

(b) A Rb-Sr age of 1915 ± 98 Ma is cited for the norite and sublayer of the Sudbury Nickel Irruptive (p. 13), which is placed on the standard (Fig. 4) at the Middle Aphebian-Late Aphebian boundary. Not cited is a more precise U-Pb concordia age of 1845 Ma (Krogh and Davis, 1974) for norite, since repeated to within 5 Ma (T.E. Krogh, personal communication, 1980). Use of this intrusion and its U-Pb age to fix this boundary would not be suitable, however, as it would limit Douglas' Late Aphebian to 45 Ma; actually, although not apparent from the standard of Figure 4, the author evidently preferred to relate the boundary (p. 13) to the onset of deformation of the Tazin Group of the Lake Athabasca district, estimated at 1925 Ma. This is the same deformation given the designation "Moranian" by Stockwell (in press), and whose close, estimated at 1890 Ma, defines the middle Aphebian-late Aphebian boundary of that author. One might therefore expect this event to appear, named or unnamed, on the standard of Figure 4.

The correlation chart of Figure 7 positions the Sudbury Nickel Irruptive below the Whitewater Group, the better to accord with the suggestion that it is cogenetic with the overlying Onaping Formation of that group. The more widely accepted view is that the irruptive is not cogenetic, and being post-Onaping (Thomson, 1957, p. 30) and presumably post-Whitewater, it is better placed above those in the chart.

(c) Douglas asserts (p. 13-14) that the rocks of the Circum-Ungava geosynclines, the Marquette Range Supergroup of Michigan, and the Animikie Group can only be classified as Apebian (undivided). However, as Stockwell (in press) has noted, the maximum age of the Marquette Range Supergroup, from a zircon concordia derived from basement rocks, is 2066 Ma, and volcanics of that sequence are dated by the same method at 1873 ± 36 and 1859 ± 30 Ma. From these, the Marquette Range rocks may be classed as Middle to Late Apebian in both Douglas' and Stockwell's scales. Presumably – equivalent Animikie strata may be similarly placed. Strata of the Circum-Ungava belts are, from strong lithostratigraphic evidence, likely of approximately similar age and in the absence of definitive isotopic ages (Stockwell, in press), are also to be excluded from the early Apebian.

(d) Coming to the later Apebian, Douglas notes that deformation in the Tazin Group north of Lake Athabasca is older than that of his "type" Hudsonian area, discussed below, but suggests that it is earliest Hudsonian (p. 13); as noted above, its estimated age of onset (1925 Ma) is used as a provisional boundary between his Middle Apebian and Late Apebian (p. 13). Note the departure from Stockwell's methodology in the use of "onset" instead of closing of orogeny. As also noted above, this is the deformation distinguished separately from the Hudsonian by Stockwell and termed "Moranian Orogeny". In that area (the Beaverlodge district) biotite post-tectonic cooling ages are, at 1790 Ma (average of 62 determinations), virtually the same as the estimated age of tectonism (1800 Ma) of Douglas' redefined Hudsonian of the Ennadai Fold Belt (see below). Accepting these figures, this suggests that the Beaverlodge block was already deformed, intruded, uplifted, cooled, and possibly shedding sediments while syntectonic plutonism was occurring in the Ennadai Fold Belt some 500 km to the northeast, and lends credence to Stockwell's concept of a separate Moranian event, particularly if, as isotopic ages suggest, the thick post-orogenic molasse of the Martin Group at Beaverlodge was nonconformably deposited, deformed, and its pitchblende deposits emplaced by about 1750 Ma (Stockwell, in press). Again accepting these figures, Douglas' alternative view might suggest that the Moranian tectonism north of Lake Athabasca is an early phase of a diachronous Hudsonian that progressed in space (500 km) and time (ca. 100 Ma) northeastward into the Ennadai region, where it produced a fold belt neither on strike nor parallel to the Moranian belt. Or, it might suggest that the Hudsonian began early in both regions and persisted longer in the Ennadai belt. More data are required to determine the tectonic relationship of these regions. Meanwhile, this writer believes Stockwell's separate treatment of these deformations to be preferable.

(e) Perhaps the most far-reaching and controversial innovation of the Douglas standard is his redefinition of the Hudsonian Orogeny and its type area (p. 12-13). Whereas the originator of these concepts defined the Hudsonian as the last important period of folding, etc., in Churchill Province as a whole (Stockwell, 1964), Douglas designates it as that tectonism affecting the Archean and Apebian rocks of a restricted part of the province identified previously (Douglas, 1971, 1974) as the Ennadai Fold Belt. The age of this tectonism is estimated to be 1800 Ma. This redefinition is done on the grounds that orogeny was diachronous in Churchill Province, that stratigraphy and other geological

relationships are reasonably well known in that belt, and that isotopic dating of supracrustal and intrusive rocks there can be reliably interpreted. Obviously no type area within Churchill Province that conforms to the original definition can be selected, even provisionally, until the province is much better known. The selection of a type area and redefinition of the orogeny by Douglas is an attempt to narrow and alter the concept of Hudsonian and to arrive at a more specific proposal is highly debatable for the following reasons:

(1) Departing as it does from fundamental principles heretofore applied to definition of all the eras, it creates a vital precedent. It also produces many differences in time and rock classification: specifically, it redefines the limits of five existing time intervals; the Apebian, the Late Apebian, the Helikian, the Paleohelikian, and the Early Paleohelikian. Conceptual change of this importance must have the strongest justification, most urgent need, and must be based on unequivocal relationships and data. The broader concept and definition of the Hudsonian having been in effect and applied widely for over 15 years, considerable confusion is bound to result. Numerous events in the range 1700-1800 Ma in various parts of Churchill Province, heretofore referred to as Hudsonian, would now have to be referred to the earliest Helikian. It would seem this move is premature.

(2) The isotopic age "reference points" used by Douglas in the Ennadai Fold Belt to set the boundary may not be as unequivocal as suggested. For example, Rb-Sr and K-Ar ages determined on Hurwitz Group volcanics and sills (1872 and 1815 Ma respectively) are taken by Douglas as metamorphic, whereas Wanless and Eade (1975) considered them to be primary. If metamorphic, they are only a minimum for Hurwitz strata, which then could be as old as early Apebian rather than younger as shown in Figure 7. If primary (depositional), the closing age of the "new" Hudsonian could then be less than 1800 Ma, even as redefined in the Ennadai Fold Belt; this boundary age results from acceptance, along with the two ages above, of a single Rb-Sr isochron (1834 ± 22 Ma) from synkinematic monzonite as definitive (p. 34), although obviously the monzonite age conflicts with the 1815 age for the pre-tectonic sill if the latter is a primary age. All in all, Douglas' Apebian boundary of 1800 Ma is based on only six isotopic ages, one of which is from Dubawnt volcanics that rest on a basement of uncertain age and relationship to the fold belt. Only the age of the anorogenic Nueltin Granite appears to be valid as a minimum for deformation in the Ennadai Fold Belt. The two isochron ages from the granite (1775 ± 59 and 1760 ± 16 Ma) fall close to Stockwell's Rb-Sr boundary (from all Churchill Province data), 1755 Ma, itself only a rough approximation. Further documentation of events in this belt is clearly required.

All things considered, it seems preferable in the present state of knowledge to retain the original definition of Hudsonian, regard its numerical age as a flexible best estimate that reflects the level of knowledge, and let selection of a type area within the Churchill Province that satisfies the definition await the accumulation of sufficient information. This type area will be manifested through isotopic dating in all the fold belts of the province. The original concept has proven generally adequate to the present for designating 1700-1900 Ma terminal tectonic events in many parts of the province; Douglas' proposal would require that they be pre-1800 Ma. The writer suggests expressing the boundary as 1750 ± 50 Ma.

Douglas, in calling for redefinition of the Hudsonian (p. 32-33), speculates about the effects of diachronism of the orogeny. To the writer's knowledge, this theoretical situation has not yet been encountered anywhere in the Canadian

Shield, but assuming that diachronism can be identified and its range quantified isotopically, the problem is non-existent if, in accordance with Stockwell's definition, the termination and the time boundary are selected at the youngest part of the range. This would obviate the suggestion of Douglas (p. 12) that Hudsonian orogeny might be Apebian in one place and Paleohelikian in another.

4. Helikian Era

In the Douglas standard, six subdivisions are created in this era, three each for the Paleohelikian and Neohelikian. There is probably no real need or indeed adequate basis for some of these subdivisions. Firstly, Douglas' "Early Paleohelikian" interval is badly defined, as its younger limit is the erroneously placed Penokean Orogeny of Southern Province, which, as mentioned above, should be moved to pre-1800 Ma in the scale (pre-Helikian). The "Early Helikian" of Figure 4 is thus untenable. Also, acceptance of the older limit of the Early Paleohelikian at 1800 Ma depends on acceptance of Douglas' redefined Hudsonian. Secondly, the "Middle Paleohelikian" appears to be largely superfluous since, as the author notes (p. 11), no rocks can definitely be assigned to it. While the purpose of a standard is to classify rocks as closely as possible, a two-fold Paleohelikian seems sufficient and all that is justified. The two-fold division may be based on terminal intrusion of the Killarney granites, taken by Stockwell to mark the close of the Killarney Orogeny at circa 1500 Ma (Stockwell, in press), rather than the suggestion of Douglas, in again departing from Stockwell's methodology, to use the Rb-Sr age of volcanism (1525 Ma) in the Croteau¹ and Petscapiskau groups of southern Labrador (see Fig. 4 and p. 11). Thirdly, a two-fold division of the Neohelikian is sufficient. The "Middle Neohelikian" of Douglas is defined (p. 8) as an interval of reversed polarity discovered in Keweenaw flows and intrusions of the western Lake Superior Basin, limited to a short span provisionally estimated from mainly K-Ar and Rb-Sr ages as 1150 to 1200 Ma. Not included in the estimate are U-Pb data indicating that the reversal ended subsequent to 1140 Ma (Silver and Green, 1972). These U-Pb data indicate also that the upper (normally polarized) Keweenaw volcanic succession was extruded in less than 20 million years, and as the interval of reversal comprises only the lower part of the Keweenaw succession, its duration may have been similarly brief. The Keweenaw rocks of the Lake Superior district, including associated intrusions, can be and have been adequately classified without the paleomagnetic data, and as shown in Douglas' Figure 5, only rocks of this subprovince (Lake Superior Basin) can presently be assigned to the Middle Neohelikian of his paper. As the time interval is only 50 Ma, it is of questionable value in long range correlation using isotopic ages, since these may have larger errors than this. It is recommended here that the "Middle Neohelikian", while possibly representing a valid magnetostratigraphic interval, be included in the Late Neohelikian. The latter can be defined geologically and isotopically as the time between the close of the Elzevirian Orogeny (or, if preferred, the intrusion of the Logan Diabase, synchronous at circa 1200 Ma), and the close of the Grenvillian Orogeny at 1000 Ma.

5. Hadrynian Era

Like Stockwell, Douglas divides the Hadrynian into early and late suberas, but has employed the intrusion of the Coronation diabase sills in northern Bear Province to establish the boundary rather than the Avalonian Orogeny of eastern Newfoundland as used by Stockwell, there being no deformed Hadrynian rocks in the Canadian Shield itself. To temporally limit the intrusion of the Coronation sills and set the boundary, Douglas has adopted a best estimate of 675 Ma, obtained by averaging a range of K-Ar isotopic dates from the intrusions and arbitrarily adding 25 Ma. Stockwell's boundary of circa 620 Ma (Stockwell, in press), is derived from the difference between cooling ages of intrusions that lie outside the type area (Avalon Peninsula of Newfoundland) but are considered to be syn-Avalonian and the age of an Ediacaran-type fossil assemblage in postorogenic strata of the type area in Newfoundland that is correlated with dated Ediacaran fossils elsewhere (Anderson, 1972)². Of the two options, the Coronation sills appear to offer more potential for a well established age, particularly if further investigated isotopically by Ar-Ar and/or Nd-Sm techniques, and on that basis might provide the preferred standard. Although anorogenic, the basic intrusive suite of which the sills form a part, the Franklin intrusions, are a widespread and important feature of the northern Canadian Shield. On the other hand, the Avalonian Orogeny is an important tectonic event even if it is not demonstrably widespread, and its use maintains consistency in the structure of the time scale. In practice, use of the different standards at present affects only the classification of the sills themselves – Late Hadrynian in Douglas' scheme and early Hadrynian in Stockwell's.

Adoption by Douglas of the Russian term "Vendian" as a time term equivalent to Late Hadrynian (Figure 4) is of questionable merit. According to Keller (1979), Vendian is a rock-stratigraphic term that has had different usage in the U.S.S.R. (originally a group, subsequently comprising a number of groups). Earlier, it was also used chronostratigraphically with the rank of stage (e.g. Sokolov, 1968). Its lower boundary is roughly fixed at 650-680 ± 20 Ma by K-Ar dating of glauconite in underlying beds. While the Vendian strata appear to represent a time interval approximately equal to Douglas' Late Hadrynian, changing the term to a time term as shown in the standard and also in Figure 5 seems arbitrary, confusing and inadvisable.

Summary and Recommendations

1. The proposed standard provides extensions of the Stockwell scale that are intended to classify Precambrian rocks of Canada as closely as possible. A number of these extensions and their nomenclature are questionable or unwarranted.
2. A chronostratigraphic scale is not practicable for the Canadian Shield. The Douglas scale, except for his Middle Neohelikian, is a chronotectonic-chronomagmatic scale. The Stockwell scale is entirely chronotectonic.
3. Paleomagnetic characteristics, while appropriate on the standard and on the correlation charts, are not independent correlation tools comparable to isotopic dating or paleontology. Paleomagnetic intervals useful in long range correlation have not yet been defined.

¹ The "Croteau" of Douglas is equivalent to the upper, volcanic part of the former Croteau Group and to the "Bruce River" Group of Smythe et al. (1975).

² Tillite in lower Conception Group strata, well below the Ediacaran fossils estimated to be 610-630 Ma old, is thought to be about 700 Ma old (Anderson, 1972); tillite in northern Cordilleran Hadrynian rocks not far from the Coronation sills may be of like age. If so, the equatorial pole position of the 650 Ma old Coronation sills, noted by Douglas, has added significance; presumably, a rapid shift of the pole occurred in the interval 650-700 Ma.

4. The Archean should comprise four intervals of era rank; these should be named in the near future.
5. The terms "Keewatian", "Timiscamian", "Huronian", "Huron Supergroup", and "Vendian", as used in this paper, should be rejected.
6. No more than a two-fold subdivision of the Paleohelikian and Neohelikian suberas is justified at present.
7. The Aphebian subdivision into three intervals is important now, in order to distinguish major supracrustal sequences and intrusive bodies. Douglas' additional subdivision is not recommended.
8. Designation of the "Ennadai Fold Belt" as the type area for the Hudsonian Orogeny is not recommended. The original definition of "Hudsonian" should remain, although eventually a more specific type area may be identified.
9. The Sudbury norite, Penokean Orogeny and Moyie sills are misplaced on the standard and/or correlation charts. Their recommended positions are 1845 Ma, circa 1850 Ma, and 1430 Ma respectively. Recent U-Pb geochronologic data indicate that the Dubawnt Group is Late Aphebian rather than Paleohelikian (A.N. LeCheminant, personal communication, 1981).
10. Adopting the recommendations herein, a modified standard appears as in Table I. Subdivision of the Paleohelikian and Neohelikian suberas justifies the additional column headed "Period"; the use of "period" and "subera" is in accord with the definition of "subera" in the Glossary of the American Geological Institute (Bates and Jackson, 1980) as proposed originally by Sutton (1940).

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ALTERATION PRODUCTS OF ACCESSORY ALLANITE IN RADIOACTIVE GRANITES FROM THE CANADIAN SHIELD: DISCUSSION

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The purposes of this note are:

1. to call attention to remarkable similarities between energy dispersive spectra published by Littlejohn (1981) and those obtained by Rimsaite on allanites in 1979 from the Grenville Structural Province (Fig. 1, 2); and
2. to provide semiquantitative analyses of inclusions and various phases of heterogeneous allanites (Table 1) shown in energy dispersive spectra.

Some of the features presented in Littlejohn's paper, such as heterogeneity, alterations and isotopic ages of allanites and other minerals have already been discussed and illustrated in previous papers by Rimsaite (1978, 1980, 1981a,b, in press).

Relatively fresh allanite grains from high grade uranium ore in the Bancroft area are made up of three main phases and contain numerous inclusions (Rimsaite, 1978, 1980). The "early allanite" phase consists (in decreasing order of abundance) of silica, iron, alumina and less abundant REE, Ca, Y, Mg, Th, U, Ti, Mn and Pb. A "younger" outer zone of the allanite is enriched in Ca and REE; and some metamict patches are enriched in Y, Th, U and Fe (Fig. 1, energy dispersive spectra, EDS, 1, 2, 3, 4; Table 1, analyses 1, 2, 3). Disseminated inclusions are galena, zircon, uranophane and uraninite; unidentified material filling fractures contains Fe, REE, Th and U (Rimsaite, 1978, 1980). Cameron-Schimann (1978) also described a REE-rich phase overgrown on, and apparently replacing, allanite crystals from Baie-Johan-Beetz area, Quebec.

Grains of altered allanite from Mont-Laurier area, Quebec, have a speckled appearance and grade into banded outer rims that consist of REE-, Th-, and U-rich phases (Fig. 2; Table 1, analyses 10, 11). On backscattered electron images, various phases of heterogeneous allanite appear in various shades of grey, from white to black, depending on their mass differences. Thus heavy minerals, such as galena, appear white whereas altered biotite is dark grey to black. Disseminated specks of diverse composition occur in a matrix of altered allanite that resembles in chemical composition an early allanite phase in Figure 1 but, since it contains some potassium, probably results from reactions with adjoining altered biotite (Fig. 2; Table 1, analyses 1, allanite; 7a, matrix of altered allanite; 7, iron-rich specks in 7a; and 8, patch of altered biotite). Note different Si/Al ratios in allanite and altered allanite (ca. 4) and in altered biotite (ca. 2). The allanite grain in Figure 2 is surrounded by a prominent rim and grains composed mainly of REE (Table 1, analysis 6).

Energy dispersive spectra (EDS) of rare-earth elements, of REE-bearing carbonate and of secondary Pb-U-bearing mineral aggregates (Littlejohn, 1981, p. 98 and 102) are strikingly similar to those published by Rimsaite (1981a, Fig. 17.5D, 17.1D; 1981b, Fig. 4.3-1). In the Grenville Structural Province, REE-rich phases crystallize during the late deuteric stage as REE-bearing rims surrounding various radioactive ore minerals and occur also around and in fractures of altered accessory minerals (Rimsaite, 1980, Fig. 38.6). Although earlier crystallized minerals may contain some REE, the REE-rich phases crystallize during a later deuteric stage (Rimsaite, in press). Secondary Pb-rich uranyl-bearing mineral aggregates are common in weathered radioactive rocks and apparently form in an oxidizing

environment, such as in partly oxidized pyrite, in fractures in biotite, in zircon, in Fe-rich thorogummite and on surfaces of various minerals (Rimsaite, 1981a, Fig. 17.6, 17.7).

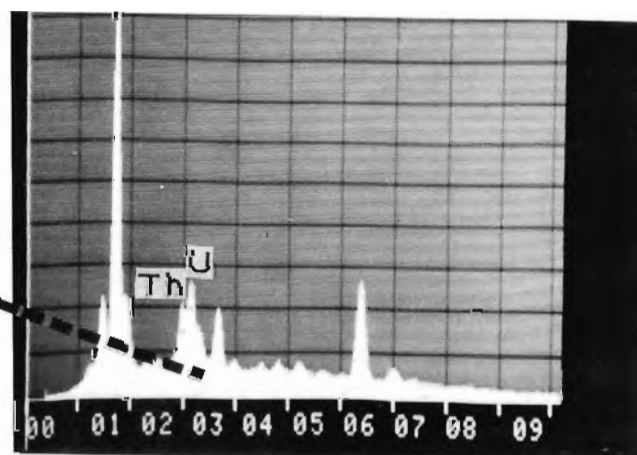
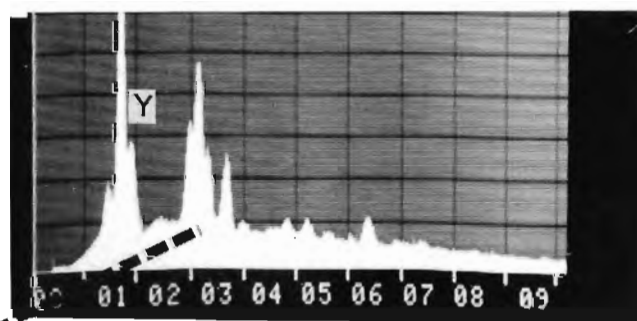
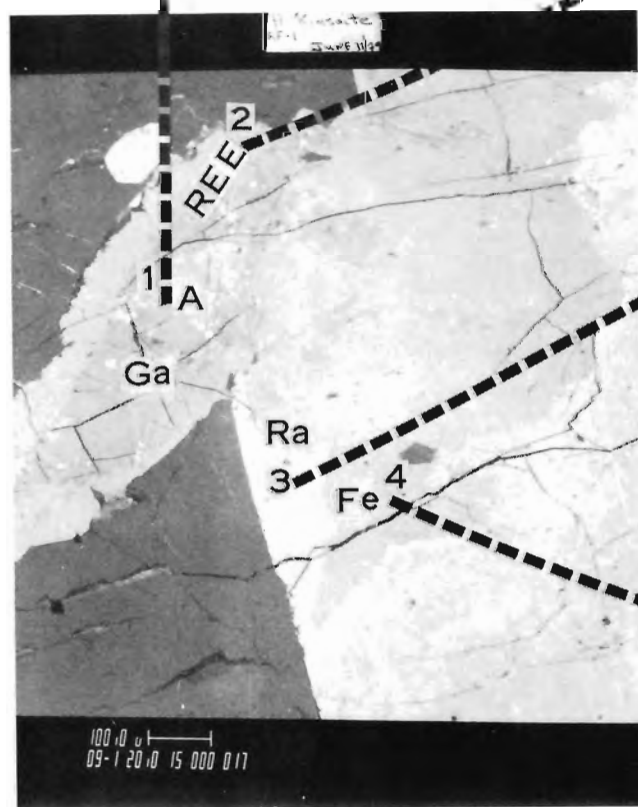
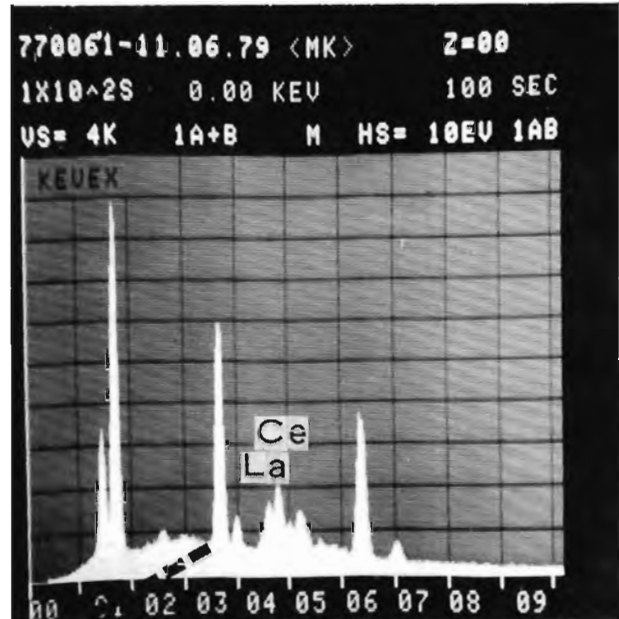
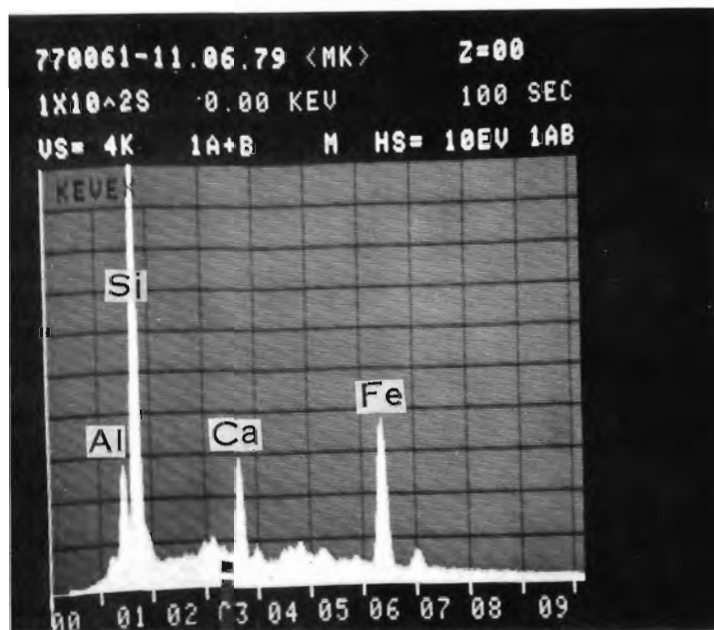
Allanite is a common host for uraninite, uranophane, zircon and other accessory minerals. Because of abundant inclusions, allanite concentrates from the Bancroft area had to be acid-treated to determine the REE content of inclusions (solution) and that of the residue of allanite grains (residue after leaching). Results indicated that the soluble fraction contained more REE, Y, Th and Pb than the residue, thus implying that most of the REE were present in inclusions (Rimsaite, 1980, Table 38.1). Consequently, in a study of alteration reactions, special care should be taken to determine which elements are derived from inclusions and which from the allanite grains themselves.

The paper by Littlejohn lacks time-perspective. I believe that Littlejohn failed to distinguish between deuteric alteration and weathering or oxidation, that he has confused inclusions and REE-bearing phases of the allanite rim and its adhering minerals with the phases derived from the inclusion-free allanite grains during alteration. He assumed also that minerals identified on X-ray patterns originated from the alteration of allanite and has presented formulae 1 and 2 to explain these changes. These formulae are in my opinion purely theoretical and speculative.

An alternative interpretation and study of alteration of accessory minerals in radioactive granites, based on the sequence of crystallization of inclusions and adhering minerals and on their mutual interaction has already been proposed by Rimsaite (1981b).

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Mineralogical and petrochemical properties of heterogeneous granitic rocks from radioactive occurrences in the Grenville Structural Province, Ontario and Quebec; in *Proceedings of the Uranium in Granites Workshop*, Ottawa, Nov. 1980, ed. Y.T. Maurice, Geological Survey of Canada, Paper. (in press)

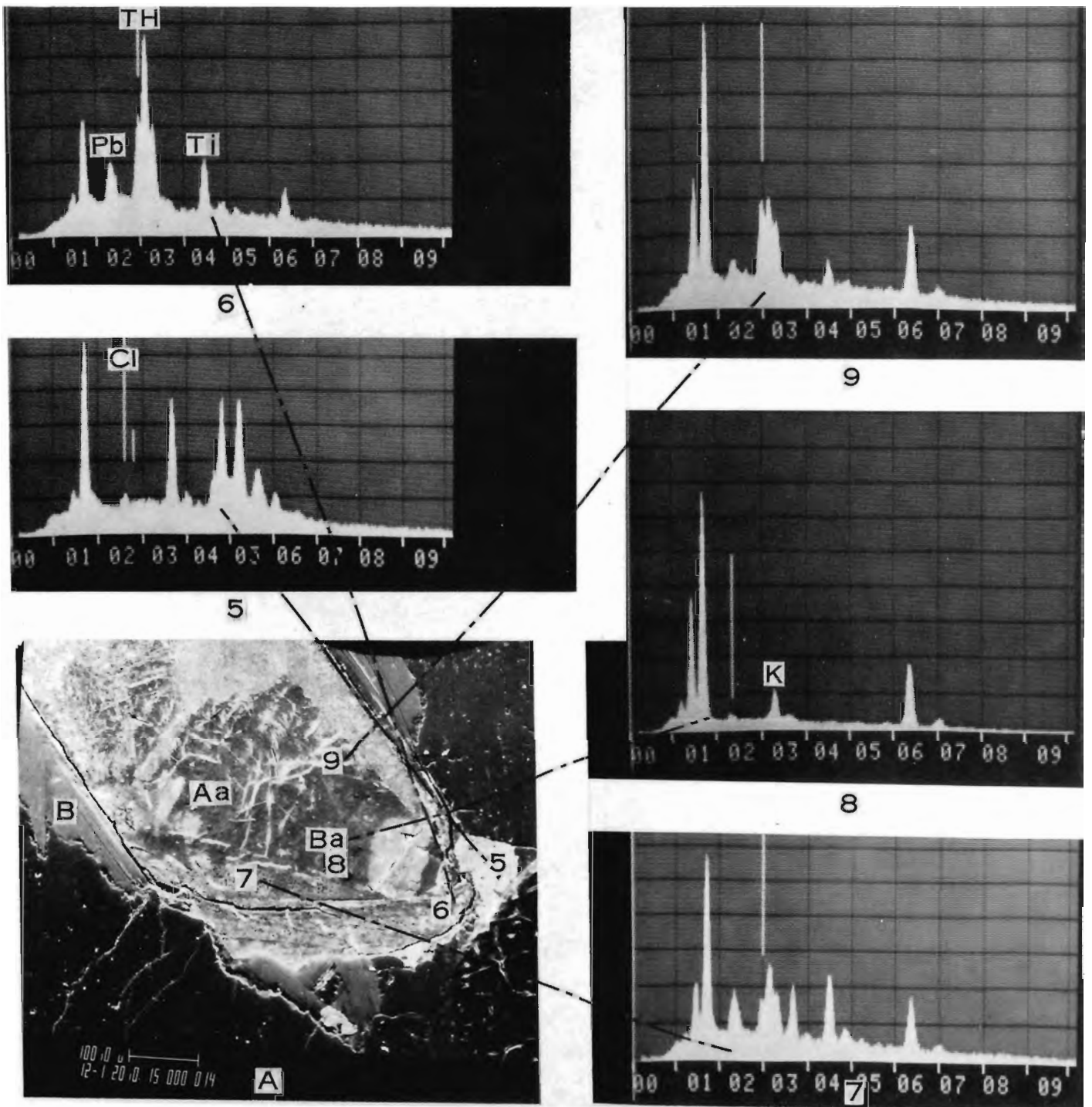


A. Backscattered electron image of allanite (A) with REE-rich rims (REE) and disseminated grains of galena (Ga, white). The allanite grades into amorphous radioactive phase (Ra, light grey) that contains variable concentrations of iron (Fe). Iron apparently substitutes for Th, U and Ca (compare spectra 3 and 4).

1. Energy dispersive spectrum (EDS) of allanite in Figure A, spot 1.
2. EDS of REE-bearing rims in Figure A, spot 2.
3. EDS of amorphous radioactive phase in Figure A, spot 3.
4. EDS of amorphous radioactive phase in Figure A, spot 4 (note Fe peak).

For additional comments and semiquantitative analyses see Table 1.

Figure 1. Heterogeneous allanite from high grade ore, RF-1, Bancroft area, Ontario. (GSC 203532-E)



A. Electron scanning image of heterogeneous, altered allanite (Aa) with adjacent patches of altered biotite (Ba) and REE- and U-, Th-Ti-, Pb-bearing phases in the rim in fractures and in disseminated specks.

5. EDS of REE-rich rim and grains adjacent to allanite in Figure A, near spot 5.
6. EDS of white specks containing Pb, Th, U and Ti in the rim of altered allanite in Figure A, near spot 6.
7. EDS of Pb-, Th-, U-, Ti-bearing phase in allanite, adjacent to altered biotite in Figure A, near spot 7.
8. EDS of altered biotite in allanite in Figure A, spot 8.
9. EDS of Th-, U-bearing material in fractures of altered allanite in Figure A, spot 9.

For additional comments and semiquantitative analyses see Table 1. The unmarked peaks in EDS are identified in Figure 1.

Figure 2. Altered allanite in biotite-rich band, Mont-Laurier area, Quebec. (GSC 203755-G)

Table 1
Partial semiquantitative microbeam analyses of heterogeneous allanite grains
from the Grenville Structural Province

Spots analyzed are close to those indicated on Figures 1 and 2**												
Oxide	1	2	3	5	6	7	7a	8	9	10	11	12
SiO ₂	37.3	31.5	30.9	-	12.1	11.1	32.2	37.	9.4	13.3	11.7	6.
TiO ₂	0.7	0.7	0.4	-	-	0.7	0.8	-	1.7	0.3	1.	-
Al ₂ O ₃	7.9	11.7	4.0	-	1.1	2.3	13.2	18.2	5.7	1.5	-	3.
Cr ₂ O ₃	-	-	-	-	-	-	-	0.3	-	-	-	-
FeO*	17.3	16.3	4.7	-	4.	38.	15.	14.5	26.5	4.	4.4	-
MnO	0.8	0.2	0.2	-	-	-	-	0.3	-	-	-	-
MgO	1.8	1.1	0.4	-	-	-	1.7	3.	-	0.7	-	-
CaO	5.4	12.	4.6	7.3	0.6	0.3	2.	0.6	1.1	0.3	2.2	3.1
SrO	-	-	-	2.4	-	-	-	-	-	-	-	-
K ₂ O	-	-	-	-	0.2	-	1.8	3.4	-	1.9	1.2	0.5
Y ₂ O ₃	3.5	0.1	9.4	-	3.6	1.3	0.8	-	0.8	1.3	0.8	-
La ₂ O ₃	1.9	6.6	0.7	16.6	0.5	0.5	1.6	-	1.7	4.6	5.4	14.
Ce ₂ O ₃	3.0	9.8	2.0	27.4	2.4	2.2	2.7	-	3.4	10.3	10.6	27.6
Nd ₂ O ₃	1.3	2.3	1.7	11.1	1.2	1.2	0.8	-	1.9	6.5	5.5	9.4
Pr ₂ O ₃	-	-	-	2.4	-	0.2	0.5	-	0.5	2.6	1.4	2.4
Sm ₂ O ₃	-	-	-	-	0.2	0.2	-	-	-	1.2	0.7	-
Gd ₂ O ₃	-	-	-	-	0.7	-	-	-	-	-	-	-
Er ₂ O ₃	-	-	-	-	0.5	-	-	-	-	-	-	-
Yb ₂ O ₃	1.2	0.2	2.9	-	-	-	-	-	-	-	-	-
PbO	0.1	0.2	0.4	-	1.5	6.	2.9	1.1	5.6	-	2.2	0.9
ThO ₂	1.5	1.1	10.2	-	33.7	4.6	1.7	0.2	9.3	-	5.2	1.8
UO ₂	1.2	0.1	5.6	-	13.6	10.2	3.8	0.9	11.3	-	4.5	0.5

Electron microprobe and semiquantitative energy dispersive spectrum analyses by G.A. Plant, J.G. Pringle and D.A. Walker, Geological Survey of Canada.

Atomic absorption analyses of allanite concentrates for REE, Y, Th and Pb in acid-soluble and acid-insoluble fractions (analyst J.G. Sen Gupta, Geological Survey of Canada) have been reported by Rimsaite (1980, Table 38.1).

* Total iron reported as FeO.

** Analyzed phases of heterogeneous allanite and energy dispersive spectra (EDS) recorded near analyzed spots:

1. Allanite in Figure 1, EDS-1.
2. Outer part of allanite 1, enriched in REE, Figure 1, EDS-2.
3. Metamict Y-bearing phase in allanite 1, Figure 1, EDS-3.
5. REE-rich grain and rim at the contact between altered allanite and host biotite, Figure 2, EDS-5.
6. U-rich specks in altered allanite which appear white on electron micrographs, Figure 2, EDS-6.
7. and 7a. Altered allanite (7a) containing Fe-rich, U-, Th-bearing specks (7), Figure 2, EDS-7.
8. Altered biotite, in allanite, Figure 2, EDS-8.
9. U-, Th-, Ti-, Fe-bearing crusts filling fractures in altered allanite, Figure 2, EDS-9.
10. and 11. Banded rim on allanite in Figure 2 that consists of two REE-bearing phases, one poor in Pb, U, Th (10), the other containing moderate quantities of Pb, U, Th (11, not shown here).
12. Fibrous REE-rich aggregates common in the rim of, and within altered radioactive ore and accessory minerals in the Grenville Structural Province (illustration in Rimsaite, 1981a, Fig. 17.5).

ALTERATION PRODUCTS OF ACCESSORY ALLANITE IN RADIOACTIVE GRANITES FROM THE CANADIAN SHIELD: REPLY

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The similarities between energy dispersive spectra published by Littlejohn (1981) and those obtained by Rimsaite in 1979 from the Grenville Structural Province are hardly remarkable since allanite is a common accessory mineral in granites (Deer, Howie and Zussman, 1962) and is particularly common in the Grenville Province (Traill, 1969). Bastnaesitization of allanite is a common feature in the Canadian Shield (Cerny and Cerna, 1972), in the United States (Adams and Young, 1961), and in the USSR (Mineyev et al., 1973). What is remarkable is that Rimsaite has failed to recognize this in the allanites collected by her and illustrated in her discussion (Fig. 1 and 2). Surely the REE-rich phases illustrated by her (Fig. 1, EDS-2 and Fig. 2, EDS-5) consist of bastnaesite and/or synchisite mixed with aluminosilicates, but there has been no attempt by Rimsaite to identify these.

As stated in the discussion, features supposedly presented in Littlejohn (1981), such as heterogeneity, alterations and isotopic ages of allanites, have already been discussed and illustrated by Rimsaite (1978, 1980, 1981a, 1981b). However, there is no mention of isotopic ages in Littlejohn (1981); there is no mention in Rimsaite (1978, 1980) of three main phases in the allanite from the Bancroft area nor of unidentified material filling fractures containing Fe, REE, Th and U; there is no illustration of altered allanite in Rimsaite (1978, 1980, 1981a, 1981b).

Rimsaite appears to be confused by the data that she has produced. She states: "The allanite grain in Figure 2 is surrounded by a prominent rim and grains composed mainly of REE (Table 1, analysis 6)". Examination of Table 1, analysis 6 and the corresponding spectra shows that there is a thorium and uranium enriched area on the rim, as does the analysis and spectra for 7 and 9. The REE enriched grains are actually analysis 5 (Table 1). The EDS spectrum (Fig. 2, EDS 5) shows a large Si peak, yet Si is not given in the analysis. EDS 6 and 7 show a significant Ti peak but Table 1, analysis 6 has no Ti reported, and analysis 7 has only 0.7% TiO₂ reported. Rimsaite has made no attempt to explain these discrepancies. She states in Table 1 of her discussion that the energy dispersive spectra were recorded near analyzed spots. If this is the case then there is no point in displaying spectra which bear no relationship to the analyses and thus lead to confusion for the reader. Rimsaite's spectra are probably those from mixtures as comparison with Figure 12.3 in Littlejohn (1981) suggests. If Rimsaite had increased the magnification of the SEM image from 120X to 400X as in Fig. 12.3C (Littlejohn, 1981) she might have observed that the Ca-REE phase (identified (Littlejohn, 1981) by XRD as synchisite) is replacing quartz which could account for the presence of Si in her spectrum obtained at a lower magnification. The thorium enriched areas (EDS-6, 7, 9 of Figure 2 of Rimsaite) could well be domains where thorite has formed in a matrix of aluminosilicate and titanium oxide. However, not having examined this section, and given the discrepancies between the displayed spectra, the analyses and Rimsaite's statements in her discussion, it is impossible to interpret these data meaningfully.

The only similarity in the EDS spectra published by Rimsaite (1981a, Fig. 17.5D and 17.1D; 1981b, Fig. 4.3-1) and by Littlejohn (1981, p. 98 and 102) is in the presence of the same elements. Even casual examination shows that the

relative peak heights are entirely different. This is not surprising since Rimsaite's spectra were obtained from the alteration of entirely different minerals and at different magnifications. Her Figure 17.5D was obtained from altered monazite and xenotime (she does not state which); her Figure 4.3-1 shows the same spectrum was obtained from three different environments (transecting a monazite grain, along the cleavage of biotite, coating zircon). I find it hard to believe that an identical spectrum can be obtained from these entirely different environments since some electron-beam overlap is to be expected with these extremely fine grained intergrowths.

The distinction between deuteric alteration and weathering is quite clear. According to Adams and Young (1961), Cerny and Cerna (1972) and Mineyev et al. (1973) the formation of bastnaesite from allanite is a deuteric process and not due to weathering. Watson (1917) and Hata (1939) studied the oxidation of allanite in the surficial environment. They both found that allanite weathers to a reddish crusty material consisting of a mixture of ferric oxides, alumina and silica; there is a depletion of REE. These are not the products of altered allanite in the study by Littlejohn (1981).

Rimsaite's assertion that most of the REE in allanite are actually contained in inclusions is clearly questionable. The total amount of REE in the residue of an allanite concentrate after acid leaching, according to Table 38.1 of Rimsaite (1980), is 2433 ppm or approximately 2.2% of the total REE in the original concentrate. This amount of REE hardly qualifies the residue to be called allanite (Deer, Howie, and Zussman, 1962). In addition Hutton (1951) studied the effect of acid on allanite and found that metamict allanite is easily dissolved; the solubility of nonmetamict allanite is dependent on composition. Rimsaite presents no evidence that the allanite in her concentrate did not dissolve; on the contrary her data suggest that the allanite was highly soluble.

What is the nature of the inclusions if they contain more than 105846 ppm REE (Table 38.1; Rimsaite, 1980)? They cannot be zircon, uraninite, uranothorite or galena as Rimsaite states. The hole in the allanite aggregate shown in Figure 12.2A of Littlejohn (1981) is the area which was powdered for XRD analysis; the pattern is that of bastnaesite. If bastnaesite did not form from the breakdown of allanite, and if the REE originated externally and formed inclusions, then the original mineral was not allanite and Rimsaite is grossly in error, as well as myself and other workers. However examination of the textures shown in Figure 12.2B and 12.3D (Littlejohn, 1981) shows that these are typical exsolution textures (Ramdohr, 1980). The reactions proposed to explain the alteration of allanite, while certainly theoretical, are not speculative but based on firm petrographic evidence and consideration of the allanite structure (Dollase, 1971).

Rimsaite states that she has already proposed an alternative interpretation and study of accessory minerals in radioactive granites in Rimsaite (1981b). This paper is certainly not an alternative explanation of allanite alteration since it deals exclusively with zircons; allanite is mentioned only once. She also states, in her discussion, that REE phases crystallize during the late deuteric stage (of granite evolution) although earlier crystallized minerals may contain some REE. I fail to see the real point of her discussion since this is exactly the hypothesis put forward by Littlejohn (1981) i.e. allanite is an early REE-bearing phase; bastnaesite and synchisite formed during the deuteric stage of granite evolution. The REE in these minerals were derived from the allanite itself and, along with uranium and thorium, have been locally redistributed by deuteric solutions.

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RUBIDIUM-STRONTIUM AND URANIUM-LEAD ISOTOPIC AGE STUDIES

ETUDES DES DATATIONS ISOTOPIQUES PAR LES METHODES RUBIDIUM-STRONTIUM ET URANIUM-PLOMB

Rb-Sr and U-Pb Isotopic Age Studies, Report 4

Compiled by **W.D. Loveridge**

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INTRODUCTION

W.D. Loveridge

The following papers present the results of 5 Rb-Sr isotopic age studies predominantly on whole-rock samples, and 6 U-Pb isotopic age studies on zircon and monazite concentrates. These age determination projects are joint research studies by field geologists of the Geological Survey of Canada and members of the staff of the Geochronology Section, typically initiated by the field geologists in an attempt to solve specific geological problems. In most cases the results are known to the geologist in question but not to the geological community at large. In some instances preliminary reference has been made in scientific reports to the resultant ages but the analytical data and other details have not been published.

Over the years the Geochronology Section has accumulated a number of such unpublished studies. This current group of papers is the second in a series to be presented in "Current Research, Part C". It is our intention to bring out all currently inactive, unpublished age determinations in this manner in a complete form including all relevant analytical data plus a description of the geological problem by the scientist who initiated the study. If, however, this is not possible the age and analytical data from some studies may be presented unsupported by geological discussion. Active studies, or portions of continuing studies, may also be included if the results are of satisfactory quality and of sufficient geological interest.

Rb-Sr Whole-Rock Isochron Studies

The analytical work leading to the Rb-Sr age results published in this collection of papers was performed from 1971 to 1981. Analytical procedures have progressed and been modified over the years, but are based on those described by Wanless and Loveridge (1972). The analytical uncertainty to be associated with the isotopic ratios has also changed with time, and is listed individually in each data table.

All whole-rock isochron data were subjected to regression analysis using the program published by Brooks et al. (1972). The published ages and initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and their associated errors (95% confidence interval) are those determined by the York 1 portion of the program. The value obtained for the Mean Square of Weighted Deviates (MSWD) from the McIntyre portion of the program is used to classify the linear array of experimental points as an isochron or an errorchron. In the terminology of Brooks et al. (1972) an errorchron is the result of regression of data that possess scatter in excess of experimental error whereas the data points forming an isochron are collinear within experimental error. Thus the excess scatter of points forming an errorchron is attributable not to lack of laboratory precision, but to inherent properties of the rocks under investigation.

Brooks et al. (1972, p. 557) pointed out that the uncertainties associated with errorchron results at a given confidence interval contain none of the predictive aspects that characterize true confidence intervals. For an isochron with error figures given at the 95 per cent confidence level, 19 of 20 repeats should fall within those error figures. This is not the case with an errorchron; it is possible that no repeats would fall within the stated error limits. Therefore the error limits associated with errorchrons may be indicative of the uncertainty in the age and initial ratio results, but are by no means definitive. For this reason the error limits associated with errorchron results are mentioned only in the text but omitted from the diagrams to prevent inadvertent misuse by the casual reader.

The ^{87}Rb decay constant used in this series of papers is $1.42 \times 10^{-11} \text{a}^{-1}$ as recommended by the IUGS Subcommittee on Geochronology (Steiger and Jaeger, 1977). Previously published Rb-Sr ages referred to in these papers have been adjusted to this constant.

U-Pb Age Studies on Zircon Concentrates

The preparation of zircon concentrates and analytical procedures for U-Pb age determination work was discussed by Sullivan and Loveridge (1980). The U-Pb results presented in this group of papers were determined during the period 1978 to 1981 and consist of two to four analytical points defining a chord, the upper intercept of which, with the concordia curve, determines the geological age. The analytical points defining the two point chords are sufficiently close to the concordia curve in these studies to define the upper intercept age with the desired precision.

The concordia intercept ages are calculated by a computer program which first determines the parameters of the chord through a least squares linear regression and then finds the intercepts by an iterative method. The uncertainties associated with concordia intercept ages of two point chords are determined from the error envelope defined by the analytical uncertainties associated with the individual analytical points. However, the uncertainties associated with concordia intercept ages of chords derived from three or more analytical points are based on the standard error of the slope of these chords. These uncertainties are determined from the intersection with concordia of hypothetical chords with slopes equal to that of the original chord plus and minus its standard error, projected from a point which is the average of the analytical points.

The uranium decay constants used are: ^{238}U , $1.55125 \times 10^{-10} \text{a}^{-1}$ and ^{235}U , $9.8485 \times 10^{-10} \text{a}^{-1}$ as recommended by the IUGS Subcommittee on Geochronology (Steiger and Jaeger, 1977). Descriptions of zircon concentrates are by R.D. Stevens.

I would particularly like to thank O. van Breemen and R.D. Stevens for critically reading all papers in this section and J. MacManus for drafting many of the diagrams.

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1. Rb-Sr STUDY OF THE MICHAEL GABBRO, LABRADOR

W.F. Fahrig and W.D. Loveridge

Fahrig, W.F. and Loveridge, W.D., Rb-Sr study of the Michael Gabbro, Labrador; in *Rb-Sr and U-Pb Isotopic Age Studies, Report 4; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C*, p. 99-103, 1981.

Abstract

The Michael Gabbro extends westward 200 km from the south-central coast of Labrador and straddles the boundary between the Grenville Province and the Makkovik Subprovince. Whole-rock K-Ar ages on chilled margins of two intrusions from the Smokey Islands average 1335 Ma in conflict with the 700-800 Ma age indicated by the paleomagnetic pole position of 163°E, 10°N, $\alpha_{95} = 2^\circ$. A Rb-Sr whole rock study on material from diabase of the Smokey Islands, thought to be part of the Michael Gabbro, yields an errorchron age of 1461 ± 96 Ma, initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7033 ± 0.0021 , and MSWD 5.8. We interpret this age as approximating the time of emplacement of the Michael Gabbro, confirming the older age indicated by the K-Ar results.

Geological Setting

The Michael Gabbro (Fahrig and Larochelle, 1972) forms a belt of undulating, predominantly south-dipping sheets extending westward 200 km from the south-central coast of Labrador (Fig. 1). The belt is at least 40 km wide and most of it lies within Makkovik Subprovince (Taylor, 1971). The intrusions were described in early reports (for example Kranck, 1953) from coastal areas and more recently their inland extent has been mapped by Stevenson (1970). In the interior, much of the gabbro forms linear outcrop ridges surrounded by drift. These ridges are shown connected in Figure 1, in order to give some idea of the probable extent and regional pattern of the intrusions, but in so doing the extent of the intrusions may be somewhat exaggerated.

The material sampled in this Rb-Sr study is from diabase dykes in the Smokey Island archipelago. This diabase is thought to be part of the Michael Gabbro which, to the west, is predominantly subophitic.

Intrusions along the southern edge of the belt have been metamorphosed (cf. Stevenson, 1970, p. 17) presumably during the Grenvillian Orogeny. It is not certain whether the strikes and dips of the sheets in the main part of the belt, i.e. north of this zone of metamorphism, are entirely primary or were modified during later deformation. It is clear, however, that the present orientation of the sheets pre-dates their magnetization (Fahrig and Larochelle, 1972).

The Michael Gabbro has some distinctive chemical characteristics. A number of the intrusions west of the Smokey Islands exhibit anomalously high alumina for basaltic rock and on the whole the Michael Gabbros are low in silica. Abundant olivine, up to 20 per cent, has been noted by a number of authors (Kranck, 1953; Stevenson, 1970).

One of the chief reasons for carrying out radioisotopic age determinations and paleomagnetic work on the Michael Gabbro is because of its position straddling the boundary between the Grenville Province and the Makkovik Subprovince. Debate about the exact position of structural province boundaries is usually not particularly rewarding as workers do not accept strict geological or geophysical

criteria for the definition of such boundaries. In the present instance, however, there seemed to be a unique opportunity to use the Michael Gabbro to define the northern limit of the Grenvillian Orogeny in this part of the Shield.

K-Ar Ages

Contacts of Michael intrusions are rarely exposed in the interior but are commonly observed along the coast, for example, on the Smokey Islands (Fig. 2). Because chilled basic rock (as opposed to coarsely crystalline material) yields fairly reliable whole-rock K-Ar ages, analyses were carried out on samples from the chilled boundaries of two dykes from the Smokey Islands. These yielded ages of 1323 ± 52^1 (Wanless et al., 1973, p. 103) and 1348 ± 120^1 Ma (Wanless et al., 1972, p. 89) which are in agreement with the Rb-Sr result within analytical uncertainty.

Table 1

Chemical composition of diabase and of a felsic veinlet in a diabase dyke, Michael Gabbro, Smokey Islands

	A	B
SiO ₂	47.7	72.6
TiO ₂	2.36	.25
Al ₂ O ₃	15.4	14.3
Fe ₂ O ₃	2.6	.3
FeO	11.5	1.2
MnO	.19	.05
MgO	6.16	.29
CaO	8.0	1.44
Na ₂ O	3.0	5.5
K ₂ O	1.21	2.02
H ₂ O _T	.90	.8
CO ₂	.20	.3
P ₂ O ₅	.42	.07
A = average composition of samples of Michael diabase from dykes in the Smokey Islands.		
B = composition of felsic veinlet in a diabase dyke in the Smokey Islands.		

¹ These K-Ar ages have been recalculated using the new constants recommended by Steiger and Jaeger (1977).

* Note added in proof. A paper by C. Brooks, R.J. Wardle and T. Rivers, "Geology and geochronology of Helikian magmatism, western Labrador"; Canadian Journal of Earth Sciences, v. 18, no. 7, 1981, p. 1211-1227, was received after the present study had reached the proof stage. These authors have presented Rb/Sr and Sm/Nd data suggesting that the age of intrusion of the Shabogamo Gabbro is about 1400 Ma. Considering the uncertainties related to both the Michael and Shabogamo results, there is still a strong possibility that these two suites are time equivalents. The reference in the above paper (p. 1223) to "unpublished data quoted in Fahrig and Larochelle (1972); Wanless and Loveridge (1978)" is to the results published herewith.

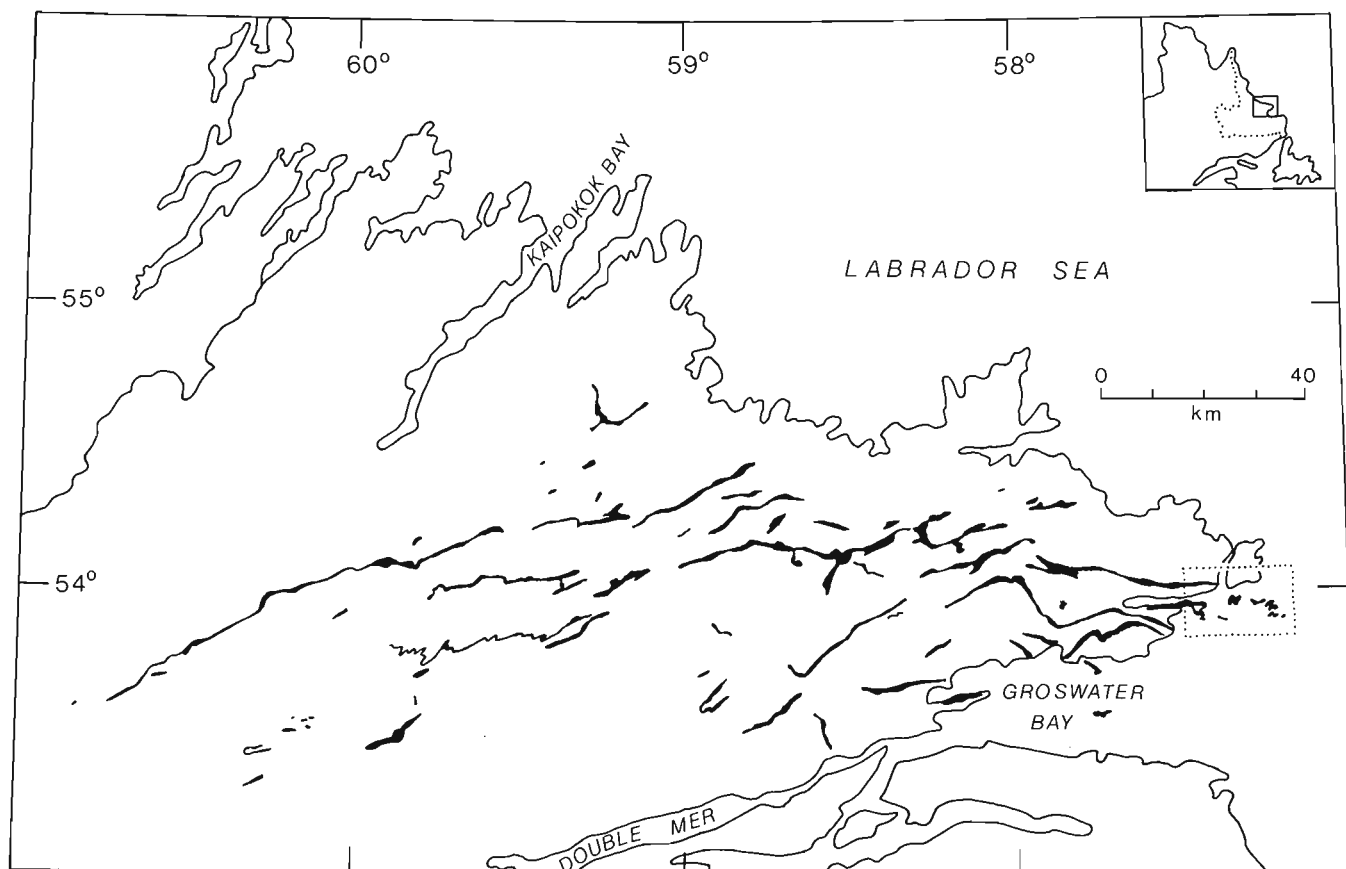


Figure 1. General distribution of the Michael Gabbro, largely after Stevenson, 1970. Rectangle outlined in dots indicates Smokey Island archipelago, which is shown in greater detail in Figure 2.

Table 2
Sample numbers and localities, Michael Gabbro, Labrador

Sample No.		Rock Type	Locality		NTS
This report	Field		Latitude	Longitude	
1	FA-68 0067	coarse grained diabase from centre of 20 m thick dyke	54°26'	57°27'	13 I/II
2	FA-68 0017	chilled margin of a 30 m thick olivine diabase dyke	54°27.5'	57°14.5'	13 I/II
3	FA-69 0606	felsic veinlet in a thick (>50 m) coarse grained olivine diabase dyke	54°27'	57°14'	13 I/II
4	FA-69 0596	felsic veinlet in a (>40 m) coarse grained olivine diabase dyke	54°27'	57°14.5'	13 I/II
5	FA-69 0636	felsic veinlet in olivine diabase dyke	54°28.5'	57°15'W	13 I/II

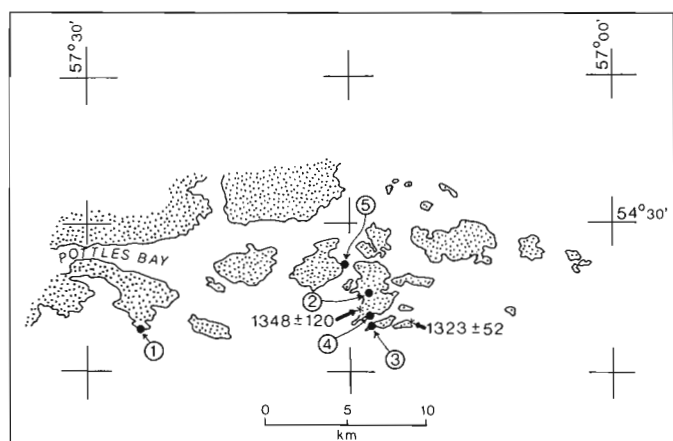


Figure 2. Smokey Island archipelago. Location of five samples used in the errorchron are indicated with black dots. Whole-rock K-Ar ages are shown and the locations from which the K-Ar samples were obtained are indicated by asterisks.

Rb-Sr Age

The ages of basic intrusions are notoriously difficult to determine by standard whole-rock Rb-Sr analysis because of the deficiency of Rb in these rocks. More refined methods combining whole-rock and mineral analysis have proved much more effective, for example, Patchett et al. (1978). A technique which is applicable to some basic intrusions involves sampling late magmatic material which represents a concentration of lower temperature constituents. In diabase dykes these constituents sometimes form pink or white veins, transverse to the dyke walls, and restricted to the dyke. They always taper off as the dyke walls are approached. These veins apparently form because the early, high temperature, silicate framework is rigid enough to fracture and form tensional joints as a result of cooling and contraction. Volatile low temperature constituents are sucked into these tension joints. Vein-joint material was collected from Michael diabase dykes of the Smokey Islands archipelago in the hope that sufficient Rb would be concentrated to provide a suitable range for a whole-rock isochron. Samples 3, 4 and 5 (Fig. 3) are from veinlets occurring in three separate Michael dykes.

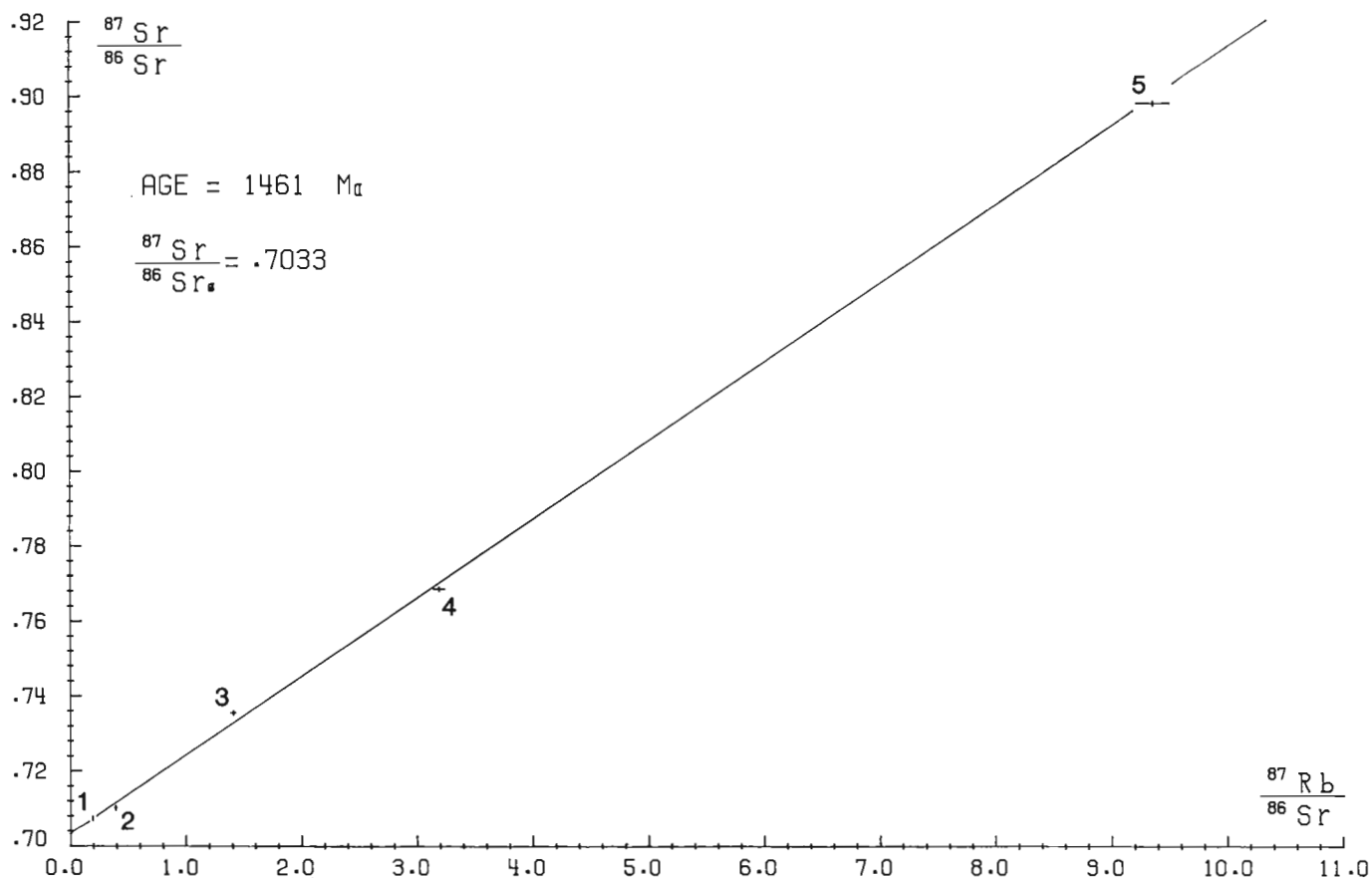


Figure 3. Rb-Sr isochron diagram of the Michael Gabbro, Labrador.

Table 3
Analytical data, whole rock samples, Michael Gabbro, Labrador

Sample No.	Rb ppm	Sr ppm	$^{87}\text{Sr}/^{86}\text{Sr}$ unspiked	$^{87}\text{Sr}/^{86}\text{Sr}$ spiked	$^{87}\text{Sr}/^{86}\text{Sr}$ average	$^{87}\text{Rb}/^{86}\text{Sr}$
1	24.87	372.4	0.7071	0.7078	0.7075 ± 0.0011	0.193 ± 0.004
2	38.57	287.6	0.7102	0.7103	0.7103 ± 0.0011	0.388 ± 0.008
3	116.5	239.9	0.7355	0.7354	0.7355 ± 0.0011	1.406 ± 0.042
4	170.4	154.6	0.7688	0.7680	0.7684 ± 0.0012	3.191 ± 0.096
5	104.0	32.18	0.8979	0.8977	0.8978 ± 0.0013	9.357 ± 0.028

Table 4
K-Ar age determination

Eskimo Paps; whole-rock K-Ar age 1222 ± 101 Ma

$K = 0.218\%$, $^{40}\text{Ar}/^{40}\text{K} = 0.10149$, radiogenic $^{40}\text{Ar} = 94.8\%$

Whole-rock from a chilled margin of a 20 metre thick fresh diabase dyke that cuts Mealy Mountain anorthosite at Eskimo Paps, a small point on the south shore of Hamilton Inlet, Labrador. The diabase is dark grey, aphanitic, with plagioclase and pyroxene microphenocrysts in an opaque matrix presumably of recrystallized glass. Field number FA-744804; lat. $53^\circ 39.5'$, long. $59^\circ 23'$.

Table 1 gives the chemical composition of one such vein and the average composition of samples from five Smokey Islands dykes. These analyses indicate that the felsic veinlets are notably higher in silica and alkalis, and lower in iron, magnesium and calcium than are the enclosing diabase.

Analytical Techniques, Data and Results

Analytical procedures for Rb-Sr isotopic analyses of whole-rock samples were based on those described by Wanless and Loveridge (1972). The isotopic compositions of five whole-rock samples (see Table 2) are listed in Table 3 and depicted on an isochron plot, Figure 3. The five analytical points are reasonably collinear but yield an errorchron age of 1461 ± 96 Ma, initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7033 ± 0.0021 and MSWD 5.8. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is in the general range of those found for mantle derived rocks of this age but the relatively high associated analytical uncertainty does not preclude a considerable period of crustal residence.

Discussion

The Michael Gabbro proved very amenable to paleomagnetic work and 21 sites extending from the Smokey Islands west for 200 km yielded a pole at 163°E , 10°N , $\alpha_{95} = 2^\circ$ (Fahrig and Larochelle, 1972). The material from many of the sites proved unusually stable for gabbroic rock, a feature which may relate to its anorthosite affinities (Stevenson, 1970). A few sites along the southern edge of the belt were unstable and this was attributed to possible effects of the Grenvillian Orogeny. The Michael pole did pose a problem because its position indicated an age in the range 700-800 Ma (Berger and York, 1980). This was in conflict with the whole-rock K-Ar ages from the Smokey Islands, as experience suggests that K-Ar whole-rock ages on basic rocks are seldom greater than the age of magnetization. Further to this, Gittins (1972), reported a K-Ar whole-rock age of about 1100 Ma on dykes from Eskimo Paps, Lake Melville, a location about 30 km south of the Michael belt.

This age was supported at the Geological Survey of Canada dating laboratory where an age of 1222 ± 101 Ma, (Table 4) was obtained on chilled material from an Eskimo Paps dyke. For this reason, Fahrig and Larochelle (1972) suggested that the Michael Gabbro had not been remagnetized but that the Makkovik Subprovince had been rotated in a clockwise direction sometime during the period 1500-1100 Ma ago.

The Rb-Sr age of 1461 Ma given in the present report probably provides a good estimate of the age of the Michael Gabbro. As previously noted, it is in agreement within analytical uncertainties with the two whole-rock K-Ar ages on chilled dyke material from the Smokey Islands. The high MSWD may reflect a mild disturbance of the system during the Grenvillian Orogeny; the relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of sample 3 may be due to contamination with radiogenic strontium from the country rock.

The Michael Gabbro bears many resemblances to the Shabogamo Gabbro (Fahrig, 1967; Fahrig et al., 1974). Both occur along the Grenville Front, parts of both belts were metamorphosed during the Grenvillian, both probably have ages between 1400 and 1500 Ma, and both have anorthositic affinities.

The main bodies of anorthosite of the Grenville Province were also intruded during this period so there is a distinct possibility of a genetic relationship between the Michael and Shabogamo gabbros and the Grenville anorthosites.

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2. A Rb-Sr STUDY OF A NEW QUEBEC ARCHEAN GRANODIORITE

F.C. Taylor and W.D. Loveridge

Taylor, F.C. and Loveridge, W.D., A Rb-Sr study of a New Quebec Archean granodiorite; in *Rb-Sr and U-Pb Isotopic Age Studies, Report 4, in Current Research, Part C, Geological Survey of Canada, Paper 81-1C*, p. 105-106, 1981.

Abstract

A Rb-Sr study of granodiorite from a pluton lying south of the Cape Smith Fold Belt has yielded an errorchron of 2685 ± 183 Ma confirming an Archean age. This pluton intrudes north-trending metasedimentary and metavolcanic rocks suspected to be much older than the east-trending more southerly Archean rocks; however, this difference in age has not been demonstrated by the present study.

Introduction

A massive to weakly foliated pluton composed of granodiorite, granite, and tonalite lying south of, and unconformably lower than, the Cape Smith Fold Belt, was sampled for a Rb-Sr age determination to confirm an Archean age and possibly obtaining an early Archean age. A geological sketch map with sample locations is presented in Figure 1. The pluton intrudes north-trending metasedimentary and metavolcanic rocks; the north-trending rocks and associated plutonic rocks in this part of the Superior Province may be older than the more common east-trending rocks in the southern parts of the province.

The sampled rocks are an equigranular, medium grained, predominantly pinkish grey granodiorite (with pale orange to dark greenish grey phases) containing about 5 per cent mafic minerals. Whereas biotite is present in all samples, hornblende occurs in only four of the seven. Spene is well developed locally and muscovite occurs rarely.

Previous isotopic age determinations from this and similar plutons consist of a K-Ar biotite age of 1750 Ma (Kretz, 1961), from 55 km to the west of the present samples and two Rb-Sr whole-rock ages on single samples of 2400 ± 72 and 2855 ± 135 Ma¹ (Beall et al., 1963) from approximately 15 km to the east. The K-Ar age shows the influence of the Hudsonian orogeny on the Archean rocks beneath the unconformity. The Rb-Sr whole-rock ages were able to penetrate through the Hudsonian metamorphic effect and show an Archean age for one of the two samples analyzed.

Analytical Procedures and Results

Analytical procedures for Rb-Sr isotopic analyses of whole-rock samples were based on those described by Wanless and Loveridge (1972) with the exception that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were obtained only from spiked Sr analyses in this study; no unspiked Sr measurements were performed.

Seven samples were analyzed isotopically and the results are listed in Table 2 and plotted in Figure 2. Six of the seven sample points form a linear trend with sample point number 5 falling somewhat below this trend. Regression analysis of the six points yields an errorchron of age 2685 ± 183 Ma, initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7020 ± 0.0029 and MSWD 12.9. Although the age given by an errorchron cannot be regarded as definitive, an Archean age is generally indicated by these results. However, the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, due to its relatively large uncertainty, may not be used to differentiate between a mantle genesis or a crustal prehistory for this pluton.

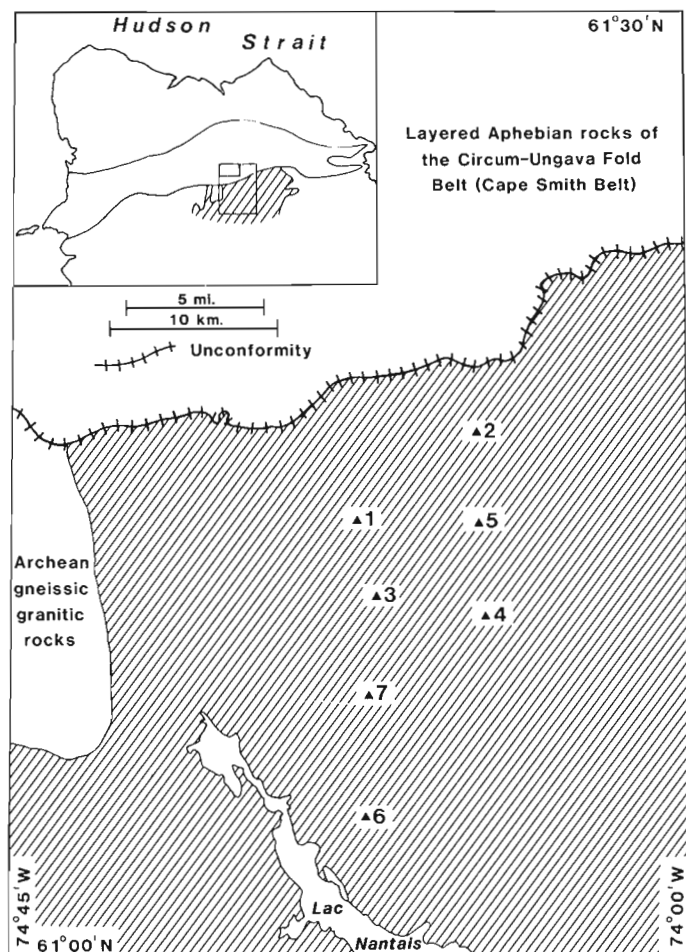


Figure 1. Geological setting and sample locations, New Quebec Archean granodiorite.

Table 1

Sample numbers and localities,
New Quebec Archean granodiorite

Sample No.		Locality		N.T.S.
This work	Field	Latitude	Longitude	
1	GV 73-325	61° 16' 41"	74° 13' 54"	35 G
2	GV 73-314	61° 13' 40"	74° 13' 58"	35 G
3	GV 73-324	61° 10' 42"	74° 13' 36"	35 G
4	GV 73-316	61° 04' 20"	74° 21' 32"	35 G
5	GV 73-315	61° 08' 06"	74° 21' 14"	35 G
6	GV 73-322	61° 11' 23"	74° 20' 30"	35 G
7	GV 73-323	61° 13' 46"	74° 21' 47"	35 G

¹ Rb-Sr ages adjusted to a ^{87}Rb decay constant of $1.42 \times 10^{-11} \text{ a}^{-1}$.

Table 2
Analytical data, whole rock samples, New Quebec Archean granodiorite

Sample No.	Rb ppm	Sr ppm	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
1	97.43	466.6	0.7247 ± 0.0004	0.604 ± 0.012
2	161.8	542.4	0.7370 ± 0.0004	0.862 ± 0.017
3	144.6	387.7	0.7423 ± 0.0004	1.078 ± 0.022
4	178.5	472.8	0.7460 ± 0.0004	1.091 ± 0.022
5	235.2	391.2	0.7589 ± 0.0005	1.738 ± 0.035
6	235.7	209.9	0.8299 ± 0.0005	3.246 ± 0.065
7	228.4	202.1	0.8251 ± 0.0005	3.267 ± 0.065

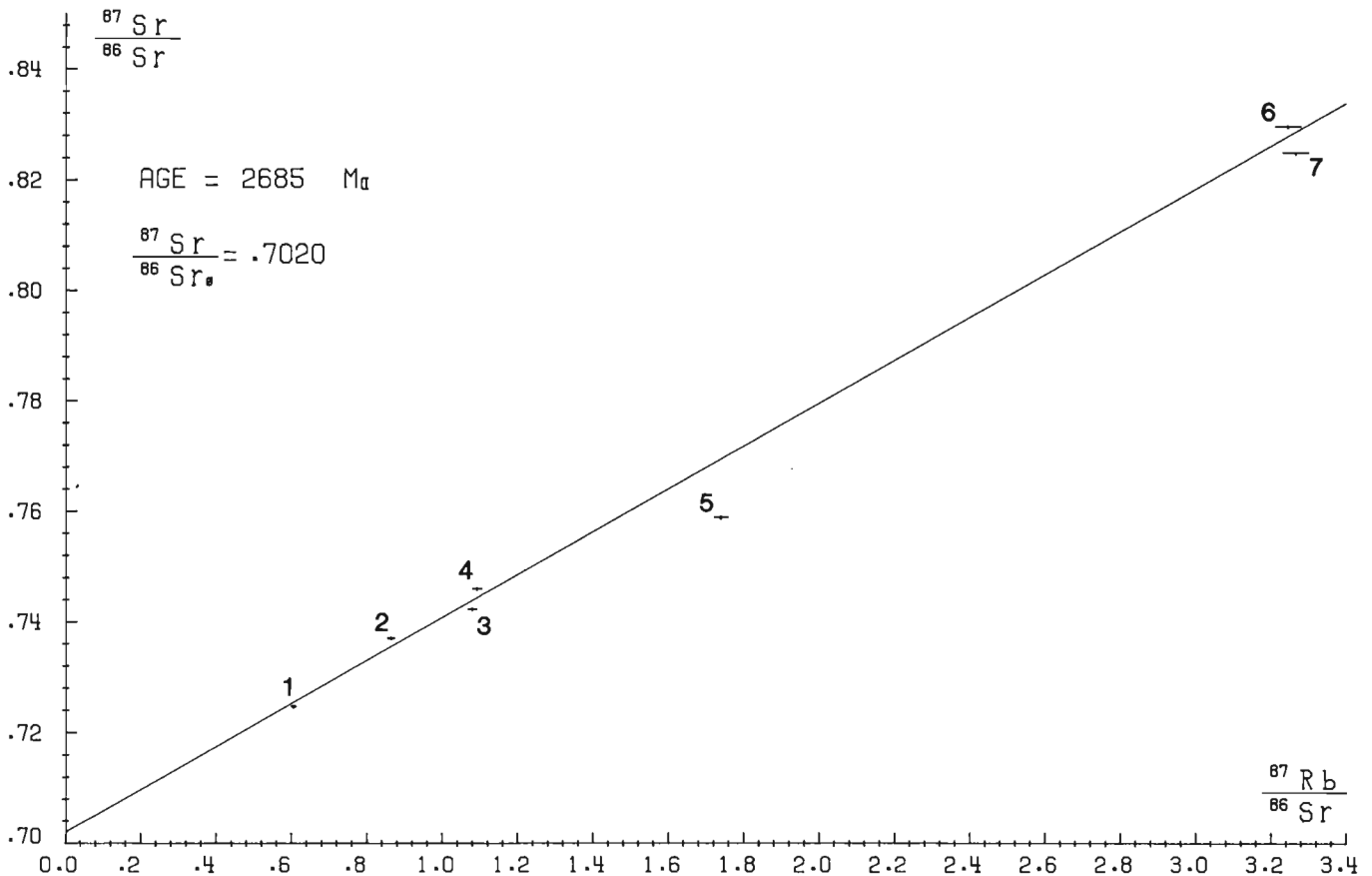


Figure 2. Rb-Sr isochron diagram, New Quebec Archean granodiorite.

Interpretation

The present errorchron age, 2685 ± 183 Ma, confirms the Archean age of this pluton and shows the intrusion to have occurred during the Kenoran orogeny. Whether or not the northerly trend of the layered rocks in the area south of the Cape Smith Belt is a different age than the easterly trend in the southern part of the Superior Province has not been resolved, as the Archean age indicated by this study dates only the intrusive rocks; development of the northerly trend may predate the intrusion.

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3. U-Pb ZIRCON AGES OF TONALITIC METACONGLOMERATE COBBLES AND QUARTZ MONZONITE FROM THE KAPUSKASING STRUCTURAL ZONE IN THE CHAPLEAU AREA, ONTARIO

J.A. Percival, W.D. Loveridge, and R.W. Sullivan

Percival, J.A., Loveridge, W.D., and Sullivan, R.W., *U-Pb zircon ages of tonalitic metaconglomerate cobbles and quartz monzonite from the Kapuskasing Structural Zone in the Chapleau area, Ontario; in Rb-Sr and U-Pb Isotopic Age Studies, Report 4, in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 107-113, 1981.*

Abstract

U-Pb analyses of zircon provide an upper concordia intercept age of 2664 ± 6 Ma for tonalitic cobbles in a polymictic, stretched-pebble metaconglomerate from the boundary area between the Kapuskasing Structural Zone and the Wawa Subprovince. Massive to faintly lineated quartz monzonite that intrudes the metaconglomerate locally, has an upper concordia intercept of 2633 ± 11 Ma. If these dates are the primary crystallization ages, they would bracket deformation and metamorphism of the metaconglomerate.

Introduction

As shown in Figure 1 the easterly-trending Abitibi and Wawa subprovinces (Stockwell, 1970) of the Superior Province are separated by a north-northeast-striking zone up to 45 km wide of high grade metamorphic rocks, the Kapuskasing Structural Zone (KSZ; Thurston et al., 1977). The massive-to weakly-foliated Ivanhoe Lake Pluton (Percival, 1981a) of the Abitibi Subprovince has a northeast-striking gneissosity within 1 km of the KSZ (Percival and Coe, 1981). This local overprinting relationship and regionally discordant nature suggest that the Kapuskasing zone is a late mobile belt. Interpretation of the age of ductile deformation in the KSZ ranges from late Archean or early Proterozoic (Watson, 1980) to Hudsonian (Brooks, 1980) or Grenvillian (McCall, 1977). The latter two suggestions are not consistent with a K-Ar date of 2505 Ma^1 (Watkinson et al., 1972) on hornblende from the KSZ which suggests that the latest metamorphism was Archean. The purpose of this project is to constrain, by zircon geochronology, the timing of plutonic and deformational events in the KSZ and adjacent subprovinces in the Chapleau-Foley area, as part of a program of regional geological synthesis (Card et al., 1980; 1981). Preliminary maps of the area are included in Card et al. (1981) and Percival (1981b).

Geological Setting

The Abitibi Subprovince in the study area is made up of a supracrustal succession intruded by massive and foliated plutons. The supracrustal pile is dominated by mafic metavolcanic rocks, with subordinate felsic metavolcanic and local metasedimentary units. Foliation and lineation are generally steep. The plutons are massive to weakly foliated granodiorite and quartz monzonite with minor foliated to gneissic tonalite. In the Timmins-Kirkland Lake area, 100 km east of the present study area, volcanic rocks yield U-Pb zircon ages between 2725 ± 2 and 2702 ± 2 Ma (Nunes and Pyke, 1980). A U-Pb zircon age of 2680 ± 3 Ma has been obtained on the Ivanhoe Lake Pluton (Percival, 1981a).

The Wawa Subprovince in the study area is composed mainly of gneissic and plutonic rocks. Tonalite-granodiorite gneiss contains up to 20 per cent mafic xenoliths and is exposed in domal structures, 10-20 km in diameter. Gneissic rocks are metamorphosed to amphibolite facies. Stretched-pebble metaconglomerate occurs in association with paragneiss and amphibolitic rocks in the Borden Lake area.

The Kapuskasing Structural Zone, separating the Wawa and Abitibi subprovinces, is characterized by alternating ENE-striking units of paragneiss, mafic gneiss, anorthositic, dioritic and tonalitic rocks. All units are deformed and metamorphosed to upper amphibolite or hornblende granulite

facies. Tonalitic leucosome segregations are common in all rock types except anorthosite. Gneissosity and fold axial surfaces strike ENE and dip moderately northwest. Rodding, mineral and fold axis lineations plunge gently ENE or WSW.

The boundary between the Abitibi Subprovince and the KSZ is a sharp break coinciding with the Ivanhoe Lake cataclastic zone, comprising cataclasite- and mylonite-series fault rocks. The Ivanhoe Lake Pluton has a gneissic zone, 1 km wide, on the Abitibi (east) side of the Ivanhoe Lake cataclastic zone.

The Wawa Subprovince-KSZ boundary is irregular and gradational, and is marked at different locations by one or more of (1) lithological differences, (2) an amphibolite-granulite facies isograd, and/or (3) a change in structural style and orientation from domal in the Wawa Subprovince to ENE belts in the KSZ (Percival, 1981a).

Geology of the Borden Lake Area

The southern part of the Borden Lake area (Fig. 2) is underlain by xenolithic tonalitic gneiss with recumbent early folds, refolded about upright easterly axial surfaces. To the north is a body of foliated to gneissic granodiorite which occurs locally as layers in the tonalitic gneiss and intrudes the southern margin of a 1-2 km wide belt of amphibolite and hornblende. Layers and bodies of massive- to faintly-lineated, coarse grained quartz monzonite intrude amphibolite and locally cut metaconglomerate to the north. Metaconglomerate and paragneiss are conformable units but contacts with amphibolite or gneissic rocks are not exposed. Clasts in the metaconglomerate are, in order of decreasing abundance, felsic metavolcanics, metasediments, tonalite-granodiorite, plagioclase-porphyritic meta-andesite, and amphibolite, with rare hornblende and vein quartz. The matrix, comprising up to 10 per cent of the rock, is garnet-hornblende-biotite-quartz schist. Clasts have length to width ratios up to 20:1 although the tonalite cobbles are generally equant. The rock has a weak, gently north-dipping foliation. The rodding lineation trends $80-100^\circ$ and plunges $0-10^\circ$ easterly. Associated metasedimentary rocks contain coarse muscovite, and mafic rocks have hornblende-plagioclase assemblages, locally with clinopyroxene. The assemblages indicate metamorphism to the amphibolite facies. The metaconglomerate has similar structural characteristics to the granulite facies rocks of the Kapuskasing zone.

Analytical Procedures and Results

Techniques for the concentration of zircon and the extraction and analysis of Pb and U are described in Sullivan and Loveridge (1980). U and Pb isotopic analyses were

¹Recalculated using revised decay constants (Steiger and Jaeger, 1977)

performed on three fractions from each of the two zircon concentrates. The analytical results are listed in Table 1 and displayed on a concordia diagram, Figure 3. A description of sample lithology is presented in Appendix 1 and of the zircon morphology in Appendix 2. Sample localities are shown in Figure 2 and listed in Appendix 1.

Sample ZR-4 is a composite sample of tonalitic cobbles from metaconglomerate north of Borden Lake. The three zircon fractions analyzed yielded data points, collinear within analytical uncertainty, defining a chord which cuts the concordia curve at 2664 ± 6 and 109 ± 94 Ma. An unusual feature of these results is that the $-149 \pm 105 \mu\text{m}$, more

magnetic fraction yields the most concordant data point and the $-149 \pm 105 \mu\text{m}$, relatively less magnetic fraction is most discordant (c.f. Silver, 1963). As the more magnetic fraction has the highest U content, the zircon fraction with the highest U content is most concordant, and that with the lowest U content is most discordant.

Hornblende and biotite concentrates were prepared from metaconglomerate matrix and dated by K-Ar analysis. Ages of 2594 ± 151 Ma and 2263 ± 29 Ma were obtained from the hornblende and biotite respectively (publication numbers GSC 80-167 and GSC 80-166, Stevens et al., in press).

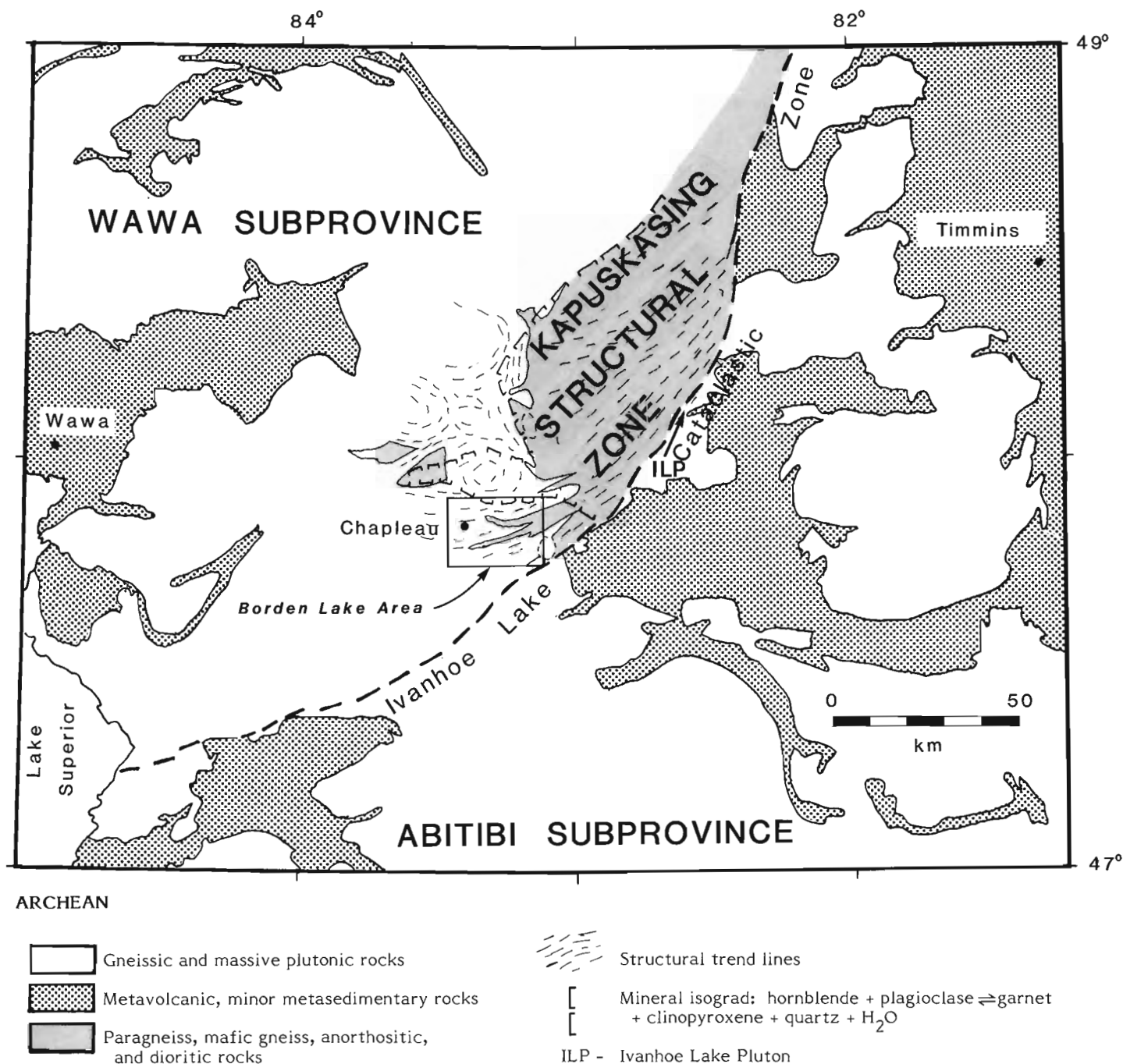


Figure 1. Generalized geological map of the Timmins-Wawa area of Ontario. Modified from maps of Percival (1981b) and K.D. Card (unpublished).

Sample ZR-3 is from massive to faintly lineated, coarse grained quartz monzonite south of location ZR-4 in the Borden Lake area. The data points from the three fractions analyzed are collinear within analytical uncertainty and define a chord which cuts the concordia curve at 2633 ± 11 and 362 ± 91 Ma.

The data points from ZR-3 are more discordant than those from ZR-4 implying a greater relative lead loss. This is probably due to additional radiation damage to the zircon crystal structure associated with the higher U content in ZR-3; 1143 to 1503 ppm as compared with 381 to 507 ppm in ZR-4.

A K-Ar age of 2198 ± 28 Ma (Table 2) was obtained on biotite from the quartz monzonite. This age, together with the K-Ar ages on biotite (2263 ± 29 Ma) and hornblende (2594 ± 151 Ma) from ZR-4, suggest partial or complete resetting of the K-Ar systems by a post-Archean event.

Discussion

The 2664 Ma zircon age on tonalitic cobbles in the metaconglomerate may represent either the primary age of crystallization of the source tonalite or a reset age, possibly of metamorphic origin. The main factor supporting the primary age hypothesis is the collinearity of the data points, which suggests a simple single stage history. It has been established empirically in other areas that amphibolite facies metamorphism does not bring about total lead loss from zircon (e.g. Davis et al., 1968; Krogh and Davis, 1973). Presence of primary Pb in reset zircons would probably have caused the data points to deviate from the observed collinear pattern.

Regional deformation in the southern Superior Province post-dates volcanism and pre-dates the intrusion of massive plutons. In the Abitibi and Wawa subprovinces, the youngest U-Pb zircon ages on metavolcanic rocks are respectively

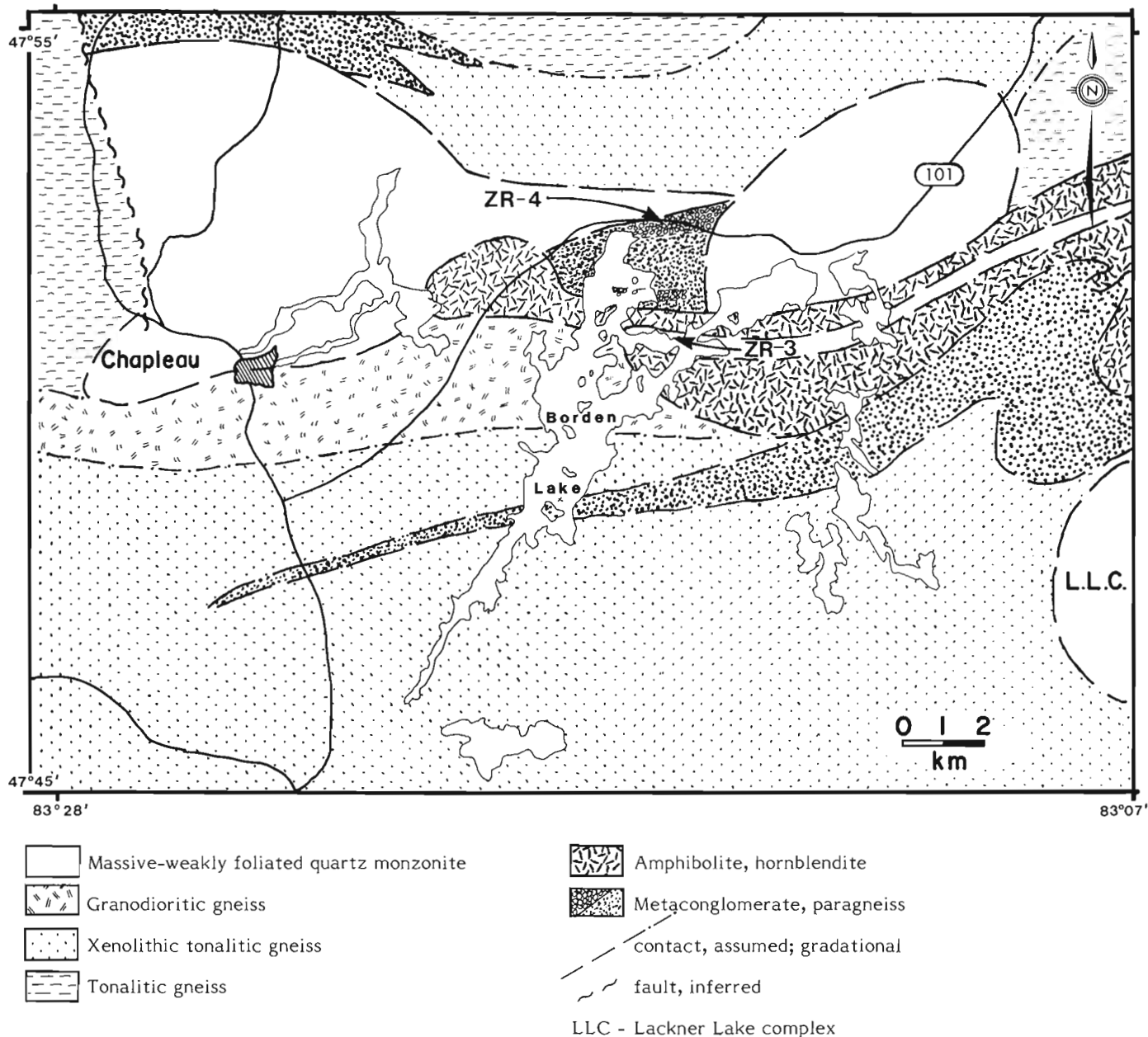


Figure 2. Geology of the Borden Lake area (after Percival, 1981b) and sample locations.

Table 1

Analytical data for zircon fractions from quartz monzonite ZR-3 and from tonalitic clasts in metaconglomerate ZR-4, Borden Lake area, Ontario

Sample number	ZR-3			ZR-4		
Fraction number	1	2	3	1	2	3
Fraction size, μm	-149 +105	+149	-149 +105	-149 +105	+149	-149 +105
Magnetic or non-magnetic	nm	-	mag.	mag.	-	nm
Weight, mg	3.11	3.15	1.77	1.59	2.44	1.32
Total Pb, ng	487.8	687.3	467.0	147.4	171.0	77.42
Pb Blank, %	0.1	0.1	0.1	1.2	0.7	1.1
Observed $^{206}\text{Pb}/^{204}\text{Pb}$	4904	3516	4293	3748	5755	3957
*Abundances ^{204}Pb ($^{206}\text{Pb} = 100$) ^{207}Pb ^{208}Pb	0.0182 17.730 14.566	0.0266 17.732 14.579	0.0210 17.540 14.760	0.0042 18.130 15.476	0.0037 18.133 18.955	0.0049 18.068 18.926
Radiogenic Pb, ppm	553.7	566.3	651.0	281.7	249.5	187.0
" " %	99.13	99.73	99.00	99.80	99.83	99.77
Uranium, ppm	1143	1218	1503	507.3	457.8	380.8
Atomic ratios $^{206}\text{Pb}/^{238}\text{U}$	0.42826	0.41212	0.38321	0.48348	0.46233	0.41722
$^{207}\text{Pb}/^{235}\text{U}$	10.338	9.8901	9.1310	12.053	11.531	10.360
$^{207}\text{Pb}/^{206}\text{Pb}$	0.17506	0.17404	0.17280	0.18079	0.18088	0.18008
Ages, Ma $^{206}\text{Pb}/^{238}\text{U}$	2298	2225	2091	2542	2450	2248
$^{207}\text{Pb}/^{235}\text{U}$	2466	2425	2351	2609	2567	2467
$^{207}\text{Pb}/^{206}\text{Pb}$	2607	2597	2585	2660	2661	2654

*Corrected for Pb blank

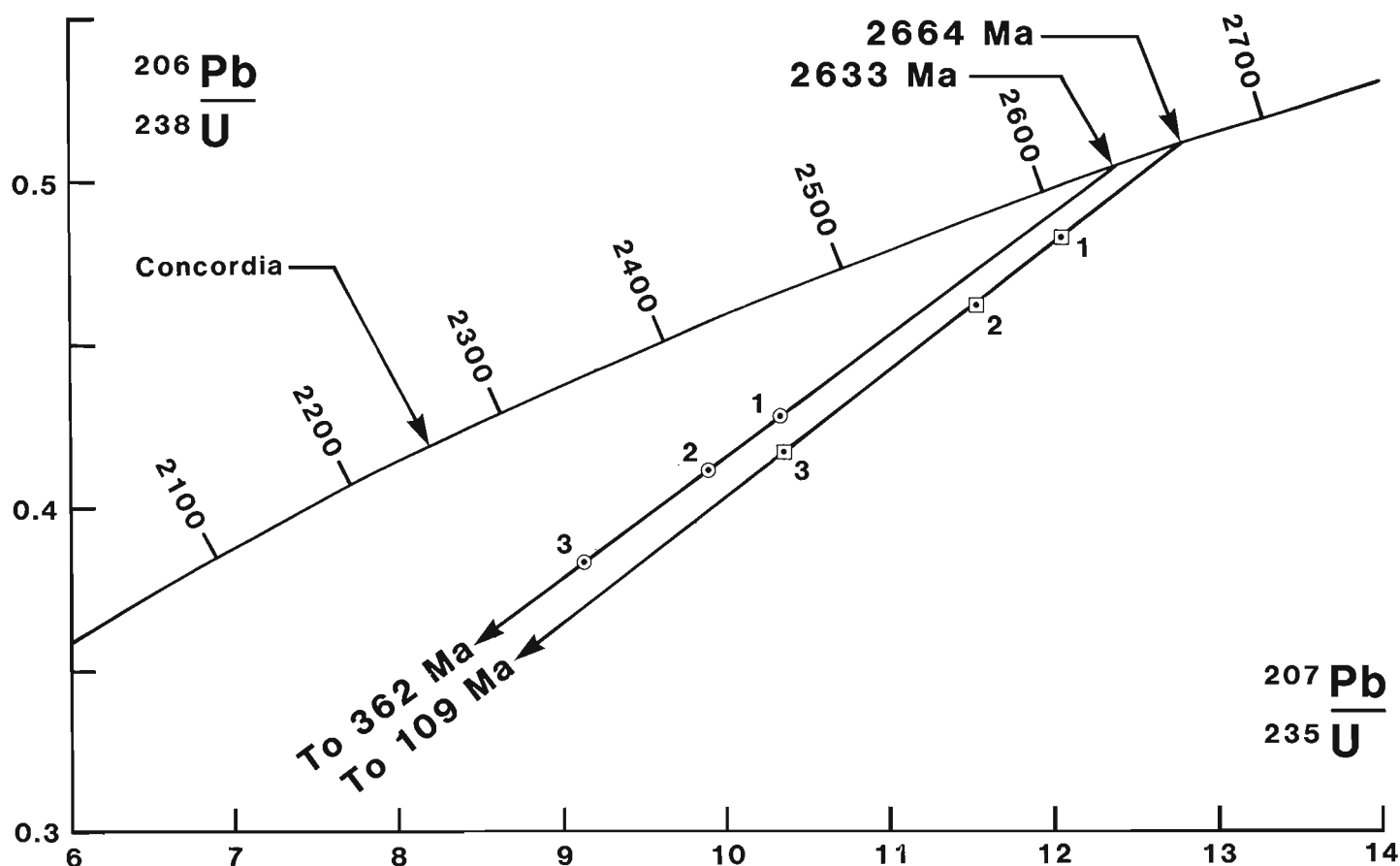


Figure 3. Concordia diagram showing the results of analyses of zircon concentrates from tonalitic clasts from metaconglomerate ZR-4, squares, and quartz monzonite ZR-3, circles, Borden Lake area, Ontario.

Table 2
K-Ar age determination

ZR-3, biotite from quartz monzonite; K-Ar age = 2198 ± 28 Ma
K = 6.16%, $^{40}\text{Ar}/^{40}\text{K} = 0.2496$, radiogenic argon = 99.6%
Concentrate: Biotite, light brown with approximately 15-20% chlorite alteration.

2702 ± 2 Ma (Nunes and Pyke, 1980) and 2708 Ma (Turek et al., 1981). Massive plutons, including the Ivanhoe Lake Pluton, yield U-Pb zircon ages ranging from 2685 to 2668 Ma (Krogh and Turek, in press; Percival, 1981a; T.E. Krogh, personal communication, 1981). Therefore, regional deformation occurred in the interval between ca. 2702 and ca. 2685 Ma.

The interpretation of the 2664 Ma result as a primary age implies that the tonalitic clasts are temporally related to the massive, "post-tectonic" plutons of the Abitibi and Wawa subprovinces. The metaconglomerate, therefore, would represent a second cycle of sediment deposition and should lie unconformably on gneissic rocks. Deformation and metamorphism of conglomerate would have occurred between 2664 ± 6 and 2633 ± 11 Ma ago. Lineations in the KSZ are similar in style and orientation to the rodding lineation in metaconglomerate, suggesting that minor structures in the two areas formed during the same event. The time constraints for deformation in the KSZ (2664 – 2633 Ma) would be consistent with contemporaneous structural overprinting of the 2680 ± 3 Ma Ivanhoe Lake Pluton.

Rocks in the Borden Lake area, including metaconglomerate, are on the low grade side of the amphibolite-granulite isograd. If metamorphism is the same age on both sides of the isograd, this would imply burial to moderate depth of the eastern Wawa Subprovince and deeper burial of the KSZ in the interval 2664 ± 6 to 2633 ± 11 Ma. Reactivation at this time is in marked contrast to the tectonic stability in most of the Superior Province which followed intrusion of the late plutonic suite.

A less likely alternative is that the zircon population crystallized during metamorphism. The coarser zircons have rounded, irregular shapes (Appendix 2) resulting from either dissolution during metamorphism or low-temperature crystallization (Pupin, 1980). Thus zircon may have crystallized for the first time during metamorphism, as suggested for the Adamant Pluton, B.C. (Shaw, 1980). The interpretation of the 2664 Ma date as a secondary age on cobbles of possibly synvolcanic origin leads to a simpler geological explanation. The conglomerate could be part of a supracrustal succession, deposited and buried before the regional deformation. Thus a second cycle of deposition, deep burial and tectonism would not be implied.

Ductile deformation in the KSZ is inferred to post-date intrusion of the Ivanhoe Lake Pluton and is therefore late than the regional deformation. An upper limit on the age of the late deformation is the age of the Ivanhoe Lake Pluton, 2680 ± 3 Ma. A lower limit is provided by the date on quartz monzonite at Borden Lake, 2633 ± 11 Ma. Thus, if the 2664 Ma date is not a primary age, the bracket on the age of the late deformation is 2680–2633 Ma.

Based on a regional structural interpretation, Watson (1980) suggested that ductile sinistral transcurrent movement and upfaulting of granulites in the KSZ accompanied the intrusion of the Matachewan dyke swarm of

age 2633 ± 91 Ma¹ (Rb-Sr whole-rock isochron study, Gates and Hurley, 1973). Uplift of granulite slivers by faulting (Watson, 1980) would not account for the gradational nature of the Wawa Subprovince-KSZ boundary. However, the bracket on the age of deformation provided by this study (2680 or 2664–2633 Ma) supports the timing of ductile deformation suggested by Watson.

Conclusions

U-Pb ages on zircon from tonalitic metaconglomerate cobbles, 2664 ± 6 Ma, and crosscutting quartz monzonite, 2633 ± 11 Ma, suggest that deformation in the KSZ which also affected the immediately adjacent Wawa and Abitibi subprovinces, occurred between 2664 and 2633 Ma ago. Interpretation of the 2664 Ma result as a primary crystallization age leads to a complicated model of geological evolution in which a period of deposition, metamorphism and deformation in the KSZ followed tectonic stabilization in the remainder of the Superior Province. A simpler model regards the conglomerate cobbles as synvolcanic (>2702 Ma) in age, with their zircon populations crystallizing (or recrystallizing) during metamorphism 2664 Ma ago, thereby being totally reset at that time. The results of this study support the primary crystallization age hypothesis but are also compatible with the possibility of a secondary age.

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APPENDIX 1

Sample lithology and localities

1. ZR-3, quartz monzonite: The rock is a pink, coarse- to medium-grained, equigranular to porphyritic (K-feldspar phenocrysts) quartz monzonite. It consists of biotite (10%), plagioclase (25%), K-feldspar (45%) and quartz (20%) with accessory sphene and zircon. A faint lineation, produced by biotite alignment, is present in parts of some outcrops.
Locality: 47°50'25"N, 83°16'30"W. N.T.S. 41 O
 2. ZR-4, tonalite clasts from metaconglomerate: Metaplutonic cobbles comprise 5-10% of the clasts in metaconglomerate. Massive to weakly foliated texture and generally equant cobble shapes indicate that tonalite is a relatively competent rock type. Matrix-clast contacts are diffuse in thin section. The clasts are of medium grained, equigranular, foliated biotite tonalite. Major minerals are biotite (15%), plagioclase (60%) and quartz (25%), with minor K-feldspar, apatite, sphene and zircon.
Locality: 47°52'05"N, 83°16'25"W. N.T.S. 41 O
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APPENDIX 2

Description of zircon fractions

1. ZR-3 quartz monzonite: After the initial purification a -149 +105 μ m size fraction was separated into relatively magnetic (fraction 3) and non-magnetic (fraction 1) subfractions by repeated passes through a Frantz isodynamic separator. Both subfractions were hand picked to yield 100% pure zircon for chemical preparation and isotopic analysis.

Fraction 1: The -149 +105 μ m non-magnetic sample consisted of 100% dark brown, slightly rounded euhedral zircon crystals and obvious crystal fragments varying in shape from stubby (1.5:1) to moderately elongated (4:1). Most crystals show distinct evidence of concentric zoning, and distinct cores were observed in some; in others bright red-brown zones were noted. Some crystals displayed unusual "nipped" terminations suggesting resorption or deformation. Rare rod-shaped inclusions and transverse cracks were observed.

Fraction 2: A +149 μ m "coarse" fraction was hand-picked to yield 100% pure zircon but was not subjected to magnetic subdivision. Crystals range in shape from stubby (1.5:1) to elongated prismatic (> 4.6:1) and are dark brown with distinct concentric zoning. Internal fractures and opaque inclusions were noted in many. Most of the zircons are of somewhat rounded euhedral outline, and some are fragments of originally larger crystals.

Fraction 3: The -149 +105 μ m magnetic sample was essentially similar to the non-magnetic (fraction 1). Rod-shaped and opaque inclusions were more abundant and large cavities were observed in some crystals indicating the effects of resorption or corrosion of some kind. One twinned crystal was noted in this subfraction.

2. ZR-4, tonalitic clasts from metaconglomerate: A -149 +105 μ m size fraction was separated into relatively magnetic (fraction 1) and non-magnetic (fraction 3) subfractions by repeated passes through a Frantz isodynamic separator. Both subfractions were hand picked to yield 100% pure zircon.

Fraction 1: The -149 +105 μ m magnetic sample consisted of 100% zircon grains, crystals, and crystal fragments ranging from very rounded equidimensional to elongated (3.6:1) prismatic euhedral crystals ranging from reddish brown, through pale purple-brown to essentially colourless. Rod, bubble and dust inclusions were common in many grains and zoning was pronounced in some.

Fraction 2: The +149 μ m fraction was not magnetically subdivided. Zircon grains of this sample were generally equidimensional and rounded, to only moderately elongated (2.6:1) with rounded or truncated terminations. Colour ranged from quite dark reddish brown to almost colourless with some internal zoning. Some grains also have a frosted or dappled surface which renders them almost translucent. Rod and bubble inclusions were observed and many grains show intense irregular internal fractures and irregular external faces indicative of moderate abrasion or corrosion.

Fraction 3: Zircon of the -149 +105 μ m non-magnetic sample was essentially similar to that of the non-magnetic subfraction (fraction 1) but was more intensely coloured and rarely contained distinctly red zone boundaries and more distinct cores. There were fewer pale coloured grains and only rare colourless ones.

4. GEOLOGICAL RELATIONSHIPS AND Rb-Sr AGE STUDIES OF THE CAUCHON LAKE TONALITE AND THE PIKWITONEI ENDERBITE IN THE NORTHWESTERN SUPERIOR PROVINCE, CENTRAL MANITOBA

W. Weber¹ and W.D. Loveridge

Weber, W. and Loveridge, W.D., *Geological relationships and Rb-Sr age studies of the Cauchon Lake tonalite and the Pikwitonei enderbite in the northwestern Superior Province, Central Manitoba; in Rb-Sr and U-Pb Isotopic Age Studies, Report 4, in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 115-121, 1981.*

Abstract

Rb-Sr studies produced errorchrons of age 2803 ± 492 , initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7037 ± 0.0015 and MSWD 48.6 on enderbite from the Pikwitonei granulite domain and of age 2879 ± 193 , initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7009 ± 0.0008 and MSWD 4.3 on tonalite from the adjacent lower metamorphic grade Gods Lake subprovince. These results are not sufficiently definitive to establish whether the enderbite is the high grade metamorphic equivalent of the tonalite. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are compatible with some crustal prehistory suggesting assignment of the primary ages of the two rock suites to the plutonic episode at 2.9-3.0 Ga which is widespread in the northeastern Superior Province.

Geological Setting

The Pikwitonei granulite domain² at the northwestern margin of the Superior Province (Fig. 1) consists of granulite facies metaplutonic, gneissic and supracrustal rocks. It represents a portion of the lower crust (Weber, 1980). The corresponding higher portion of the crust is the adjacent lower grade (amphibolite to greenschist facies) greenstone-metaplutonic terrane of the Gods Lake subprovince (Weber and Scoates, 1978; Hubregtse, 1980a). The contact between the two domains is an orthopyroxene isograd (Fig. 1) which is the result of Archean prograde metamorphism increasing in grade towards north and northwest. Plutonic units from both sides of the orthopyroxene isograd were chosen for dating to test the hypothesis that enderbites from the Pikwitonei granulite domain are the higher grade equivalents of tonalites from the Gods Lake greenstone-metaplutonic terrane and, possibly, to date the high grade metamorphism. Along the Churchill-Superior boundary zone (Fig. 1) a lower grade metamorphism was superimposed during the Hudsonian orogeny (Weber and Scoates, 1978). The orthopyroxene isograd is oblique to the regional lithologic layering and structures (Hubregtse, 1980a). It crosses greenstone strata at Cauchon Lake and High Hill Lake (Fig. 1), and the southwestern extension of the Cross Lake greenstone belt (Hubregtse, 1980b).

The rocks identifiable as the oldest in the field are supracrustal rocks, both in the granulite and the adjacent lower grade terrane. These supracrustal rocks are intruded by granitoid rocks on both sides of the isograd and they are also seen as inclusions of variable size in metaplutonic rocks. There is no evidence that the granulites are basement to the greenstones, in contradiction to the "basement granulite hypothesis" suggested by Bell (1971, 1978) on the basis of a granodiorite/conglomerate relationship at Cross Lake (see Hubregtse, 1980a). Since field relations suggest only one period of felsic plutonism prior to the high grade metamorphism (M_1 , Hubregtse, 1980a) it is assumed that these plutons were emplaced at the same time on both sides of the orthopyroxene isograd. However, the high grade metamorphism may have obliterated a more complex plutonic history.

During a major episode of plutonism at 2.7-2.8 Ga, greenstones in the northwestern Superior Province, also believed to be of age 2.7-2.8 Ga, were intruded by tonalitic to granodioritic plutons. U-Pb zircon ages of 2.71 to 2.76 Ga were obtained from plutonic and volcanic rocks in the Uchi Lake area (Krogh and Davis, 1971³; Nunes and Thurston, 1980) and from plutonic rocks in the Rainy Lake

area (Krogh and Davis, 1971³; Hart and Davis, 1969³) and also in the Berens belt (Krogh et al., 1974). Available Rb-Sr data suggest a similar age for plutonism and volcanism in the Oxford-Knee Lake area, near the area of the present study and the Rice Lake area (see Weber and Scoates, 1978).

An earlier plutonic episode, at 2.9-3.0 Ga, is also widespread in the northeastern Superior Province, and was identified in the Berens belt (Krogh et al., 1974), in the English River Belt (Krogh et al., 1976; Clark et al., 1981) and the Favourable Lake area (Krogh and Davis, 1971³; Corfu et al., 1981). Zircon ages from volcanic rocks in the Uchi Lake, North Spirit Lake and Favourable Lake areas (Nunes and Thurston, 1980; Nunes and Wood, 1980; Corfu et al., 1981) indicate that volcanism had taken place during the same time interval, between 2.9-3.0 Ga.

Age relationships between the 2.9-3.0 Ga old plutonic rocks and supracrustal rocks are generally inconclusive; contacts between plutonic and volcanic rocks in this age bracket are not exposed. In analogy to the 2.7-2.8 Ga episode it is probable that the 2.9-3.0 Ga plutonic episode is related to the volcanism of similar age, is largely intrusive into the volcanic rocks, and is not a basement. Such a relationship is supported by trace element data from early Archean rocks from many Shield areas which suggest that tonalitic plutons probably resulted from partial melting of metamorphosed mafic juvenile crust, compositionally similar to mafic volcanics in the greenstone belts (e.g. O'Nions and Pankhurst, 1978).

Thus, the age of the metaplutonic rocks in the northern Gods Lake Subprovince probably lies either between 2.7 and 2.8 Ga or 2.9 or 3.0 Ga. An age for the plutonic rocks in the Cauchon Lake and Pikwitonei area would also provide a minimum age for the supracrustal rocks.

Cauchon Lake Tonalite

Sample locations are shown in Figure 1, petrographic descriptions are listed in Table 1, and the results of chemical analyses are given in Table 2. Most of the samples were collected from metatonalite material over an area of 10 km² near the south shore of Cauchon Lake. A late (D_3 ?) mylonite of regional extent runs along the south shore. Samples 1 and 3 are from near this mylonite and are cataclastic. Samples 5 and 6 are from metatonalite which was "soaked" by younger granitic melts related to a late-tectonic (post M_2) intrusion.

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²Tectonic terminology after Manitoba Mineral Resources Division Geological Map of Manitoba, 1979, scale 1:1 000 000 (Map 79-2)

³Results as published were recalculated using U decay constants of Jaffey et al. (1971)

Table 1
Sample numbers, description, mineralogical composition and localities, Cauchon Lake tonalite, N.T.S. 63P

This work	Sample No.		Rock description	Mineralogical composition	Location			
	Field				Latitude	Longitude		
1	02-6-22-1		Cataclastic, foliated (late S ₂ or S ₃) coarse grained, biotite tonalite; quartz, and to a lesser degree plagioclase, granulated; biotite cataclastic, exsolution of opaques, some epidote associated with biotite aggregates.	Qtz. Plag. Bio.	20-30% 50-60% 10-15%	Ep. +Opaq. 5%	55° 20' 00"	96° 38' 50"
2	02-6-24-1		Weakly cataclastic, foliated biotite tonalite; quartz partly granulated.	Similar to 02-6-22-1			55° 20' 20"	96° 38' 40"
3	02-6-19-1		Similar to 02-6-22-1	Similar to 02-6-22-1			55° 20' 10"	96° 38' 30"
4	02-6-33-1		Weakly S ₂ -foliated, medium grained, biotite tonalite; interstitial microcline, aggregates of biotite + epidote + sphene (derived from hornblende?)	Qtz. Plag. Bio. Ep. Micr.	20-30% 40-50% 15-20% 2-3% 2-5%	Cpx. +Opaq. 2%		
5	02-6-42-2		Weakly foliated, fine grained to coarse grained, inequigranular biotite tonalite. Microcline interstitial to blastic. Coarse grained, inequigranular portions of stained slab contain more microcline than fine grained portion. Biotite largely chloritized.	Qtz. Plag. Micr. Bio, Chl.	30-40% 40-50% 5-15% 10-15%	Opaq. 2-5%	55° 21' 30"	96° 23' 30"
6	02-6-42-1		Weakly foliated, fine- to medium-grained granodiorite; microcline myrmekitically replacing plagioclase.	Qtz. Plag. Micr. Bio.	30-40% 40% 10% 10%	Ep. +Opaq. 2%	55° 21' 30"	96° 13' 30"

Pikwitonei Enderbite

Sample locations are shown in Figure 1, petrographic descriptions are listed in Table 3, and the results of chemical analyses are given in Table 4. The enderbite samples are honey-coloured, medium to coarse grained, foliated and in part layered, and contain hypersthene and biotite. Hypersthene is fresh or only slightly altered.

Analytical Results

Cauchon Lake Tonalite

The results of Rb and Sr analyses on the Cauchon Lake tonalite are given in Table 5 and depicted in an isochron diagram, Figure 2. The six sample points are not collinear

within analytical uncertainty and thus do not define an isochron. The results of regression analyses on the six points, and also two groupings of five points, are listed below:

Samples included	Age (Ma)	Initial $^{87}\text{Sr}/^{86}\text{Sr}$	MSWD
1-6	2879 ± 193	0.7009 ± 0.0008	4.3 (errorchron)
1-5	3066 ± 317	0.7004 ± 0.0011	3.2 (errorchron)
1-4,6	2803 ± 133	0.7011 ± 0.0005	1.7 (isochron)

Figure 2 depicts the regression results on all six samples, the first of the three above combinations. One grouping of five points, 1 to 4 plus 6, yields an MSWD low enough to be considered an isochron.

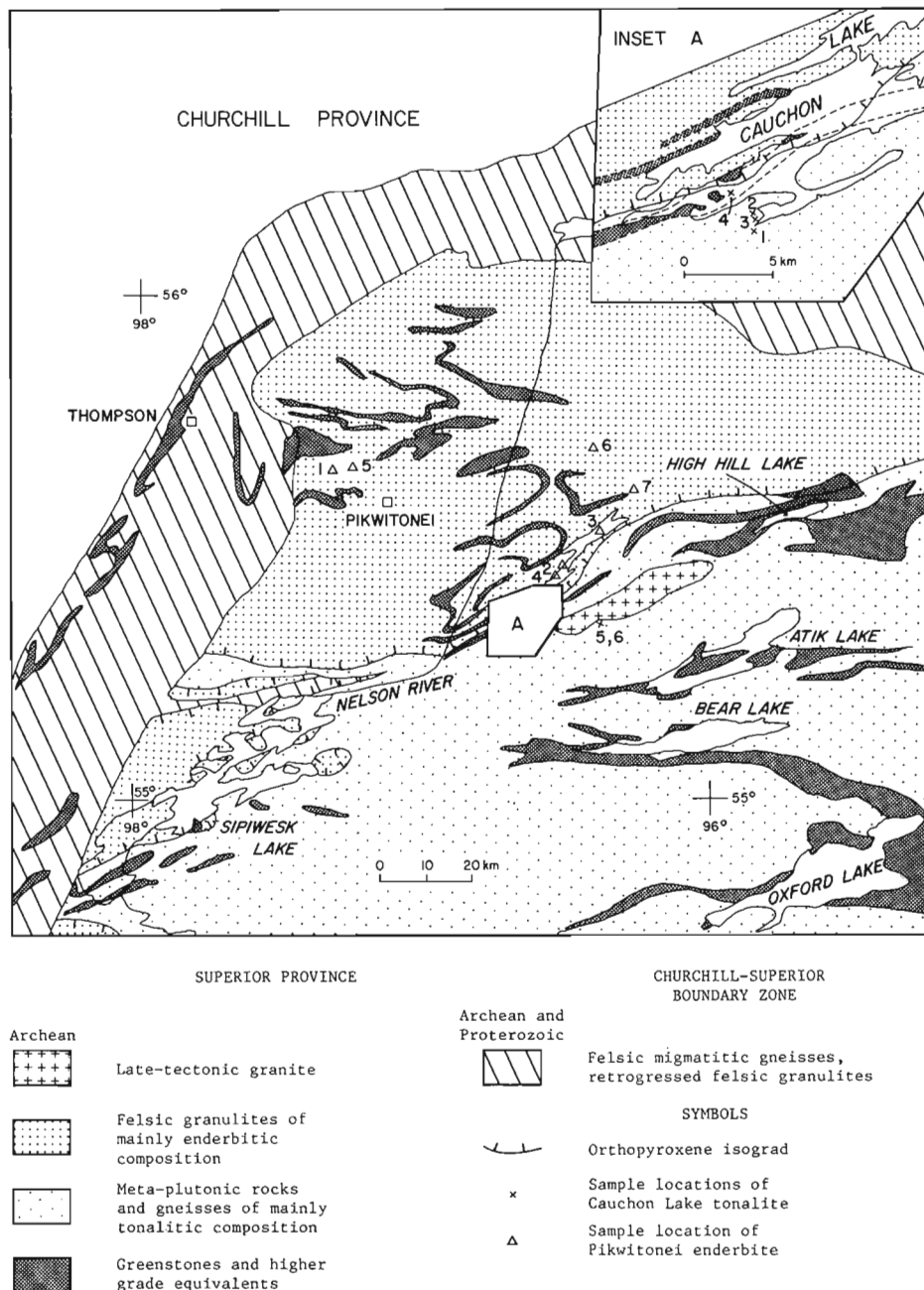


Figure 1. Geology and sample locations, northwestern Superior Province, central Manitoba.

Table 2
Chemical analyses (wt.%), Cauchon Lake tonalite

Sample no. (this work)	1	2	3	4	5	6
SiO ₂		70.84				
Al ₂ O ₃		16.26				
TiO ₂	0.21	0.23	0.18	0.23	0.21	0.35
Fe ₂ O ₃		0.84				
FeO		1.06				
MnO	0.03	0.04	0.03	0.04	0.03	0.05
MgO	0.65	0.74	0.57	0.70	0.50	0.90
CaO	0.86	3.19	3.27	3.27	2.28	2.25
Na ₂ O	5.50	5.19	5.54	5.19	4.91	4.58
K ₂ O	1.14	1.38	1.23	1.31	1.51	2.06
P ₂ O ₅		0.05				
CO ₂		0.15				
H ₂ O _t		0.35				

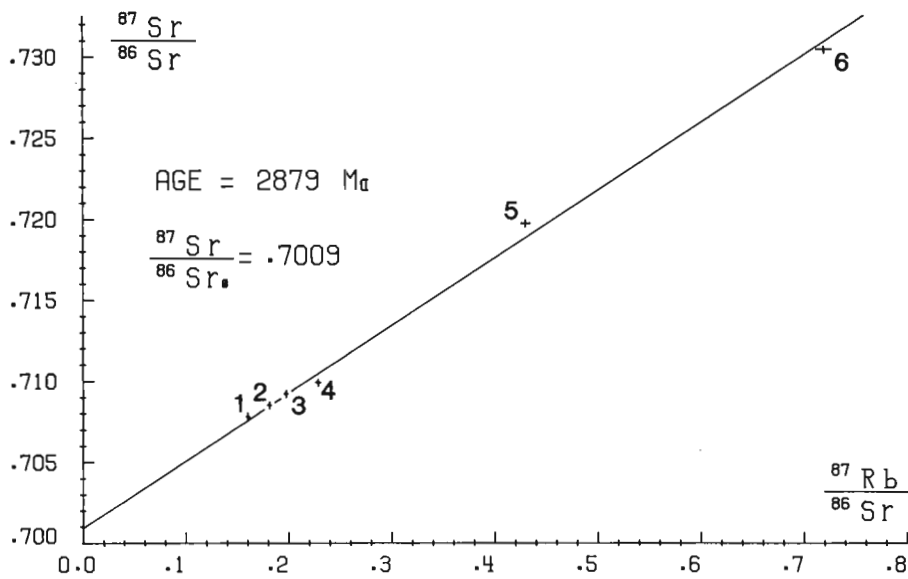


Figure 2. Rb-Sr isochron diagram, Cauchon Lake tonalite, central Manitoba.

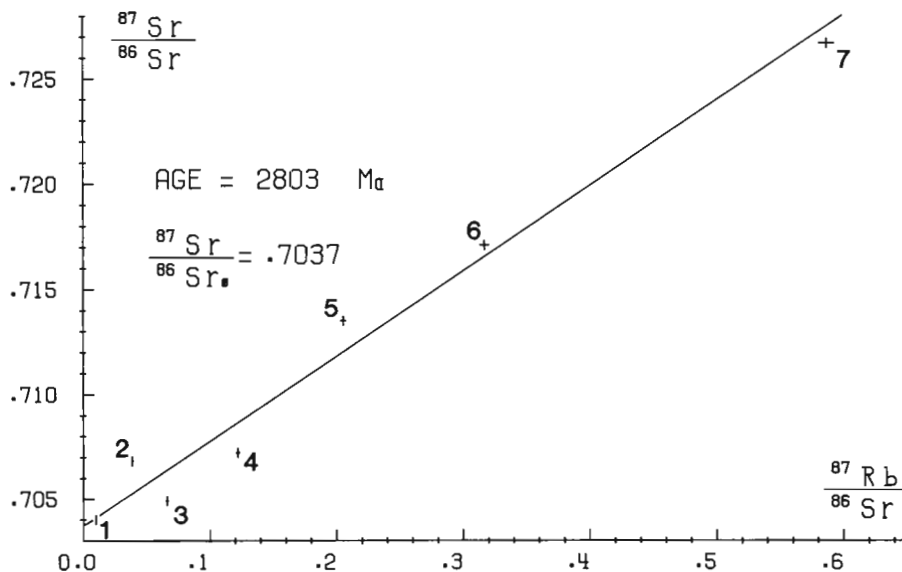


Figure 3. Rb-Sr isochron diagram, Pikwitonei enderbite, central Manitoba.

Table 3

Sample numbers, description, mineralogical composition and localities, Pikwitonei enderbite, N.T.S. 63P

Sample No. This work	Field	Rock description	Mineralogical composition	Location	
				Latitude	Longitude
1	02-6-313	S ₂ -foliated, coarse grained enderbite; small biotite as exsolutions in hypersthene; hornblende, clinopyroxene rimming hypersthene.	Qtz. Plag. Hyp. Hb.+Cpx. Opaq. 30% 40% 20% 5% 5%	Bio. < 5% 55° 39' 40"	97° 19' 20"
2	02-6-10-2	S ₂ -foliated, medium grained enderbite; some of the hypersthene marginally altered to fine grained biotite.	Qtz. Plag. Bio. Hyp. 30-35% 40-45% 5-10% 5-10%	Opaq. 2-5% Hb. +Cpx. < 1% +Serp. 55° 28' 00"	96° 30' 30"
3	02-6-59-1	S ₂ -foliated, coarse grained enderbite; porphyroblastic hypersthene.	Qtz. Plag. Hyp. Opaq. 20-30% 40-50% 10-20% 5-10%	Bio. < 5% 55° 32' 30"	96° 22' 30"
4	02-6-12-1	S ₂ -foliated and layered medium- to coarse-grained enderbite. On same outcrop intrusion of pegmatite leads to alteration of hypersthene and recrystallization to hornblende.	Qtz. Plag. Bio. Hyp. Opaq. 30% 40% 10% 10% 5%	Hb. +Cpx. < 2% 55° 27' 30"	96° 31' 00"
5	02-6-316-2	S ₂ -foliated, coarse grained enderbite; some microcline in feldspar mosaic. Part of outcrop is retrogressed and sheared enderbite.	Qtz. Plag. Hyp. Hb.+Cpx. Opaq. 30% 40% 10-20% 5% 5%	Bio. < 2% 55° 39' 30"	97° 16' 30"
6	02-6-44-1	S ₁ -foliated and layered, S ₂ -folded, medium- to coarse-grained enderbite. Most hypersthene marginally slightly altered to biotite or hornblende.	Qtz. Plag. Bio. Hyp. 30% 40% 10-15% 5-10%	Opaq. 2-5% Cpx. < 1% +Hb. +Apat. 55° 42' 00"	96° 23' 25"
7	02-6-77-1	S ₁ -foliated, medium grained, homogeneous enderbite, microcline disseminated in feldspar fabric.	Qtz. plag. Hyp. Opaq. Micr. 30% 40% 5-10% 5-10%	Cpx. < 2% +Bio. 55° 36' 20"	96° 16' 50"

Table 4
Chemical analyses (wt. %), Pikwitonei enderbite

Sample no. (this work)	1	2	3	4	5	6	7
SiO ₂					70.60		
Al ₂ O ₃					15.20		
TiO ₂	0.52	0.48		0.45	0.29	0.34	0.35
Fe ₂ O ₃					0.75		
FeO					1.44		
MnO	0.07	0.06		0.06	0.03	0.03	0.05
MgO	1.73	1.79		1.64	1.05	1.04	0.79
CaO	4.84	5.76		4.65	3.18	3.67	3.08
Na ₂ O	4.29	4.81		4.38	4.06	4.68	3.39
K ₂ O	0.82	0.85		1.32	2.62	1.68	2.79
P ₂ O ₅					0.06		
CO ₂					0.36		
H ₂ O _t					0.42		

Table 5
Analytical data, whole-rock samples.
Cauchon Lake tonalite, Manitoba

Sample No.	Rb ppm	Sr ppm	⁸⁷ Sr/ ⁸⁶ Sr	⁸⁷ Rb/ ⁸⁶ Sr
1	33.54	601.1	0.7078 ± 0.0004	0.1600 ± 0.0032
2	37.05	591.8	0.7085 ± 0.0004	0.1810 ± 0.0036
3	36.55	535.6	0.7092 ± 0.0004	0.1973 ± 0.0039
4	45.22	572.3	0.7099 ± 0.0004	0.2284 ± 0.0046
5	46.14	310.3	0.7197 ± 0.0004	0.4299 ± 0.0086
6	80.23	322.5	0.7304 ± 0.0004	0.7192 ± 0.0144

Table 6
Analytical data, whole-rock samples,
Pikwitonei enderbite, Manitoba

Sample No.	Rb ppm	Sr ppm	⁸⁷ Sr/ ⁸⁶ Sr	⁸⁷ Rb/ ⁸⁶ Sr
1	1.75	504.4	0.7040 ± 0.0004	0.0100 ± 0.0002
2	4.30	320.5	0.7068 ± 0.0004	0.0388 ± 0.0008
3	9.37	408.6	0.7049 ± 0.0004	0.0663 ± 0.0013
4	20.90	494.5	0.7072 ± 0.0004	0.1222 ± 0.0024
5	31.58	444.2	0.7135 ± 0.0004	0.2055 ± 0.0041
6	43.12	393.4	0.7171 ± 0.0004	0.3169 ± 0.0063
7	48.36	238.3	0.7267 ± 0.0004	0.5867 ± 0.0117

Pikwitonei Enderbite

The results of isotopic analyses on seven whole rock samples are listed in Table 6 and displayed in an isochron diagram, Figure 3. The seven sample points show considerable scatter about any line drawn through them, well outside analytical uncertainty. The results of regression analyses on all seven samples are: age, 2803 ± 492 Ma; initial ⁸⁷Sr/⁸⁶Sr, 0.7037 ± 0.0015; MSWD, 38.6.

Interpretation of the Rb-Sr Data

Cauchon Lake Tonalite

Although Sample 6 may have been affected by a younger granitic melt, only the inclusion of this sample and exclusion of sample 5 from the regression calculation produces an isochron, of age 2803 ± 133 Ma. The initial ratio of 0.7011 ± 0.0003 does not preclude a previous crustal history

of this rock, prior to 2803 Ma. Since sample 6 essentially controls the slope of this isochron, it is possible that the age of 2803 Ma is too young, and that sample 6 records a younger event.¹

The original age of the tonalite may be better reflected by excluding sample 6. If, instead, sample 5 is included the resulting errorchron indicates an age of 3066 ± 317 Ma with a low initial ratio of 0.7004 ± 0.0011.

Statistically the two ages are not different from each other and the results do not provide a clear distinction between the 2.7-2.8 and the 2.9-3.0 Ga age groupings. We interpret the Rb-Sr data of the Cauchon Lake tonalite as being compatible with emplacement during the 2.9-3.0 Ga plutonism but not definitive of this age range.

Pikwitonei Enderbite

The points for the enderbite show considerable scatter and do not define an isochron.

A representative age of 2803 ± 492 Ma (initial ratio of 0.7037 ± 0.0015) is statistically not different from the age of the tonalite. The high initial ratio suggests that this "age" is not an original age but could indicate the time of granulite facies metamorphism. Thus the enderbite may also be older than the 2.7-2.8 Ga plutons in the northwestern Superior Province. It is possible that the scattered distribution of the points does not represent an age but rather a disturbance to the Rb-Sr system by the granulite facies metamorphism. We interpret the Rb-Sr systematics of the Pikwitonei enderbite as indicating a disturbed system reflecting the effects of granulite facies metamorphism, probably at 2.7-2.8 Ga, on pre-existing rocks of a possible 2.9-3.0 Ga age.

The results of these two studies are not sufficiently definitive to establish whether the Pikwitonei enderbite is the high grade metamorphic equivalent of the Cauchon Lake tonalite, but they do not contradict that hypothesis.

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¹ Rb-Sr ages of granitic melts in the northwestern Superior Province have yielded ages of approximately 2.5 Ga (see Weber and Scoates, 1978)

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5. U-Pb AGES OF ZIRCONS FROM THE FOOT BAY GNEISS AND THE DONALDSON LAKE GNEISS, BEAVERLODGE AREA, NORTHERN SASKATCHEWAN

L.P. Tremblay, W.D. Loveridge, and R.W. Sullivan

Tremblay, L.P., Loveridge, W.D., and Sullivan, R.W., U-Pb ages of zircons from the Foot Bay gneiss and the Donaldson Lake gneiss, Beaverlodge area, northern Saskatchewan; in Rb-Sr and U-Pb Isotopic Age Studies, Report 4, in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 123-126, 1981.

Abstract

U-Pb studies on zircon have yielded concordia intercept ages of $2513 \pm 36/-22$ Ma for the Foot Bay gneiss and 2179 ± 12 Ma for the Donaldson Lake gneiss of the Tazin succession of Tremblay (1972). Alternative interpretations of these results prevent resolution of the controversy as to whether or not there is an unconformity and considerable age difference between the two units.

Geological Setting

The zircon concentrates dated here are from the two lowermost stratigraphic units of the Tazin succession as mapped by Tremblay (1972) in the Beaverlodge area, Saskatchewan. However, as the lower limit of the Tazin Group has not yet been mapped or recognized, the stratigraphic position of these two units within the Tazin succession is somewhat arbitrary. Figure 1 is a geological sketch map of the Donaldson Lake - Foot Bay area showing sample locations.

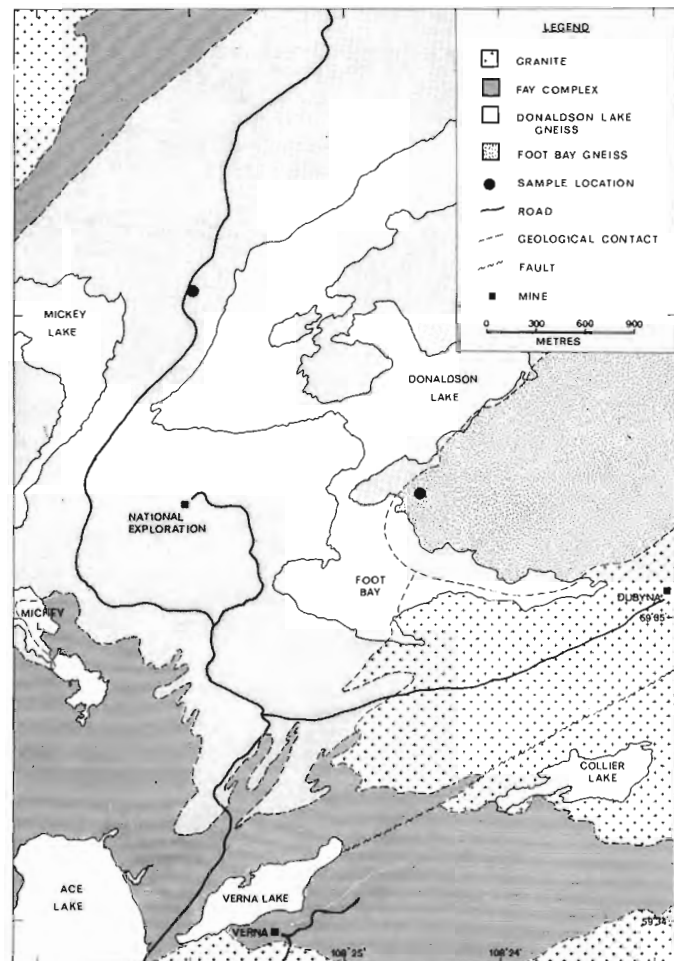


Figure 1. Geological sketch map of part of the Beaverlodge area showing sample locations.

The lowermost unit, called the Foot Bay gneiss (Tremblay, 1972), is a dark red, fine- to medium-grained, layered gneiss with 10 to 30 per cent mafic minerals, mainly chlorite as an alteration product after hornblende. This rock also contains much epidote in clusters and seams. It has the composition of a greywacke and is assumed to be a metasediment. It exhibits in thin section lenses and layers of cataclastic products.

The other unit sampled, the Donaldson Lake gneiss, is a white, coarse grained, poorly layered, granitoid rock much resembling a granite. It is seen interlayered with glassy quartzite beds and amphibolite bands and is also assumed to be a metasediment, probably a recrystallized arkose or feldspathic quartzite. Its composition resembles that of a biotite granite. It also shows cataclastic effects but these are more diffuse than in the Foot Bay gneiss.

The Foot Bay gneiss is believed by Tremblay (1972) to be in conformable contact with the overlying Donaldson Lake gneiss whereas Krupicka and Sassano (1972), based on the number of deformations that they have recognized in thin sections of samples derived mainly from the rock succession in the underground workings of the Eldorado Fay-Ace-Verna Mine, state that this contact is an unconformity implying that the Donaldson Lake gneiss is younger than the Foot Bay gneiss and that it belongs to a different geological period. However it is possible that these thin sections were from samples not belonging to the above two lowermost units of the Tazin Group as mapped by Tremblay but from rock samples resembling those units in the mine workings, where deformation is extreme and of several generations.

Analytical Techniques and Results

Techniques used in the concentration of zircon and the extraction and analysis of lead and uranium are described in Sullivan and Loveridge (1980). Analytical results are presented in Table 1 and displayed on a concordia diagram, Figure 2. A description of the zircon morphology of the analyzed fractions is presented in the appendix.

Foot Bay gneiss: The zircon concentrate was sieved to produce a $+105\mu\text{m}$ fraction which was separated into relatively magnetic and nonmagnetic subfractions by a Frantz isodynamic separator. These subfractions were purified to approximately 100 per cent zircon by hand picking. The two analytical points fall near the upper end of a chord which intersects the concordia curve at 2513^{+36}_{-22} and 913^{+144}_{-111} Ma.

Donaldson Lake gneiss: The zircon concentrate was sieved to produce a $+149\mu\text{m}$ fraction and a (much finer) $-62 + 44\mu\text{m}$ fraction. Both were separated into relatively magnetic and nonmagnetic subfractions by a Frantz

isodynamic separator. Both of the +149 μm subfractions were purified to approximately 100 per cent zircon by hand picking prior to analysis but only the more magnetic -62 + 44 μm subfraction was analyzed and it was not hand picked due to the fine grain size. We estimate impurity levels to be less than 5 per cent in this subfraction. The three analytical points are collinear within analytical uncertainty and define the upper end of a chord which cuts the concordia curve at 2179 ± 12 and 491 ± 69 Ma.

Interpretation

The Foot Bay gneiss has been mapped as Archean (Tremblay, 1972, 1978; Krupicka and Sassano, 1972) and the age of 2513 ± 22 Ma obtained on zircon concentrates from this rock is not inconsistent with this conclusion. (2513 Ma may be considered Archean or very early Aphebian depending on the time scale employed; e.g. Stockwell (1973) places the Archean - Aphebian boundary at 2560 Ma, whereas Sims (1980) suggests 2500 Ma).

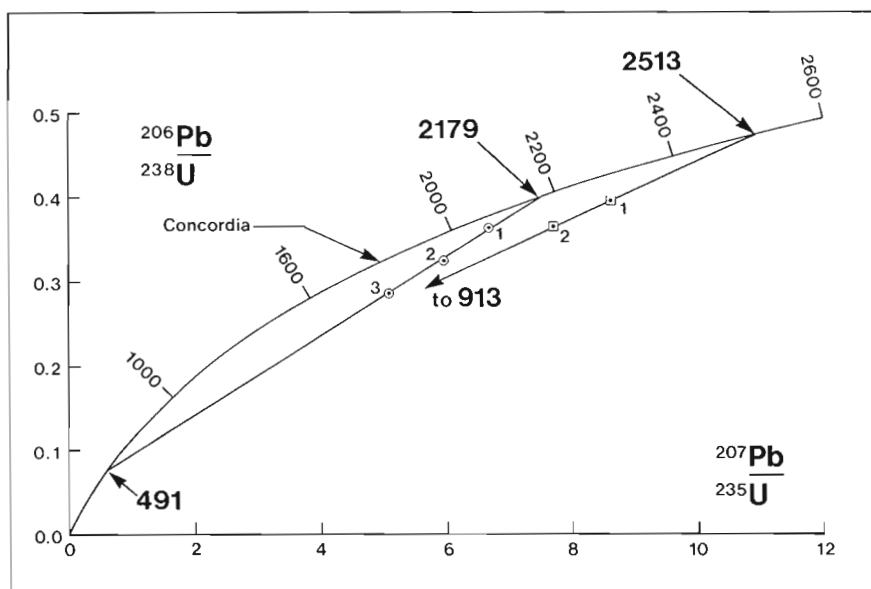


Figure 2. Concordia diagram showing the results of U-Pb analyses of two zircon fractions from the Foot Bay gneiss (squares) and three zircon fractions from the Donaldson Lake gneiss (circles).

Table 1
Analytical data, zircon fractions. Foot Bay gneiss and Donaldson Lake gneiss, northern Saskatchewan

	Foot Bay gneiss		Donaldson Lake gneiss		
Fraction number	1	2	1	2	3
Fraction size (μm)	+105	+105	+149	+149	-62+44
Magnetic or nonmagnetic	nm	m	nm	m	m
Weight (mg)	4.14	3.21	4.53	5.24	4.07
Total Pb, ng	272.5	269.1	231.9	389.0	296.5
Pb blank, %	0.4	0.9	0.4	0.2	0.4
Observed $^{206}\text{Pb}/^{204}\text{Pb}$	5848	3016	2230	2679	3128
Abundances* ^{204}Pb	0.0107	0.0178	0.0376	0.0331	0.0260
($^{206}\text{Pb} = 100$) ^{207}Pb	15.827	15.430	13.894	13.640	13.174
^{208}Pb	12.608	12.534	16.187	13.116	8.420
Radiogenic Pb ppm	212.1	270.5	139.2	194.8	262.6
%	99.5	99.1	98.1	98.3	98.6
Uranium ppm	481.4	674.1	345.9	551.3	881.5
Atomic ratios $^{206}\text{Pb}/^{238}\text{U}$	0.40020	0.36679	0.36410	0.32794	0.28769
$^{207}\text{Pb}/^{235}\text{U}$	8.6591	7.6888	6.7288	5.9719	5.0905
$^{207}\text{Pb}/^{206}\text{Pb}$	0.15691	0.15202	0.13403	0.13206	0.1283
Ages (Ma) $^{206}\text{Pb}/^{238}\text{U}$	2170	2014	2002	1828	1630
$^{207}\text{Pb}/^{235}\text{U}$	2303	2195	2076	1972	1835
$^{207}\text{Pb}/^{206}\text{Pb}$	2423	2369	2151	2126	2075

* After subtraction of Pb blank.

The age obtained may be that of the original rock unit from which the sedimentary precursor to the Foot Bay gneiss was derived. However, as this sedimentary precursor was probably a greywacke, the period of time between emplacement of the parent unit and formation of the greywacke may be minimal. The internal euhedral zonation of the zircons suggests an igneous origin supporting the hypothesis that the age is that of the parent unit. Therefore, the age of 2513 Ma is probably a maximum for the Foot Bay gneiss although it may very closely represent the age of the gneiss itself or its sedimentary precursor.

The Donaldson Lake gneiss is regarded by Tremblay (1972) as part of the Tazin Group and of Archean age. Krupicka and Sassano (1972), because they assume that an unconformity exists between the Foot Bay gneiss and the Donaldson Lake gneiss, regard the Donaldson Lake gneiss and its overlying rocks (that is most of the Tazin Group) as Aphebian. The date obtained here on the zircon concentrate from the Donaldson Lake gneiss is 2179 ± 10 Ma, which supports the contentions of Krupicka and Sassano.

Three alternative interpretations may be derived from the data at hand:

1. It is possible that the 2179 Ma age does not represent the primary age of the Donaldson Lake gneiss but rather indicates a period of new zircon growth during the anatexis that is widespread in the Beaverlodge area. This anatexis has affected both the Foot Bay gneiss and the Donaldson Lake gneiss but it has affected the Donaldson Lake gneiss more strongly; it is now a medium- to coarse-grained, granitoid and granite like rock. If this is the case, the sedimentary predecessor to the Donaldson Lake gneiss must have been zircon free as there is no evidence of detrital zircon cores or of an older lead component in the zircon fractions analyzed from the Donaldson Lake gneiss. If this suggestion is correct and the age of the Foot Bay Gneiss is 2513 Ma as measured, the Tazin Group may indeed be an Archean or very early Aphebian sedimentary succession.
2. The 2179 Ma age may date the anatexis of the Donaldson Lake gneiss or the unit from which the sedimentary predecessor of this gneiss was derived. If the 2513 Ma age associated with the Foot Bay gneiss dates the unit from which the Foot Bay gneiss was derived, then both gneisses could be Aphebian in age and the Tazin succession of Tremblay may be entirely an Aphebian sedimentary succession.

3. Both the 2513 Ma age for the Foot Bay gneiss and the 2179 Ma age for the Donaldson Lake gneiss may be taken at face value. In that case the geological history of the Tazin Group corresponds with that propounded by Krupicka and Sassano.

Although it is apparent that the geological and geochronological information on hand is insufficient to provide a clear choice between these three hypotheses, Tremblay favors the first interpretation. We plan further geochronological studies in an attempt to define more precisely the geological history of the Tazin succession of the Beaverlodge area.

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APPENDIX I

Description of zircon concentrates as analyzed.
Foot Bay gneiss and Donaldson Lake gneiss

Foot Bay Gneiss

Fraction 1, + 105 μ m, non magnetic, hand picked

The analyzed material consisted of clear to translucent, essentially colourless often rounded zircon grains and fragments of grains with clearly visible euhedral internal zoning, transverse fractures and rod- and bubble-like inclusions. Grain surfaces are irregular and "pitted". Grain elongation is variable from 1.5:1 to about 3:1. Interference colours are "mottled" and extinction imperfect.

Fraction 2, + 105 μ m magnetic, hand picked

The analyzed material was essentially identical to fraction 1 above except that the zircon grains and fragment also contained some black inclusions.

Donaldson Lake Gneiss

Fraction 1, + 149 μ m, non magnetic, hand picked

The analyzed material consisted of very pale tinged pink to purple zircon crystals and crystal fragments. Crystal terminations vary from euhedral to slightly rounded. Some of the crystals have a plate-like morphology. Bubble-like inclusions were observed and lines and fractures are common. Zoning is evident in some crystals and most tend to be translucent. Interference colours are "mottled", extinction is imperfect, and some crystals tend towards being isotropic, possibly indicating metamictization. Grain shape varies from equidimensional to an elongation of about 2.5:1.

Fraction 2, + 149 μ m, magnetic, hand picked

Essentially similar to the nonmagnetic fraction, but more (25%) of the zircon crystals contain dark inclusions. Most are euhedral, light purple-tan, and commonly prismatic but occasionally platy. Internal zoning is clearly visible.

Fraction 3, -62 + 44 μ m, magnetic

The very fine grained material for this analysis consisted of more than 90 per cent clear, rounded zircon grains and about 5 per cent zircon fragments. Essentially similar to the +100 mesh material but possibly more rounded and with occasional length to breadth ratios up to 4:1. Internal zoning is clearly visible and internal fractures and inclusions are visible, but being very much finer the grains appear clearer and colourless.

6. Rb-Sr ISOCHRON AGE OF WEATHERED PRE-ATHABASCA FORMATION BASEMENT GNEISS, NORTHERN SASKATCHEWAN

W.F. Fahrig and W.D. Loveridge

Fahrig, W.F. and Loveridge, W.D., Rb-Sr isochron age of weathered pre-Athabasca Formation basement gneiss, northern Saskatchewan; in *Rb-Sr and U-Pb Isotopic Age Studies, Report 4; in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 127-129, 1981.*

Abstract

The Athabasca Formation of northern Saskatchewan, consisting largely of quartz sandstone, is known to have been deposited within the time interval 1800-1220 Ma ago. A number of major uranium deposits at the base of this formation have prompted an extensive exploratory drilling program. The data presented in this paper are from measurements on specimens of core from the Rumble Lake drill hole representing deeply weathered basement rocks. We interpret the results obtained, age 1632 ± 32 Ma, initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7067 ± 0.0015 , MSWD 1.77 as dating a time of homogenization due to weathering or diagenesis, approximately the age of Athabasca sedimentation.

Geological Setting

The Athabasca Formation occupies a great basin covering about 98 000 km², chiefly in northern Saskatchewan (Fahrig, 1961; Fraser et al., 1970). It has been correlated with other Helikian sedimentary units of the western Shield; the Thelon, Hornby Bay, and Tinney Cove formations (Fraser et al., 1970). The Athabasca is cut by dykes of the Mackenzie swarm (Fahrig and Jones, 1969) which are 1220 Ma old (Patchett et al., 1978) and lies on gneisses of the Churchill Province, which in this region have been undisturbed except for uplift and faulting since at least 1800 Ma ago (Baadsgaard and Godfrey, 1972; Stockwell, in press). The Athabasca was therefore deposited in the interval 1800-1220 Ma ago.

The formation consists largely of quartz sandstone although it contains a significant shaly interval (Lerand, 1980). Deposition was in shallow water and was mainly fluvial.

A number of major uranium deposits, the first being the Rabbit Lake deposit, have been discovered at essentially the interface between the Helikian Athabasca Formation and the underlying Aphebian gneisses. This early find initiated an extensive exploratory drilling program in the late 1960s and early 1970s and material from some of these drill holes (unoriented in the horizontal plane) was used in an initial paleomagnetic study of the Athabasca Formation (Fahrig et al., 1978). This study suggested an age of at least 1550 Ma for Athabasca sedimentation and also suggested the occurrence of polarity reversals during deposition of the formation.

Table 1
Basement chemistry

Depth*	MnO	TiO ₂	CaO	K ₂ O	SiO ₂	Al ₂ O ₃	MgO	FeO	Fe ₂ O ₃	Na ₂ O	P ₂ O	H ₂ O _T
4776-4792 (4)	.06	1.27	0.1	1.6	63.6	17.1	1.2	0.6	9.1	0.1	.17	5.2
4797-4816 (5)	.06	.54	0.1	2.8	62.3	21.4	1.7	1.3	3.2	0.1	.08	5.9
4821-4835 (5)	.06	.79	0.2	5.1	63.0	17.2	2.7	2.0	5.3	0.1	.10	3.9
4840-4854 (4)	.04	.35	0.1	3.8	72.2	15.8	2.0	1.4	2.8	0.1	.02	2.8
4859-4873 (4)	.02	.24	0.1	2.6	78.7	12.0	1.6	0.4	3.6	0.1	.08	2.5
4877-4888 (6)	.02	.14	0.1	4.4	71.6	16.0	2.4	0.5	1.8	0.1	.03	3.1
4910-4925 (4)	.09	.72	2.4	4.4	60.4	15.7	5.2	4.0	1.7	1.3	.16	3.5
Proterozoic Average	.09	.64	3.2	3.4	65.8	15.5	2.0	3.5	1.3	3.4	.18	0.8
Eade & Fahrig, 1971												
CHEMISTRY OF SAMPLES ANALYZED FOR Rb/Sr DATING												
1 - 4915	.06	.61	3.0	3.1	62.3	16.9	4.5	4.4	2.0	3.2	.13	1.8
2 - 4920	.06	.58	1.3	4.7	67.8	13.7	3.2	4.5	1.4	1.8	.09	2.2
3 - 4878**	.02	.06	0.1	4.6	68.5	17.6	1.8	0.1	1.9	0.1	.01	3.9
4 - 4797	.04	.38	0.0	1.2	51.2	26.9	0.7	0.5	2.4	0.1	.10	8.8
5 - 4910	.14	1.24	3.3	5.5	53.7	16.5	7.3	4.0	1.1	0.1	.17	4.8
6 - 4816	.05	.48	0.2	4.6	63.7	14.9	2.6	1.5	4.2	0.1	.08	5.7

* Note that number 3 at 4878 feet was not used in calculating the isochron age.

** The numbers under the depth column give the intervals in feet represented by the chemical analysis and the numbers in brackets indicate the number of analyses averaged for each interval. The bottom set of analyses are for the specific samples used in the isochron and numbers on the left are those used to identify samples on the isochron (Fig. 1).

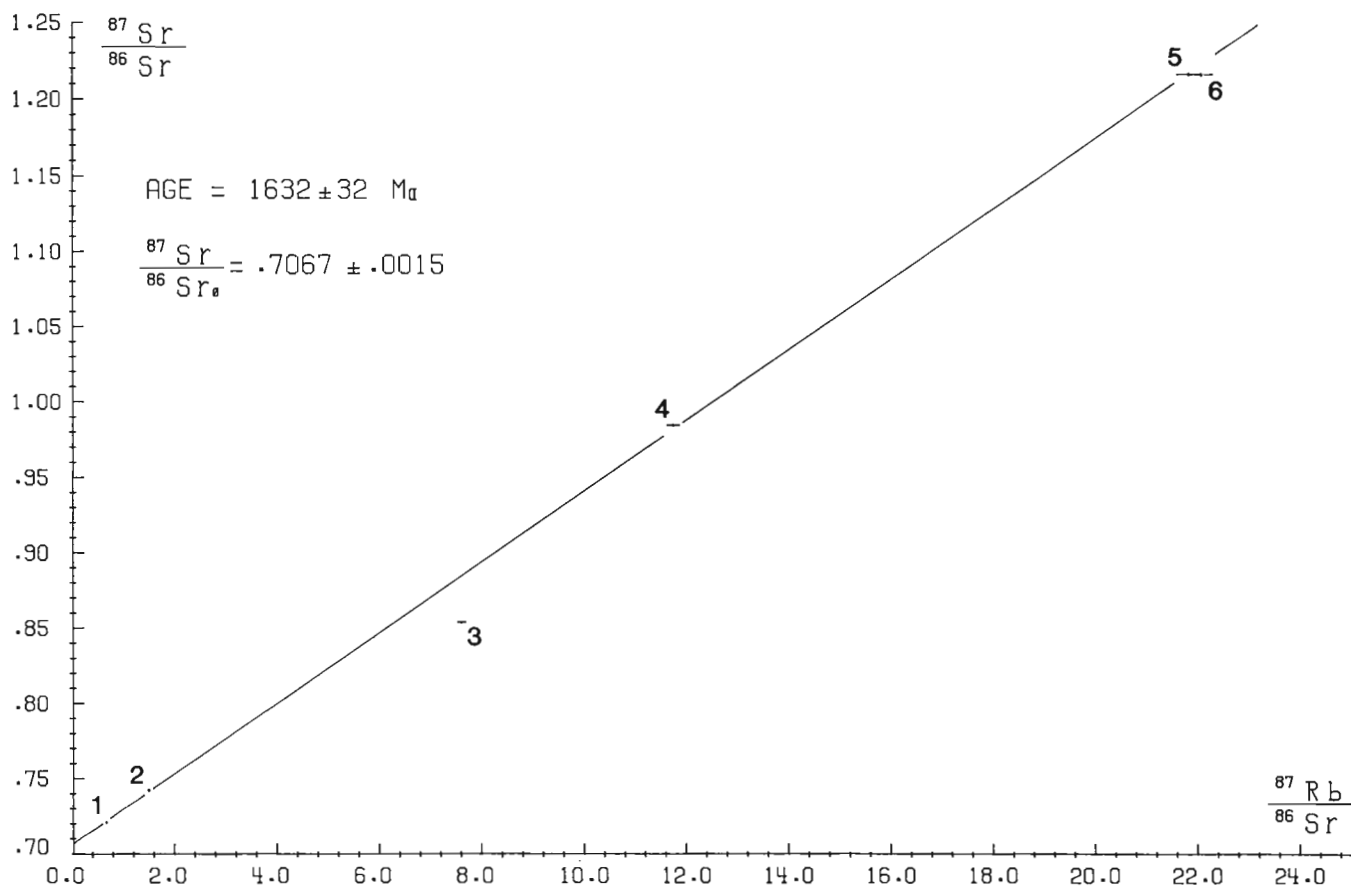


Figure 1. Rb-Sr isochron diagram, weathered pre-Athabasca Formation basement gneiss, northern Saskatchewan.

Chemistry of the Basement

The determination of the age of sedimentation in basins such as the Athabasca has been, and remains, a serious problem. In some cases strontium isotopes are apparently homogenized in fine grained sediments during diagenesis so that source-area effects are negligible. In a case such as this where sedimentation has been rapid (Fraser et al., 1970), diagenesis may closely follow sedimentation and Rb-Sr whole-rock ages may provide a good estimate of the age of sedimentation. For example, Moorbath (1969) determined the probable age of Torridonian sedimentation with whole-rock Rb-Sr work on shales and siltstones. This corroborated the presence of a significant unconformity which had been indicated by earlier paleomagnetic work (Irving and Runcorn, 1957).

In a similar way, it seems reasonable that strontium homogenization could take place in deeply weathered basement rocks which would be susceptible to diagenetic changes during and after sedimentation and might then give a reasonable estimate of the age of sedimentation. With this in mind, 31 samples from basement core of the Rump Lake drillhole at $106^{\circ}35'W$, $58^{\circ}19'N$ (Lerand, 1980) were chemically analyzed. These analyses represent 149 feet (46 m) of core and from the 31 analyses, 7 averages were calculated; each derived from 4 to 6 samples and representing 11 to 19 feet (3.4 to 5.8 m) of core (Table 1). These averages smooth out the effects on chemistry of the rapid, large scale variation in primary lithologies of the gneisses.

The overall chemical results are much as expected in a deeply weathered profile of complex gneiss. Calcium and sodium are dramatically reduced in the upper 112 feet (34 m) of core; either one or the other is absent in ten samples from this zone. Potassium is significantly lower in the higher zones, and the ferric/ferrous ratio is significantly higher. Water is also higher in the zone beneath the unconformity.

It is interesting that TiO_2 is high in the zone below the unconformity and a highly hematitized sample from immediately below the unconformity contained 2.47 per cent TiO_2 and 18.4 per cent Fe_2O_3 . The high MgO value in the lowest zone is due partly to one particularly magnesia-rich sample but could reflect enrichment in this element during the weathering process.

These basement analyses are compared (Table 1) with those obtained for present-day, unweathered, surface crystalline rocks of comparable age for the Canadian Shield (Eade and Fahrig, 1971). The Proterozoic average is notably higher in CaO , K_2O , and Na_2O , and lower in H_2O . The large amount of water found in even the lowest sub-Athabasca zone suggests that hydration during weathering extended to below this level.

The chemistry of these basement rocks indicates that they have undergone significant element re-arrangement during pre-Athabasca weathering. The least altered, of course, is the lowest zone, but even this may have been altered enough to make it susceptible to Sr homogenization during diagenetic changes.

Table 2
Analytical data, whole-rock samples,
pre-Athabasca Formation gneiss

Sample no.	Rb ppm	Sr ppm	$^{87}\text{Sr}/^{86}\text{Sr}$ spiked	$^{87}\text{Rb}/^{86}\text{Sr}$
1	94.63	427.7	0.7208 ± 0.0011	0.640 ± 0.013
2	126.5	249.4	0.7422 ± 0.0011	1.467 ± 0.029
3	61.44	23.43	0.8539 ± 0.0013	7.585 ± 0.152
4	87.40	21.55	0.9843 ± 0.0015	11.73 ± 0.23
5	242.8	32.20	1.2174 ± 0.0018	21.81 ± 0.44
6	179.7	23.57	1.2173 ± 0.0019	22.05 ± 0.44

Analytical Techniques, Data and Results

Analytical procedures for Rb-Sr isotopic analyses of whole-rock samples were based on those described by Wanless and Loveridge (1972) with the exception that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were obtained only from spiked Sr analyses in this study; no unspiked Sr measurements were performed.

Six samples were analyzed isotopically and the results are listed in Table 2 and plotted in Figure 1. Five of the six sample points are collinear yielding an isochron age of 1632 ± 32 Ma, initial $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7067 ± 0.0015 and MSWD 1.77. Sample point number 3 falls well below the line defined by the other five points. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7067 ± 0.0015 is higher than that usually seen for rocks of this age derived directly from the mantle and supports our interpretation of the age as that of a secondary event.

Interpretation

Five of the six samples of weathered basement yield an excellent isochron age of 1632 ± 32 Ma. The possibility exists that this reflects a homogenizing event older than the pre-Athabasca weathering; for example, the intense regional metamorphic and/or igneous activity indicated elsewhere in the western Churchill Province. This is unlikely because the age is so much younger than other Rb-Sr ages obtained for gneisses in this region (Baadsgaard and Godfrey, 1972). It is also conceivable that the age is that of the basement gneiss, overprinted as a result of weathering and diagenesis. The collinearity of the five points makes this "mixed" age unlikely. We believe the best possibility is that this age of 1632 ± 32 Ma reflects homogenization during weathering or diagenesis. This age is therefore a minimum for the age of Athabasca sedimentation and, since the Athabasca was probably rapidly deposited, is probably close to the age of sedimentation.

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7. ZIRCON AND MONAZITE AGE OF A GRANITIC CLAST IN TOBY CONGLOMERATE (WINDERMERE SUPERGROUP), CANAL FLATS, BRITISH COLUMBIA

W.D. Loveridge, G.B. Leech, R.D. Stevens, and R.W. Sullivan

Loveridge, W.D., Leech, G.B., Stevens, R.D., and Sullivan, R.W., *Zircon and monazite age of a granitic clast in Toby conglomerate (Windermere Supergroup), Canal Flats, British Columbia; in Rb-Sr and U-Pb Isotopic Age Studies, Report 4, in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 131-134, 1981.*

Abstract

U-Pb age measurements on zircon and monazite from a granitic pebble collected from the Toby Formation, which lies at the base of the Windermere Supergroup, yielded a concordia intercept age of 1781 ± 9 Ma. The provenance of these granitic pebbles is not immediately apparent; it probably lies to the west or southwest, perhaps within the Shuswap Metamorphic Complex.

Geological Setting

The Toby Formation is the basal unit of the Windermere Supergroup (Hadrynian) and rests unconformably on various formations of the Purcell Supergroup (Helikian). The Horsethief Creek Group overlies the Toby conformably. The Toby Formation is predominantly a diamictite, consisting typically of pebbly or conglomeratic mudstones, with subordinate conglomerate, sandstone and argillite. The clasts were derived mainly from Purcell formations. One of the distinctive features of the Toby Formation is the local variation in amount and lithology of clasts and the way its content varies in accord with the underlying strata. This, together with the large size and angularity of some of the clasts and the typically poor sorting and local changes in thickness, suggests rapid deposition and local provenance. The depositional conditions are equivocal. Aalto (1971) favoured marine glacial sedimentation and Reesor (1973) a nonglacial origin.

In contrast to the undoubted Purcell provenance of most of the Toby fragments, there are rare occurrences of granitic clasts for which no local source is known (Leech, 1959; Reesor, 1973). A concentration of these clasts occurs in the western part of the floor of the Rocky Mountain Trench, 4 km southwest of Canal Flats. The material reported here is from that occurrence (lat. $50^{\circ}08'N$, long. $115^{\circ}51'W$). Another local concentration, in the west wall of the Trench 10 km south-southwest of Canal Flats, contains boulders up to about 90 cm across. Granitic clasts also occur locally in Toby Formation along the west edge of the Trench at sites 10.5 km and (Reesor, 1973) 34 km northwest of Canal Flats.

At the locality 4 km southwest of Canal Flats several outcrops contain granitic clasts in Toby Formation almost immediately below conformable Horsethief Creek strata. One exhibits a layer-lens of coarse clasts, a high proportion of which are granitic, in otherwise more mudstone material (Figure 1). Many of the clasts in this diamictite, sedimentary ones in particular, are flattened and aligned with cleavage (Figure 2) in a manner characteristic of the Toby Formation in this area. Unfortunately this splendid and important outcrop was vandalized by a collector of granitic material soon after it was reported.

The strata of the entire region are allochthonous. The Toby strata, folded, penetratively cleaved, and metamorphosed to the chlorite zone of the greenschist facies, are parts of thrust fault slices. Thrust faults, including the major Hall Lake fault in the Canal Flats area, are cut by the White Creek batholith intruded between 130 Ma and 110 Ma (Wanless et al., 1968, recalculated using decay constants recommended by Steiger and Jaeger, 1977). Transport of the Toby and other strata from a depositional site that must have been west of the present Kootenay Arc (Price, 1981) commenced before that time. Reesor (1973) concluded that a regional greenschist metamorphism also predates intrusions such as the White Creek batholith.

Lithology

The granitic clasts at the locality 4 km from Canal Flats are too altered for K-Ar age determination but a pebble yielded the zircon and monazite whose age is reported here. The main components of the rock as estimated in thin section are: K feldspar (orthoclase, perthitic microcline) 40%; albite, 30%; quartz, 20%; chlorite, 5%; fine grained



Figure 1. Toby Formation 4 km southwest of Canal Flats. Many of the lightest coloured clasts are granitic. GSC 151064.



Figure 2. Closer view of Toby conglomerate. The fractured granitic boulder (hammer head) appears in lower right of Figure 1. GSC 151062.

Table 1
Analytical data, zircon and monazite fractions,
Toby conglomerate, Canal Flats, British Columbia

Fraction	1	2	3	4
Fraction size, μm	unsized	unsized	unsized	unsized
Mineral	monazite	monazite	zircon	zircon
Weight, mg	n.a. ¹	0.04 ²	0.11 ²	n.a. ¹
Total Pb, ng	60.71	375.0	26.77	14.74
Pb Blank, %	2.6	2.0	7.0	10.5
Observed $^{206}\text{Pb}/^{204}\text{Pb}$	442.6	839.4	115.1	94.12
³ Abundances ^{204}Pb	0.07386	0.02416	0.6861	0.7752
($^{206}\text{Pb} = 100$) ^{207}Pb	11.837	11.226	18.653	19.417
^{208}Pb	326.5	226.1	90.02	99.54
Radiogenic Pb, ppm	n.a. ¹	30780 ²	670 ²	n.a. ¹
" " %	98.9	99.5	78.0	76.3
Uranium, ppm	n.a. ¹	36200 ²	6560 ²	n.a. ¹
Atomic ratios $^{206}\text{Pb}/^{238}\text{U}$	0.30750	0.29174	0.064786	0.052216
$^{207}\text{Pb}/^{235}\text{U}$	4.5932	4.3841	0.81652	0.61987
$^{207}\text{Pb}/^{206}\text{Pb}$	0.10833	0.10898	0.091402	0.086092
Ages, Ma $^{206}\text{Pb}/^{238}\text{U}$	1728	1650	405	328
$^{207}\text{Pb}/^{235}\text{U}$	1748	1709	606	490
$^{207}\text{Pb}/^{206}\text{Pb}$	1772	1782	1455	1340

¹n.a. = not available; see text

²Analytical uncertainty high due to small sample size; see text

³Corrected for Pb blank

"white mica", 5%. Minor remnants of biotite occur in the chlorite. Some relatively large (1 mm) muscovite may be primary; most of these grains seem to be embayed and replaced by, or intergrown with, quartz and albite. The (albitized) plagioclase grains are subhedral, <5 mm, exhibit slight relict zoning and are replaced, especially internally, by fine grained white mica. Microcline and orthoclase grains, <5 mm, and the quartz grains, <3 mm, are anhedral. Microcline surrounds and replaces plagioclase.

Analytical Procedures and Results

The pebble was cleaned of all matrix material, and two thirds of it was crushed and processed in small stages through methylene iodide (SG 3.3) and the resulting small concentrate of heavy minerals was hand picked to produce subconcentrates of monazite with minor zircon (fraction 2) and zircon with minor monazite (fraction 3). The remaining one third of the pebble was subsequently subjected to the same procedure and even more carefully hand picked to yield 30 crystals of zircon (fraction 4) and 20 crystals of monazite (fraction 1). A description of zircon and monazite morphology is presented in the Appendix.

Procedures for the isotopic analysis of Pb and U are discussed in Sullivan and Loveridge (1980). Analytical results are presented in Table 1 and displayed on a concordia diagram, Figure 3. The two zircon and two monazite fractions yield four sample points, collinear within analytical uncertainty, defining a chord which intersects the concordia curve at 1781 ± 9 and 140 ± 18 Ma. The two zircon points, when treated separately, give results in excellent agreement

with the zircon-monazite chord, yielding concordia intercepts of 1782 and 141 Ma. The monazite results alone yield a chord with an upper intercept of 1767 Ma and a negative lower intersection which is geologically uninterpretable. The average of the two monazite $^{207}\text{Pb}/^{206}\text{Pb}$ ages is 1777 Ma, in satisfactory agreement with the 1781 ± 9 Ma age of the four point chord.

This was the first time that such small sample sizes (20 grains of monazite in fraction 1, and 30 grains of zircon in fraction 4) had been processed in the Geological Survey of Canada geochronology laboratories. For this reason, the procedures were developmental and, due to the technique employed, satisfactory weights for these two fractions were not obtained. Therefore U and Pb concentrations of fractions 1 and 4 could not be calculated. Although fractions 2 and 3 were considerably larger than fractions 1 and 4, their measured weights were only 0.04 and 0.11 milligrams respectively. Analytical uncertainties in obtaining these weights, due to balance limitations, are estimated to be about 0.02 milligrams, causing uncertainties of about $\pm 20\%$ in the stated U and Pb concentrations of fraction 3 zircon, and $+100$, -33% in these quantities for fraction 2 monazite.

The two monazite points are reasonably close to being concordant, showing relative Pb loss of less than 10%, whereas the two zircon points are highly discordant with relative Pb loss greater than 70%. Despite the high analytical uncertainty associated with the measurement of the U concentrations, it is apparent that both the zircon and the monazite have relatively high U contents. The dark colour of the zircon crystals together with the high U content suggests extensive metamictization and consequently high loss of radiogenic Pb. However the remarkably high U content of the monazite has not resulted in pigmentation of the monazite crystals or comparable Pb loss.

We interpret the upper intercept age, 1781 ± 6 Ma, as the primary age of the igneous precursor of the granitic clasts. The lower intercept age, 141 ± 18 Ma, is in agreement, within analytical uncertainty, with the age limits associated with the emplacement of the White Creek batholith and may reflect this event or penecontemporaneous activity, if it has geological significance at all.

Provenance of the Granitic Clasts

The granitic clasts are lithologically markedly unlike the Hellroaring Creek stock, 70 km south of Canal Flats. That is the only Precambrian granitic intrusion in the Purcell anticlinorium, whose west flank and north end the Toby Formation outlines. This stock yielded an Rb-Sr whole rock isochron age of 1220 ± 50 Ma (recalculated using ^{87}Rb decay constant of $1.42 \times 10^{-11} \text{a}^{-1}$) with an unusually high $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio of 0.81 ± 0.06 (Ryan and Blenkinsop, 1971). Neither the lithology nor the apparent age of the Hellroaring Creek stock accord with it being the source of the Toby Formation clast that yielded a concordia intercept age of 1781 ± 9 Ma.

Alternative sources are the (now covered) Canadian Shield basement to the east and granitic rocks to the west in what is now the Shuswap Complex.

The age of the granite from which the Toby pebble was derived is compatible with its origin in a Hudsonian Shield terrane. Furthermore, the granitic debris is in the

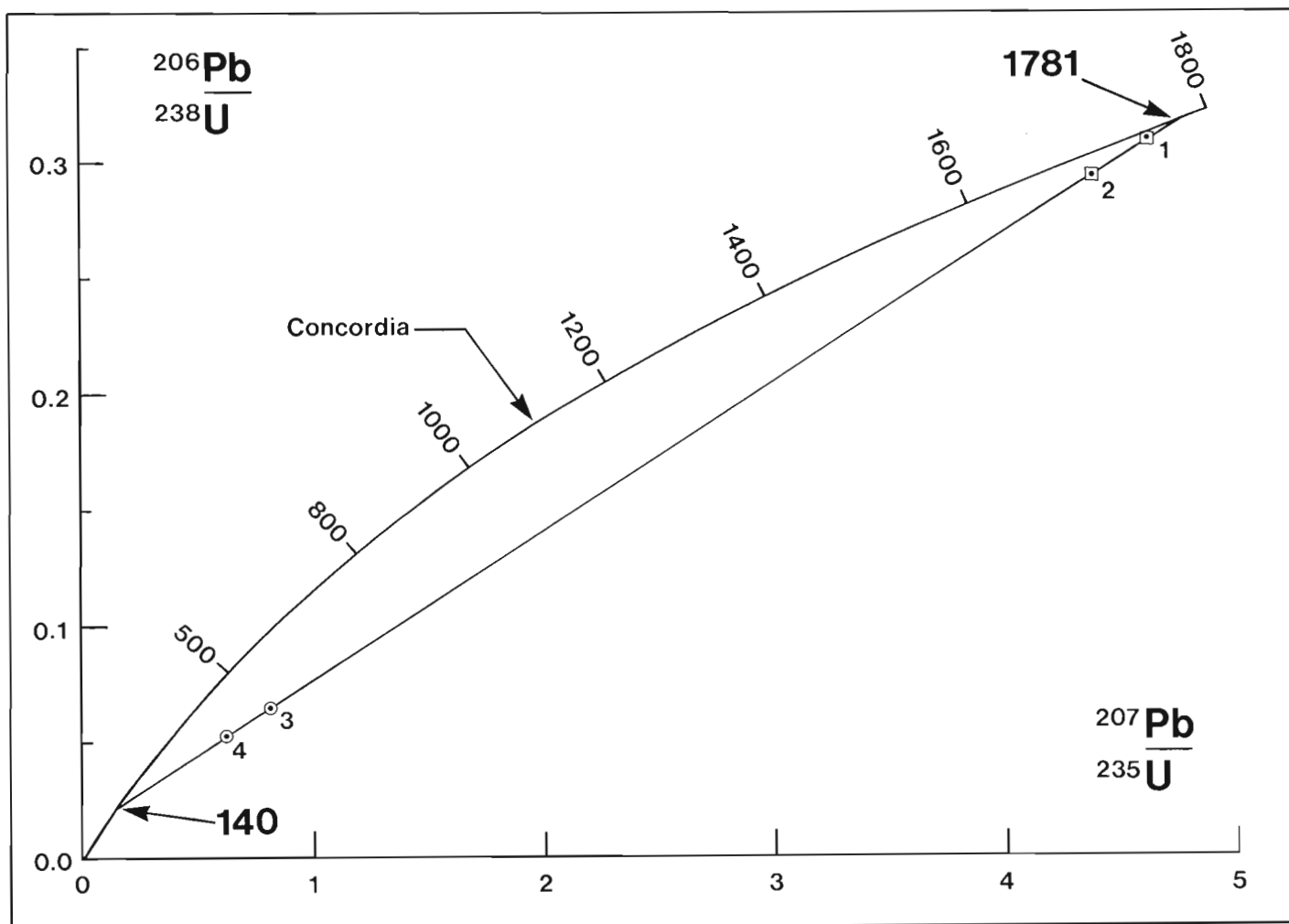


Figure 3. Concordia diagram showing the results of analyses of two zircon fractions (circles) and two monazite fractions (squares) from the Toby Formation, Canal Flats, British Columbia.

easternmost Toby exposures, closest to the Shield. If the Toby Formation has glacial affinities the presence of granitic boulders and large cobbles could be accounted for by ice-rafting from the Shield. On the other hand, the apparently limited distribution of the granitic clasts (they have been reported nowhere in the Toby Formation outside the Canal Flats – Windermere Lake district), and their occurrence in pronounced local concentrations within that district, suggest a relatively restricted source rather than the widespread granitic terrane of the Shield. As noted earlier, the bulk of the Toby clasts are Purcell Supergroup fragments suggestive of a relatively local source with high relief. The apparently limited distribution and the coarseness of the granitic clasts accord with such provenance.

The strata are allochthonous and the site of Toby deposition was far west of Canal Flats, somewhere west of the present Kootenay Arc according to Price's palinspastic reconstructions (Price, 1981). In the Shuswap Metamorphic Complex, which now occupies a position west of the Kootenay Arc, the core of the Thor-Odin gneiss dome has yielded a zircon age of 1960 ± 35 Ma (Wanless and Reesor, 1975). Other domes may have cores of similar age (op. cit.). We conclude that rocks which now make up the cores of these Mesozoic structures are a likely source of the granitic clasts in the Toby Formation.

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APPENDIX

Description of zircon and monazite concentrates as analyzed

Monazite, fractions 1 and 2

The monazite crystals were very pale red and orange to colourless, and more or less spherical in outline although irregular fragments were also observed. Grain diameters generally were ca. 50 to 75 μm .

Zircon, fractions 3 and 4

The zircon crystals were very dark, some 75% being essentially opaque and appearing to have a black coating on them. The remainder were very dark purple and slightly transparent. Crystal form was generally rounded euhedral with elongation ratios of about 3:1 on average. Crystal lengths ranged from ca. 150 to 300 μm .

8. AGE AND GEOLOGICAL SIGNIFICANCE OF A TONALITE PEGMATITE FROM EAST-CENTRAL BAFFIN ISLAND

J. R. Henderson and W. D. Loveridge

Henderson, J.R. and Loveridge, W.D., Age and geological significance of a tonalite pegmatite from east-central Baffin Island; in *Rb-Sr and U-Pb Isotopic Age Studies, Report 4, in Current Research, Part C, Geological Survey of Canada, Paper 81-1C, p. 135-137, 1981.*

Abstract

The age of the most recent penetrative deformation in the Home Bay region of east-central Baffin Island was found to be $1806 \pm 15/-8$ Ma, by the U-Pb method on zircon from a synkinematic tonalite pegmatite. This age is younger than two K-Ar cooling ages on hornblende obtained from the host amphibolite (1927 ± 47 Ma and 1876 ± 45 Ma).

Introduction

In order to establish the age of the most recent penetrative deformation in east-central Baffin Island (F₄, Henderson and Tippet, 1980) we obtained a U-Pb age on zircon from a synkinematic tonalite pegmatite that had crystallized in an extension fracture within an amphibolite layer. Figure 1 shows the sample locality and the regional geology. Figure 2 shows the relationship of the pegmatites to the enclosing rocks. In the Home Bay region (Fig. 1) the F₄ folding is characterized by upright, open to tight folds with northwest striking axial surfaces and gently inclined axes. The F₄ folds are developed north of a line extending east from Cape Hooper. The folds may have resulted from dextral rotation of east-northeast striking F₃ folds in a wide transcurrent shear zone. The pegmatite, composed of plagioclase (An₂₂), quartz and biotite, is little deformed compared with the foliated and lineated granodiorite gneiss composing the bulk of the country rock.

Analytical Procedures and Results

Techniques for the concentration of zircon and the extraction and analysis of lead and uranium are described in Sullivan and Loveridge (1980). Analytical results are listed in Table 1 and displayed on a concordia diagram (Fig. 3). A description of the zircon morphology of the fractions as analyzed is presented in the Appendix. The three fractions analyzed yielded data points, collinear within analytical uncertainty, defining a chord which cuts the concordia curve at 1806^{+15}_{-8} and 47^{+509}_{-47} Ma. K-Ar ages on hornblende concentrates from the host amphibolite (1927 ± 47 Ma) and from an amphibolite inclusion in the pegmatite (1876 ± 45 Ma) are documented in Table 2.

Interpretation

We interpret the upper concordia intercept age of the zircon, 1806^{+15}_{-8} Ma, as the time of crystallization of the pegmatite in an extending crack in the amphibolite during F₄. We suggest the older K-Ar hornblende ages from the host amphibolite may indicate that the rocks had cooled below 500°C at least 10 Ma before the pegmatite crystallized. We also suggest that if biotite tonalite magmas crystallize above 650°C as indicated by the experimental studies of Winkler (1979), it would appear that the pegmatite dated in this study crystallized from an aqueous solution rather than a silicate magma.

Table 1

Analytical data, zircon fractions from tonalite pegmatite, east-central Baffin Island

Fraction number	1	2	3
Fraction size	+149	-149+105	-149+105
Magnetic or non magnetic	-	n.m.	mag.
Weight, mg	4.68	3.61	2.44
Total Pb, ng	147.2	113.8	98.89
Pb Blank, %	1.0	1.3	1.5
Observed $^{206}\text{Pb}/^{204}\text{Pb}$	5123	3535	1718
*Abundances ($^{206}\text{Pb}=100$)			
^{204}Pb	0.00213	0.00559	0.03190
^{207}Pb	11.066	11.096	11.468
^{208}Pb	12.855	13.782	14.796
Radiogenic Pb, ppm	103.8	102.6	109.3
%	99.88	99.70	98.30
Uranium, ppm	309.2	307.2	330.4
Atomic ratios			
$^{206}\text{Pb}/^{238}\text{U}$	0.31520	0.31159	0.30828
$^{207}\text{Pb}/^{235}\text{U}$	4.7969	4.7348	4.6910
$^{207}\text{Pb}/^{206}\text{Pb}$	0.11037	0.11020	0.11035
Ages, Ma			
$^{206}\text{Pb}/^{238}\text{U}$	1766	1749	1732
$^{207}\text{Pb}/^{235}\text{U}$	1784	1773	1766
$^{207}\text{Pb}/^{206}\text{Pb}$	1805	1803	1805

*Corrected for Pb blank

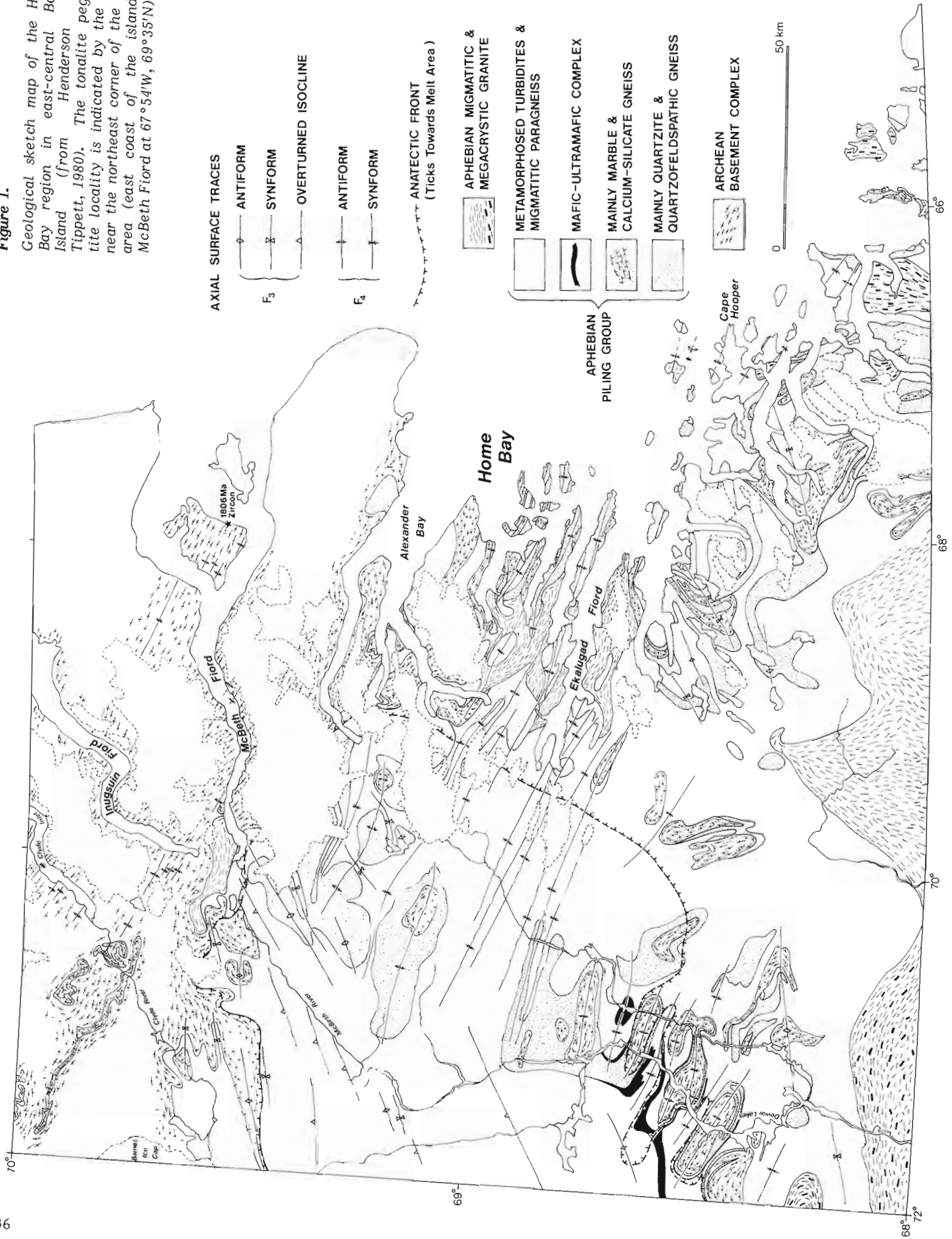
Table 2

K-Ar age determinations

Hornblende concentrate from host amphibolite: Age = 1927 ± 47 Ma K = 1.19%, $^{40}\text{Ar}/^{40}\text{K}$ = 0.2001, radiogenic ^{40}Ar = 97.5% Concentrate: Clean, fresh, unaltered, pleochroic light brown to dark green hornblende with no contamination.
Hornblende concentrate from amphibolite inclusion in pegmatite: Age = 1876 ± 45 Ma K = 1.22%, $^{40}\text{Ar}/^{40}\text{K}$ = 0.1918, radiogenic ^{40}Ar = 94.9% Concentrate: Clean, fresh, unaltered, pleochroic brown to dark green hornblende with no contamination.

Figure 1.

Geological sketch map of the Home Bay region in east-central Baffin Island (from Henderson and Tippett, 1980). The tonalite pegmatite locality is indicated by the star near the northeast corner of the map area (east coast of the island in McBeth Fiord at 67°54'W, 69°35'N).



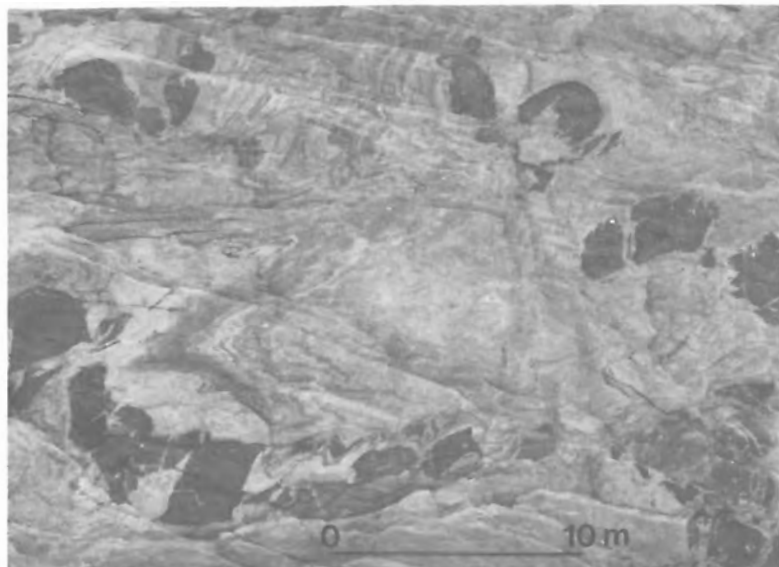


Figure 2. Vertical cliff exposure of boudinaged amphibolite (black) with white pegmatite in extension fractures. Granodiorite gneiss composes the bulk of the country rock. The tonalite pegmatite dated in this study was collected from an analogous setting nearby.

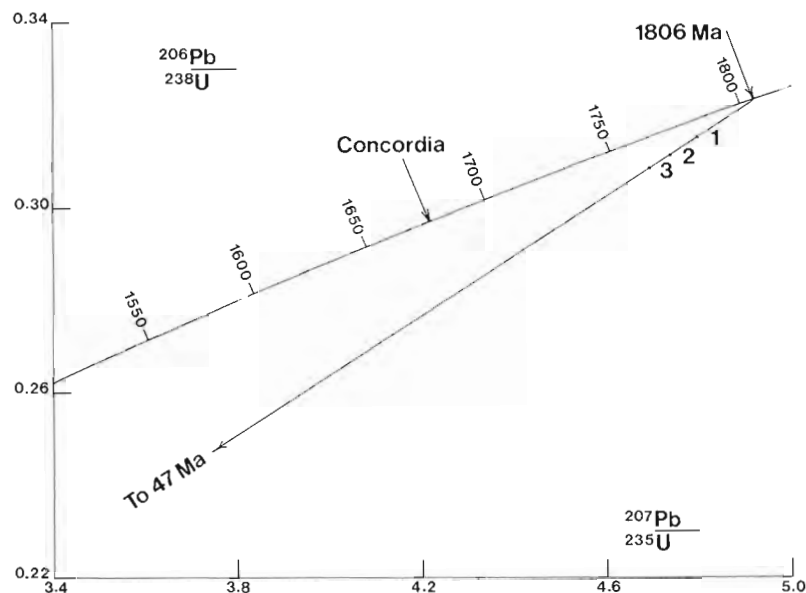


Figure 3. Concordia diagram showing the results of analyses of zircon concentrates from tonalite pegmatite, east-central Baffin Island.

APPENDIX

Description of zircon concentrates as analyzed, tonalite pegmatite from east-central Baffin Island.

Fraction 1, +149 μ m: The original, somewhat impure concentrate was sieved to separate the +149 μ m fraction which was then hand picked to approximately 100 per cent pure, clear, very pale purple zircon grains. Most of the crystals were euhedral, equidimensional to only moderately elongated (2.3:1), and of generally rounded appearance. All grains were very pure and free of inclusions. No zoning was observed.

Fractions 2 and 3, -149 +105 μ m: The -149 +105 μ m fraction was separated into relatively magnetic and nonmagnetic subfractions by repeated passes through a Frantz Isodynamic separator. Both subfractions were hand picked to yield approximately 100 per cent pure zircon samples. The nonmagnetic split, Fraction 2, consisted of very clear essentially colourless equidimensional to moderately elongated crystals some of which contained one or more fine crystalline inclusions and very small "dusty" central spots.

This description applies equally to Fraction 3, the more magnetic split, except for the presence of a tinge of very pale purple.

We thank R.D. Stevens for supplying the zircon descriptions.

Conversely, the older K-Ar hornblende ages may be the result of excess argon in the hornblende. Further geochronological work is planned to discriminate between these two possibilities.

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