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MID-PROTEROZOIC CHEMICAL WEATHERING, REGOLITH, AND
SILCRETE IN THE THELON BASIN,
NORTHWEST TERRITORIES

by

© JEFFREY ROBERT CHIARENZELLI, Hns. B.Sc.

A thesis submitted to the Faculty of
Graduate Studies in partial fulfilment
of the requirements for the degree of
Master of Science

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Ottawa, Ontario

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ABSTRACT

Chemical weathering, interrupted by erosion and Pitz Formation volcanism, created several regoliths and silcretes in the Thelon Basin between 1.8-1.65Ga. The best developed (>50m thick) regoliths are saprolitic, mineralogically zoned, and display Fe-enrichment similar to modern lateritic profiles. Diagenetic changes include dehydration and ordering of clay minerals, replacement of metastable phases, and illitization of kaolin. The clay mineralogy suggests burial temperatures of 150-200°C. The regolith formed during hot, humid climatic conditions which became evaporitic during Pitz Formation volcanism. Mineralogical changes which occurred as a result of Pleistocene weathering were documented in two drill cores.

Polygenetic silcretes developed in truncated regolith profiles, Pitz rhyolites, and Thelon Formation sediments. Spatial association of silicification with Pitz volcanic rocks suggests that weathering of rhyolites provided most of the silica. Clasts of silcrete and regolith in the Thelon Formation establish a Proterozoic age for the weathering and silcretization.

The sub-Thelon and sub-Athabasca regolith display similar mineralogical zones, thickness, and diagenetic alteration; and probably formed at the same time under equivalent conditions.

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I would also like to acknowledge the following people at Carleton University: D. Marshall and R. Taylor for excellent technical help; R. Conlon and Dr. G. Chao for x-ray diffraction assistance; P. Jones for microprobe work; and J. Hogg, S. Hill, and E. Lambton for administrative assistance.

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INTRODUCTION

PHYSIOGRAPHY

The study area (Figure 0-1) is located just below the Arctic Circle, well within the belt of continuous permafrost. The area is subject to sub-arctic climatic conditions; the larger lakes do not break up until late June or early July. Although fresh water abounds in numerous lakes and rivers (making up approximately 35% of the land surface), the annual precipitation is sufficiently low to create "Arctic desert" conditions. Mid-summer temperatures range from a few degrees below freezing to greater than 30°C. Sunlight lasts for more than twenty hours per day throughout June and July. The land surface shows numerous relics of the last ice age: till plains, drumlins, eskers, moraines, felsenmeer, erratic boulders, patterned ground, raised beaches, glacial striae and polished outcrops. Because the area is above the northern extent of the tree line, vegetation consists predominantly of grasses, low shrubs, moss, lichen and wild flowers. Caribou, fox, wolves, arctic hares, hawks, falcons and waterfowl were frequently seen on traverse. Signs of a sparse, but persistent, past Inuit inhabitation are common. Their encampments and stone inukoks are particularly abundant on hills that overlook major waterways.

The Aberdeen Lake (NTS 66B) map area is dominated by Aberdeen Lake, part of the Thelon River system, which extends east-west centrally across the map area. Another major river Figure 0-1. Location map of the study area, Thelon Basin,

Topographic map of the Baker Lake Basin area in the Northwest Territories, Canada. The map shows the Baker Lake Basin and Thelon Basin. Key features include Baker Lake, MacQuoid Lake, Thirty Mile Lake, Tebesiuk Lake, Tulemalu Lake, Kamliukak Lake, Dubawnt Lake, and several smaller lakes like Amer, Deep Rose, and Schultz. A 'STUDY AREA' is outlined around Baker Lake and the surrounding lakes. The map includes a grid with coordinates (e.g., 66°N, 106°W) and a scale of 1:2,500,000. The map is labeled 'National Topographic System Map Sheets'.

system, the Dubawnt River, joins the Thelon system west of Aberdeen Lake in the Beverly Lake map area. Both the southern and northern parts of the map area are relatively flat, and outcrops are scarce. To the north of Aberdeen Lake, flat-lying sandstone and minor conglomerate of the upper Thelon Formation crop out almost exclusively; the exception is a small outlier of Ordovician limestone. To the south, Archean metasediments and gneisses are exposed in widely separated outcrops. The Marjorie Hills run approximately east-west, parallel to the southern shore of Aberdeen Lake. They provide the major relief in the area, rising to a maximum height of 275m near the western edge of the map area. Rhyolitic volcanic rocks of the Pitz Formation crop out there, as well as east of Marjorie Lake. Several large post-Thelon diabase dikes, belonging to the north-northwest-trending Mackenzie swarm, crop out in several parts of the map area.

The Schultz Lake (66A) map area is dominated by Schultz Lake, in the northwest quadrant of the map sheet. About half the size of Aberdeen Lake, Schultz Lake also is part of the Thelon River system.

The Innuit village of Baker Lake with a population of 1500, is on the northwestern shore of the lake bearing the same name. A gravel runway provides access from Winnipeg and Yellowknife. Air traffic is fairly regular throughout the summer. Accommodation is generally available at one of two tourist lodges. Most supplies were flown into the area from Churchill, Manitoba via Baker Lake. The Thelon River flows into Baker Lake southwest of the village. Baker Lake is only

2.5m above sea level; the eastern end is slightly saline.

A series of east-west step faults dominate the central part of the map area, producing scarps up to 50m high. From the northwest corner of the Schultz Lake map area, a faulted triangle of basement rocks with a rugged but low topography extends into the adjacent Aberdeen Lake map area. The northern and southern parts of the map area are dominated by low irregular hills, with a maximum elevation of 250m.

PREVIOUS GEOLOGICAL INVESTIGATION

During helicopter reconnaissance work of operations Keewatin, Baker, and Thelon in 1952, 1954 and 1955 respectively, the Dubawnt Group was named and its areal extent mapped (Lord, 1953; Wright, 1955, 1957, 1967). Wright (1955) divided the Dubawnt Group into four subdivisions: 1) the Lower Dubawnt Conglomerate; 2) the Dubawnt Sandstone; 3) the Dubawnt Porphyries and 4) the Upper Dubawnt Conglomerate. During 1963, 1965 and 1966, Donaldson continued field work on the Dubawnt Group, studying the stratigraphy and sedimentology of the sedimentary units (Donaldson, 1965, 1966a, 1966b, 1967a, 1967b, 1969). Donaldson (1965) separated the Dubawnt Group into six formations: 1) South Channel; 2) Kazan; 3) Martell Syenite; 4) Christopher Island; 5) Pitz and 6) Thelon. Subsequent mapping led to addition of the Angikuni Formation at the base of the Dubawnt Group (Blake, 1980), and of the Kunwak Formation, representative of the magmatic hiatus between the Christopher Island and Pitz Formations (LeCheminant et al, 1979). Recent work has led to the informal recognition of a lower unit of the

Thelon Formation in the Aberdeen Lake map area and a sequence of sediments beneath Pitz Formation volcanic rocks (See Table 0-1) (LeCheminant et al, 1983; this study).

Additional mapping at 1:250,000 scale in the Thelon and Baker Lake Basins was undertaken in the Baker Lake (56D) map area (Schau and Hulbert, 1977; Schau and Ashton, 1979; Heywood and Schau, 1981; Schau et al, 1982), Kamilukuak Lake (65K) map area (Tella and Eade, 1980; Tella et al, 1981), MacQuoid Lake (55M) map area (Reinhardt and Chandler, 1973; LeCheminant et al, 1976; LeCheminant et al, 1977), Tebesjuak Lake (65O) map area (LeCheminant et al, 1979; LeCheminant et al, 1980; LeCheminant et al, 1981), Thirty Mile Lake (65P) map area (LeCheminant et al, 1976; LeCheminant et al, 1977; LeCheminant et al, 1979), Tulemalu Lake (65J) map area (Eade, 1976; Eade and Blake, 1977; Blake, 1980). A compilation of much of this work is shown in Figure 0-2. During the 1982 field season, two Geological Survey of Canada field parties worked in the Thelon Basin at Aberdeen Lake (66B) and Deep Rose Lake (66G) map areas (LeCheminant et al, 1983; Tella et al, 1983). A stratigraphic and sedimentological study of the lower unit of the Thelon Formation and the sub-Thelon regolith was started by the author in conjunction with G.S.C. operations at Aberdeen Lake.

Several undergraduate and graduate theses have been written on Dubawnt Group rocks. Cecile (1973) subdivided the Thelon Formation into four lithofacies on the basis of petrographic examination of samples from the basin. Macey (1973) compared the South Channel and Kazan Formations to the Martin

Table 0-1. Stratigraphic evolution of the Dubawnt Group (1955-1983).

Wright (1955)	Donaldson (1965)	Compilation (1982)	Aberdeen Lake (This study)
Upper Dubawnt Conglomerate	Thelon Formation	Thelon Formation	Upper Thelon Formation Lower Thelon Formation
Dubawnt Porphyries	Pitz Formation Martell Syenite	Pitz Formation (2)	Pitz Formation (1) (2)
		Kunwak Formation	Kunwak Formation
	Christopher Island Formation	Christopher Island Formation	(3)
Dubawnt Sandstone	Kazan Formation	Kazan Formation	(3)
Lower Dubawnt Conglomerate	South Channel Formation	South Channel Formation	(3)
		Angikuni Formation	(3)
		Sub-Dubawnt Regolith	(3)
Basement	Basement	Basement	Basement

- (1) Includes several stratigraphically distinct regolith and silcrete horizons, interflow sediments, and a basal lithic sandstone.
- (2) The intrusions in the Dubawnt Group range widely in age and composition. They include lamprophyres, syenites, fluorite-bearing granites, and others.
- (3) Not present in the Aberdeen Lake map area.

ORDOVICIAN

Fossiliferous limestone

HELIKIAN

Basalt

Dolostone, siliceous dolostone

Thelon Formation: sandstone, pebbly sandstone, conglomerate

APHEBIAN

Porphyritic diorite-gabbro (includes McRae Lake dyke)

Granite: in part fluorite bearing and rapakivi textured

Gyenite: quartz syenite; includes Martell intrusive suite

Pitz Formation

Christopher Island and Kunwak formations

South Channel and Kazan formations

Dubawnt Group

Amer group and equivalents

APHEBIAN and/or ARCHEAN

Orthoquartzite, quartz pebble conglomerate, muscovite schist and phyllite,

Basic metavolcanics, metapelites, calc-silicate rocks, graphitic schists, meta-greywacke, iron formation

Migmatite, irregularly layered granitoid gneiss, paragneiss; amphibolite, cataclastic augen gneiss

ARCHEAN

Ultramafics: komatiites, serpentinites

Basic metavolcanics, massive and pillowed flows, pyroclastics, tuff, and agglomerate; meta-greywacke, mixed basic and intermediate

metavolcanics, amphibolite, carbonate iron formation, calc-silicate rocks; quartzite, quartz pebble conglomerate, black slate, includes some gabbro

Fault

Figure 0-2. Geology of the Thelon and Baker Lake Basins
(from Miller and LeCheminant, in press).

Formation in Uranium City, Saskatchewan. Mole (1977) studied the South Channel, Kazan, and Christopher Island Formations in the Angikuni Lake area. Waunch (1977) studied the heavy minerals in the Thelon Formation and found sparse populations of resistant zircon, rutile, tourmaline, and apatite.

In contrasting the sedimentology of the South Channel and Kazan Formations with that of the Thelon Formation, Donaldson (1967b) found a number of differences largely related to tectonic activity. Dean (1967) studied the matrix of Thelon Formation sandstones and found a variety of kaolins, mica, septechlorite, anatase, hematite, quartz, feldspar, and an unidentified phosphate mineral similar to goyazite. A maximum thickness of 2000m for the Thelon Formation in the center of the basin was calculated by Overton (1971). An investigation of fossils from the Ordovician outlier north of Aberdeen Lake proved them to be of exceptional quality (Bolton and Nowlan, 1979).

Extrusive volcanic rocks of the Pitz and Christopher Island Formations and the associated Martell Intrusions have received relatively little attention. Blake (1980) studied the volcanic rocks of the Dubawnt Group in the Baker Lake - Angikuni Lake area. Smith et al (1980) proposed that the Martell Syehite was derived by partial melting of an ultramafic source at mantle pressures, based on geochemical and structural relationships within the volcanic sequence.

Uranium has been the resource receiving the most economic attention in the Thelon Basin. Similarities of the Thelon Basin to the productive Athabasca Basin account for this.

Miller (1979, 1980) found uranium occurrences in the Dubawnt Group to be related to red beds above unconformity surfaces, fractures, and epigenetic mineralization in the thermal aureole of fluorite-bearing granites and syngenetic syenites. Miller (1983) reported on phosphate occurrences in the Thelon Formation and suggested a playa lake rather than marine source based on geochemical and lithological arguments.

Geochronological data for the Dubawnt Group and its basement rocks are limited. A whole rock Rb-Sr isochron for five Dubawnt volcanic rocks gives an age of 1.725 ± 4 Ga. This is in good agreement with the mean K-Ar age of 1.698 ± 45 Ga for micas from these same samples (Donaldson, in Wanless and Loveridge, 1972). These ages are very similar to those obtained from the Nueltin Lake Granite thought to be related to the Dubawnt volcanics (Stockwell, in Wanless and Loveridge, 1972). The Rb-Sr isochron discussed above yields an age of 1.786 ± 26 Ga when converted to the decay constant recommended by the 1976 International Commission on Stratigraphy (Blake, 1980). A 1.834 ± 23 Ga Rb-Sr isochron for a syntectonic quartz monzonite gives the timing of some of the last activity of the Hudsonian Orogenic event in the area (Wanless and Eade, 1975). A 1.66 Ga Pb-Pb age from apatite cement at the base of the Thelon Formation may give a minimum age for cementation of these sandstones (Miller, 1983).

Park et al (1973) used paleomagnetic data (including some from samples of Kazan Formation red beds and Christopher Island Formation volcanics) to develop a polar wander curve for $1.750 - .700$ Ga. The derived curve suggests that the Dubawnt Group was

deposited at equatorial latitudes. Fraser et al (1970) discussed similarities of Proterozoic basins in northwestern Canada, and suggested that at one time, the cratonic cover was considerably more extensive than now preserved. They concluded that the Thelon Formation, Athabasca Formation, and the Hornby Bay Group are erosional remnants of this cover (Figure 0-3).

A wealth of information also has been obtained by private companies working in the area. In the Aberdeen and Schultz Lake map areas, numerous drill holes have been cored by Anaconda, Cominco, Essex, Marline, and Urangesellschaft. Although the intensive prospecting that went on early in the 1970's has been discontinued due to the low price of uranium, Anaconda and Urangesellschaft continued operations in the study area during the summer of 1982.

OBJECTIVES

The study was suggested by my thesis advisor J.A. Donaldson, who during his work in the Thelon Basin, recognized a thick and widespread regolith developed on basement rocks beneath the Thelon Formation. In the summer of 1982, the author worked in conjunction with A.N. Lecheminant and A.R. Miller out of their G.S.C. field camp on Aberdeen Lake, and was visited by J.A. Donaldson in the field for two weeks.

The original objectives of this study were to:

- 1) map the areal extent of regolith exposures within the study area;
- 2) determine the relationship of the Thelon Formation to the

Figure 0-3. Correlative Mid-Proterozoic Basins in northwestern Canada (from Ross, 1983b). AA-Athapuscow Aulacogen, BF-Bathurst Fault, CG-Coronation Gulf, GBL-Great Bear Lake, GSL-Great Slave Lake, HB-Hudson's Bay, KB-Kilohigok Basin, LA-Lake Athabasca, MF-MacDonald Fault

underlying regolith;

- 3) assess the geochemical conditions responsible for the observed alteration;
- 4) describe the regolith's textures and infer pedogenetic processes;
- 5) infer the climatic conditions existing during weathering;
- 6) investigate the possibility of using the regolith for stratigraphic purposes;
- 7) determine if any diagenetic events or subsequent weathering affected the Proterozoic weathering profiles; and
- 8) compare the profile developed in the study area to other Proterozoic weathering profiles described in the literature.

As a result of the summer's field work, additional problems and relationships became apparent: the Thelon Formation was separated into two mappable units on the basis of distinctive lithologic characteristics; synsedimentary faulting during Thelon Formation deposition was documented; drill cores revealed important subsurface information, particularly about the regolith, which was found to be polygenetic, involving several stratigraphically distinct weathering horizons; and both silcretes and silicification of the lower Thelon Formation were found to be particularly abundant within the study area.

The major goals of this thesis are:

- 1) to describe textures, structures, mineralogy, and diagenetic alteration observed in the sub-Thelon regolith for comparison with future studies of Proterozoic weathering profiles;
- 2) to document the occurrence of Mid-Proterozoic laterites and

silcretes, and to provide criteria for their recognition;
3) to incorporate the results of this study with the previous work in the Thelon Basin and hopefully provide a contribution to the geology of the Dubawnt Group.

INVESTIGATIVE TECHNIQUES

Samples investigated in this study were taken on traverse as well as from available drill core. Drill core profiles were sampled at one meter intervals, and wherever further sampling was required. Typically, 5-10cm long samples were taken for later laboratory work. Core was sawn in half on a trim saw and one half was coated with clear epoxy for binocular examination.

Initial examination with a binocular microscope proved particularly useful for examining textures and determining which samples would be cut for thin-section. Thin-sections were cut for approximately every 2-3m of core in homogeneous areas, and at closer intervals where particular features warranted. Three of the best developed profiles were petrographically studied in detail.

Modal mineralogy was determined by 500-1000 point counts per thin-section. Results were found to be reproducible. Table 2-2 lists the point counting error.

X-ray diffraction was performed on over 25 whole rock samples for modal mineralogy at the G.S.C.. In addition, two profiles (81-1 and 82-23) were picked for clay mineral studies to supplement petrographic techniques. Uncoated splits of drill core samples were trimmed and scrubbed to remove any surficial contamination. Then they were wrapped in paper towel

and gently disaggregated to pea size fragments with a rock hammer. Imersion and gentle aggitation in distilled water provided a suspension of the finest particles. The suspension was centrifuged according to tables in Jackson (1958) to separate the 2-10u and <2u fractions. Samples were then resuspended and allowed to dry and orient on a glass mount. Several mineralized fracture fills and fifty oriented clay mounts were analyzed with copper radiation on the Philips x-ray powder diffraction system at Carleton University consisting of x-ray generator PW 1730/10, automated powder diffractometer PW 1050/81, microprocessor PW 1710, and printing recorder PW 8210. The presence of swelling clays was checked for by running glycolated samples. Infra-red determination of nacrite from dickite and identification of polytypes of kaolin, illite, and chlorite by random orientation was done by Dr. Kodama's lab in the Soil Chemistry and Mineralogy Branch of Agriculture Canada.

Chemical analyses of regolith, silcrete, and several other relevant samples were done by X-ray fluorescence at the G.S.C.. Trace element compositions of mudstones were obtained at the G.S.C. by emission spectroscopy.

Energy dispersive analysis was done on silicified evaporites on the microprobe at Carleton University.

Field samples were slabbed and thin-sections cut for selected samples. The sodium cobaltinitrite staining method was used on relevant samples to determine the proportions of quartz, plagioclase, and potassium feldspar.

CHAPTER I STRATIGRAPHY OF THE DUBAWNT GROUP

INTRODUCTION

The study area is located between latitudes 64-65° and longitudes 96-100°, (NTS map sheets 66B Aberdeen Lake and 66A Schultz Lake), in the District of Keewatin. Rocks of the Dubawnt Group (Wright, 1955; Donaldson, 1965) are exposed here overlying with angular unconformity basement gneisses and schists deformed during the Hudsonian Orogeny. The oldest member of the Dubawnt Group recognized in the study area is the Kunwak Formation (Figure 1-1). Although nowhere is the entire stratigraphic sequence preserved, a vertical slice taken south from the Marjorie Hills to Wharton Lake (A-A' on Figure 1-1) shows the stratigraphy (Figure 1-2). Also apparent is the northward tilting of lithologies composing the Marjorie Hills.

KUNWAK FORMATION

The Kunwak Formation, originally named for exposures in the Thirty Mile Lake (65P) map area extends through adjacent map areas in the Baker Lake Basin, south and east of the present study area. The Kunwak Formation formed in isolated basins and is related to renewed faulting after the Christopher Island volcanic event. This relationship is shown by reverse stratigraphic clast composition (initially Christopher Island Formation, then basement dominated). LeCheminant et al (1979) proposed the name Kunwak Formation for this

Figure 1-1. Geological map of the Marjorie Hills, Aberdeen Lake (66B) map sheet (from Lecheminant *et al*, 1983).

Figure 1-2. Geological cross-section through the Marjorie Hills.

sequence of continental, volcanoclastic, and alluvial sediments deposited between the cessation of the Christopher Island volcanic event and the initiation of Pitz Formation volcanism.

The Kunwak Formation is assumed, on stratigraphic evidence from the Tebesjuak Lake (650) map area, to be older than the start of intensive chemical weathering. In one area, intensely weathered, initially arkosic Kunwak Formation conglomerates and arenites up to 50m thick are sandwiched between fresh arkoses and Pitz Formation tuffaceous sediments containing fresh biotite (personal communication; A. N. LeCheminant, 1983). In the study area the Kunwak Formation does not display the intense alteration which occurred after its deposition because of its geographical location in the southwestern part of the map sheet. In this area the regolith has not been preserved on any rock types.

An older regolith preserved beneath the basal member of the Dubawnt Group (South Channel Formation) has been described in the Kamilukak Lake map area (Tella et al, 1981), Tebesjuak Lake map area (LeCheminant et al, 1981), Thirty Mile Lake map area (LeCheminant et al, 1977), and the Tulemalu Lake map area (Eade and Blake, 1977). Inconsistent development of this regolith (in most places less than 2m thick) suggests that it is of less significance than the regolith of the present study. It shows hematitic alteration rather than hematite plus clay displayed by the later weathering event (A.R. Miller, personal communication, 1983).

PITZ FORMATION

Regolith has been observed on Archean and/or Archean basement rocks, beneath and between rhyolitic flows and crystal tuffs of the Pitz Formation, and along the top of the Pitz Formation and its unconformable contact with the overlying Thelon Formation. This indicates that chemical weathering started before deposition of the Pitz Formation and continued during deposition of the basal parts of the Thelon Formation. Stratigraphically, this makes alteration events difficult to separate, especially in areas which were never covered by the volcanic rocks of the Pitz Formation. In the northern part of the study area, rocks of the Pitz Formation probably were never deposited, and whether the regolith there is older, synchronous, or younger than the Pitz Formation is unclear. Here the Thelon Formation rests with angular unconformity on a truncated regolith profile which may have been affected by one or more weathering events. Figure 1-3 shows the distribution of regolith and silcrete.

In areas such as the Marjorie Hills, where the Pitz Formation covers the initial regolith, this profile can be stratigraphically separated from other profiles developed within or on the Pitz Formation. Drill cores show a maximum thickness of at least 50m of regolith (Table 1-1). This initial event of deep weathering appears to have been of great areal extent, and perhaps was basinwide (regolith has been observed in many parts of the basin: personal communication, J.A. Donaldson, A. LeCheminant, A.R. Miller, 1983).

Silcrete in the study area has four main stratigraphic

Figure 1-3. Distribution of regolith and silcrete outcropping in the Marjorie Hills.

Table 1-1 Regolith thicknesses determined from drill core and measured sections.

Cominco property northeast of Aberdeen Lake (1)

DH 81-1	68m+	Feldspathic Metawacke
DH 81-2	26m+	Feldspathic Metawacke
DH 81-3	55m+	Feldspathic Metawacke
DH 81-4	5m+	Protolith unknown

Anaconda property Marjorie Hills (2)

DH 81-14	109m+	Feldspathic Metawacke
DH 81-16	81m+	Feldspathic Metawacke
DH 81-17	104m+	Feldspathic Metawacke
DH 82-23	43m+	Granodiorite Gneiss

Scarp north of Lone Gull	50m+	Feldspathic Metawacke
Northeast shore of Marjorie Lake	10m+	Granitic Gneiss
North of Marjorie Lake	20m+	Pitz Formation Rhyolite
Central Marjorie Hills	10m+	Augen Gneiss

(1) Thicknesses listed are believed to represent minimums as the tops of profiles are truncated and often fresh protoliths are not reached in drill core or are covered.

(2) The unusual thicknesses shown by regolith in drill holes 81-14, 81-16, and 81-17 is probably related to their location in the Marjorie Hills where the thickest accumulation of Pitz Formation volcanics occur in the map area. All three profiles show faulting, fracturing, and explosive intrusives which may have allowed weathering to greater depths.

occurrences: 1) on top of truncated regolith profiles developed on basement rocks; 2) within or on top of Pitz Formation volcanic rocks; 3) within or on top of faults and tectonic breccias exposed to pedogenic processes and 4) hosted within sediments of the lower Thelon Formation. Because several stratigraphically distinct silcretes have been noted (see Chapter 3) their potential use for correlation purposes is questionable.

Silcretes capping truncated regolith profiles are thought to have formed extensive, thin, flat sheets over the highly weathered, planated paleotopography. Their present outcrop distribution is related to later faulting and incision which provided a significant proportion of detrital material to the Thelon Formation. Silcretes intercalated with Pitz Formation volcanics were probably of more limited extent. Those hosted in lower Thelon Formation sediments reach considerable thickness, but are restricted in occurrence to areas proximal to Pitz Formation volcanics. Silcretes are discussed in detail in Chapter 3.

The lack of preserved upper soil horizons (in places the regolith retains original protolith textures, such as foliation, up to its truncation surface) is evidence that in many areas the original regolith profile was truncated by erosion. East of Marjorie Lake and along the Dubawnt River, conglomerates and quartz arenites have been documented resting on truncated profiles (LeCheminant et al, 1983). One location visited by the author (MP on Figure 1-1) exposes approximately 13m of regolithic granitic gneiss consisting of

illite, kaolin, partially altered feldspars, quartz, chlorite and minor epidote. This is sharply truncated by a coarse, angular, immature conglomerate containing numerous regolith fragments. In all other regolith profiles examined, the presence of chlorite is indicative of the basal "chlorite zone" of the profile (for description of the chlorite zone see chapter 2). A significant portion of the profile must have been removed to expose this zone. Overlying the conglomerate are potential eolian quartz arenites (LeCheminant et al, 1983) which are in turn overlain by Pitz Formation rhyolites.

These arenites with minor conglomerate have been included within the Pitz Formation. It should be noted that they contain no Pitz Formation clasts and are thought to pre-date volcanism in this area. Their similarity with quartz arenites of the Thelon Formation may cause confusion or misidentification in areas where they are not separated by the volcanic rocks of the Pitz Formation. They are distinguished from the underlying Kunwak Formation by their stratigraphic position above the regolith. The extent of this lower unit of the Pitz Formation can be traced by in sparse outcrops across half of the Aberdeen Lake map area. Stratigraphic control on its thickness is poorly constrained, but probably is at least 150m (LeCheminant et al, 1983).

Volcanic rocks of the Pitz Formation crop out primarily in the Marjorie Hills and on a large hill east of Marjorie Lake in the Aberdeen Lake map area. In the Schultz Lake map area they occur primarily around Pitz Lake. Fluorite-bearing

rapakivi granites in the Kamilukuak Lake map area (65K) are dated at ~1.8Ga (personal communication, A. R. Miller, 1983). This complex is continuous with similar granites in the Tebesjuak Lake (650) map area which intrude Pitz Formation rhyolites and are interpreted as coeval (LeCheminant et al, 1979). The Pitz Formation consists primarily of rhyolitic lavas, crystal tuffs, and intercalated sediments. Commonly these rhyolites are weathered. Phenocrysts which were originally orthoclase and/or sanidine are pseudomorphed by mixtures of chert, kaolin, and minor illite. The recrystallized groundmass is deeply stained by disseminated hematite which give the rock its characteristic deep maroon coloration. Thicknesses vary considerably from suspected eruptive centers to the distal areas of volcanic accumulation.

Evidence that the weathering event which produced the Pitz regolith continued throughout deposition of the Pitz Formation is shown by several exposures in which a relatively fresh flow is sandwiched between intensely kaolinized and hematized rhyolite. The intense alteration is taken to represent interflow subaerial weathering. In many areas a distinctive reddish brown chert infills fractures and joints in the rhyolite below weathered surfaces. These infills were found down to 138m in drill core. They indicate that the rhyolites were fractured and jointed, which enabled alteration during relatively short hiatuses. In most areas where the Thelon Formation/Pitz Formation unconformity is exposed, basal Thelon sediments rest on regolithic Pitz Formation lithologies.

THELON FORMATION

Lower Thelon Formation sediments contain abundant volcanic detritus of the Pitz Formation, so that the upper and lower parts of the Thelon Formation can be mapped as distinct units as shown in Figure 1-1. The lower Thelon unit consists largely of coarse, highly rounded (may be a function of spheroidal weathering rather than long transportation distances), rhyolite cobble para- and ortho-conglomerates, lithic sandstones and mudstones. These sediments characteristically are deep maroon, brown, or red and are highly silicified. Silicification is sufficiently intense and related to pedogenic processes to warrant designation as silcretes. Both debris flows and fluvial channels have been documented (Chiarenzelli et al, 1983).

The transition between the lower and upper units in most places is gradational (Chiarenzelli et al, 1983). It occurs over a significant portion of section (10-50m) in which the red coloration, silicification, and amount of Pitz Formation fragments gradually decrease upwards, coupled with an increase in kaolin cement and basement clasts (quartzite, vein quartz, silcrete, silicified regolith, quartz arenite, lower Thelon Formation, iron formation). The higher percentage of silica-rich clasts is distinctive, and gives the upper Thelon unit a white speckled appearance. In some areas, synsedimentary faulting has resulted in locally unconformable contacts. A 600m section of the lower Thelon unit near Hailstone Lake (denoted as HS on Figure 1-1) is tilted 45° and truncated by the upper Thelon unit. Synsedimentary faulting

also was documented in the Schultz Lake map area by the author. There the basal Thelon sediments are inclined up to 40° and are overlain by a debris flow, which in turn is overlain by upper Thelon sediments inclined at 5° .

The upper Thelon unit has a much greater areal extent than the lower Thelon unit, which is proximal to its source. In the study area the upper unit is characterized by pale tan, yellow, and pink colours. It consists of quartz arenite, commonly with a matrix and/or cement of kaolin and/or illite. First cycle diagenetic quartz arenites contain illite and/or kaolin pseudomorphs after feldspar grains similar to those described from the Athabasca Basin by Ramaekers (1981). It commonly is coarse grained, and the grains are subangular to subrounded. Conglomerates and mudstones at the base of the formation are common, especially north of the extent of the lower unit. The deep red mudstones are probably distal fine-grained equivalents of the lower unit. The lack of common labile minerals such as biotite and feldspar indicate that this upper unit is derived from a weathered source or paleosols (Cecile, 1973) and/or underwent intensive diagenetic alteration. The resistant clast composition of coarser grained sediments indicate significant transport, weathering, breakdown of less resistant clast types, or a combination of these factors.

SUMMARY

In summary, a stratigraphic column for the Dubawnt Group in the Aberdeen and Schultz Lake map sheets contains, in

ascending order, the Kunwak Formation, Pitz Formation, and the Thelon Formation (Figure 1-1). The Pitz Formation contains an undetermined number of interflow weathering profiles and silcretes. The geological history of the Dubawnt Group in the area began with alluvial sedimentation of the Kunwak Formation, followed by deep chemical weathering interrupted by erosion, deposition of the basal Pitz Formation sediments, and rhyolitic volcanism of the Pitz Formation. Silcretes of different ages formed as resistant cappings on weathered regolith profiles or in permeable host lithologies. The lower unit of the Thelon Formation is a proximal fan derived exclusively from Pitz Formation volcanic detritus. In places it conformably grades into the upper Thelon unit. Syn-sedimentary faulting has created angular unconformities between the units in other areas. The upper Thelon unit is composed of compositionally mature and texturally immature sandstones and conglomerates. It forms an extensive blanket deposit over the entire Thelon Basin. Both units are distinctive and can be mapped separately on a 1:250,000 scale in the study area.

CHAPTER 2

CHEMICAL WEATHERING OF THE SUB-THELON REGOLITH

INTRODUCTION

As discussed in the previous chapter, there is abundant evidence that residual products of chemical weathering developed before, during, and after Pitz Formation volcanism. Field sampling and the availability of drill core profiles through this regolith has enabled its characterization.

The study of paleoweathering profiles is best undertaken by a multidisciplinary approach. Geomorphologists are concerned with the surficial aspects, topographical expression, and the weathering processes under which paleosols formed. Pedologists are primarily interested in the upper (A and B) soil horizons which can give information about the paleoclimate, biological activity, and the evolution of pedogenic processes. Geologists consider chemical weathering of rocks and minerals and the significance of weathering in terms of ancient terrestrial environments, erosion, and sedimentation.

TERMINOLOGY

A complete weathering profile (from overlying soil to fresh rock below) consists of a number of distinct zones or horizons. Soil science terminology divides the weathering profile into four major horizons: A, B, C, and D. Horizons A and B, the upper parts of the profile, are collectively termed the solum. Horizon A undergoes leaching and depletion

of certain components and is known as the zone of eluviation. Horizon B undergoes accumulation and is termed the zone of illuviation. Horizon C is a transitional zone of partially altered rock and horizon D is unaltered rock. The solum is usually an order of magnitude or more thinner than horizon C. Individual layers within a horizon may be subdivided and denoted by a subscript numeral (e.g. A₀).

Soil terminology used in various fields of earth science differs slightly. Geomorphologists use the term "regolith", which is a general term for residual and transported weathered material, to denote the complete weathering profile. "Saprolite" is a more specific term which refers to weathered igneous or metamorphic rock retaining original textures. In profiles where the surface has been disturbed by restructuring, movement, or deposition of additional material, the term regolith is preferred. Paleosols ("fossil" soils) are soils formed before the present, commonly under different environmental conditions than exist now in the area where they are located.

The present study is concerned with chemical, mineralogical, and textural changes during formation of the sub-Thelon regolith, and with separation of these from later diagenetic or weathering events. Many of the following ideas on chemical weathering are taken from Loughnan (1969) and Ollier (1969). A suitable definition for chemical weathering is taken from Loughnan (1969), after Reiche (1950); it reads:

"Chemical weathering is a process by which atmospheric, hydrospheric, and biologic agencies act upon and react with

the mineral constituents of rocks within the zone of influence of the atmosphere, producing relatively more stable, new mineral phases."

Essentially, chemical weathering involves the chemical reactions which take place when rocks, containing minerals formed at elevated temperatures and pressures within the Earth, are exposed to H_2O , CO_2 and O_2 at the Earth's surface.

FACTORS CONTROLLING CHEMICAL WEATHERING

A number of interrelated factors play important parts in chemical weathering. Of greatest importance is leaching. Chemical weathering operates by three simultaneous processes: 1) breakdown of parent minerals and release of cations and silica; 2) removal by solution of some cations; and 3) the residue reacting with O_2 , H_2O , or CO_2 to form new metastable or stable minerals (Loughnan, 1969). These processes work toward destroying minerals unstable under the present chemical conditions, removing mobile cations, and concentrating immobile cations and resistant minerals. Cations show various mobilities over the pH range of natural waters (Table 2-1).

If leaching is minimal, mobile cations may concentrate and influence the composition of the new mineral formed. Loughnan (1969) gives the following example where leaching of potassium ions is incomplete, so that illite forms preferentially to kaolin, (which would have formed if all the potassium cations were removed).

Table 2-1. Mobilities of common rock-forming cations (from Loughdan, 1969).

Ca^{++} Mg^{++} Na^{+}

Readily lost under leaching conditions.

K^{+}

Readily lost under leaching conditions, but rate may be retarded through fixation in the illite structure.

Fe^{++}

Rate of loss dependent on the redox potential and degree of leaching.

Si^{4+}

Slowly lost under leaching conditions.

Ti^{4+}

May show limited mobility if released from the parent mineral as $\text{Ti}(\text{OH})_4$; if in the TiO_2 form, immobile.

Fe^{3+}

Immobile under oxidizing conditions.

Al^{3+}

Immobile in the pH range of 4.5 - 9.5.

partial leaching of potassium
 potash feldspar illite
 $3\text{KAlSi}_3\text{O}_8 + 2\text{H}_2\text{O} \text{ --- } \text{KAl}_2(\text{Al}, \text{Si}_3)\text{O}_{10}(\text{OH})_2 + 6\text{SiO}_2 + 2\text{KOH}$

complete leaching of potassium
 potash feldspar kaolinite
 $2\text{KAlSi}_3\text{O}_8 + 3\text{H}_2\text{O} \text{ --- } \text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4 + 4\text{SiO}_2 + 2\text{KOH}$

In rocks undergoing extreme leaching, desilicification can reach completion and aluminum hydroxides such as gibbsite ($\text{Al}_2(\text{OH})_6$) can replace kaolin as the ultimate weathering product. This explains the dissolution of silica commonly shown in bauxites and laterites, which form under intensive leaching conditions in areas of extreme rainfall. A sequence of weathering products from 2:1 clays to 1:1 clays to sesquioxides of aluminum, iron, and titanium occur as leaching increases. This can be seen as vertical zonation in some weathering profiles and soils. Leaching is inhibited in basal parts of the profile which are below the water table, but is complete in the uppermost horizons. The degree of leaching is determined by climate, topography, runoff, the water table, and other factors (Figure 2-1).

Climate plays a key role in chemical weathering. A comparison of weathering profiles with decreasing latitude shows that the thickest profiles develop in tropical regions where the climate is hot and humid, and the thinnest profiles develop in arid regions, hot or cold (Figure 2-2). This suggests that the combination of warm temperatures and high precipitation promotes the formation of deep weathering profiles. High temperatures tend to speed chemical reactions

Figure 2-1. Factors influencing chemical weathering.

Figure 2-2. Weathering profiles as a function of latitude (from Strakhov, 1967).

2-3 per 10°C , according to van't Hoff's rule (Loughnan, 1969). However, the influence of leaching is apparent, because podsol profiles in temperate latitudes can develop to significant thicknesses in well-leached areas of moderate precipitation.

Topography has a direct relationship to leaching. In areas of high relief, much of the precipitation available for leaching may be lost to surface runoff. Erosional losses may be severe and inhibit soil formation. Flat, swampy, poorly drained areas inhibit sub-surface removal of mobile cations released by weathering processes, and hence suppress additional weathering. The ideal conditions for chemical weathering are achieved with rolling topography and well-developed subsurface drainage.

The effect of the protolith on chemical weathering is varied and greater when the overlying soil is immature. The development of the great soil groups (zonal soils) over large areas of similar environmental conditions suggests that under the proper conditions the influence of the protolith is minimized. The common occurrence of intrazonal soils that differ from the prevailing zonal soils indicates that local conditions such as protolithology play an important part in soil genesis.

Two factors are very significant when considering the influence of the protolith: texture and mineralogy. Texture determines porosity and permeability, both of which affect leaching. Hence permeable sandstones weather faster and deeper than granites in the same area because they are much

more susceptible to leaching. For the same reason highly jointed rocks weather faster and more severely than massive ones.

Original mineralogy can strongly influence weathering products. Minerals show different susceptibilities to weathering related to their bonding energies, fracture, cleavage and chemical composition. A soil overlying quartzite will be very different from a soil developed on basalt in the same area. The chemical composition of the minerals weathered determines which cations are available for formation of weathering products. The degree of leaching determines how much influence the original chemical composition of the mineral may have. Generally, the greater the leaching a rock undergoes, the less influence the protolith has on the resulting soil profile. This is shown by the genesis of bauxites, laterites, silcretes, and calcretes, all of which can form on widely differing lithologies if other conditions for their formation are met.

Important chemical controls are placed on weathering by pH, Eh, and organic material. The solubilities of many substances are affected by changes in pH. For example, the solubility of iron is 100,000 times greater at a pH of 6 than at a pH of 8.5 (Ollier, 1969). Acidic pH values are generated by the oxidation of organic material and by extreme leaching conditions. Alkaline pH values occur when alkali cations are released from the parent material faster than they are leached away. Changes in pH can play an important role in the precipitation and dissolution of minerals. Cor-

rosion of successive generations of quartz and carbonate cement in sedimentary rocks reflects changing pH conditions. Eh controls whether oxidized or reduced valence states of certain cations will exist under the prevailing conditions. In iron-rich solutions of high Eh, hematite will form preferentially over goethite. When Eh is plotted against pH, an estimation of which phases will be stable under the existing conditions can be made. Organic material places constraints on the pH of certain soils and is an important consideration in oxidation and reduction reactions.

DEEP WEATHERING

Chemical weathering to depths of 100m or more is widespread in areas of hot, humid climatic conditions with extreme leaching, tectonic stability, and minimal erosional losses. Parts of Australia, Africa, and South America display these conditions today. Weathering profiles in these areas commonly show a highly irregular basal surface related to differential weathering of various lithologies and along fractures (Thomas, 1966). Alteration along fractures penetrating the Russian Platform occurs at depths greater than 1000m (Ollier, 1969).

Deep weathering profiles can be subdivided into a number of zones. An ideal sequence from Ollier (1969) contains: 1) the upper soil horizons A and B; 2) structureless regolith; 3) saprolite retaining protolith textures; 4) saprolite with rounded corestones; 5) saprolite with angular, interlocking corestones; and 6) fresh rock. Corestones consisting of

slightly altered or fresh rock surrounded by highly weathered rock represent blocks bounded by fractures which escaped complete destruction of the primary mineralogy.

METASTABILITY

In weathering profiles, initial products formed by the breakdown of primary minerals may be metastable. Feldspar only rarely alters directly to a highly crystalline polymorph of kaolin. It generally alters to a variety of poorly crystalline clay minerals or amorphous substances, and passes through a number of intermediate phases before developing into structurally ordered kaolinite. With time and burial, hydrous phases may become dehydrated, amorphous substances may crystallize, and poorly crystalline material may recrystallize. Iron hydroxides alter with time and burial to hematite, opal dehydrates to quartz, and illite's crystallinity index increases. With increasing maturity the weathering profile may change, because continued leaching can remove more soluble cations and concentrate resistant ones. Some clay minerals stable at the surface may alter upon burial or exposure to different climatic conditions, while others will not. Hence illite formed under moderate leaching conditions may alter to kaolin if leaching is increased, but kaolin will not alter to illite if leaching is decreased (Singer, 1980). In summary, weathering profiles are highly susceptible to diagenetic processes; the chemical conditions of their formation may last only as long as the profile is actively weathering at the surface.

LATERITES

Laterites are important features of the landscape in many tropical regions. They are economically important as building stone, and for the concentrations of nickel, chromium, vanadium, and manganese that some contain. What a laterite is and how it is formed are topics of considerable debate. A definition given by Schellmann (1981) reads:

"Laterites are products of intense subaerial rock weathering whose iron and/or aluminum content is higher and silica content is lower than in merely kaolinized parent rocks. They consist predominantly of mineral assemblages of goethite, hematite, aluminum hydroxides, kaolin minerals, and quartz".

Bauxites are commonly considered a type of laterite in which aluminum predominates over iron.

Laterites are only one horizon of a lateritic profile. They form near the surface and are influenced by pedogenic processes as displayed by their pisolitic, vermicular, and vesicular textures. Characteristically they display red, brown, and yellow shades of coloration related to the iron and manganese hydroxides which compose them. Some laterites, used as bricks, can be cut easily when in place, but permanently indurate upon exposure to air.

Laterites are composed of goethite, hematite, gibbsite, kaolin, quartz, and other rarer iron and aluminum hydroxides and amorphous material. Illite is commonly present as degraded muscovite (Maignien, 1966). Chemically they consist of iron, aluminum, silicon, resistant trace elements, and little else. Iron/aluminum ratios can vary widely from

bauxites to iron-rich laterites containing minimal amounts of aluminum.

Description of a characteristic profile for lateritic weathering is of questionable value because many variations are described in the literature. For example, Leprun (1981) noted over 100 profiles in western Africa which consist of the following sequence: laterite, ferruginous soil, mottled zone, red zone, pallid zone, saprolite, and fresh rock; however, a more typical sequence reported in the literature is ferruginous soil, laterite, mottled zone, pallid zone, saprolite, and fresh rock. Thicknesses of most profiles are substantial but vary considerably; the range of the lateritic horizon is 0-10m.

Climatic inferences concerning laterites suggest that they form under tropical climates, with or without a dry season. The extreme leaching they display argues in favor of considerable precipitation. In rocks of Tertiary age in Australia laterites predominate in wetter coastal areas whereas, silcretes occur in the interior. The many varied examples suggest that their formation is not highly restricted.

Genesis of laterites involves the removal of most cations other than iron and aluminum by the breakdown of silicate minerals and leaching of the mobile components. The presence of a mottled zone, the development of concentric pisolites, and their thin, sheet-like occurrence suggest that reduced iron may be brought to the surface of the water table where it is oxidized which would represent a component of

absolute accumulation. The extreme leaching implies low (acidic) pH values. Maignien (1966) and McFarlane (1976) give recent reviews of research on laterites.

An important point made by many authors is that a distinction should be drawn between true laterites (and lateritic profiles sensu stricto) and tropical soils. Many tropical soils have deep weathering profiles which are similar to lateritic profiles in that they show ferruginous, mottled, and pallid zones. Lateritic profiles, as pointed out in Schellman's definition, should have higher aluminum and iron concentrations than kaolinized parent materials. In addition, a concretionary laterite horizon is diagnostic.

SUB-THELON REGOLITH

PROFILE DEVELOPED ON FELDSPATHIC METAWACKE DH 81-1

This hole was originally logged by R.A. Stanton who worked at the Aberdeen Lake Cominco camp during the summer of 1981. It was drilled vertically and penetrated 157m of Thelon Formation coarse-grained pebbly sandstones and poly-mictic conglomerates before regolith was intersected. The contact between the Thelon Formation and the regolith is horizontal and erosional. The regolith retains a foliation oblique to the unconformity, and regolith clasts are abundant in the overlying conglomerate.

The regolith, because of retention of original metamorphic textures, can be termed a saprolite. Alteration extends from 157m down to ~ 225m where fresh feldspar and

sulfides (pyrite and chalcopyrite) were observed. A gradual colour change from deep red through light maroon to green is apparent. The change reflects mineralogical variations with depth of alteration. Foliation, $50-60^{\circ}$ to the core axis, is well developed throughout the regolith, and the core commonly parts parallel to it. The regolith decreases in induration downwards with decrease of the secondary silica/clay ratio. In the first few meters, irregular quartz veining is abundant, and dolomite veining occurs parallel to the foliation from 195-200m.. In the first 20m, minor reduction spotting occurs.

The regolith has been divided into three zones. The boundaries of each are approximate and gradational. Smaller subzones of a stratigraphically higher zone occur as lenses in the next lower zone. Separate designations have not been made for these because they are considered to be part of the transition zone described below.

Zone 1 is distinguished as the hematite zone (Plate 1A). It extends from the Thelon Formation/regolith contact at 157m to 171m. It is characterized by a deep red to intermediate maroon coloration, gradually losing intensity downwards. Pinkish reduced areas, with little unleached hematite present, occur in ellipsoidal spots parallel to foliation and along fractures (Plate 1E). The darker appearance of this zone is related not only to a greater amount of hematite, but also to the dispersion into, and staining of, the clays by hematite. Several smaller fractures are incompletely filled with vein quartz and criss-crossing platelets of specular

hematite. At 157m, apatite occurs with specularite and illite along fractures, accounting for ~ 5% of the sample. The contact with the next zone is arbitrarily set at 171m, where ellipsoidal spots of remnant chlorite give a greenish tinge to the regolith.

Zone 2, the transitional zone, extends from 171-195m (Plate 1C). It is characterized in the upper half by a maroon and white speckled coloration with faint green isospheroidal chlorite blebs (5cm in diameter). In the lower half the maroon layers are intercalated with 15-66% green and white speckled, lensoidal layers parallel to foliation. The alternation of these layers give the rock a striped appearance. Some anomalies exist. At 172m a bleached zone in which the rock is off white in colour was noted. Its mineralogy is somewhat anomalous in that neither hematite nor chlorite is present. At 190-195m (at the base of the transitional zone), dolomite veining and replacement are common, the rock has a deep red coloration, and the foliation is disturbed and broken. This zone probably represents oxidation along a fracture or a small fault. The transition zone grades in less than 2m into the lower chlorite zone, where the appearance of the rock is drastically different.

Zone 3, the chlorite zone, extends from 195-225m (Plate 1D). It is characterized by abundant green and white speckled layers alternating with dark green chlorite-rich layers up to 1cm thick. At 199m the rock takes on the appearance of a proto-mylonite. It has coarser grained clay aggregates, 5-10% less quartz, slightly more chlorite and

muscovite, and a greater amount of mineralogical layering than the overlying lithology. These changes may be related to the beginning stages of mylonitization. However, weathering effects on both lithologies are identical. Dolomite appears to be associated with the fractures and fault zone at the boundary of the transitional and chlorite zones, and is not common deeper in this profile.

The protolith of the regolith was a well-foliated quartz-feldspar-muscovite-chlorite-biotite schist (feldspathic metawacke). As would be expected in a metasediment of this type, modal mineralogy probably varied by at least several percent from sample to sample. However, mineralogical variations are masked by the alteration. Differences in original grain sizes are apparent, as shown by comparison of sizes of clay pseudomorphs after feldspar.

MINERALOGY

Quartz in the regolith consists of elongate polycrystalline grains (up to 1mm in diameter) parallel to foliation. It shows undulose extinction. Grain boundaries commonly show scalloped edges, and fractures are replaced by mcq¹. Quartz content is thought to be fairly constant throughout the protolith (except at 198m, where about 25% of the original framework quartz has been replaced by dolomite). Commonly the quartz grains are strained and give biaxial interference figures.

¹ - mcq is the abbreviation used for microcrystalline quartz

Feldspar is virtually absent from the upper parts of the regolith. A few highly corroded microcline grains were seen in the two lowermost thin-sections, largely replaced by dolomite (Plate 4D). Kaolin and illite pseudomorphs after feldspar are common, and frequently appear as discrete pockets in the quartz-rich layers parallel to foliation. In other areas a pseudolattisepic texture (Brewer, 1964) has developed, with kaolin and illite grain orientations pseudomorphing microcline twinning directions. Albite twinning, is also pseudomorphed yielding clay pockets which appear striated. This has been noted in other Proterozoic weathering profiles (Williams, 1968; Chown and Caty, 1983). At 225m, most of the feldspars appear fresh.

Muscovite occurs as elongate plates, parallel to foliation, up to 2mm long. Alteration of muscovite occurs in the hematite zone by swelling and splitting of grains parallel to cleavage. In extreme cases the muscovites have split and curled in random directions, creating a matted texture (Plate 2F). This is most common in the top of the profile. Within the split muscovites, kaolin has filled the swelled areas and pseudomorphed the muscovite, producing "bleached" muscovites. In other areas, both kaolin and illite have totally replaced muscovite. The illite is thought to be degraded muscovite, and the kaolin fills the interstices of the illite laths. Secondary mcq replaces split muscovites, either alone or in association with kaolin and hematite. Hematite and leucoxene are common inclusions in the minerals that fill spaces in the split micas. The physical and chem-

ical destruction of muscovite decreases downwards. In pink reduction spots within the hematite zone, muscovite is destroyed. However, in green reduction spots they are commonly fresh. In the transition zone muscovite usually is slightly altered to muscovite-chlorite-hematite books in which each mineral occurs as a stripe parallel to cleavage. In the chlorite zone, fresh muscovites and chlorites are commonly interlayered or attached. Three muscovite and chlorite alteration zones can be established: 1) Muscovites are physically swelled and replaced by secondary mineralization and chlorite is altered to hematite (157-182m); 2) Muscovite occurs interlayered with hematite and chlorite; some may be replaced by secondary mineralization, and chlorite is only partially altered to hematite (183-190m); and 3) Muscovite and chlorite are fresh.

Chlorite is commonly associated with muscovite as interlayers or as adjacent grains. It shows similar orientation parallel to foliation, and is the same size. It is completely altered to hematite in the hematite zone. In the transition zone it can be altered, partially altered (Plate 2B), or fresh. In the chlorite zone it is fresh (Plate 2C). Some shows anomalous "Berlin Blue" interference colours, and is thought to be iron-rich, however x-ray diffraction results indicate most chlorite is an iron-rich clinocllore. Smaller, irregular grains and veinlets showing bright green pleochroism may be of iron-rich composition and of diagenetic origin.

Hematite occurs as pseudomorphic replacements after

chlorite and as inclusions in kaolin, illite, or secondary silica. In the hematite zone where it gives the regolith its deepest red colour it is well dispersed, staining the clays (Plate 2A). Abundance and dispersion decreases downwards. Specular hematite is common in fractures in the matrix of the overlying Thelon Formation sediments, and in fractures at the top of the regolith.

Kaolin occurs within split and bleached micas, in isolated pockets pseudomorphing feldspars, and with illite in clay pockets between quartz grains. It commonly shows a book-like structure and a sweeping extinction parallel to cleavage. Typically it occurs in elongate book-like stringers parallel to foliation (up to 1cm). In many places, illite has pseudomorphed the original kaolin growth morphology.

Illite forms from degraded muscovites split and swelled by kaolin or mcq replacement between cleavage fragments. Its almost invariable association with kaolin in pockets, along fractures, in worm-like stringers, and its inverse relationship with kaolin abundance suggests that much of the illite unassociated with weathered muscovite may be diagenetic. Grim (1953) states that illite differs from micas in four ways: 1) aluminum replacement of silica in illites is 1:6, in micas it is 1:4; 2) Ca, Mg, and H ions partially replace K; 3) the stacking of layers is more random; and 4) illites generally occur as fine particles whereas micas are relatively coarse. In thin-section illite was distinguished from muscovite by its smaller size and lower birefringance.

Mcq occurs as 40u to submicroscopic, anhedral, equant, to amoeboid shaped, grains. Optical properties are masked by the cryptocrystalline grain size. It shows sweeping extinction, possibly related to an earlier fibrous silica precursor. Larger grain sizes are obtained in the middle of sheltered pockets. It occurs replacing quartz along grain boundaries and fractures, in fractures near the top of the hematite zone with apatite and specularite, within clay pockets and replacing them, replacing muscovites and between cleavage fragments, in hematite blebs, and in monomineralic pockets. It commonly contains inclusions of hematite, illite, and kaolin, and may contain titanium-bearing secondary minerals.

Dolomite occurs in the chlorite and transitional zones. It is most concentrated between the boundary of the two, where it occurs in fractures (Plate 3B) and as discrete areas of replacement (Plate 4D). It replaces all previous mentioned minerals, but chlorite and muscovite seem to be the least susceptible to replacement by dolomite. Above the zone in which dolomite fracture fills are common, only minor amounts occur. The dolomite occurs in large (up to 2mm) blocky grains, or as tiny inclusions in minerals being replaced. In one fracture it displays sigmoidal shape related to displacive growth in a fracture (Watts, 1980). Apatite occurs as euhedral crystals .1mm - 1mm in size. It is commonly associated with specular hematite in veins near the unconformity. It can also occur as isolated grains replacing secondary minerals between quartz grains. At 157m, about 5%

apatite is present (Plate 6A). Apatite and specularite are commonly associated with matrix fills in Thelon conglomerates, or as replacement of feldspars in Thelon Formation red beds (Miller, 1983).

Figure 2-3 shows the estimated modal mineralogy of the regolith. Values are based on 500 or 1000 point counts per thin-section. The uncertainty of point counting is listed for several minerals in Table 2-2. On the average it is <7%, which is similar to the variations in mineralogy from two thin-sections cut from the same sample.

Mineralogical trends are as follows:

- 1) Quartz shows slight dissolution (scalloped edges due to replacement by mcq), except at 193 and 198m where dolomite replaces it. At 186m, where dolomite is abundant, it occurs largely in veins and replaces clays preferentially over quartz.
- 2) Muscovite shows an increase from 5% at the top to 15% at the bottom of the profile. This is thought to be largely a function of the greater swelling and degradation of muscovites to illite in upper layers. This is also shown by a decrease in the previously described matted texture downwards.
- 3) Chlorite and hematite occur in an inverse relationship to each other, directly related to oxidation (Plate 2D). Chlorite is absent from the hematite zone, whereas hematite is absent from the chlorite zone.
- 4) Mcq decreases downwards from 25% in the hematite zone

Figure 2-3. Modal mineralogy of the regolith profile developed on feldspathic metawacke with depth (DH 81-1).

Table 2-2. Point counting error.

	Quartz	Hematite	Illite	Kaolin
<u>81-2 203m</u>				
1st 500 counts	150	128	23	121
2nd 500 counts	135	136	19	142
Total	285	264	42	263
1st - 2nd = x	15	8	4	21
% Error = 100x/Total	5.2%	7.6%	10.0%	12.3%
<u>81-2 221m</u>				
1st 500 counts	166	60	16	80
2nd 500 counts	154	72	28	88
Total	320	132	44	168
1st - 2nd = x	12	12	12	8
% Error = 100x/Total	3.8%	9.1%	27.3%	4.8%
<u>81-2 226m</u>				
1st 500 counts	152	37	87	128
2nd 500 counts	151	32	98	126
Total	303	69	185	254
1st - 2nd = x	1	5	11	2
% Error = 100x/Total	0%	1.0%	2.2%	.4%

Error Range = 0 - 27%

Mean Error = 7% for 500 point counts per thin-section

to 0% in the transitional zone. Decreasing induration and a greater tendency to part along foliation downwards is related to the absence of silicification at depth.

5) Apatite, illite, mcq, and specular hematite are present in fractures in the top few meters and sporadically at depth.

6) Dolomite has an inverse relationship with quartz, clays, and muscovite. Only chlorite and muscovite seem to be relatively stable in its presence.

7) Hematite, illite, and kaolin show an inverse relationship to muscovite and chlorite, indicating their origin as weathering products.

8) Kaolin is largely absent from the top 6m, indicating it has been replaced by illite during diagenesis of the Thelon Formation (Plate 2E).

9) Illite and kaolin seem to be preferentially replaced by dolomite, and show the closest inverse relationship to it.

INTERPRETATION

The original weathering of a coarse to medium grained quartz-feldspar-muscovite-chlorite-biotite protolith (feldspathic metawacke) eventually yielded a stable secondary assemblage of kaolin after feldspar, hematite after chlorite, and illite from physically and chemically degraded muscovite.

The original deep weathering profile represented on Figure 2-4 experienced erosion, truncation, and formation and superpositioning of zones sometime before Thelon deposition. This is shown by: 1) the lack of upper soil horizon preservation; 2) the regolith commonly retains foliation to the

Figure 2-4. Interpretation of the regolith profile developed on feldspathic metawacke (DH 81-1).

1. Original Weathering Profile
2. Erosion and Superposition of Zones
3. Silicification and Silcrete Genesis
4. Erosion, Thelon Deposition, and Diagenesis

erosional surface, and is thus a saprolite, which typically forms in the lower part of a weathering profile; 3) clasts of regolith are common in the overlying Thelon Formation; and 4) green remnant chlorite blebs, common in the transitional zone, have original chlorite preserved and are surrounded by hematitic alteration, indicating that they are the last remnants of the original mineralogy (Plate 3F).

Silcretes developed on the truncated regolith profile in the study area. This is shown by quartz veining in the upper hematite zone (Plate 3A) and by the replacement of clays by mcq. The most intense silicification is shown at the top of the profiles, and decreases with depth. An alternative explanation is that this is a function of Thelon diagenesis. However, the amount of silica which would have to be released and leached from above is great, and the predominance of silicified regolith clasts in basal Thelon sediments indicate that silicification associated with silcrete formation is more plausible. Evidence that these clasts were not silicified in situ in the Thelon Formation includes their highly rounded shapes, association with extremely resistant quartzite and vein quartz clasts, the lack of less resistant clasts in the conglomerates, and the similarity of clast textures to textures in the underlying silcretes. Some of the later mcq in fractures associated with specularite and apatite may be related to the diagenesis of the Thelon Formation.

In areas where silcrete did form, it was largely removed by a later erosional event, as is shown by its abundant

occurrence as clasts in the basal Thelon conglomerates. This truncated, silicified surface is thought to serve as basement to the Thelon Formation throughout most of the Aberdeen and Schultz Lake map areas.

Diagenetic changes to the regolith are discussed later in this chapter.

PROFILE DEVELOPED ON GRANODIORITE GNEISS DH 82-23

This core was drilled by Anaconda during the summer of 1982. The author sampled it less than one week after it was retrieved. It was drilled vertically through semi-lithified till, the same till which is common in other drill core profiles as fracture fills down to 40m within the bedrock. At 17m, a dark brown regolith was penetrated and the hole progressed through at least 40m of regolith. Gradually the dark brown coloration gives way to yellows and finally green. The hole was drilled on the southern shore of Aberdeen Lake on the Anaconda property, but the exact location remains unknown.

The protolith of the regolith is a plagioclase (oligoclase-andesine)-quartz-hornblende-biotite-potassium feldspar granodiorite to tonalite gneiss. Mafic minerals originally ranged from 15-35%. Xenoliths of amphibolite are common, and were encountered in at least two of the core samples (Plate 1A). Although replaced and highly altered, it is apparent that they mostly consist of amphibole and plagioclase, the plagioclase being slightly subordinate. The upper 10m of the regolith is highly friable and was impregnated for thin-

section preparation.

MINERALOGY

Quartz occurs as highly elongate, strained grains up to 2mm in diameter. Many grains show biaxial interference figures. Some anhedral to subhedral quartz occurs as floating lenses in larger fractures surrounded by kaolin and illite. It is thought to be microcrystalline, because no interference figures could be obtained. In the upper few meters of the profile, quartz shows extreme dissolution and replacement by goethite.

Feldspar occurs as slightly more equant and larger grains pseudomorphed by kaolin and illite. The pseudomorphs in places are stained by iron oxides (goethite) or contain minor amounts of chlorite at depth. Pseudomorphs commonly show relic twinning marked by secondary clay growth along microcline or albite twins. In addition, optical orientation of clays in pseudomorphed feldspar can mimic Carlsbad twinning or perthitic intergrowths. Most fragments of relic feldspar in the pseudomorphs are irregular or elongate parallel to cleavage. Rhombohedral fragments and elongate strips commonly reflect shapes of the original plagioclase grains. Fresh to slightly altered sericitic plagioclase occurs at 45m in a corestone which has relatively little alteration.

Goethite occurs in the upper 25m of the profile as rod-shaped pseudomorphs after amphiboles or as anhedral, shapeless inclusions. Many of these rods outline in skeletal form the amphibole cleavages. Between the rods chlorite, illite,

kaolin, and secondary silica can be found. Goethite also occurs in books replacing chlorite and interlayered with it. Staining of clays by goethite gives the rock a dark brown colour. In the upper layers where goethite is abundant, the rock smears easily and its slippery feel indicates the presence of considerable amounts of water. Without goethite, at depth the rock becomes considerably better indurated.

Chlorite occurs in fresh rock as pseudomorphs after biotite, as slim remnant cleavage fragments in goethite books, at depth as one of the clays pseudomorphing feldspars, and as tiny veinlets following quartz grain boundaries. At the base of the section where chlorite is abundant and fresh, it shows distinct green pleochroism and anomalous purple to brown birefringence. The pleochroism may be related to high iron content (Deer, Howie, and Zussman, 1966).

Illite and kaolin occur as pseudomorphs after feldspar, in the interstices of goethite rods pseudomorphing amphiboles, and as veins in fractures up to 1cm wide. Illite appears to replace kaolin. Kaolin dominates in the top few meters where goethite is particularly abundant, suggesting neoformation during Pleistocene times related to intense leaching.

Hornblende occurs in the fresh rock in amounts up to 15%. It is quickly weathered to chlorite, because as it is rare to find both in the same area in a thin section. Pyrite also occurs as euhedral cubes in the fresh rock, and is pseudomorphed by goethite in the upper part of the profile. Its presence in veins as cubes floating in illite and kaolin

indicate that it formed during weathering as well as being an accessory in fresh rock. This may indicate dissolution followed by illuvation and precipitation beneath the water table under reducing conditions.

A wide variety of relic accessory minerals were noted. The most stable appear to be zircon and apatite. Original opaques such as magnetite and ilmenite show alteration to goethite, and break apart along fractures. Some euhedral sphene is pseudomorphically replaced by leucoxene (Plate 4B).

On the basis of petrography, this profile has been subdivided into two zones. The upper zone, extending from 17-40m, has been designated as the goethite zone (Plate 1F). It is characterized by brown coloration, distinct friability in the upper parts, and more than 5% goethite. From 41-61m is characterized by green colors, lack of goethite, and an increase downwards in chlorite. It is correlated with the chlorite zone of DH 81-1 (Plate 1H). The boundary between the two zones is diffuse and arbitrarily placed at 40m.

Figure 2-5 shows the modal mineralogy in profile 82-23 with depth. Before explaining the observed trends, a number of anomalies should be pointed out. At 23m, a distinct change in the mineralogy is noted. This core sample is part of an amphibolite xenolith. Fresh rock occurring at 45-47m is interpreted as a corestone floating in altered rock resulting from downward migration of the weathering front along fractures (Plate 1G). Any trends considered must take these anomalies into account. The extreme alteration at

Figure 2-5. Modal mineralogy of the regolith profile developed on a granodiorite gneiss with depth (DH 82-23).

depth (complete removal of feldspar) is thought to be related to fracturing observed as clay fracture fills from 53-61m.

The observed mineralogy trends are as follows:

- 1) Feldspar is almost completely pseudomorphed except in corestones. A preferential destruction of plagioclase over potassium feldspar is noted.
- 2) Kaolin and illite content is directly related to the state of feldspar preservation.
- 3) Goethite and chlorite show a direct inverse relationship to one another. Goethite decreases downwards from the surface as chlorite increases. This relationship suggests that the goethite is an alteration product of chlorite, as well as of hematite.
- 4) Pyrite occurs only at depth, although pseudomorphs of it can be seen higher in the profile. This suggests that it was not stable during the original weathering event.
- 5) Hornblende, an original component of the protolith, occurs only in relic corestones where alteration is insignificant.
- 6) Quartz content increases slightly down profile. Quartz grains in upper parts of the core are somewhat corroded, indicating that this trend can attributed to dissolution.
- 7) Illite and kaolin have an inverse relationship with feldspar.
- 8) Remnant hematite in upper parts of the profile show evidence of its replacement by goethite.

INTERPRETATION

The proposed sequence of events is diagrammatically illustrated in Figure 2-6. Original weathering created a profile, similar to DH 81-1, on granodioritic gneiss containing amphibolite xenoliths. The profile was deeply eroded, perhaps as a result of uplift during faulting in the area. The lack of secondary silica indicates that the effect of any surface silicification has been removed. The profile was covered by Thelon Formation. Most inherited diagenetic effects have been masked due to later alteration. The profile was exhumed sometime after Thelon deposition, perhaps during Ordovician time (an Ordovician outlier rests on the Thelon Formation 30km to the north). Alternatively, the profile may have been exhumed during Pleistocene erosion of the Thelon Formation. The truncated profile was covered by Pleistocene till, and most likely was affected by ground water flowing through the overlying till. Mineralogical changes would have occurred as the regolith's mineralogy equilibrated to the new environmental conditions.

The principal changes appear to be reweathering and hydration of well-ordered diagenetic clay minerals and the alteration of hematite and chlorite to goethite. The goethite alteration is thought to be of Pleistocene age rather than original, because in both profiles where it is occurs (82-20 and 82-23) it is overlain by a semi-lithified till. The regolith at the top of these profiles is extremely soft and preservation since Precambrian time is unlikely; the lack of goethite and other hydrated phases in

Figure 2-6. Interpretation of the regolith profile developed on granodiorite gneiss (DH 82-23).

ANACONDA DH 82-23

Protolith (Granodioritic-Tonalitic Gneiss)

1. Original Weathering Profile

2. Erosion and Superposition of Zones

3. Deposition of Thelon Formation

4. Erosion of the Thelon Formation, Deposition of Pleistocene Till, and Recent Weathering

other profiles agrees with the proposition that hydrated phases are metastable and would alter in time (in the case of goethite, to hematite).

Xenoliths occur in this profile and can be identified by mineralogical and textural differences in an intensely altered rock. Well preserved corestones have been reported by Chown and Caty (1983) in the Aphebian Mistassini regolith.

ADDITIONAL PROFILES

A number of additional profiles were sampled and observed in the field. Figure 2-7 represents the modal mineralogy with depth of DH 81-2. This profile was drilled close to DH 81-1, but, shows few signs of silicification. The original lateritic alteration is identical to that displayed by DH 81-1. The protolith of this profile was probably a metawacke. The unconformity with the overlying Thelon Formation is irregular in that in the core it is inclined at 70° to horizontal (Plate 1A). A 1-2 cm rim of recrystallized specularite outlines the contact. No redox spotting occurs. The upper few meters of this profile show destruction of the original foliation, and may indicate development of a laterite, now removed.

Several additional profiles from the Anaconda property in the Marjorie Hills were slabbed and studied under the binocular microscope. Abundant breccias, fractures, and intrusions are thought to be related to faulting and volcanic activity synchronous with the Pitz Formation. These features destroy any original zonation, and hematitic coloration is

Figure 2-7. Modal mineralogy of the regolith profile developed on a feldspathic metawacke with depth (DH 81-2).

prevalent throughout (although remnant bleached areas and chlorite-rich zones are sporadically preserved). In two profiles a distinctive fault breccia was observed, both at depths greater than 100m. It consists of hematitic regolith fragments in hematitic chert showing multiple generations of brecciation and silicification. In some of these profiles intrusive dikes apparently have exploded, presumably when they hit the water table. Angular regolith fragments are commonly incorporated in silicified intrusive material.

In DH 82-7 a rapikivi intrusion (Plate 3A) shows differential alteration of plagioclase rims to dolomite and potassium feldspar cores to kaolin and illite. The groundmass has altered to a fine-grained mixture of illite, kaolin, chlorite, silica, and hematite. Pitz Formation rhyolites show similar alteration. Large .5 - 2cm phenocrysts of potassium feldspar are pseudomorphed by kaolin and illite. Some clay pseudomorphs have been subsequently replaced by mcq. In Thelon Formation conglomerates, rhyolite cobbles contain both recessive-weathering pseudomorphed phenocrysts and resistant phenocrysts related to this silicification. Regolith developed in different lithologies is shown on Plate 5.

FRACTURE MINERALOGY

Occurrence of a number of minerals atypical of regolith mineralogy occurs in fractures. Barite, fuchsite, gypsum, apatite, and dolomite have been noted. Spectacular fracture fills of coarse clays (nacrite and dickite) in the regolith

probably represent filtering of material down through the Thelon Formation. These occurrences represent diagenetic accumulations related to localized geochemical anomalies or may be associated with evaporitic climatic conditions discussed in Chapter 7. Figure 2-8 shows diffractograms for several of these minerals. Figure 2-9 shows the infra-red pattern for nacrite from a fracture in silcrete.

CLAY MINERALOGY

X-ray diffraction, petrographic techniques, and infra-red spectra were utilized in the identification of the clay mineralogy of the regolith. Oriented clay mounts were used to determine basal spacings of modal clay minerals in whole rock samples of <2u and between 2-10u. Random samples ran at the Soil Chemistry and Mineralogy Lab at Agriculture Canada were investigated by Dr. H. Kodama to determine clay polytypes. Hydrated, oriented samples were used to check for the presence of "swelling clays". Infra-red techniques were applied to determine kaolin polymorphs and the presence of kaolin in a kaolin-chlorite mixture.

DH 81-1

Over 35 diffraction patterns were run for DH 81-1 to determine the clay mineralogy. The results are as follows:

- 1) little difference in intensity and no difference in composition was found between <2u and 2-10u populations for the same sample. This may be related to the large grain size of most of the clays observed; the <2u fraction being

Figure 2-8. X-ray diffraction patterns for barite (B), dolomite (D), and Nacrite (N) from fractures in the regolith.

Figure 2-9. Infra-red spectra from sub-Thelon nacrite in fracture. A, B, and C patterns for dickite, nacrite, and kaolinite, respectively, (from Kodama and Oinuma, 1963).

Table 2-3. Comparison of sub-Thelon nacrite, dolomite, and barite x-ray diffraction patterns with JCPDS standards.

Sub-Thelon Nacrite (oriented)		JCPDS standard 16-606 (random)		
d-space	I	d-space	I	hkl
7.9	24	-	-	beta peak
7.1	100+	7.18	100	002
4.38	22	4.36	80	202
4.13	21	4.13	70	200
3.98	26	3.94	20	111
3.57	100+	3.59	80	004
3.34	30	3.41	20	112
2.70	30	-	-	-
2.52	27	2.53	20	021
2.40	100+	2.404	40	006
1.79	65	1.797	6	008

Sub-Thelon Dolomite		JCPDS standard 11-78		
d-space	I	d-space	I	hkl
3.69	17	3.69	5	012
2.88	100+	2.886	100	104
2.66	16	2.670	10	006
2.53	12	2.540	10	015
2.40	10	2.405	10	110
2.19	15	2.192	30	113
2.01	12	2.015	15	202
1.80	13	1.804	20	018
1.78	13	1.781	30	009

Sub-Thelon Barite		JCPDS standard 24-1035		
d-space	I	d-space	I	hkl
4.44	13	4.44	16	110
4.33	14	4.339	30	101
3.90	18	3.899	50	111
3.57	100	3.577	30	200
3.44	100	3.445	100	021
3.31	32	3.319	70	210
3.09	35	3.103	95	121
2.84	18	2.836	50	211
2.73	21	2.729	45	130
2.32	10	2.325	14	022
2.30	9	2.305	6	310
2.21	10	2.211	25	122
2.12	18	2.121	80	311
2.10	18	2.106	75	212

composed primarily of fragments broken off larger grains during preparation; 2) the clays identified were kaolins, illites, and chlorites; 3) most of the kaolin was kaolinite, but some nacrite was identified in the present study, and dickite has been reported before (Cecile, 1973). This indicates that pressures and temperatures stayed within the range in which kaolinite can exist, and only locally did environmental conditions produce nacrite (fractures); 4) X-ray work interpreted by Dr. H. Kodama indicates the presence of the 1M polymorph of illite. However, the large grain-size and low crystallinity index (highly crystalline), suggests that the 2M polymorph is present also. Dean (1967) found 2M illites in Thelon sandstones from various parts of the Thelon Basin; and 5) Aluminous chlorite has been reported in Thelon sandstones by Cecile (1973) and Miller (1983). MacDonald (1980) also found dioctahedral aluminous chlorite in the sub-Athabasca regolith. In the sub-Thelon regolith at least two types of chlorite probably are present: 1) highly crystalline, iron-rich magnesium chlorite; and 2) iron chlorite. The iron-rich magnesium chlorite is clinocllore and may be original chlorite which has been unaffected by the weathering event. Its petrographic relationships (parallel to foliation) and large grain size concur with this interpretation. It was identified by random x-ray diffraction samples. The other chlorite shows brilliant green pleochroism, smaller grain size, and textural relationships indicating occurrence as an alteration product. It is common as small veinlets and associated with clay pseudomorphs of

feldspars. This chlorite may be similar to that described from feldspar pseudomorphs by Miller (1983). Its presence is based on petrographic examination, because the more abundant "remnant chlorite" masks its x-ray properties.

The possibility exists that other minor clay components occur in this regolith profile. A peak at 19A in the x-ray diffraction pattern for 157m may represent a smectite/illite interlayered clay incompletely altered to illite. However, no other such peaks were observed (Figure 2-10). Hydration of these clays caused no change in the diffraction patterns. The amorphous hump observed in this profile is caused by the high content of iron, which is excited by copper radiation.

DH 82-23

X-ray diffraction patterns were run for more than 35 clay size samples of this profile. Its clay mineralogy proved to be more complex because of the effects of recent weathering. In the lower part of the profile the clay mineralogy is essentially the same as that of profile 81-1. The major difference is the presence of a fine-grained white mica (sericite) formed as an alteration product on plagioclase grains.

The upper part of the profile is dominated by illite, kaolin and/or chlorite, goethite, and possibly vermiculite. clay minerals. This assemblage is identical to that described by Hill and Tedrow (1961) occurring in arctic soils developed on till. The goethite staining and fine grain size makes petrographic techniques of limited value in this part

Figure 2-10. Oriented x-ray diffraction patterns from clay-sized material from the hematite (H), transitional (T), and chlorite (C) zones of DH 81-1.

of the regolith. Illite shows various crystallinities as its inclined pattern in Figure 2-12 indicates. Weathering of some of the highly crystalline, diagenetic illite may account for this. Both kaolin and chlorite are probably present.

The presence of an intense peak at 14A, in several of the samples run from the goethite zone, is problematic. The 002 chlorite peak, which would occur at 7A, should not be so weak compared to the 001 peak. This indicates the presence of another clay mineral. Glycolation of the samples with this "oversized" 001 peak showed no change. The 14A peak because is probably vermiculite. In addition, the presence of a small percentage of expandable clays is expected, as some samples swell when wet. They were not identified in x-ray patterns, however. Figure 2-11 shows x-ray diffraction patterns for samples from DH 82-23.

DIAGENETIC ALTERATION

Diagenesis of the Thelon Formation has been discussed by Cecile (1973) and Miller (1983). Cecile outlined three diagenetic stages occurring in the following order: 1) lithification, development of quartz overgrowths and partial disintegration of clay-silica aggregates (thought to be derived from underlying paleosols); 2) reaction of quartz grains and quartz overgrowths with the surrounding matrix and recrystallization of clay phases; and 3) the growth of muscovite (illite) as rims around detrital grains, microlites within the matrix, kaolin pseudomorphs, and in veins. Authigenic K-feldspar was observed in several samples

Figure 2-11. Oriented diffraction patterns for clay-sized material of the goethite zone (G), chlorite zone (C), and corestone (CS) of DH 82-23.

studied in thin-section. The muscovite was considered by Cecile to be secondary due to its replacing relationship with kaolinite. He concluded that muscovite formed from kaolinite during burial.

Miller (1983) observed that diagenesis of the Thelon Formation produced multiple generations of quartz, kandite group clays, muscovites, aluminous chlorite, fluorapatite, and goyazite. Phosphates, illite, and specular hematite were also observed in vein systems in the underlying regolith (Plate 6D). The apparent chemical maturity of much of the Thelon Formation was concluded to be a result of diagenetic replacement of feldspar by kaolin + quartz + illite in non-phosphatic units and by fluorapatite + illite in phosphate-bearing units. Quartz cements are locally replaced by hematitic fluorapatite. Miller suggested that very low Na_2O contents in the Thelon Formation sediments reflect diagenetic feldspar dissolution and the absence of smectite clay minerals. On the basis of work by Nriagu (1976) he concluded that the common illite-fluorapatite assemblage indicates a pH of 6.8-7.5, which would explain the observed dissolution of quartz.

Diagenesis of the sub-Thelon regolith is complicated by several factors: 1) mineralogical zonation in the original profile would influence cation concentrations, resulting in different interstitial water compositions; 2) how much influence diagenetic waters from the overlying Thelon Formation have had is open to speculation; 3) the influence of silicification in the regolith from silcretes which

developed above it is unknown, but it probably reduced the permeability of the regolith; 4) differing lithologies would have various permeabilities; and 5) some of the regolith may be as much as 200my older than the Thelon Formation which overlies it, and may have undergone any number of diagenetic or regressive diagenetic events before being covered by the Thelon Formation.

The original mineralogy of the weathering profile was discussed previously in this chapter. It is thought to have been inhomogeneous, with vertical variations reflecting different leaching intensities. Although the profile was truncated, superposition of zones resulted in vertical variations within the saprolite.

Any consideration of diagenetic processes must take into account the presence of hematite and illite which are rare in modern lateritic profiles, and the highly crystalline nature of many of the clay minerals in the profile.

Goethite and other iron hydroxides make up the chief iron-bearing phases in laterite horizons. Argument exists over whether or not hematite may form as a primary component of laterite horizons or red beds (Chandler, 1980). However, with time and the increased temperatures of burial, the stable iron-bearing phase in a lateritic profile would be hematite.

Kaolinite is stable in acidic burial conditions but unstable in alkaline conditions. The stability of kaolinite is controlled more by pH and the presence of organic matter than by temperature. Kaolinite can persist at depths

corresponding to 80-180°C (Dunoyer de Segonzon, 1969). Hoffman and Hower (1979) observed that kaolinite alters to dickite in the range of 100-150°C in sandstones. Nacrite formation is thought to depend on pressure, as its common occurrence in slickensided fractures suggests. With increased temperature, all kaolin phases will recrystallize to illite or chlorite in the presence of K^+ and Mg^{++} . Illitization of kaolin is common with depth in bore holes (Dunoyer de Segonzon, 1970).

Smectite eventually alters by dehydration and cation substitution to illite with burial through the formation of a number of smectite/illite interlayered clays. The proportion of illite layers in the smectite/illite clays is considered the most sensitive indicator of diagenetic grade in shales (Hoffman and Hower, 1979). Boles and Franks (1979) noted that smectites disappear at temperatures greater than 70°C. The complete transition of smectite to illite is thought to occur between 175-200°C. In addition, the diagenetic alteration of smectite to chlorite by the loss of water has been observed (Dunoyer de Segonzon, 1970).

Chlorite in diagenetically altered weathering profiles can be of primary origin or a recrystallization product of kaolin or smectite. From his investigation of chlorite polymorphs, Hayes (1970) established the sequence of type $Ib_d \rightarrow Ib \rightarrow IIb$ with increasing temperature. Type Ib_d chlorites do not exist at temperatures above 70-80°C, and the transition of Ib to IIb occurs coincidental with the change from $IM \rightarrow 2M$ illite at 175-200°C. MacDonald (1980) concluded

that the Mg-dioctahedral chlorite in the Athabasca regolith was a product of diagenetic recrystallization of Mg-smectite.

The transition of illite to a well-crystallized mica has been documented in many studies. The Crystallinity Index of illite is taken to represent the depth of burial. The Crystallinity Index is measured as one-half of the intensity of the 001 reflection, and is conventionally plotted against the intensity ratio of 002/001. The 002/001 ratio is a measure of the chemical composition of illite (Al/Fe+Mg) (Esquevin, 1969). The CI Index of illite is dependent on the transition of IM_d to IM and 2M illite. Low grade metamorphic rocks have a stable clay mineralogy of illite + chlorite (Hoffman and Hover, 1979).

Diagenetic alteration to the Pitz regolith involved the following changes:

- 1) the alteration of goethite and other iron hydroxides to hematite (Walker, 1974).
- 2) partial illitization of kaolinite, localized formation of dickite where acidic conditions prevailed during diagenesis, and formation of nacrite in fractures.
- 3) dehydration of any smectite clay minerals to illite or chlorite.
- 4) transition of smectite/illite clays to illite.
- 5) the recrystallization of illite from IM_d to IM and 2M polymorphs.
- 6) the recrystallization of chlorite from Ib_d to Ib and IIb polymorphs.
- 7) the precipitation of phosphate-bearing phases, illite, and

specularite in fractures.

8) partial recrystallization of hematite to specularite at the regolith/Thelon Formation unconformity in voids and fractures.

9) minor redox spotting and reduction along fractures in the hematite zone of the regolith.

10) replacement, and filling of fractures by dolomite.

Recrystallization of illite is apparent from Figure 2-12. The data from DH 81-1 compares well with data from the Athabasca regolith from MacDonald (1980). The elongate horizontal pattern with a relatively low Crystallinity Index (highly crystalline) is considered representative of the lower zone of Late Diagenesis where the 2M form of illite predominates. Petrographic observations confirm the highly crystalline nature of the illite as many illite crystals are > 20 μ in size and orientated illite stringers are found. The identification of 1M illite in the sub-Thelon regolith may suggest burial temperatures were not sufficient to complete the 1M to 2M transition.

Dolomite replacement and fracture fill is thought to represent one of the last diagenetic events to affect the regolith. Its occurrence replacing all other minerals with the exception of chlorite and in late fractures suggests this. Dolomite has not been reported in the Thelon Formation. This may indicate that its occurrence is restricted to the regolith. An example of baroque dolomite lining a fracture was noted in thin-section. Baroque

Figure 2-12. Comparison of Illite Crystallinity Indices from the sub-Athabasca and sub-Thelon regoliths.

dolomite is common as a void-filling or replacement mineral (Rudke and Mathis, 1980), and is thought to be indicative of temperatures of 50-150°C.

Diagenesis of the sub-Thelon regolith was preceded by a number of overlapping reactions related to temperature and depth of burial (Figure 2-13). Postulated temperatures between 150-200°C are consistent with the loss of all smectite minerals, absence of smectite/illite clays, the partial recrystallization of kaolin to illite, and the crystallinity index of the illite present. Assuming a geothermal gradient of 30°C/km and an initial surface temperature of 30°C, burial estimates for the regolith vary from 4-5.6km for diagenetic temperatures of 150-200°C, respectively. Geophysical work by Overton (1971) suggested that the Thelon Formation is ~2km thick near the center of the basin, which would suggest that 2km of the Thelon Formation have been removed by erosion from this area.

Although the influence of diagenetic waters of the Thelon Formation was significant in fractures, and caused redox spotting in the hematite zone of the regolith, the influence on diagenesis within the regolith was minimal. Thelon diagenetic waters were slightly alkaline, thought to vary in Eh from reducing to oxidizing, and contained significant concentrations of PO_4 , K, and Fe. Regolith diagenetic waters are thought to have been vertically stratified due to original mineralogical differences. In the hematite zone they were probably less alkaline than in the underlying chlorite zone.

Figure 2-13. Clay mineral diagenetic sequences as a function of burial temperatures (after Hoffman and Hower, 1979; Hayes, 1970; Dunoyer de Segonzac, 1969; Rudke and Mathis, 1980).

Profile 82-23 was affected by a later regressive diagenetic event (surficial weathering) during Pleistocene times. Evidence for this includes the pattern displayed by illite crystallinities in Figure 2-12, broadening of the chlorite peaks indicating a loss in crystallinity, alteration of hematite to goethite, and the presence of swelling clays. In general the reverse diagenetic trends shown are hydration and decrease in ordering of the clay minerals.

The pattern shown on Figure 2-12 is identical to that proposed by Dunoyer de Segonzac for the upper zone of Late Diagenesis where three polymorphs of illite may exist together (IM_d , IM, 2M). The weathering conditions imposed on the regolith during the Pleistocene would promote weathering of diagenetic illites and the lowering of their degree of crystallinity. Illites lower in the profile (unaffected by recent weathering events) would retain the high degree of crystallinity attained during diagenesis.

CONCLUSIONS

1. In the study area, intensive chemical weathering affecting rocks of the Dubawnt Group and underlying basement occurred over a period of at least 150 million years. This weathering resulted in formation of a number of superimposed weathering profiles interrupted by sedimentation, Pitz Formation volcanism, and erosion. The oldest profile which developed in Aphebian and/or Archean basement rocks is the best developed and thickest. Although no complete profiles with intact upper soil horizons were found, thickness of 20-

2

60m are typical. Some extremely thick profiles (> 100m) are interpreted as being influenced by faulting and alteration along fault zones. Subsequent profiles developed on Pitz Formation volcanics are thinner and show little evidence of zonation.

2. In well-developed profiles on basement rocks, distinct mineralogical zones can be separated on the basis of colour. A typical sequence shown in most profiles studied includes, from the top down, a hematite or red zone, a transitional zone, and a chlorite or green zone. The formation of these zones is secondary and related to the transposition of geochemical zones downward during erosional truncation of the profiles. The evidence supporting this includes the lack of preserved upper soil horizons, erosional contact with the overlying Thelon Formation, abundant regolith clasts in the Thelon Formation, and remnant chlorite blebs in areas thoroughly invaded by hematite. Mineralogical zonation did not develop in profiles on Pitz Formation rocks, where the entire profiles are hematitic. This is related to the shorter time these profiles had to develop, high suspected erosional losses, and a rapid loss of water through abundant fractures and joints in the rhyolites, causing intensive leaching and lowering of the water table.

Many profiles also show silicification linked to development of nongenetically related silcretes above the truncated weathering profiles. Silicification takes two forms: replacement of clays and hematite by mcq, and abundant veining parallel and perpendicular to foliation and

in fractures. Depending on pre-Thelon erosion losses to the profiles, silicification and silcretes may or may not be preserved.

3. Leaching, climate, topography, and protolithology each had a direct influence on deep weathering in the study area. The climate provided abundant rainfall and tropical temperatures to facilitate chemical weathering. The paleotopography was most likely flat with some hills up to 50m in height, which would provide adequate drainage without extreme erosional rates. The depth of weathering varied in different lithologies. Highly fractured and strongly foliated rocks were most conducive to weathering, whereas quartzites altered only slightly along fractures. The combination of paleotopography and paleoclimate produced extreme leaching conditions.

The severity of leaching is shown by the depth of the profiles and the almost complete removal of most components except quartz, iron, and aluminum. This suggests that acidic pH conditions prevailed throughout weathering. The absence of organic material would provide additional free oxygen (whether this additional oxygen was significant is unknown).

4. Diagenesis within the weathering profiles has been documented. In profiles of this age amorphous material, hydrated phases, and certain other minerals would be metastable. Dolomite has been observed in most profiles as filling fractures and as a replacement mineral. In DH 81-1 at 198m, replacement dolomite accounts for almost 30 % of the modal mineralogy. The common association of apatite, illite,

and specularite is noted in fractures in the regolith, and also in the overlying Thelon Formation (Miller, 1983). The relationship between the dolomite and the previous association mentioned is uncertain. The dolomite may represent a unique event, because dolomite is absent from lower Thelon Formation rocks, or it may reflect a stratified diagenetic sequence, in which dolomite was precipitated only in areas containing magnesium, such as the chlorite zone (magnesium is thought to have been thoroughly leached from the Thelon Formation). The Illite Crystallinity Index for regolith samples is compared to data from MacDonald (1980) for the Athabasca Basin in Figure 2-10. Both groups of samples show identical results, indicating that much of the illite in the regolith is recrystallized and diagenetic in origin. This is backed up by the large size ($> 20\mu$), oriented morphology, and mineralogical association of many of the illites. How much of the illite is diagenetic is unknown, but kaolin probably dominated the original clay mineralogy. The presence of baroque dolomite in several of the sections studied indicates temperatures of $50-150^{\circ}\text{C}$ (Rudke and Mathis, 1980), comparable to those determined for the Athabasca Formation on the basis of fluid inclusion studies (Pagel, 1975). Formation waters affecting diagenesis in the regolith and lower Thelon Formation probably had a slightly alkaline pH (by comparison with stability field data of Nriagu (1976) for apatite, illite, quartz, and specularite). In addition, bleached zones in the Thelon Formation and redox spotting in the upper parts of the

hematitic zone of the regolith are common, indicating iron mobility and local reducing conditions.

5. The effects of Pleistocene weathering on the initial mid-Proterozoic weathering profile are displayed in DH 82-23. The major changes include formation of goethite from hematite and chlorite, dissolution of primary quartz, a significant decrease in induration, abundant water and presence of unidentified clay minerals (most likely smectites) not present in other profiles, and a predominance of yellow to brownish colours. Evidence for a Pleistocene origin for this secondary weathering event includes: the only two profiles displaying it are overlain by Pleistocene till, the presence of goethite as the primary iron-bearing phase, hydrated clays, and the low probability of such poorly indurated material being preserved for more than a few thousand years.

6. Geochemical studies of ancient weathering profiles can be substantially enhanced by detailed petrographic studies. The effects of diagenesis and recent alteration have been documented in this study, which place severe limitations on the interpretation of any geochemical studies. In profiles such as these, development on inhomogeneous metasedimentary rocks also severely limits any advantages gained from mass balance studies (cf. Gresens, 1967) as original variation in aluminum, generally assumed for comparison purposes to remain stable throughout weathering, can be considerable. A similar conclusion was reached by MacDonald (1980), who carried out the Gresens procedure on samples from the Athabasca Basin Regolith. Isovolumetric

investigations as described by Gardner et al (1970) also have limited value in studies where little restructuring or volume changes have occurred. However, in profiles where upper horizons as well as saprolite are preserved, post-burial compaction or original expansion may necessitate substantial corrections for volume change.

7. The deep weathering profiles of the study area are considered lateritic. Even though preserved laterite horizons are lacking due to erosional truncation, the amount of iron in the hematite zone, concentration of Cr, V, and Ni from mudstone directly above the truncated regolith (DH 81-4 364m, Table 7-1) and the profile zonation all are suggestive of lateritic affinities.

A number of Tertiary analogs to the profile described in DH 81-1 are recorded in the literature. Jessup (1960) described laterite profiles overlain by silcrete from Australia. Melfi et al (1981) described Tertiary profiles from Brazil which show truncation of a lateritic profile, subsequent development of silcrete, silcification of the underlying profile, and in some areas, removal of the silcrete so that only the silicified lateritic profile remains.

Plate 1. Mineralogical zonations in the sub-Thelon regolith.

A. Unconformable contact between sandstone of the Thelon Formation and hematite zone of the regolith. Dark rim of specularite outlines the contact. The steep nature of the unconformity is suggestive of a fracture fill.

B. Hematite zone of the regolith. Note absence of foliation and hematite-stained clay pseudomorphs of feldspar.

C. Transitional zone of the regolith. Dark areas contain hematite which has replaced the chlorite-rich lighter areas. Note retention of the original foliation and its apparent control over hematitic alteration.

D. Chlorite zone of the regolith. Darker layers are chlorite, and large white areas are feldspar pseudomorphs composed of illite. Protolith is a mylonitic gneiss.

E. Amphibolite xenolith in regolith developed on granodiorite gneiss. Dark coloration due to goethite staining during Pleistocene alteration.

F. Goethite zone of regolith developed in granodiorite gneiss.

G. Fresh corestone of granodiorite gneiss.

H. Chlorite zone of regolith unaffected by Pleistocene weathering. Vertical fracture filled by alternating layers of illite, mcq, and minor pyrite.

Plate I

Plate 2. Photomicrographs of textures in DH 81-1.

A. Hematite zoné of the regolith. Composed of hematite (black), quartz (white), and illite (light-coloured matrix). (plane polarized light, scale bar = 1mm)

B. Transitional zone of the regolith. In top half of photograph chlorite is fresh; in bottom half chlorite is altered to darker hematite. (plane polarized light, scale bar = 1mm)

C. Chlorite zone of the regolith. Illite pseudomorphs (P) of feldspar are present. (crossed nicols, scale bar = 1mm)

D. Chlorite grain split along cleavage and altered to hematite. Light area is mcq. (plane polarized light, scale bar = 50mu)

E. Diagenetic replacement of kaolin (dark) by illite (light) in transitional zone of the regolith. Slight replacement of a muscovite grain (M) by illite is also apparent. (plane polarized light, scale bar = 250mu)

F. Matted texture of swelled and split muscovite in hematite zone of the regolith. (plane polarized light, scale bar = 250mu)

Plate 2

Plate 3. Secondary alteration of the regolith.

A. Hematite zone of the regolith cut by quartz veins sub-parallel and sub-perpendicular to foliation.

B. Dolomite fracture-fill parallel to foliation in transitional zone of the regolith. Increased hematite content along fractures gives the rock a dark coloration.

C. Regolith developed in a rapakivi-textured intrusive porphyry. The darker, outer rim on the largest phenocryst is dolomite that has pseudomorphed the plagioclase rim. Lighter, inner area is composed of illite and kaolin that have pseudomorphed K-spar.

D. Silicified breccia zone containing fragments of hematitic chert (C) and regolith 104m below the Thelon Formation/regolith unconformity.

E. Hematite zone of the regolith displaying ellipsoidal reduction spots, devoid of hematite and parallel to foliation.

F. Remnant chlorite bleb in transitional zone of the regolith.

G. Reduction parallel to foliation and fractures in the hematite zone of the regolith.

H. Thelon Formation mudstone showing reduction parallel to bedding and fractures, and in small spots.

Plate 3

A

B

C

D

E

F

G

H

81-1 163m

81-1 190m

81-1 159m

81-2 182m

Plate 4. Photomicrographs of diagenetic features of the regolith.

A. Amphibolite xenolith displaying Pleistocene goethitic alteration of hematite rods which pseudomorphed original amphibole cleavage. (plane polarized light, scale bar = 1mm)

B. Sphene pseudomorphed by leucoxene in regolith developed in granodiorite gneiss. (crossed nicols, scale bar = 1mm)

C. Interlayered geopetal dolomite (dark) - illite (light) fracture fill in the transitional zone of the regolith. (crossed nicols, scale bar = 1mm)

D. Microcline grain partially replaced by dolomite. (crossed nicols, scale bar = 250mu).

E. Silicified, ferruginous paleosol containing pedorelics (P). Differences in iron-oxide content of a pedorelic (glæbule) are apparent in the lower right corner. (plane polarized light, scale bar = 1mm)

F. Folded kaolin fracture fills in regolith developed in a Pitz Formation rhyolite. The lower fracture has been replaced by illite (white). (crossed nicols, scale bar = 1mm)

Plate 4

Plate 5. Regolith developed in different lithologies.

A. Regolithic augen gneiss. Note slight rotation of clay-pseudomorphed feldspar augens and quartz grains, and hematitic mud matrix.

B. Saprolite developed on a fine-grained psammitic rock.

C. Regolithic granitic gneiss.

D. Kaolin-filled ortho-joint planes developed in a Pitz Formation rhyolite. Field of view=8cm. Photomicrograph on Plate 4F.

E. Silicified, ferruginous paleosol with pedorelics. Note erosional contact (E) separating intact and reworked paleosol. Photomicrograph Plate 4E.

F. Regolith developed in Pitz Formation rhyolite with flattened clay-pseudomorphed phenocrysts. The flattening is related to compression or creep while the regolith was soft.

Plate 5

Plate 6. Diagenetic apatite in the regolith and Thelon Formation.

A. Apatite (high relief), in hematite zone of the regolith, in clay pseudomorphs of feldspars. (plane polarized light, scale bar = 250 μ)

B. Zoned apatite crystals associated with illite (white) and specularite (dark) in a void between clasts of Thelon Formation conglomerate. (crossed nicols, scale bar .5mm)

C. Apatite-rimmed clasts and void-fills in Thelon conglomerate. Note thicker rims on the right side of some clasts (R) and the initial darker layer.

D. Same sample as previous photograph showing reworked sandstone clasts (SS) coated by apatite; void-fill also composed of apatite. Concentrations of hematite give the darker colour to early rims. (plane polarized light, scale bar = 1mm)

E. Thelon mudstone showing desiccation crack at base of core, infilling of sand, and redox spotting. Mudstone rip-ups occur near top of the photograph.

F. Thelon conglomerate containing a large rounded clast of silcrete (S).

G. Thelon conglomerate containing clasts of regolith (R), Christopher Island volcanic rock (C), and iron formation (I).

H. Thelon conglomerate with clasts of quartzite (Q) and silcrete (S) which show two irregular concentric rims. The sheltered pore-fill contains hematite and clay.

Plate 6

CHAPTER 3 SILCRETES AND SILICIFICATION

INTRODUCTION

Silcretes are indurated, brittle, surficial and near-surface pedogenic accumulations of silica. Silcrete crusts display a wide range of colours with yellow, grey, purple, red and brown shades common. Fracture are conchoidal or irregular due to variations in crystallinity of the silica phases. A number of textures and structures have been observed: blocky, bulbous, massive, pisolitic, columnar, polygonal, platy, botryoidal, pillowy (Wopfner, 1978), and concentric, pillared, nodular, and piped (Watts, 1978).

Conglomeratic, brecciated, and fine-grained varieties have been reported, many of them showing several generations of silicification and brecciation. Most silcretes form flat, extensive, thin (<5m) sheet-like cappings that commonly are associated with weathering profiles and planation surfaces.

The mineralogy of silcretes is dominated by opal, chert (=mcq, microcrystalline quartz), chalcedony, and mega-quartz, but rarer forms of silica may also be present. The silica displays various replacive, cementing, and/or void-filling textures. Most silcretes are composed of greater than 90% silica, with minor amounts of other elements. Some silcretes are sufficiently enriched in titanium to show visible inclusions of anatase, leucoxene, or rutile in the siliceous matrix, and they also commonly contain anomalously high concentrations of other resistate elements, such as zirconium

and niobium (Hutton et al, 1978).

A wide variety of rocks serve as hosts to silcretes. The most common hosts are sandstones in which all non-siliceous material has been replaced and detrital quartz grains show dissolution features (fretting, embayment) and partial replacement by authigenic silica.

TERMINOLOGY

Confusion over the term "silcrete" arises from its extensive use for a variety of silicified rocks. Additional problems occur from overlapping terminology used by geologists, geomorphologists, and pedologists regarding soils and sediments. The "silcrete" of the geomorphologist may be the "quartz-overgrowth cemented arenite" of the geologist. "Duricrust" is a term applied broadly to all surficially cemented crusts, including calcrete and ferricrete, as well as silcrete.

Any definition for the word silcrete must encompass several of its distinctive properties: 1) formation by absolute accumulation of silica rather than by relative accumulation as a result of selective leaching of other components (Ollier, 1978); 2) formation at or near the surface under the influence of pedogenic processes (Hutton et al, 1978); 3) development in a host material which commonly consists of sedimentary or weathered rock (Williamson, 1957; Langford-Smith and Watts, 1978); and 4) subaerial occurrence. In the present study silcrete is defined as a non-marine rock formed by absolute accumulation of silica,

within a host material, within the zone of influence of pedogenic processes. This definition eliminates silica replacements of marine origin resembling silcretes in texture and mineralogy (Wilson, 1966) and silicified material having no relationship to pedogenetic processes. Included are silica-overgrowth cemented arenites which can be shown (by evidence of reworking, symsedimentary brecciation, or pedogenetic features) to be a product of near-surface silica accumulation. Possible exceptions include silcretes developed beneath basalt flows, and skin silcretes.

OCCURRENCE AND STRATIGRAPHY

In the study area, silcrete outcrops form elongate ridges that commonly dominate the subdued topography. Most have east-west or southeast-northwest trends due to northward tilting during later block faulting. Some silicified ridges are extensive, such as the scarps near Lone Gull (Chapter 5); others occur as discontinuous erosional remnants. Much of the silcrete is thought to have originally formed extensive sheets which were eroded before and during deposition of the Thelon Formation. Thicknesses range from 1m to more than 10m. Some silcretes developed in lithic arenites of the lower Thelon Formation may be considerably thicker, and are of polygenetic origin. Their unusual thickness (up to 50m) would imply that they were cemented at considerable depths, but geological evidence refutes this, and thin-section characteristics are identical to other sandstone-hosted silcretes described in the literature.

Silcretes rest on truncated regolith profiles and occur within Pitz Formation volcanics and lower Thelon Formation sediments, suggesting that they formed over a significant time period. Silcretes displaying multiple events of brecciation and silicification provide evidence of polygenetic origin. A number of stratigraphically distinct silcretes occur. In the study area silcrete has four main occurrences: 1) in the upper, preserved portion of truncated regolith profiles; 2) within the weathered upper boundary to Pitz Formation rhyolitic flows and crystal tuffs; 3) in near-surface and surficially exposed faults, fractures, and tectonic breccias and 4) within lithic sandstones and conglomerates of the lower Thelon Formation (Plates 7&8).

DESCRIPTION OF SILCRETES

Type 1 Silcretes

As previously mentioned, almost all regolith profiles in the study area show evidence of truncation. It is on such truncated surfaces that many of the silcretes defined here as "type 1" are developed (similar occurrences are described in Australia by Jessup (1960) and in Brazil by Melfi et al (1981)). The type 1 silcretes are up to 10m thick and display a whitish coloration mottled by red, silicified regolith clasts. Clasts of silicified, ferruginous soil, previously formed silcrete, and quartzite are less common but ubiquitous, ranging in size up to 20cm angular blocks (Plate 7B&7E). The underlying regolith profile shows widespread evidence of authigenic silica replacement of clays and

original quartz grains. In some profiles, the regolith has been extensively replaced by mcq (filled fractures and silicification along foliation), accounting for more than 70% of the rock. Many of these silcretes are ferruginous because they replace hematite-rich sediments and regolith. Their silica mineralogy and cementing, replacement, and void-filling textures show wide variability depending on host lithology.

Microscopically, type 1 silcretes consist of detrital grains and clasts coated by hematite in an mcq matrix. Most grain boundaries show fretting and recrystallization to mcq. Within the mcq matrix, tiny inclusions of hematite, illite, kaolin, and rare leucoxene may be seen. In some type 1 silcretes, repeated brecciation events are evident, and the cavities formed contemporaneously with brecciation are filled with several generations of silica. Cavities are lined by drusy quartz crystals and filled by euhedral mega-quartz.

Silcrete developed within saprolite commonly shows a mineralogy of quartz, mcq, and hematite in thin-section. Mcq occupies pockets between quartz grains replacing clay pseudomorphs after original feldspars in metasediments. Allogenic quartz is commonly fretted and replaced by mcq along grain boundaries and fractures. Veins sub-parallel and sub-perpendicular to foliation are abundant and filled with euhedral quartz and/or mcq. Hematite occurs as inclusions in the mcq.

Type 2 Silcretes

Silcretes of type 2, best exposed in the Marjorie Hills, are hosted by silicified, weathered rhyolites (Plates 7D&7F).

They commonly are red, purple, and brown due to abundant hematitic inclusions and clasts of Pitz Formation rhyolites. They tend to be thinner (1-2m) than other types. The effects of silicification are apparent in the resistant nature of replaced kaolin pseudomorphs after original K-spar phenocrysts, both in weathered bedrock and clasts of rhyolite. They are developed at interfaces between weathered rhyolites and overlying Thelon Formation sediments where abundant joints and fractures allowed easy access to silica-rich groundwater. Several thin silcretes may be developed in the same profile, separated by coherent rhyolite regolith. Also included in type 2 silcretes are detrital quartz grains that have fallen into fractures from above, and chalcedony fragments, some of which may have fractured in situ. Type 2 silcretes are neither as continuous or as sheet-like as type 1 silcretes. In one sample, limonite staining as inclusions in chalcedony was apparent.

In thin-section, type 2 silcretes show abundant void-fillings and geopetal textures. Some are essentially Pitz Formation rhyolite breccias cemented by chalcedony. In a few samples, a fibrous zeolite occupies some cavities, and many samples display sequential cavity fillings recorded by several generations of silica.

Type 3 silcretes

These silcretes occur above and within faults, fractures, and tectonic breccias (Plates 10B&10C). Best preserved on the scarps near Lone Gull, they occur in a variety of colours including white, pink, cream, and red depending mostly on the

abundance and composition of contained clasts. They are composed of angular crushed fragments of quartzite, silicified regolith and minor silicified volcanic rock or soil fragments. Thicknesses range up to 10m or more, being fairly consistent on the scarp south of Lone Gull. A pseudopedal texture commonly is developed in zones showing multiple episodes of brecciation (Plate 3-1) and silicification. Clay minerals and hematite occur as inclusions in the mcq matrix.

In thin-section, type 3 silcretes contain from 30-90% mcq. Varieties of silica other than original quartz are generally absent. Quartzite and regolith fragments are extensively recrystallized/replaced by mcq. Many small remnant grains are highly angular and difficult to distinguish from the mcq matrix, except where the latter is inclusion-rich. Some fragments display evidence of several generations of microbreccia, with rebrecciation indicated by crosscutting fractures filled by mcq.

Type 4 silcretes

Type 4 silcretes are developed exclusively in lower Thelon Formation sediments (Plates 7C, 8A, 8B). They form resistant cappings in the Marjorie Hills, where they typically show colours similar to Type 2 silcretes because of similar clast composition. Host sediments range from rhyolite cobble conglomerates to lithic sandstones to sparse hematitic mudstones. Most appear massive because silicification has obscured original sedimentary textures, but bedding is commonly recognizable. Thicknesses up to 50m were observed, although silcretes of this well developed are probably of

polygenetic origin. One outcrop of type 4 silcrete shows synsedimentary brecciation similar to that described by Selleck (1978) in the Potsdam sandstone, as well as abundant rip ups and reworked silicified arenite clasts.

Under the petrographic microscope these silcretes are seen to contain extremely fine-grained mcq cement. Pitz rhyolite fragments, embayed volcanic quartz, and other detrital components all show dissolution and replacement by mcq. Although commonly accounting for less than 50% of the total volume, mcq can replace all but 20% of the detrital material in 1-2cm zones parallel to bedding. Some mcq shows ghost laminations due to differences in grain size or in abundance of inclusions. These laminations fill spaces between detrital grains, and commonly overlap them.

SILICA MINERALOGY

The silcretes studied contain a variety of silica phases: mega-quartz, chalcedony, mcq, and possibly opal. In addition, several textural variants of quartz can be recognized: mosaic, euhedral, flamboyant, optically continuous overgrowths, spiked overgrowths, and chalcedonic spherules. Relative abundances of silica phases in silcrete types are given in Table 3-1.

Opal

Opaline silica is isotropic under crossed polars and yellow to brownish in plane-polarized light. Individual grains are submicroscopic. These properties also

Table 3-1. Relative abundances of silica phases in silcrete varieties.

Type 1	Type 2	Type 3	Type 4	
-	*	-	-	Opal
-	**	-	*	Chalcedony
***	**	***	***	Chert (mcq)
**	*	**	**	Mega-quartz (all forms)

*** - very abundant

** - abundant

* - sparse

- - rare

Type 1 - silcretes developed in truncated lateritic weathering profiles.

Type 2 - silcretes developed in regolithic Pitz Formation rhyolites.

Type 3 - silcretes developed in faults, fractures, or tectonic breccias exposed to pedogenic processes.

Type 4 - silcretes developed in lower Thelon Formation sandstones and conglomerates.

have been described for poorly crystalline mcq (Milliken, 1979), and may be the phase identified as opal in this study. Williamson (1957) reported both opal and mcq in the silcretes of Australia. The rare occurrence of this phase in the silcretes studied did not warrant this additional investigation. Where observed, the "opal" is exclusively associated with cavity fills. It occurs as the first silica phase precipitated in a cavity on the outer wall. Commonly it is botryoidal and may alternate towards the center of a cavity with chalcedony (Plate 8E). Its location in cavities may have served to preserve it. Many type 4 silcretes are thought to have been originally cemented by isopachous opal rims (ghost laminations are preserved in some of the mcq, suggesting dehydration of opal to more stable mcq with time, as would be expected in rocks of this age. This relationship is most common in type 2 silcretes, which have the greatest amount of cavity fills.

Chert (mcq)

Mcq ranges in size from submicroscopic to an arbitrary upper boundary of 20u (Milliken, 1979). It can show a variety of pale colours in plane polarized light, due to abundant inclusions. Depending on grain-size, it can appear isotropic or display the normal interference colors of quartz. It tends to appear clearer with increasing grain-size. Grains show a hazy or sweeping extinction, and interference figures are usually unobtainable. Grain shape ranges from equant to anhedral or highly irregular with amoeboid

appendages, possibly due to the coalescence of several grains. Mcq, forming an estimated 55% of the authigenic silica in silcretes of the study area, shows void-filling, cementing, and replacement textures. It can show geopetal laminations due to grain-size or inclusion content. It is the only silica phase which has replaced detrital quartz, and in turn can be replaced by chalcedony, chalcedonic spherules, mosaic quartz, and euhedral quartz. It forms the outer rims of many voids with mosaic quartz or single quartz crystal interiors. Where associated with chalcedony in a void, it usually is central. It forms the cementing agent binding the type 4 silcretes developed in sandstones. Large interstitial areas incompletely filled by mcq in these sandstone-hosted silcretes are filled by mosaic quartz which may be a recrystallization product or lateral variation of mcq.

Chalcedony

Chalcedony in the silcretes appears yellow to clear in ppl, and shades of yellow, brown, grey, and white under crossed nicols. The yellow and brownish interference colours may be anomalous, and probably are related to inclusions of water, which can give yellowish colors under ppl (Folk and Weaver, 1952). The chalcedony occurs commonly as geopetal, botryoidal cavity fills, with its fibrous nature showing spectacular sweeping extinction (Plate 8C). In a few cases, pseudo-uniaxial cross extinction was noted. Most of chalcedony is length fast, but in several cavities, length fast chalcedony alternates with length slow chalcedony on the

same rim. No evidence of replacement of evaporites (Folk and Pittman, 1971) was observed in association with the length slow chalcedony. Chalcedony is most common in type 2 silcretes, as is opal. It is the most abundant silica phase in void fills, and also occurs as spherules replacing mcq matrix. Some of these spherules grade to elongate megaquartz (Plate 8D). Some type 4 silcretes originally may have contained [^]chalcedonic overlay rims, now replaced by mcq.

Mosaic quartz

Mosaic quartz is similar to mcq, except that it has a grain size greater than 20u. It shows similar interference colours and straight to shadowy extinction. It forms in the center of voids surrounded by mcq. In some cases it may represent a larger, interlocking, recrystallized equivalent of mcq. It is common in fracture fills, with two- or three-grain thick layers alternating with mcq.

Euhedral quartz

Euhedral quartz is coarse grained and shows straight extinction and the typical optical properties of quartz. It occurs as drusy cavity linings and in fractures. Concentric growth laminations are commonly present (Plate 8E).

Flamboyant quartz

Only one example of this type of quartz was observed. Its elongate nature and undulose extinction make it hard to visually separate individual grains. It grades laterally

into mcq, and may represent the beginning stages of the development of spherules (Milliken, 1979).

Crystallographically continuous quartz overgrowths

These overgrowths are common in type 4 sandstone-hosted silcretes. Where well developed, they show euhedral terminations superimposed on detrital grains. Inclusions may be present; most have been partially recrystallized to mcq.

Spiked quartz overgrowths

These overgrowths form as tiny euhedral crystals (40-100u) with crystallographic axes normal to grain boundaries (Plate 8A). Their common association with hematite-rimmed grains suggests that they form when detrital grains are coated and continuous overgrowths cannot form. In several thin-sections, crystallographic-continuous overgrowths contain inclusions of spiked overgrowths, suggesting that some spiked overgrowths formed early and were replaced.

CEMENTATION, VOID-FILLING, AND REPLACEMENT

Introduction

As noted previously, several types of silica are present in silcretes and silicified rocks of the study area. These show a variety of replacement, cementation, and void-fill textures. Although Wilson (1966) studied silicification in marine limestones, many of the textures and crystallization sequences noted in his work can be applied to this study.

Krauskopf (1959) has shown that amorphous silica is

metastable and has a solubility of 140ppm at 25°C, quartz has a solubility of 6-14ppm at low temperatures, and opal and chalcedony have intermediate values. Kastner (1979) has shown that temperature, pressure, and pH strongly affect the transformation rate of amorphous silica and opal to quartz. During this transformation, porosity may be significantly reduced (siliceous sediments to chert, 70-80% to 10%), and brecciation may occur (Kastner, 1979). Because many Tertiary silcretes show evidence of originally having significant percentages of opaline silica, this transformation may explain the brecciation commonly observed. Because of their age, Dubawnt Group silcretes are characterized by a stable silica mineralogy, largely as a result of dehydration of amorphous or opaline precursors.

Cementation

Any silica phase may serve as a cement; the phase precipitated depends on the concentration of silica in the groundwater of the system and other replacement conditions. Single-phase silica cements or multiphase cements can develop if silica concentrations in groundwater change during cementation. All of the mcq cements in type 4 silcretes are thought to have originally been opaline or amorphous. In some, early optically continuous quartz overgrowths were formed, indicating an initial low silica content followed by increasing concentrations.

Void-Filling

Voids may be filled by any type of silica in virtually any sequence. Monogenetic and polygenetic void-fills can be distinguished. Typically voids in type 2 silcretes show monogenetic sequences where lower solubility silica phases show concentric growth away from higher solubility phases on the periphery. Void-fills commonly show an alternation of silica groundwater concentrations reflected in repetition of paragenetic sequences.

Replacement

Any silica phase can be replaced. Replacement involving dissolution and/or recrystallization of quartz by mcq is by far the most common replacement fabric in Dubawnt Group silcretes, and also has been documented as such in marine carbonates of the mid-Proterozoic Dismal Lakes Group (Kerans, 1982). Quartz grains show evidence of original dissolution before replacive mcq precipitation. This suggests fluctuating geochemical conditions, possibly due to repetitive changes in pH groundwater values from acidic to alkaline (Plate 8B).

Replacements also may be of a strictly recrystallization type. In these types of replacements, dehydration of an original metastable silica phase may cause recrystallization to a stable phase. Opal recrystallizing to mcq in type 4 silcretes is representative of this type of transition.

DISCUSSION

Age and Geography of Silcretes

Woolnough (1927) suggested that all Australian silcretes (siliceous duricrusts) formed at the same time. This idea has subsequently been shown to be oversimplified, because stratigraphically distinct silcretes can be observed in the same area. Although as much as 90% of Australian silcretes may have formed an extensive, flat sheet of equivalent age (Langford-Smith, 1978) it is clear that silcrete genesis occurred during other periods of Earth history. Pre-Tertiary silcretes have been reported from the Cambrian sedimentary sequences of northern New York (Selleck, 1978; Rubin and Friedman, 1981), the Cambrian or late Precambrian of Michigan (James et al, 1968), the Mid-Proterozoic of the Belcher Islands, NWT (Donaldson and Ricketts, 1978) and the Mid-Proterozoic of the Thelon Basin, NWT, (Chiarenzelli et al, 1983). Geographically silcretes are widespread, with reported occurrences from Africa, Australia, Europe, North America, and South America.

Silcrete and Climate

The climatic significance of silcrete has long been debated. The distribution of laterite on the humid coastline of Australia and silcrete in the arid interior (eg. Young, 1978) led earlier workers to conclude that silcretes form under arid conditions. The association of silcretes with evaporites, either capping them or replacing, and calcretes

seems to support this conclusion (Langford-Smith and Watts, 1978; Smale, 1978; Watts, 1978; Rubin and Friedman, 1981; Chapter 6, this study). However, the association of silcretes with deeply weathered profiles which require severe leaching to develop, suggests humid climatic conditions. Paleontological evidence suggests that much of Australia experienced warm, humid, tropical conditions throughout the Tertiary when much of the silcrete probably developed (Hutton et al., 1978). Silcrete occurrences which seem to suggest opposing climatic interpretations have led some authors to change their minds and refute earlier interpretations (for example see Wopfner et al., 1974; Wopfner, 1978). A general consensus is that both silcretes and laterites form in humid, tropical climates, with laterites favouring greater precipitation (Langford-Smith, 1978). Ollier's (1978) conclusion that silcrete forms under varying climatic conditions seems to have the most merit. Seasonal precipitation may also be important. Summerfield (1983) has found distinct petrographic and geochemical differences between silcretes formed during arid and humid climatic periods.

Silcretes and Deep Weathering

The common association of silcrete and weathering profiles has been noted in the literature (Langford-Smith and Watts, 1978; Wopfner, 1978; Stephens, 1971). Wopfner (1978) concluded that the deep kaolinitic profile beneath the silcretes he studied is related to a high groundwater table that supplied silica for silcrete formation. Other workers such

as Senior (1978), have shown a polygenetic origin for silcretes developed on lateritic profiles. Langford-Smith and Watts (1978) have shown that ferruginous weathering products have accumulated in broad synclinal warps on pre-existing silcrete. Downward migration of iron-rich solutions along cracks and fractures has stained the silcretes, rendering the relationship between the two profiles difficult to interpret. Exxon et al (1970) has shown that silcrete can develop without a profile.

The common association of silcretes with deep weathering is related to their formation on level, planed, highly eroded, and stable surfaces. Because silcretes represent the accumulation of authigenic silica, weathering profiles and sediments capping weathering profiles offer an ideal, highly permeable medium in which to develop. As Ollier (1978) noted, silica accumulates in the absence of suitable bases for clay mineral formation. In extensively leached weathering profiles, these bases would be scarce or absent.

Host Lithology

Silcrete researchers have also debated the importance of host lithologies. Woolnough (1927) suggested that host lithology determines whether or not silcrete or laterite will form: laterites on granites, and silcretes on sediments. This extreme view is negated by silcretes that have developed on many different rock types (for example Watts, 1975). Wopfner (1978) concluded that host rock influence during silcrete genesis is minimal. I agree because most silcretes develop on

leached weathering profiles that differ significantly from the host rock, or in sediments above the bedrock. Although silcrete can develop on fresh igneous rock, the zone of silicification must have been weathered initially to permit penetration of silica-rich groundwaters into an impermeable protolith. Non-permeable rocks when subjected to conditions of silcrete genesis might preferentially develop case hardened skin silcretes, as described by Hutton et al (1978) on Precambrian sandstones. In the study area, silcrete developed in fresh rock is restricted to fractures.

Source of Silica

Silica can be delivered by groundwater to the place of precipitation by either vertical or lateral transport. Strictly downward vertical transport would imply cannibalism of upper parts of a weathering source or release of silica from weathering silicate phases above. Smale (1978) cites Spilsby Island silcretes as having formed by the downward reprecipitation of silica produced when feldspars weathered to kaolin within the same profile. A number of authors have noted embayment and dissolution features in detrital quartz in and above silcretes. Watt's (1978) volume calculations indicate that only 25% of the authigenic silica in the silcretes he studied could have been derived from cannibalism. Frankel and Kent (1937) suggested that upward migration and evaporation of silica-rich solutions could provide the silica. This silica may come from weathering profiles directly beneath the silcrete (Wopfner, 1978) or be trans-

ported laterally first (Stephens, 1971). Weathering of basalts is commonly cited as the source of silcretes known as "grey billies" formed under basalts (Gunn and Galloway, 1978). As Davis (1964) has shown, weathered and unweathered volcanic rocks may provide the greatest amount of silica to groundwater.

Silica transported laterally, from outside the profile, may provide the bulk of authigenic silica in most silcretes. Stephens (1971) suggested that laterization of large parts of the Australian continent provided the silica for silcrete formation. Under different circumstances, silica in silcretes can come from a variety of sources. Both in situ and external sources of silica are considered valid.

Silcrete Genesis

Many relevant theories have been proposed for the origin of silcretes. A number of factors which may play an important role are: 1) precipitation under conditions of low (acidic) pH which would tend to leach all major cations except silica and titanium (Valeton, 1972); 2) sufficient water to serve as a transporting medium for silica; 3) a barrier, be it the water table, an early-formed impermeable silicified layer or sedimentary bed; 4) fluctuating geochemical conditions, ranging from an original environment where silica can be removed in solution to a depositional environment where silica is precipitated and all other elements besides titanium are removed (fluctuating geochemical conditions are perhaps related to seasonal precipitation

which enables the dissolution and replacement of original detrital quartz grains in many silcretes); 5) a weathering source area to provide a bulk of the authigenic silica; 6) a porous, fractured, or weathered substrate in which the silica can penetrate and be precipitated; 7) warm to hot climatic conditions which increase the solubility of silica and expedite chemical reactions and weathering by a factor of 2-3x for every 10°C (Loughnan, 1969).

Recognition of Ancient Silcretes

Recognition of ancient silcretes has been hampered by several factors: 1) lack of preservation as extensive sheets by incision of planated surfaces with silcrete cappings; 2) subsurface occurrence of most ancient silcretes (Langford-Smith, 1978); 3) destruction of flat, extensive sheets by tectonic activity (folding or faulting); and 4) discrepancies in terminology and lack of awareness concerning silcretes (they are absent in most countries where standard geology texts are written, Smale, 1978).

In Australia and South Africa silcretes are highly visible because they form flat, extensive, surficial cappings. Their preservation in this form is related to their geologically recent age (Tertiary) and to the lack of tectonic activity in these areas. In many areas they have been eroded and incised, but enough of the original areal extent is often preserved at the surface to make an impression on even the casual observer. Ancient silcretes on the other hand are more apt to be buried, highly eroded, and

their original continuity broken by folding and faulting. In the study area, for example, silcretes are exposed as discontinuous linear ridges due to their inconsistent preservation and exposure on tilted fault blocks. Although silicification beneath the Thelon Formation is indicated by silicified clasts within it (Cecile, 1973), possibly none of the silcrete would be exposed now had not post-Thelon faulting occurred.

As discussed before, silcretes are near-surface accumulations of silica; so are quartz arenites. What makes a quartz arenite a silcrete? Sandstones which show evidence of extensive silicification at or near the surface are considered silcretes. Without careful observation and knowing what to look for, evidence of this type may be easily overlooked.

DUBAWNT GROUP SILCRETES

Classification

Dubawnt Group silcretes have been subdivided into four types on the basis of host lithology which reflects occurrence and mineralogical differences. This classification is simplified, because some of the silcretes may be polygenetic. However, this classification is useful in field and petrographic studies. Type 1 silcretes are developed in 2-10m extensive, flat sheets within or superimposed on the truncated lateritic regolith profiles developed on Aphebian and/or Archean schist, paragneiss, and orthogneiss. Type 2 silcretes are developed as irregular elongate lenses between

or capping regolith developed on Pitz Formation rhyolite. These silcretes typically contain chalcedony in significant amounts as botryoidal cavity fillings. Type 3 silcretes occur within or above tectonic breccias and faults. They are characterized by extremely brecciated host rocks and extensive replacement of allogenic quartz by mcq. They also show evidence of multiple fracturing and silicification (rebreciated breccias). Type 4 silcretes develop in lower Thelon sediments and are cemented by mcq. Their extreme thickness (up to 50m) is unusual but characteristic.

Age

The Dubawnt Group silcretes are at least 1.6 Ga in age. Their stratigraphic position is well defined between the Thelon Formation and the truncated regolith profile above which they developed. Within this bracketed time span of up to 200 million years, several generations of silcretes have formed. Type 1 silcretes may be substantially older than the 1.66 Ga age obtained from apatite from the base of the upper Thelon Formation (Miller, 1983). Type 2 and 4 silcretes, developed on Pitz Formation volcanics and in lower Thelon Formation sediments respectively, are thought to be significantly younger than type 1 silcretes. The age of type 3 silcretes is related to the thrusting, slump, or faulting event which produced them. Arguments that silcretes near Lone Gull developed synchronously with Pitz volcanism is given in Chapter 5.

Climate

Type 1 silcretes may have developed in a time-transgressive relationship with lateritic weathering (ie. silcretes developed on truncated profiles, while weathering continued in undisturbed areas). In this case they would have developed under the more humid conditions associated with original lateritic weathering. Later silcretes (types 2,3, and 4) may have formed under arid conditions (See Chapter 6) and are associated with silicified evaporites.

Silica Source

The source of silica for silcrete genesis is as follows: Type 1- original weathering forming lateritic regolith; Type 2- weathering of Pitz Formation rhyolitic volcanics; Type 3- weathering of Pitz Formation rhyolites and cannibalism of parts of the breccia zone; and Type 4- weathering of Pitz Formation rhyolites and cannibalism of detrital quartz in sandstones higher in the profile.

Type 1 silcretes exist on erosional surfaces which pre-date Pitz volcanism. This surface is exposed on the eastern shore of Marjorie Lake, where a silicified regolithic gneiss is overlain by up to 500m of lower Pitz Formation sediments containing no volcanic clasts, which is in turn overlain by Pitz volcanic extrusive rocks. Although the silicified gneiss is not a true silcrete in that it contains less than 90% silica, it shows a highly resistant 2-3m zone along its contact with the overlying sediments, which probably developed as a result of downward migration of silica-rich

solutions.

Both type 2 and 4 silcretes obviously received silica from the weathering and devitrification of Pitz Formation rhyolites. Type 2 silcretes are hosted by Pitz rhyolites, and type 4 silcretes are developed in sediments which may be time equivalents of part of the Pitz Formation volcanics, and are synchronous with their weathering. Type 4 silcretes have been shown to have dissolution features indicating cannibalism and downward profile migration of silica. Two sandstone sequences were observed directly beneath rhyolitic volcanics. In the sequence described above from east of Marjorie Lake, no silcrete developed in the underlying sandstone. In a sequence north of Marjorie Lake, less than 10km away, silcrete development was extensive. Silica availability is not the only factor to be considered.

Type 3 silcretes show extreme dissolution of brecciated fragments and replacement by mcq. Their time of formation indicates that the weathering of Pitz Formation volcanics may have provided an abundant external source of silica.

Genesis

The genesis of Dubawnt Group silcretes is complex. Each type has a unique history. All profiles show evidence of external as well as cannibalistic sources of silica. Silica was delivered in solution to permeable host lithologies. At the site of precipitation it encountered geochemical conditions which enabled its accumulation at the expense of other elements. Fluctuating geochemical conditions, perhaps related to seasonal

precipitation or groundwater table variations, allowed dissolution and reprecipitation of allogenic quartz.

Conclusions

The Dubawnt Group silcretes are made up of a number of stratigraphically separate and areally distinct horizons 1.6 - 1.8 Ga in age. The source of silica was multiple. Lateritic weathering of basement lithologies, weathering and devitrification of the Pitz Formation rhyolites, and cannibalism of siliceous Thelon sediments all provided significant amounts of silica. In comparison to Australian silcretes they are enriched in hematite and do not show titanium enrichment. This implies that they formed quickly, and had an abundant supply of silica that inhibited the total leaching of the iron and accumulation of titanium. Their stable silica mineralogy is indicative of their age. Their development on a highly weathered and continually weathering planated surface is characteristic of most silcretes throughout time.

SILICIFICATION BENEATH UNCONFORMITIES AND AT DEPTH

Two additional types of silicification warrant mention due to their abundant occurrence. Silicification of unconformities by a distinctive red, hematitic chert is widespread. The best example is exposed in the scarp north of Lone Gull, where regolith is truncated by Thelon Formation sediments. Here the unconformity stands out as a prominent ridge. At a depth of 12m, the regolith contains up to 15% authigenic silica. Red, hematitic chert in fractures and joint fillings

commonly mark these unconformity surfaces throughout the study area. In many exposures, coherent regolith grades to angular, silicified interlocking blocks by intersections of chert-filled fractures parallel and perpendicular to foliation.

In Anaconda drill core 82-7 and 81-17, a breccia composed of angular fragments of regolith in a hematitic mcq matrix was sampled at 106 and 138m respectively. The breccias show multiple episodes of movement and silicification, because the hematitic mcq is brecciated and successive generations are different shades of red (Plate 3B). These samples are interpreted as silicified fault breccias. Many other occurrences of silicification not directly related to silcrete formation are evident in drill core. In Anaconda drill hole 81-14, silicification and brecciation occurs from 40-110m, commonly associated with porphyry intrusions or fractures. Some of the porphyries also are highly silicified. Deep silicification is thought to be related to penetration of silica-rich groundwaters in fractures.

Plate 7. Silicification and silcretes.

A. Extensively quartz-veined regolith. Several episodes of fracturing are apparent from cross-cutting fractures.

Hammer for scale in circle.

B. Silcrete containing recessively-weathering dark coloured blocks of regolith (Photograph - A. Miller).

C. Silcrete developed in granitic detritus.

D. Siliceous coating on Pitz Formation rhyolite cobbles in conglomerates of the lower Thelon Formation.

Phenocrysts (white) of feldspar, originally pseudomorphed by kaolin, are now silicified.

E. Silcrete developed in breccia composed of regolith fragments. Note incompletely filled fractures (F).

F. Silcrete developed in Pitz Formation rhyolite. Note fine laminations in chalcedony (C).

Plate 7

MH-12

ML-12

LG-36

MH-24

Plate 8. Photomicrographs of silcrete textures.

A. Thelon sandstone showing spiked quartz overgrowths. Biotite grain replaced by hematite; mcq is in the center of the field of view. (crossed nicols, scale bar = 1mm)

B. Silcrete developed in Thelon sandstone displaying initial dissolution and replacement of quartz grains by mcq and later partial quartz overgrowths. Dark groundmass is finely crystalline mcq. (crossed nicols, scale bar = 1mm)

C. Chalcedony in silcrete developed in Pitz Formation rhyolite showing concentric botryoidal growth laminations. Note chalcedonic overlay (O) rimming original void boundary. (crossed nicols, scale bar = 1mm)

D. Mega-quartz (M) spherules formed from the recrystallization of an original fibrous precursor. Note chalcedonic overlay rim (O) and mcq (center of field of view).

E. Botryoidal chalcedony with inner opal (?) rims (dark). Zoned quartz, showing overlapping growth laminae, fills a fracture. (plane polarized light, scale bar = 1mm)

F. Silicified evaporites above the Pitz Formation regolith. Note hexagonal zoned crystal (H) which may be a gypsum hemipyramid (cf. Arakel, 1980, p. 1321). (plane polarized light, scale bar = 1mm)

Plate 8

CHAPTER 4 THELON FORMATION

INTRODUCTION

Donaldson named the Thelon Formation and compared its sedimentology and stratigraphy to that of the underlying Kazan and South Channel Formations (Donaldson, 1966, 1967b). Cecile (1973) divided the Thelon Formation into four lithofacies on the basis of petrographic characteristics. Recent work in the Aberdeen Lake (66B) map area has provided a basis for informal division of the Thelon Formation into lower and upper units on the criteria of distinctive lithological characteristics (Chiarenzelli et al, 1983).

LOWER THELON FORMATION

The lower unit of the Thelon Formation is typified by brown, deep red, and mauve rhyolite cobble conglomerates, lithic arenites, minor siltstone and hematitic mudstone. This unit, more than 600m thick in the Marjorie Hills south of Aberdeen Lake, is extensively silicified, and is the host for well-developed, unusually thick, silcretes.

Conglomerates in this unit commonly contain > 90% well rounded, porphyritic, volcanic clasts of the Pitz Formation (Plate 9B). Phenocrysts of feldspar within the clasts have been pseudomorphed by kaolin, and subsequently the clasts have been extensively silicified. The highly rounded, commonly spherical shapes are considered to be a function of spheroidal weathering, intensified by short but rigorous

transportation. Clast-supported orthoconglomerates show evidence of deposition in fluvial channels (cut-and-fill structures, well-sorted lenses, and imbrication of tabular clasts). The matrix of the conglomerates, also extensively silicified, consists of fragments of Pitz volcanics, chalcedony, silicified volcanic rocks, and volcanic quartz grains in a continuum of grain-sizes from granules on down (Plate 9D). Abundant, dispersed, fine-grained, hematitic inclusions give the rock its dark coloration. Paraconglomerates containing mudstone and arenite clasts derived from underlying lithologies up to 2m in diameter display isotropic fabrics. Clasts float in silicified, hematitic, massive, mudstone. These units are interpreted as debris flows (Plate 9E&9F).

Lithic arenites with planar lamination and trough cross-bedding (sets up to 1m) are associated with hematitic mudstones containing abundant sand-infilled desiccation cracks and mud-chip conglomerates. These lithologies have been interpreted as braidplain/overbank and playa deposits, respectively (Chiarenzelli et al, 1983a).

UPPER THELON FORMATION

The upper Thelon Formation in the study area consists primarily of buff, pale pink, grey, cream, and minor red coarse-grained to pebbly granule sandstones and conglomerate. Minor mudstone and siltstone was noted in drill core. Conglomerates are composed chiefly of highly resistant, rounded clasts (< 30cm) of quartzite, vein quartz, silicified rego-

lith, and silcrete (Plates 6E-H,9A). Other typical lithologies include iron formation, Pitz Formation volcanics, and mudstone (see Table 4-1). Conglomerates are common at the unconformity above the regolith, but only rare conglomeratic channels exist higher in the section.

Coarse-grained to granule sandstones containing scattered quartz pebbles compose most of upper Thelon Formation rocks in the study area. The matrix of these sandstones commonly is kaolin and illite. In sandstones and conglomerates immediately above the regolith, specularite and apatite are sporadically abundant in the matrix. The angular to irregular outlines of many grains in these sandstones is suggestive of diagenetic origin, possibly through the alteration of feldspars. In thin-section the widespread occurrence of clay pseudomorphs after feldspar support this interpretation. Trough cross beds are abundant in sets up to 5m thick. The presence of conglomeratic channels, trough cross bedded pebbly sandstones, and the scarcity of fine-grained sediment are suggestive of deposition in a braidplain environment.

STRATIGRAPHIC RELATIONSHIPS BETWEEN THE UPPER AND LOWER THELON FORMATION

The contact between the upper and lower Thelon Formation is gradational in most places, but in a few localities it is unconformable. In the area north of the northern arm of Marjorie Lake, a gradational contact over 100m of section is fully exposed. The transition is marked by decreased silic-

Table 4-1. .Clast composition of Thelon Formation conglomerates.
(Pebble counts of drill core samples, summed and recalculated
to give mean number per cent estimates).

		Q	S	R	IF	CI	MS	P	FS	LT
81-1	154m	7					3	1	2	
	157	21		5	1				2	1
81-2	114m	14	1	1	1		1			
	125	15	2							
	147	6	1	1	2		2	2		
	157	5		2						
	162	14	1		1	1			2	1
	170	13	9	3						
81-3	97m	5		2						
	98	10	1					1		
81-4	369m	23	3					1		
	375	4		3						
Total		137	18	17	5	1	6	5	6	3
%		69.0	9.0	9.0	2.5	.5	3.0	2.5	3.0	1.5

Q - Quartzite
S - Silcrete
R - Regolith
IF - Iron Formation
CI - Christopher Island Formation
MS - Mudstone
P - Pitz Formation
FS - Ferruginous Soil
LT - Lower Thelon Formation

ification (and hence induration), waning contribution of Pitz Formation detritus, and abrupt reduction in hematite content, coupled with gains in matrix clays, quartz grains, and quartzite pebbles. Transitional rocks show intermediate compositions.

Five km southeast of Hailstone Lake and 10km east of Lone Gull (Figure 4-1), an angular unconformity can be observed between the upper and lower units. The correlation of the basal lithologies in the latter locality with the lower unit exposed in the Marjorie Hills is discussed in the following section. In both areas, the lower units are inclined up to 45° , and are erosionally truncated by the upper unit which contains clasts derived from the lower unit. Because the two units are elsewhere conformable this relationship indicates that synsedimentary faulting occurred during deposition of the Thelon Formation.

CORRELATION OF SURFACE AND SUBSURFACE LITHOLOGIES

Figure 4-1 shows the location of Cominco drill sites and the distribution of lower and upper Thelon units. Stratigraphic columns for drill core profiles are presented in Figure 4-2. In general, the drill cores show basal conglomerate immediately overlying the regolith, the local presence of hematitic mudstones overlying the unconformity, large-scale fining-up cycles, and the dominance of coarse, trough crossbedded, sandstones.

The basal conglomerates in these holes are dominated by resistant clast types (Table 4-1) more typical of the upper

Figure 4-1. Geological map showing the distribution of Pitz Formation, lower Thelon Formation, and upper Thelon Formation rocks, and drill hole localities. Rose diagrams summarize paleocurrent data for Thelon Formation south of Aberdeen Lake.

Figure 4-2. Stratigraphic sections from drill cores of the Aberdeen Lake map area. Cominco drill cores were originally logged by R. Stanton; sections compiled by J. Murray.

Thelon unit. The position of these conglomerates immediately above the unconformity, as well as significant content of Pitz volcanic detritus, renders assignment to either unit questionable. Some hematitic mudstones show an unusual clotted texture due to unequal concentrations of hematite in rounded blebs (Plate 9C). These blebs are of unknown origin, but are not thought to be related to redox spotting, because many also show unequivocal redox spotting. This feature, along with abundant desiccation cracks, are atypical of the upper Thelon Formation (Plate 6E). For this reason they are considered to be distal equivalents of the lower Thelon Formation. The basal conglomerates may be wadi channels cutting mudstones of playa lakes or, where they directly overlie the unconformity, older residual deposits covering the regolith. The hematitic mudstones are overlain by typical upper unit sandstones.

Basal, hematitic, silicified arenites unconformably overlain by typical upper Thelon sandstones, as near Lone Gull, are considered to be time correlative lateral equivalents of the lower Thelon unit. The lack of Pitz Formation detritus is attributed to their location far from any outcrops of Pitz Formation lithologies. Rocks along the shore of Schultz Lake near Whale Bone Hill show characteristics similar to the Lone Gull rocks.

CONCLUSIONS

A comparison of the upper and lower units of the Thelon Formation reveals some significant differences (Table 4-2).

Table 4-2. Comparision of the lower and upper units of the Thelon Formation.

	<u>Lower Unit</u>	<u>Upper Unit</u>
Colour	brown, deep red, mauve	buff, pink, grey cream
Cement	silica	kaolin, illite silica
Lithologies	conglomerate *** lithic arenite *** siltstone ** mudstone **	conglomerate * quartz arenite *** siltstone * mudstone *
key:	* rare ** abundant	*** extremely abundant
Sedimentary Strucures	cut-and-fill mud cracks graded bedding debris flows mud-chip conglomerates	trough crossbedding chaannels
Extent	Marjorie Hills	Thelon Basin
Paleocurrents	Northeastward (18 measurements)	Westward (85 measurements)
Environmental Interpretation	Arid alluvial fan Playa lakes	Alluvial fan and braidplain
Provenance	Pitz Formation	Varied

The lower unit is characterized by silicification, lack of clays, dark coloration, northward-flowing paleocurrents, the presence of debris flows and hematitic mudstone, and a dominance of Pitz Formation provenance. The upper unit is characterized by paler colours, matrix clays, westward-directed paleocurrents, the absence of debris flows and fine-grained sediments, distinct resistant clast lithologies (quartzite, vein quartz, and silcrete), and a wider provenance. This indicates that the lower Thelon Formation formed a northward dipping sequence derived almost exclusively from the erosion of volcanic accumulations of the Pitz Formation. Proximal facies are well exposed in the Marjorie Hills; hematitic mudstones in drill core from holes 20-30km north of the Marjorie Hills may be distal equivalents. Some basal sandstones exposed in areas distal to Pitz Formation outcrops may be time equivalents. Figure 4-3 shows the interpreted stratigraphic relationships. Interpretation of the upper unit has been documented by Donaldson (1967a), and is beyond the scope of this study.

Figure 4-3. Proposed geological evolution as seen in cross-section from the Marjorie Hills northward to drill hole localities. A. Formation of regolith on Aphebian and/or Archean basement. B. Deposition of Pitz volcanics, interflow weathering, and silcrete formation. C. Erosion, and deposition of the lower Thelon Formation and silicification. D. Synsedimentary faulting and deposition of the upper Thelon Formation.

Plate 9. Rocks of the Thelon Formation.

A. Conglomerate of the upper Thelon Formation containing clasts of quartzite (light) and sandstone (darker with faint bedding).

B. Conglomerate of the lower Thelon Formation containing clasts of Pitz Formation rhyolite.

C. Clotted texture in Thelon Formation mudstone. (plane polarized light, scale bar = 1mm)

D. Granulestone, lower Thelon Formation, composed of Pitz Formation detritus. Note detrital sand-sized specularite in the upper 2 cm and the chalcedony void-fill (C).

E. Debris flow paraconglomerate with blocks of underlying quartz arenite. Clipboard for scale.

F. Photomicrograph of debris flow showing hematitic mud matrix. Volcanic fragment, with feldspar phenocryst (P) is derived from underlying Dubawnt Group volcanics. (plane polarized light, scale bar = 1mm)

Plate 9

CHAPTER 5 GEOLOGY OF THE LONE GULL AREA

INTRODUCTION

Lone Gull, a Urangesellschaft uranium camp, is located in the west central part of the Schultz Lake Map Sheet (Figure 5-1). Aside from a preliminary 1:250,000 map by Donaldson (1966), most work in the area has been done by exploration companies. In the Lone Gull area exposure is primarily along two parallel, east-west trending fault scarps located 10km apart. Along these scarps up to 50m of section is preserved; between, the land is flat and swampy with isolated outcrops.

The oldest rocks in the area are a sequence of foliated metasediments. The dominant rock type of these is metagreywacke. In other areas metagreywacke occurs in successions with graphitic schist, iron formation, sericitic schist, metavolcanics, and minor quartzite. These rocks are considered to be of Archean or Lowest Proterozoic Age (Miller and LeCheminant, in press). Field evidence suggests a tectonic contact between the older metasedimentary sequence and Amer equivalent sediments. This evidence includes: 1) unique structural trends; 2) cataclastic zones along contacts; and 3) repeated stratigraphy. This tectonic contact has been observed as thrusting in the Baker Lake Map Sheet by Schau et al (1982). Transport was to the northwest from a belt of Amer equivalent sediments to the southeast (Schau et al, 1982). The timing of the major thrusting event

Figure 5-1. Location and geological map of the Lone Gull area. Mapped by A.R. Miller, R. Albert, and J.R. Chiarenzelli, June 1983.

is uncertain but is thought to be contemporaneous with the Hudsonian Orogeny.

Along both fault scarps, Amer equivalent quartzites rest stratigraphically on the regolith profile developed on older meta-sediments. The quartzite is exposed on the northern scarp as discontinuous, intensely brecciated patches resting on silicified, saprolitic regolith and also as larger, more continuous outcrops with sheared contacts against fresh meta-sediments on the upthrown, southern block. Exposed along the southern scarp, a silicified quartzite breccia zone occurs as a 10-20m sandwich between regolith developed on metasediments similar to those exposed along the northern scarp. The breccia zone was traced laterally for 5km, the extent of the mapping. The presence of fragments of ferruginous paleosols and silicified volcanics within the breccia indicate that the movement which caused brecciation occurred either synchronous with, or after, Dubawnt Group volcanism.

Brecciation post-dates the weathering event which produced the regolith, as shown by truncation of the weathering profile by the quartzite breccia zones and by regolith clasts in the breccia. It pre-dates deposition of the Thelon Formation. This movement may be directly related to thermal doming accompanying local Pitz Formation rhyolitic volcanism, or to northward compression due to horizontal extension in the Baker Lake Basin, 50km to the southeast. Minor displacement (~10km) of a fluorite-bearing granite along the fault provides evidence of east-west strike-slip movement along the northern scarp (A.R. Miller, personal communication, 1983).

The timing of the strike-slip motion relative to vertical movement is unknown.

After brecciation, erosion and silicification occurred. Breccia zones in the area were probably shallowly buried and partially exposed by erosion. The brecciated, intensely fractured quartzite allowed silica-rich groundwater to penetrate along fractures and silicify the underlying regolith and fractured basement lithologies. Along the northern scarp, silcrete developed capping the exposed, highly brecciated quartzite. In several areas the silcrete is dominated by silicified ferruginous soil fragments which may represent paleosols incorporated into the breccia by an advancing thrust or slump.

The intense silicification and silcrete formation implies that groundwaters were silica-rich. The source of some of this silica could be cannibalism of the Amer. equivalent quartzite breccia zones. The intense brecciation would provide multiple pathways and sufficient surface area to supply silica. Undoubtedly some of the secondary silica was derived from this source. However, the close spatial relationships with Pitz Formation rhyolites and silicification in other areas devoid of evidence of brecciation indicate that the major source of secondary silica was weathering of rocks of the Pitz Formation.

Erosion after silcrete formation dissected the silcrete, quartzite breccia, and regolith, creating a topography with significant (>50m) local relief. This erosion is reflected

in the almost exclusive occurrence of quartzite, silcrete, and silicified regolith clasts in basal Thelon Formation sediments and also by unconformable contact of the Thelon Formation with each of these units along the northern scarp. Synsedimentary faulting within the Thelon Formation is described in the previous chapter.

NORTHERN SCARP

Along the northern scarp, up to 50m of section through regolith, quartzite breccia, silcrete, and Thelon Formation sediments is exposed. The regolith is developed on a sequence of strongly foliated metasediments. These metasediments consist of metagreywacke grading to fine-grained quartzites. In metagraywacke sequences, lenses of grayish-green quartzite indicate that primary bedding may be parallel to foliation. The metasediments are cut by numerous syenite and aplite dikes, which have also been altered. In lichen-covered areas, syenite dikes were easily located by their high scintillometer response, presumably due to high original potassium feldspar content. All of the previously mentioned lithologies are cut by quartz veining, which commonly parallels foliation. Abundant networks of quartz veins up to 1m wide cross-cut the regolith.

The regolith here has a deep red coloration, is extremely indurated, and forms most of the exposed scarp. At no point along the northern side of the scarp does fresh rock outcrop, although alteration decreases downwards. The transitional and chlorite zones typical in drill core from other areas are

not exposed at the surface, although they may occur at depth. The regolith here is composed of approximately 50% quartz, 15% muscovite, 15% mcq silica (at 12m, content decreases with depth), 15% kaolin, and 5% hematite. These percentages vary with depth, and in accordance with mineralogical composition of the protolith.

Equigranular, equant, quartz with individual grains up to .5mm form lensoid aggregates parallel to foliation. Smaller quartz grains (.1mm), slightly flattened parallel to foliation, spread in ~~mill~~-like wisps from the lenses. Most quartz grains have straight extinction. Muscovite occurs as fine-grained (.05mm) elongate crystals subparallel to foliation. It is concentrated in rhythmic alternation with quartz in bands up to 3mm wide.

Secondary silica, largely mcq, replaces preferentially along muscovite-rich layers and along fractures (~~Plate 10B~~). It varies in grain-size, but most is less than .05mm. Pockets of secondary silica in quartz lenses probably replace kaolin pseudomorphs of feldspars. Quartz grains are also recrystallized to mcq along grain boundaries and fractures.

Kaolin occupies pockets between quartz grains and occurs with mcq in fractures. Some occurs in book-like masses, which commonly contain swelled cleavage fragments of muscovite, indicating a replacive origin. Replacement of kaolin by mcq is a later event, related to silicification.

Hematite occurs along fractures and as irregular blotches parallel to foliation, tiny inclusions between muscovite grains and inclusions within the secondary silica.

Fresh rock from the nearest outcrop (.5 km south of the scarp) shows an identical foliation. It is composed of quartz, muscovite, potassium feldspar, biotite, and chlorite. Quartz and muscovite grains show the same textural relationships as present in the regolith.

Near the top of the scarp an extremely siliceous ridge marking the truncated regolith profile can be traced for considerable distances (10 km). The unconformity here shows three distinct relationships: 1) In most areas breccias, conglomerates, and coarse pebbly sandstones of the Thelon Formation rest with angular unconformity on the regolith. These sediments are intensely silicified, particularly in the first few meters above the ridge; 2) In several areas quartzite breccia rests on top of regolith and is overlain by silcrete. The stratigraphic positioning of the silcretes on the quartzite breccia is indicative of a unique event separate from the original deep weathering which formed the regolith; and 3) In one area a profile starting in coherent regolithic syenite was observed to grade upwards, through a section of jig-saw breccia, into breccia in which angular fragments have been rotated such that they no longer can be pieced together. This breccia is highly silicified and cemented by a yellow to deep red mcq. The yellow coloured mcq consists of chert-like veins containing opaline pellets (Plate 10A). Rounded pellets up to .3mm in diameter are thought to be areas that were not recrystallized to mcq. They have a parallel direction of fracturing which indicates that they are not detrital. Where replaced by mcq, the

yellow veins become whitish. This chert, especially the hematitic variety, commonly marks unconformity surfaces as fracture fills throughout the study area. The stratigraphic relationship between the basal breccia and the silcrete is uncertain, but the regolithic breccias are thought to post-date erosion of the silcrete.

The silcrete consists of two end members which show extreme variation in secondary mcq recrystallization. Sample LG-1 is a mottled whitish rock forming a small knoll 5m above the nearest regolith exposure. In thin-section it is apparent that the protolith was quartzite (Plate 10e&f). Strained grains of dusty, inclusion-filled quartz showing undulatory extinction are cut by numerous quartz-filled fractures (up to 30% total volume). The fractures, .5-2.0 mm wide, are best visible in plane polarized light, because they contain few inclusions. The fractures are filled by quartz grains (< .5mm) in highly irregular shapes showing straight to sweeping extinction. Some veins are almost unrecognizable under crossed nicols because they have been filled by quartz in crystallographic continuity with the grains fractured. The irregular mcq also has replaced framework quartz grains and rare illitic pockets within voids associated with fractures. Some quartzite fragments are ghost-like in appearance and difficult to visually separate from the matrix.

Sample LG-2 is dark pink with white quartz fracture fills ranging in size from 1.0cm on down. It is exposed 3m above LG-1. It is composed of 70% mcq, 5% anhedral - subhedral vein quartz, and traces of clay in some fractures.

Within the mcq are highly corroded quartz framework grains (25% total volume) < 1mm in size (Plate 10D). The quartz grains show scalloped edges and rafted shards separated from the parent grains by mcq. Several generations of brecciation and silicification are evident. The first generation quartz veins have dusty inclusions like the framework quartz grains of sample LG-1. The second generation quartz veins cut the dusty veins and are inclusion free. Larger fractures have dusty quartz grains on the walls with clear quartz in the center. A few glaebole-like concentrations of iron oxide appear in the mcq, perhaps representing dissolved regolithic fragments. LG-2 is thought to have been derived by replacement of a rock similar to LG-1, and is considered to be a silcrete.

Basal Thelon Formation sediments show intense silicification and a history of synsedimentary faulting. Up to 50m of pebbly sandstone, conglomerate, breccia, and the regolith beneath are inclined up to 40°. Above that is a sequence of overlapping debris flow tongues with abundant blocks of silicified regolith, silcrete, and underlying silicified arenite floating in a silicified, hematitic, red mud matrix. The debris flow is capped by typical upper Thelon Formation pebbly sandstones inclined up to 5°. Synsedimentary faulting predates the debris flow.

SOUTHERN SCARP

For a minimum of 5km a breccia zone primarily composed of silicified, angular fragments of Amer equivalent quartzite

and regolith can be traced along the southern scarp. The breccia zone varies from 10-20m in width and consistently dips northward at 45° due to later synsedimentary faulting during Thelon Formation deposition. The breccia zone is sandwiched between regolith.

A profile was sampled along a well exposed, sloping, south facing exposure. The profile is divided into two parts: 1) (0-13m) the quartzite breccia zone; and 2) (13-25m) the overlying regolith slab. It is assumed that below the quartzite breccia zone another regolith profile is present, as this relationship is fully exposed farther along the scarp. An outcrop of extremely silicified, quartz phenocryst-rich, volcanic rock was observed stratigraphically above the upper regolith slab. It may be correlative with the Pitz Formation and possibly was deposited before the movement represented by the breccia zone occurred.

The quartzite breccia zone is characterized by abundant, highly silicified fragments of regolith and Amer equivalent quartzite in mcq matrix (Plate 10B). Many of the clasts show evidence of multiple episodes of brecciation and silicification. A single clast may be composed of several fragments of regolith, quartzite, and mcq microbreccia fractured and welded together several times.

Several visual divisions based on colour, clast composition and size, texture, and matrix/framework ratio can be made in the lower quartzite breccia zone. Measurements of the individual zonations are approximations, and gradational boundary relationships are estimated due to heavy lichen

cover.

Lower Zone

0-1m (base) Large clasts of silicified regolith (up to 10cm) float in a matrix of cream coloured mcq. Clasts of quartzite are present in subordinate amounts and sizes. The deep red, well foliated regolith clasts are penetrated by the mcq along fractures and parallel to foliation.

1-3m The most distinctive characteristics of this division are its deep red coloration, 10% irregular white quartz veining, and a system of sub-parallel fractures. These tiny fractures (<.5mm in width) have pale tan coloured clay fills, and the rock commonly fractures along them. The majority of clasts are silicified regolith. Fragments are welded together in elongate shapes parallel to the fractures.

3-5m This division shows a pale pinkish colour and pseudo-pedal texture. In the field the rounded dark fragments bordered by pale fracture fills gives the rock the appearance of a silicified pedal soil. This structure is developed by in situ brecciation of quartzite fragments and precipitation of lighter coloured secondary mcq in the fractures.

Quartzite fragments make up 90% of the rock. Each pseudoped is a microbreccia. Late, larger, incompletely filled fractures are lined, in sequence, by euhedral vein quartz, kaolin and/or illite, and specular hematite.

5-7m Large fragments (up to 5cm) of quartzite and regolith float in a buff mcq matrix. Irregular hematitic staining occurs in lensoid blebs in the matrix, possibly following

solution paths. A number of silicified ferruginous soil and volcanic fragments are present in this division. The bottom contains more regolith fragments than the top half, which contains quartzite fragments.

8-13m Large, deep red, clasts composed of quartzite, regolith, and mcq are fractured and filled by later mcq. At 10m these clasts take on an elongate shape, and fractures between them are filled with smaller fragments and a hematitic matrix which may represent an area of grinding between larger plates. From 10-13m the rock grades into a coherent protolith of the regolith.

Upper Zone

13-25m At 13m the upper zone begins. It starts within a relatively fresh, fine-grained, quartz-rich metasediment. The foliation within this rock is disturbed but generally trends northeast and is subvertical. Minor hematitic alteration parallels foliation in wispy lens-like patches. The overall colour is greyish red. From 13-25m the rock gradually grades upward into deep red regolith. As silica decreases and hematite increases the foliation becomes more disturbed until at 23m foliated blocks, slightly rotated but still subparallel, are separated by cracks up to 1cm thick filled with a deep red hematitic mud.

Alteration and silicification in the profile is identical to that observed along the northern profile.

Thin-sections from samples at the base of the breccia zone show foliated regolith clasts floating in a mcq matrix.

The mcq penetrates the regolith along foliation. It consists of 50% quartz grains, 30% mcq, 10% muscovite, 5% hematite, and 5% illite and kaolin mixtures in pockets and as micro-litic inclusions in the mcq matrix. Foliation is defined by elongate quartz grains, parallel hematite pseudomorphs of biotite, and compositional layering. Quartz grains show sharp to undulose extinction and are up to .5mm in size. Hematite occurs as biotite pseudomorphs and as inclusions in the mcq matrix. Mcq replaces quartz grains along boundaries and fractures, fills in pockets which may have originally been kaolin pseudomorphs after feldspar, and composes the matrix the clasts float in. The matrix is composed of 50% mcq and 50% silicified regolith fragments. Many framework grains show severe dissolution and fracturing.

CONCLUSIONS

In summary a complex history can be shown in the Lone Gull area (Figure 5-2). Deep chemical weathering produced, in areas, at least 50m of alteration. This alteration consisted of hematite and kaolin pseudomorphing biotite and feldspar respectively in a sequence of Archean or early Aphebian metasediments. Associated with either thermal doming during Dubawnt Group volcanism or Baker Lake Basin extensional tectonics, horizontal movement causing brecciation occurred. Whether this movement is a function of the reactivation of an earlier thrust, or simply represents a gravity slide cannot be determined at the present time.

The movement created a breccia zone composed largely of

Figure 5-2. Geological evolution of the Lone Gull area. A. Development of regolith on thrustusted Amer equivalent quartzites and older metasediments. B. Slump or thrust related to rhyolitic doming of the Pitz Formation. C. Erosion, silcrete formation, and deposition of the lower Thelon Formation. D. Synsedimentary faulting and deposition of the upper Thelon Formation.

Amer equivalent quartzite (10-20m wide) to be sandwiched between regolith. Multiple episodes of movement, cementation, and brecciation occur red. After the brecciation event the area experienced erosion which exposed the fractured quartzite and parts of the truncated regolith profile. A period of intense silicification and silcrete formation followed. Silica invaded the breccia zone replacing clasts and cementing it. The underlying regolith as well as fractured fresh basement rocks also were silicified.

A second erosional event incised the silcrete, quartzite breccia zone, and regolith, creating significant local relief. The lower unit of the Thelon Formation, dominated by detrital material from the regolith, quartzite breccia, and silcrete blanketed this unconformity surface. A period of synsedimentary faulting interrupted Thelon Formation deposition.

Late fractures in underlying lithologies are filled with a sequence of euhedral vein quartz, illite, and specular hematite. These are related to diagenesis of the Thelon Formation sediments. The basal part of the Thelon Formation is cemented by mcq and is highly silicified. It may be synchronous with silcrete formation in adjacent areas.

Plate 10. Rocks of the Lone Gull area.

A. Photomicrograph of "butterscotch" chert showing opaline pellets surrounded by mcq. The chert fills fractures in regolithic basement, (in this case a syenite).

B. Breccia containing regolith and quartzite fragments from the southern scarp. Mcq makes up most of the matrix and replaces regolith clasts along foliation.

C. Quartzite breccia showing pseudopedal texture. The fragment marked (F) is composed of a number of small fragments and is bounded by a illite - specularite rimmed void on the right.

D. Photomicrograph of silcrete which developed by dissolution and replacement of fractured quartzite. Note embayed quartz grains. (crossed nicols, scale bar = 1mm)

E. Photomicrograph showing brecciated quartzite and multiple fractures. (plane polarized light, scale bar = 1mm)

F. Same view as in previous photograph with nicols crossed.

Plate 10

CHAPTER 6 COMPARISON OF PRECAMBRIAN WEATHERING PROFILES

INTRODUCTION

During the past two decades, a substantial number of ancient weathering profiles in Precambrian sedimentary sequences have been recognized, especially in successions of Mid-Proterozoic age. However, detailed studies of these profiles are sparse. Precambrian weathering profiles and paleosols offer an opportunity to gain information about the ancient terrestrial environment (Retallack, 1981).

DISCUSSION

In comparing ancient paleosols with recent soils, a uniformitarian approach must be undertaken with caution. Furthermore, a number of additional factors must be considered in comparisons of Precambrian paleosols and recent soils, and between Precambrian paleosols of different ages.

Without exception, Precambrian paleosols have been diagenetically altered. Even before burial, alteration due to successive climatic changes is possible, especially to clay minerals (Singer, 1980). The preservation of paleosols indicates that they were buried soon after development, and in most examples, they have not been exhumed until relatively recent times. In any type of study of paleosols, diagenetic changes must be noted before interpretations of the original profile can be made. Extensively metamorphosed paleosols

recognized by distinctive chemical compositions and stratigraphic positioning also have been noted in the rock record (Rozen, 1967; Norton, 1969; Vogel, 1972; Meng and Moore, 1972).

As in sediments, diagenesis of weathering profiles can proceed soon after shallow burial, and can continue for significant periods, with or without interruption, possibly even to the present. Initially, under shallow burial conditions, metastable phases would begin to recrystallize, and further reactions may occur with changes in pore-water composition, Eh, and pH. With progressively deeper burial, recrystallization of poorly ordered clays, significant compaction and dewatering of hydrous phases, and reactions with hydrothermal cation-enriched formation waters will occur. Continued increase in temperature and pressure will lead to development of a stable metamorphic clay mineral assemblage, commonly very different from the original assemblage.

The preservation potential for paleosols is low. This is related to a number of factors including their lack of induration and potential for reworking. Almost all Precambrian paleosols lack the uppermost soil horizons, and most are saprolitic. Unfortunately, the missing horizons are the most distinctive for soil classification and paleoclimatic inferences. Whether the lack of preserved upper soil horizons is a function of nondevelopment due to faster erosion rates, or of preferential preservation, is difficult to assess. MacDonald (1980) concluded that upper soil horizons

never did form over the Athabasca regolith, because of the lack of land plants to stabilize the Precambrian terrestrial surface. Chemical trends and textures indicate that upper soil horizons may be present in some paleosols as old as 2.45 Ga (Button, 1979; Gay and Grandstaff, 1980). In the present study, the recognition of soil textures, pedorelicts, silicified ferruginous paleosols (Plate 4E), and pedogenic silcretes indicate upper soil horizons must have been present above that part of the regolith preserved in the Thelon Basin. Erosion rates during Precambrian times must have been higher than at present. The lack of fine-grained sediment and the predominance of braided river systems indicate that the terrestrial surface was continually reworked. How much higher the erosion rates were is open to speculation. The formation of upper soil horizons in Precambrian paleosols was probably the exception rather than the rule, and thus their preservation is extremely rare.

The absence of land plants during the Precambrian may have had a number of consequences on soil genesis. Unfortunately, the influence of plants on recent soils is complex and poorly understood. Basu (1981) suggested that the absence of land plants in pre-Devonian times permitted stability of K⁺feldspars in soils. He reasoned that K-feldspars are unstable in soils today because of potassium chelation by land plants, which lowers the K-ion activity. This conclusion, however, is not supported by any known study of Precambrian weathering; Graustein and Velbel (1981) have suggested that K-ion activity may actually be higher in recent

soils because of potassium cycling by plants.

Cawley et al (1969) studied the concentration of ions in rivers draining vegetated and unvegetated areas of Iceland. Their data indicate that unvegetated areas tend to chemically weather two to three times slower than do vegetated areas. The implications of this study for the Precambrian are open to question. Increased leaching due to the lack of vegetation during Precambrian times might have increased the rate of chemical weathering, given an appropriate (humid) climate.

The lack of organic material also may have influenced pH and Eh. Perhaps the extremely acidic pH values of recent soils rich in organic matter could not be reached during Precambrian times. Additional oxygen unconsumed by the decomposition of organic material might serve to raise Eh conditions in the upper horizons. The lack of chelating compounds may have tended to slow the rates of chemical weathering. All of these ideas are speculative, and require additional research.

The composition of the Precambrian atmosphere is another important factor which must be considered. On the basis of recent work on sulfur isotopes from the Huronian Supergroup, the major compositional change in the atmosphere, first suggested by Cloud (1976) to have taken place approximately 2.0 Ga ago has been substantiated (K. Hattori, personal communication, 1983). Additional lines of evidence, including the evolution of iron formations, the initiation of red-bed sedimentation, and study of detrital minerals unstable under

the present atmospheric composition seem to agree broadly with these results. Unfortunately, reliable quantitative interpretations of atmospheric evolution are not possible at the present time.

Before comparisons of paleosols can be made, a number of factors related to the genesis of each must be considered. Most importantly, paleoclimate, paleotopography, and the maturity of the paleosol in question must be investigated. For example, if all preserved paleosols of Proterozoic age were gley soils, preferentially preserved in low-lying areas, the use of paleosols to determine whether the atmosphere was reducing would not be valid. The comparison of a lateritic profile and a soil of more temperate latitudes also is questionable, as is the comparison of a mature zonal soil with an azonal soil.

Finally, a number of additional factors may be established in the future. Bacterial and algal terrestrial life in the Precambrian and its possible influence has been discussed by Campbell (1979). Although some workers have suggested that ultraviolet radiation, now almost completely blocked by the ozone layer, would have been too intense during much of Precambrian time to permit terrestrial life (Sagan, 1965), the abundance of stromatolites in intertidal and supratidal successions suggests that at least blue-green algae were not significantly retarded by this factor.

EVOLUTIONARY TRENDS IN PRECAMBRIAN WEATHERING PROFILES

Data from a number of sources, including the red beds

and sulfur isotopes previously discussed, indicate that a major change in the oxygen content of the atmosphere occurred at approximately 2.0 Ga. The mineralogy and chemistry of preserved paleosols and weathered horizons older than 2.0 Ga support this conclusion. Evidence from paleosols suggests that, although the pre-2.0 Ga atmosphere contained some free oxygen, partial pressures were significantly lower than post-2.0 Ga levels (Button, 1979; Gay and Grandstaff, 1979).

Table 6-1 lists in reverse chronological order the reported occurrences and their references.

Button and Tyler (1979) reviewed the numerous occurrences of pre-2.0Ga weathering horizons in South Africa. Profiles beneath the Huronian Supergroup in Ontario have been documented by numerous workers. These profiles show a wide range in thickness, attaining a maximum of 25m. Significant compaction in sericite-rich units has been inferred on the basis of the folding of sand-filled fractures in the profile developed beneath the Pretoria Group (Button, 1979). Widespread vertical differentiation of the units is marked by changes in mineralogy, chemistry, colour, texture, and clay/framework grain ratios. Most profiles are slightly metamorphosed (lower greenschist grade). Some high grade schists and chloritoid-bearing rocks are considered to be metamorphosed regoliths (Button, 1979; Schau and Henderson, 1983). Mineralogically these units are dominated by sericite, iron-rich chlorite, quartz, and relic microcline grains. Accessory minerals include pyrrhotite, zircon,

Table 6-1. Outline of reported Precambrian weathering profiles.

Age	Location & Association	Thickness (maximum)	Notes	References
Archean	Island Lake Manitoba Island Lake Series	100m	#1	25
2.45Ga	N. Shore Lake Huron Elliot Lake Group	16m	#1,2,3	12, 20, 22, 23 40, 44
2.2Ga	S. Africa Pretoria Group	25m	#1,2,12	2, 3
2.0Ga	Finland Bothnian Schists	-	#1	43, 45
Aphebian 2.0-1.8Ga	Coronation Gulf NWT, Canada Coronation Supergroup		#13	26
1.9Ga	Bathurst Inlet NWT, Canada Goulburn Group	3m	#13	4
1.9Ga	N. Labrador Ramah Group	12m	#5	31
1.9Ga	Nonacho Lake Nonacho Group	<2m	#6	L. Aspler
>1.8Ga	Thelon Basin NWT, Canada Dubawnt Group	5m	#5	17, 33, 34, 49
Aphebian ~1.8Ga	Mistassini Lake Quebec Mistassini Group	25m	#4	9, 10, 11
1.8Ga	Belcher Islands NWT, Canada Belcher Group	-	#6, 7	16
~1.8Ga	Thelon Basin NWT, Canada Fitz Formation	>50m	#4, 5, 6 7, 8, 9	5, 7, 8, 37
1.75Ga	Coppermine Homocline Hornby Bay Group	?	#5	G. Ross

Table 6-1 continued.

>1.65Ga	Athabasca Basin Saskatchewan Athabasca Formation	60m	#5,8,9	36,42,49
Mid Proterozoic	Kariela Russia Jatulian Series	20m	#2,5,6, 8	6,29,30,47
1.6Ga	Surinam Guiana Roraima Group	1m	#5	24,32
1.5Ga	Missouri	-	#5	1
Mid Keweenawan	Minnesota Solor Church Formation	31m	#5,9	38
Keweenawan	Michigan Bayfield Group	10m	#6,10	27,28,35
1.1-1.4Ga	Grand Canyon Arizona Grand Canyon Supergroup	10m	#5	18,46
Torridonian 1.0Ga	Scotland Torridonian Sediments	6m	#5,11	50

* in addition the following references are given for regolith developed at Precambrian/Cambrian unconformities:

Alberta	41
Grand Canyon	18,46
New Brunswick	39
Wisconsin	13

- # 1 - Loss of iron, reducing conditions implied, less oxygen than in the present atmosphere.
- 2 - Sericitic alteration abundant, usually associated with slightly metamorphosed profiles.
- 3- Gay and Grandstaff (1980) recognize both a gley and spodosol from this area.
- 4 - Dolomitization
- 5 - Hematitic alteration
- 6 - Calcrete
- 7 - Silcrete

Table 6-1 continued.

- 8 - Lateritic weathering
- 9 - Kaolin abundant
- 10 - Controversial, may be of hydrothermal origin, see Lewan (1979).
- 11 - Interpreted as a pedocal soil.
- 12 - A number of other South African weathered profiles >2.2Ga exist. They are very similar to the 2.2Ga profile described by Button (1979).
- 13 - A number of stratigraphically distinct paleosols have been documented in this supergroup and equivalents. The Tree River Paleosols include pisolitic calcretes and oolitic iron stones.

References:

1	Blades and Bickford (1976)	30	Knight and Morgan (1981)
2	Button (1979)	31	Kroonenberg (1978))
3	Button and Taylor (1979)	32	LeCheminant <u>et al</u> (1981)
4	Campbell and Cecile (1981)	33	LeCheminant <u>et al</u> (1981)
5	Cecile (1973)	34	Lewan (1977)
6	Chayka (1970)	35	MacDonald (1980)
7	Chiarenzelli <u>et al</u> (1983)	36	Miller (1983)
8	Chiarenzelli <u>et al</u> (1983)	37	Morey (1977)
9	Chown and Caty (1973)	38	Patel (1977)
10	Chown and Caty (1979)	39	Pienaar (1963)
11	Chown and Caty (1983)	40	Pugh (1973)
12	Collins (1925)	41	Ranaekers (1981)
13	Cummings and Scrivner (1980)	42	Ranakama (1955)
14	Donaldson (1966)	43	Roscoe (1969)
15	Donaldson (1967a)	44	Sederholm (1931)
16	Donaldson and Ricketts (1978)	45	Sharp (1940)
17	Eade and Blake (1977)	46	Sochava (1975)
18	Elston and Scott (1976)	47	Tella <u>et al</u> (1983)
19	Fahrig <u>et al</u> (1978)	48	Tremblay (1983)
20	Frarey and Roscoe (1970)	49	Williams (1968)
21	Fraser <u>et al</u> (1970)		
22	Fryer (1977)		
23	Gay and Grandstaff (1980)		
24	Grabert (1976)		
25	Herd <u>et al</u> (1976)		
26	Hoffman <u>et al</u> (1970)		
27	Kalliokoski (1975)		
28	Koryakin (1971)		
29	Koryakin (1977)		

pyrite, magnetite, leucoxene, sphene, epidote, and stannomelane. Most show evidence of enrichment in aluminum, titanium, and potassium, and depletion in total iron, sodium, calcium, magnesium, and silica relative to the parent lithology. Ferric iron/ferrous iron ratios have been observed to increase in some profiles, and to decrease in others. Variations in chemical trends between Huronian paleosol samples of the same stratigraphic position were interpreted by Gay and Grandstaff (1980) as representing the development of two distinct soil types. In some profiles groundwater conditions are interpreted to have been alkaline, as a result of the eluviation of silica (Button, 1979). The presence of pyrophyllite and diasporite in pre-2.2 Ga saprolites from Swaziland indicates that intense leaching and acidic conditions also did exist (Button and Tyler, 1979). On the basis of rare-earth element trends, Fryer (1977) concluded that weathering processes were substantially different than those affecting recent soils. Button (1979) reached a similar conclusion, but Gay and Grandstaff (1980) and Schau and Henderson (1983) disagree.

Post-2.0 Ga paleoweathering profiles are common in Archean and Proterozoic basins of the Canadian Shield. Additional occurrences from the U.S., Russia, Surinam, and Scotland are noted in the literature. Although most profiles are comparable in thickness to pre-2.0 Ga profiles, some (Athabasca Basin, Thelon Basin, Coppermine Homocline) are extremely thick (>50m). Their saprolitic character suggests that overlying soil horizons were eroded, or that upper soil

horizons either were very thin, or could not develop due to high erosion rates (MacDonald, 1980). Although all are diagenetically altered, most recorded examples are virtually unmetamorphosed. Well-defined vertical zonations are noted by mineralogical changes which reflect down-profile chemical changes (probably due to variations in leaching intensity). The upper zone is typically reddened as a result of enrichment in ferric iron. Kaolin and illite are by far the major clay minerals. At depth, chlorite and illite dominate, reflecting reduction of iron. Chemical trends include severe leaching of sodium, calcium, and magnesium from the upper zone, and the enrichment of aluminum, titanium, and iron. Concentration of soluble cations increases with depth and is related to less intense leaching. Potassium enrichment throughout the profile is typical. Profiles considered to have lateritic affinities are abundant, and silcrete, calcrete, and meta-bauxites have been reported (Figure 6-1). A variety of groundwater conditions may be postulated for the different types of profiles noted. No argument exists over the similarity of these profiles to modern occurrences.

Although detailed comparisons of pre- and post-2.0 Ga weathering profiles is unwarranted, some general trends may be noted. Most important is the greater amount of iron oxidation in the post-2.0 Ga profiles. The absence of illite and kaolin and the presence of sericite and chlorite in the pre-2.0 Ga profiles agrees with their metamorphic grade. Original clay mineralogy in pre-2.0 Ga profiles was probably more heterogeneous, varying with depth. Much of the sericite

in these profiles may form directly from original illitic clays, or by the alteration of illite formed by the diagenetic illitization of kaolin (Button and Tyler, 1979). The only major difference between pre-and post-2.0 Ga weathering horizons (that can be determined from the present data) is the lower oxidizing potential of pre-2.0 Ga groundwaters.

As early as 2.0 Ga, weathering profiles that were mineralogically, chemically, and texturally similar to recent examples probably could form. However, whether the wide variety of upper soil horizons (A and B) that exist today could develop in the absence of vegetation and animal life remains unresolved. Contrary to the conclusions of Schau and Henderson (1983), diagenetic processes probably are homogenizers of weathered profiles, and mask original mineralogy and chemistry. The consistent enrichment of potassium to produce illite in almost all Precambrian weathering profiles and sedimentary rocks is enigmatic, but probably occurred as a result of diagenetic metasomatism.

COMPARISON WITH THE ATHABASCA REGOLITH

As mentioned before, comparisons between ancient weathering profiles must be done with caution, primarily because of the abundance of soil types which may develop under different climatic, topographic, and parent rock parameters. Factors which could change with time, such as vegetation and atmospheric composition, also must be taken into account. Fortunately for comparative purposes, a detailed

Figure 6-1. Evolution of residual deposits, organic life, and the atmosphere.

study has been carried out on the Athabasca regolith by MacDonald (1980). The Athabasca regolith is extremely similar to the regolith of this study. Both are thought to have formed at similar times, at roughly the same latitude, and are covered by similar and perhaps correlative lithologies (Fraser et al, 1969). In addition, both are widespread throughout their respective basins and are developed on a wide range of similar lithologies. In spite of complicating factors such as diagenetic alteration due to the hydrothermal event or events responsible for the uranium mineralization in the Athabasca Basin, and the hydrothermal effects of Pitz volcanic activity in the Thelon Basin, profiles which have not been visibly affected by these events can be chosen from both basins for comparative purposes.

Profiles developed on meta-arkose and meta-semi-pelite from the Athabasca Basin were chosen for comparison with those developed on feldspathic metawackes from the Thelon Basin. The extreme thicknesses developed in both basins seem to be related to hot, humid climatic conditions which enabled extreme leaching. In addition, both profiles are saprolites, and alteration of most primary minerals is pseudomorphous. Common textures such as swelled, split, and curled micas are observed in both examples. Both show evidence of truncation and removal of any upper horizons. Distinctive mineralogy-related colour zones can be identified in both regions (Figure 6-2). Although correlation of these zones may be complicated by diagenetic alteration, the hematite, transitional, and chlorite zones of the sub-Thelon regolith are

approximately correlative to MacDonald's red, green/red, and green zones of the sub-Athabasca regolith. The original alteration products also appear to have been similar. In summary, numerous similarities exist, and these substantiate equivalent climatic conditions, protolithologies, paleogeographies, paleotopographies, diastrophic histories, genesis, and as would seem reasonable, age.

A number of differences also exist. The bleached zone noted by MacDonald marking the top of the sub-Athabasca regolith immediately below the unconformity is not present in the sub-Thelon regolith profile. However, redox spotting and bleaching along fractures is common in the uppermost zone of both regoliths. The green/red zone of the Athabasca Basin regolith shows the greatest alteration in the green layers, and remnant hematite blebs are surrounded by chlorite envelopes. Commonly green fractures cut the red interlayers indicating that the chloritic alteration was a late-stage event. MacDonald (ibid) concluded that the green zones are of diagenetic origin. In the Thelon regolith, the alteration is the reverse. Remnant chlorite blebs are surrounded by hematitic alteration, and hematitic fractures cut the chlorite zone. To explain this difference, the transitional zone of the sub-Thelon profile is inferred to have been erosionally truncated.

In addition, a number of different mineralogical trends have been noted in the Thelon regolith: 1) biotite is destroyed much deeper in the profile and never occurs above

Figure 6-2. Comparison of mineralogy with depth and division of zones in the sub-Thelon and sub-Athabasca regoliths. Diagrams for the sub-Athabasca regolith from MacDonald (1980).

the chlorite zone; 2) kaolin is present further down the profile, and not just in the upper few meters; 3) chlorite invariably is absent in the uppermost hematite zone; 4) hematite is much more abundant in the hematite zone; 5) illite is everywhere the dominant clay mineral at depth; 6) dolomite occurs in significant quantities in replacive and void-filling relationships; and 7) secondary silica from the formation of silcretes capping the profiles is present.

MacDonald's (ibid) chemical data for the Athabasca regolith shows almost complete loss of Ca and Na, a slight gain in ferric iron over ferrous iron, a loss of K and Mg in the hematite zone with increases at depth, variable behavior of Si, and an increase in Al. Data from the Thelon Basin regolith show similar trends; however, ferric iron enrichment is much greater, titanium and silica may be greatly enriched upwards, and phosphorous is increased in the hematite zone. Silica and phosphorous enrichment are thought to be related to the formation of silcretes and playa lake evaporitic conditions, respectively.

A number of conclusions can be drawn from the above comparison of the Athabasca Basin regolith with the Thelon Basin regolith. Due to the persistence of biotite, the lack of strong ferric iron enrichment in the uppermost red zone, and the greater amount of chlorite within the profile and its persistence into the red zone, the Athabasca regolith is concluded to be less intensely leached. Contradicting this is the depth of alteration; however, the alteration is less severe as is shown by the persistence of biotite and feld-

spar. Diagenetic alteration may have increased the thickness of the profile.

Although both regoliths have undergone similar diagenetic alteration, that of the Athabasca regolith has been more severe, and probably took place at higher temperatures than for the sub-Thelon regolith. This would explain the relationships observed in the green/red zone, green cross-cutting red alteration, and the bleached zone at the unconformity. This is consistent with the greater preserved thickness of the Athabasca Group, the high fluid inclusion temperatures (cited by MacDonald) and the proposed hydrothermal origin of many of the uranium ore bodies.

In addition, much of the Thelon regolith experienced different post-formational events, which probably affected its mineralogy and chemistry, significantly.

CHAPTER 7 PALEOCLIMATIC INFERENCES

Present climatic zones are determined by direct measurement of precipitation, temperature, and other factors.

Reasonably accurate isobars, isotherms, and isohuets also can be drawn for the historical past. Inferences about past climatic conditions, however, must be indirectly inferred on the basis of features preserved in the rock record, supplemented by paleogeographic reconstructions. Conclusions thus derived are qualitative, and generally expressed in terms such as hot, cool, humid, and arid. Occasionally the rock record will preserve features which impose limitations on the paleoclimatic conditions (tillites, evaporites, fossils, etc.). Preservation of such features depends on burial and lithification, and most chemical and mineralogical paleoclimatic indicators (e.g. evaporites and clay minerals) are subject to overprinting or obliteration by diagenetic or metamorphic processes. Some paleoclimatic indicators may be more susceptible to alteration or erasure, resulting in preferential preservation (and therefore bias) in the rock record. An example is the dominance of acidic, highly leached paleosols over alkaline, slightly leached paleosols in the sedimentary record (Singer, 1980). Any investigation of paleoclimatic conditions, especially those of Precambrian times, must take these factors into consideration.

Geographic location, precipitation, and temperature are of primary concern when considering the climatic conditions

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of an area. The geographic location of the study area is thought to have been equatorial throughout deposition of the Dubawnt Group. Park et al (1973), on the basis of a paleomagnetic study of samples from the Christopher Island Formation, deduced equatorial latitudes during deposition of this pre-Thelon volcanic unit. The polar wander path they were able to reconstruct using paleomagnetic data from many other sedimentary sequences in the Canadian Shield parallels the equator throughout the estimated time of deposition (1.8-1.6 Ga) of the Dubawnt Group. Irving (1979) found the paleolatitude of the Thelon Basin at ~1.66 Ga would be sub-equatorial. This information is in agreement with the probable correlative sedimentary sequences of the Hornby Bay Group (Ross, 1983a; 1983b) and the Athabasca Basin (Fraser et al, 1970).

Relatively hot climatic conditions are thought to have prevailed throughout deposition of the Dubawnt Group on the basis of paleolatitude suggested. In addition, enough indirect evidence can be obtained from the rock record to make qualitative statements about the amount of precipitation during the initial stages of the weathering event which produced the Pitz regolith. Humid conditions are inferred to have prevailed. The evidence to support this comes primarily from the alteration observed in the regolith and the depth of that alteration.

The original alteration profile has been interpreted as lateritic in Chapter 2. Laterites form from the intense leaching of most major rock-forming cations except Al^{++} and

Fe⁺⁺⁺. The lateritic profile shows enrichment of these elements upwards towards the laterite horizon (sensu stricto). The mineralogy of this horizon characteristically consists of kaolin, gibbsite, goethite, various other aluminum and iron hydroxides, and quartz. The profiles in the study area show a mineralogy consisting of quartz, mcq, kaolin, illite, and hematite in the uppermost hematite zone. The presence of mcq is interpreted as representative of overprinting during development of silcrete which in most places caps the truncated weathering profiles. Kaolin and remnant quartz grains appear to be the only minerals that have survived subsequent diagenesis. Goethite and iron hydroxides are thought to have altered to hematite.

Whether gibbsite and related aluminum hydroxides were present originally in the weathering profile is unclear. Schaufelberger (1953) recorded gibbsite in only 7 out of the 15 lateritic crusts that he investigated, so it is not considered an essential part of the laterite horizon. As discussed in Chapter 2, the hematite zone is not thought to be equivalent to the laterite horizon, because the profile shows abundant evidence of truncation. Gibbsite could also have been absent if leaching was not sufficiently intense for it to form as a result of desilicification of kaolin. For these reasons, gibbsite is not thought to be a modal component of the original weathering profile (although it may have been present, but was subsequently altered diagenetically or simply lost through erosion of an upper soil horizon).

Although some illite may have been present originally in

the hematite zone, especially as a degradation product of muscovite, most of it probably was formed by the illitization of kaolin during burial (Dunoyer de Segonzac, 1970).

Evidence for this is provided primarily by the crystallinity index of illite (all of the illite present is well crystallized). In comparison with the data of Dunoyer de Segonzac (1970), the Illite Crystallinity Index is similar to samples from the lower zone of Late Diagenesis and/or the Anchizone. At this stage of diagenesis much of the kaolin would be illitized. Remnant kaolin still is a major component of observed profiles, and occurs in irregular zones that in places make up almost 100% of the clay mineralogy. In other areas, illite has pseudomorphed kaolin growth forms, and occurs in illite stringers. This leads me to conclude that kaolin originally was much more abundant in the profile, so that with or without gibbsite, it constituted the entire clay mineralogy of the hematite zone. The hematite zone thus is the closest preserved component to a laterite horizon in composition, stratigraphic position, and occurrence.

The presence of an inferred lateritic profile and the original predominance of kaolin clays reflects intensive leaching of the protolith, which in turn implies abundant precipitation. However, leaching and precipitation were not sufficiently intense for the formation of bauxite.

The thickness of the observed weathering profiles have been substantially decreased by erosional truncation. Even so, they attain thicknesses of as much as 60m (Table 1-1.). This compares with modern tropical areas where rock is

altered to similar depths (Ollier, 1969). Alteration to depths of this magnitude is unlikely in cool or arid climates (Loughnan, 1969). This combined with inference of a lateritic profile, indicates that hot, humid climatic conditions persisted throughout development of the initial weathering profile of the Pitz regolith.

Evidence of transition to semi-arid or seasonally precipitative climatic conditions is preserved in the rocks of the Pitz Formation above the initial deep weathering profile. This includes potential eolian sandstones near the base of the Pitz Formation (cf. LeCheminant et al, 1983, p. 444,445) and within the Thelon Formation noted by Donaldson in 1966, the development of ortho-joint planes in upper horizons of the regolith; the trace element chemistry of apatite cements and mudstones at the base of the Thelon Formation, the presence of minerals associated with evaporitic conditions in fractures cutting the regolith, and silcretes hosted in sedimentary rocks of lower Thelon Formation, silicified evaporites, and geopetal apatite pisolites. Although wind-blown sand can accumulate under any climatic condition, the formation of dunes generally is limited to arid, semi-arid, or coastal areas. Eolian sandstones have been documented by Ross (1983a, 1983b) in the potentially correlative Hornby Bay Group. Although the recognition of two occurrences of Proterozoic eolian sandstone does not imply that arid conditions prevailed across the Canadian Shield during this time, it does imply that major dunes were capable of forming, and should be looked for in other basins of similar age.

Ortho-joint planes are thought to reflect swelling and shrinking events in soil horizons. The larger the separation, the more intense the shrinking event is supposed to be (Brewer, 1966). Identification of these features exclusively in regolith developed on Pitz Formation rhyolites suggests seasonal precipitation with periods of desiccation of the soil to significant depths (because most surficial horizons are thought to have been removed). Ptygmatically folded kaolin that has filled sub-vertical conjugate separations provides evidence that the regolith was soft and subjected to swelling (indicating that these features are not later clay fills of fractures which formed after the regolith was indurated (Plate 4F).

The trace-element chemistry of apatite cements at the base of the Thelon Formation, investigated by Miller (1983), show elemental enrichment similar to that found in playa lake environments, and is different from marine phosphorites (cf. Altschuler, 1980). Samples of mudstones from the base of the Thelon Formation (also interpreted as playa lake deposits) show similar enrichment trends in B, Ba, and Sr (Table 7-1). This close geochemical similarity to modern playa lakes is suggests evaporitic concentration.

Gypsum, barite, and phosphate-bearing phases occur in fractures that cut the initial lateritic profile, and hence post-date development of the laterite. However, their occurrence may be related either to diagenesis of the Thelon Formation or to evaporitic concentration in fractures developed in the lateritic profile. The latter interpretation is

compatible with trace element compositions of the mudstones and apatite cements.

The silcretes hosted in lower Thelon Formation rocks also support arid or semi-arid climatic conditions. Although the climatic interpretation of silcretes has long been debated, recent work by Summerfield (1983) favours interpretation of the weathering profiles associated with silcretes as humid, and those not associated with weathering profiles as arid. In the study area, both types of silcretes may be present: those hosted in lower Thelon Formation rocks may be of arid origin, because the hematitic mudstones interpreted in Chapter 4 to be distal equivalents show the appropriate trace-element chemistry.

Geopetal apatite pisolites indicate development in a terrestrial environment (Plate 6C&D).

The presence of silicified evaporites has been noted in sediments overlying regolith developed on Pitz Formation volcanic rocks. The large cross-cutting plates reach lengths of several centimeters and are up to 2mm in cross-section. Some show twinning parallel to the long axis. Petrographic study and results of microprobe analysis indicate complete replacement of these evaporites which probably were gypsum.

In conclusion, during the period 1.8-1.6 Ga, the study area was at equatorial latitudes and initially experienced hot, humid climatic conditions. The deeply weathered lateritic profile developed on basement rocks supports this interpretation. Eventually, precipitation dropped sufficiently to create semi-arid to arid conditions (which may

have been seasonally precipitative. Evidence for this conclusion is tentative, but nevertheless is supported by the evaporites and nature of the silcretes in the lower Thelon Formation. Within these two major trends, any number of small-scale fluctuations may have occurred. Evidence of evaporitic conditions has been partially erased by diagenesis and perhaps by the onset of wetter conditions following evaporitic ones (leading to preferential preservation of evidence of the wet intervals - i.e. enhanced leaching). The continuation of chemical weathering throughout Pitz Formation volcanism, perhaps interrupted by several periods of evaporation, is shown by the scarcity of preservation of unaltered rhyolite throughout the section. Inferences about the paleoclimate during deposition of the upper Thelon Formation cannot be made on the basis of the present study.

Table 7-1. Comparison of trace, minor, and ore-formed elements in apatite from the Thelon Formation and hematitic mudstones. Data for M#1, phosphatic lithic arkose, from Miller (1983).

	<u>MH-41F</u>	<u>81-17</u> <u>138</u>	<u>81-4</u> <u>322</u>	<u>81-2</u> <u>186</u>	<u>81-4</u> <u>364</u>	<u>M#1</u>
AG	-	-	-	-	-	-
AS	-	-	-	-	-	<5
B	54	790	290	950	180	120
BA	250	200	220	130	260	880
BE	2	-	-	2	3	*
CE	-	-	-	-	-	*
CO	190	68	59	33	50	-
CR	6	12	18	11	25	9
CU	8	9	9	6	9	8
LA	-	-	-	-	33	*
MO	-	-	-	-	-	<50
NI	-	-	-	-	19	5
PB	-	-	-	-	-	124
SB	-	-	-	-	-	*
SN	-	-	-	-	-	*
SR	360	250	610	240	270	1500
V	-	22	41	16	68	69
Y	26	26	21	22	36	*
YB	3	4	3	4	2	*
ZN	-	-	-	-	-	15
ZR	100	100	240	97	200	*

* - not tested for

All values in ppm.

CONCLUSIONS

Intense chemical weathering occurred in the Thelon Basin from 1.8 - 1.65 Ga and produced a number of distinct weathering profiles separated by erosional and depositional events. The thickest (> 50m), best developed saprolites show development of distinct mineralogical zonations which can be used for stratigraphic correlation. The enrichment of hematite in the uppermost horizon and comparison with present profiles is suggestive of lateritic weathering. Diagenetic changes to the original mineralogy indicate burial temperatures of 150 - 200°C and depths of 4 - 5.5 km.

Silcretes developed on truncated regolith profiles and in rocks of the Pitz and Thelon Formations. Although several sources of silica were available, weathering of Pitz volcanic rocks provided the bulk of the silica. These silcretes are among the oldest recorded in the world.

The Thelon Formation has been informally subdivided into a lower unit formed from the erosion of Pitz Formation rocks of the Marjorie Hills and an upper unit of wider extent and provenance. Relationships between the two units can be conformable or unconformable.

Geological mapping in the Lone Gull area suggests slumping or reactivation of a thrust fault synchronous with Pitz volcanism. This event sandwiched a slab of Amer equivalent quartzite between regolith.

The similarities between the sub-Thelon and sub-Athabasca regoliths indicate they formed under equivalent environmental

conditions and at roughly the same time. The sub-Thelon regolith appears to have experienced greater original weathering and less severe diagenetic alteration. By 1.8 Ga, weathering profiles existed which are similar to those associated with recent laterites.

The rocks of the study area preserve a paleoclimatic history that began with hot, humid conditions which enabled severe leaching and formation of the thick regolith. Silcretes formed during periods of decreased precipitation that at times were sufficiently arid to allow precipitation of evaporites. Pleistocene weathering consisting of goethitic alteration of hematite and the weathering of well crystallized diagenetic clay minerals occurs locally in the regolith overlain by till.

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