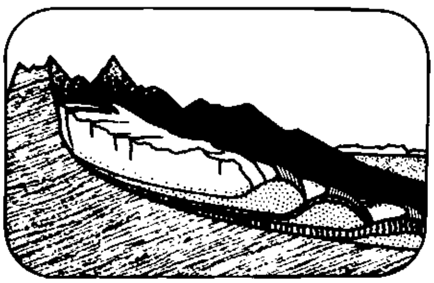
Articles



Dating Methods of Pleistocene Deposits and Their Problems: VIII, Weathering

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Summary
This paper reviews the nature and origin of selected weathering features at small and large scales, and their use as indicators of the age of Quaternary glaciated and periglaciated land surfaces. Chemically weathered forms include crystal etching, boulder weathering rinds and surface relief; mechanically weathered forms include degrees of frost-weathering and associated rubble accumulations. Examples of their use as Quaternary geochonometers are drawn mainly from studies in the North American Cordillera and the highlands of eastern Canada. Alternative explanations of weathering differences in the latter area, specifically those emphasizing lithology, weathering environment, and glacial thermodynamics, are summarily reviewed and found to be inadequate when compared to a wealth of independent evidence that affirms the value of weathering features as chronological tools in eastern Canadian highlands, a value long recognized in the Cordillera.

Introduction
Throughout its post-Huttonian history geological science has been primarily concerned with the *age* of rocks, structures, fossils, and land surfaces. Dating relied upon estimation of the *rates* at which processes operated, at least until the advent of radiometric dating methods. Even these, however, are based on rates of radioactive decay. A rock record may preserve only a fraction of the time elapsed between its beginning and end, but lacunae may often be bracketed by the ages of older and younger features. If marked by subaerial exposure as land surfaces, their duration as such may be more of less accurately fixed with reference to rocks and structures truncated by, or superimposed upon them, and by the degree of modification of their original form. These criteria were widely employed in the first half of the century to erect a chronology of pre-Pleistocene denudation in the circum-North Atlantic uplands, following elaboration of the Geographical Cycle by

Table I *Criteria of Surface Modification used in Studies of Glacial Sequence**

	Blackwelder 1931	**Kay 1931**	**Porter 1976**	**Pierce 1979**
	boulder frequency			
	boulder weathering	pebble weathering	weathering rinds	rind thickness
Weathering features	preservation of glacial polish	oxidation	soils:	obsidian hydration
			horizon development	surface weathering and pitting
			vertical distribution of clay	
	till weathering	leaching of carbonates	vertical distribution of magnetic minerals	extent of soil development
	soil characteristics	soil depth		
			B-horizon thickness	
			structure	
			colour	
	weathering of cirque	amount of subaerial erosion	cirque modification	muting of morphology
	postglacial valley deepening	cutting of valley	gullying	presence/thickness of loess mantle
Erosional and depositional features	lowering of lake outlet	strength of shoreline features	tributary stream erosion	
	filling of lake		mass wasting	
	bulk of talus		slope angle	
	boulder sandblasting		eolian sedimentation	
	morainal modification			
	thickness of eolian deposits			

*Selected from lists which may include other geomorphic relative-age criteria

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Davis (1899). Similarly, criteria of modification were early employed to differentiate deposits and land surfaces associated with multiple Pleistocene glaciations (e.g., Leverett 1909, Penck and Brückner 1909), and to estimate the length of time represented by them.

The use of such 'relative' methods of dating glaciated landscapes has declined with the increasing use of radiometric dating techniques since mid-century, particularly in lowland areas of the humid middle latitudes, where dateable organics are abundant. In drier, high-relief terrain the dearth of organics has assured the continued profitable use of relative methods. At the same time, radiometric methods have furnished datums for relative chronologies, and have allowed rates of landform modification to be determined.

As criteria of relative age, surface modifications resulting from weathering, erosion and deposition occupy a broad spectrum of scale, from the crystal to the terrane levels. Table I lists criteria used in two 'pioneer' and two 'modern' applications to glaciated landscapes. Use of these criteria has not been challenged since the seminal studies of Matthes (1930) and Blackwelder (1931) in the California Sierra. In the northeastern highlands of the continent, from Acadia to Innuitia, interest in the chronological significance of weathering contrasts has waxed and waned in tune with changing concepts of the form and extent of the Laurentide Ice Sheet at the last glacial maximum. The same is true in Scandinavia, for similar reasons. In these highlands, glacial style and paleoclimate conspired to suppress the morainal evidence of the limits of glaciations more than in the Cordillera. Thus, it was altitudinally arranged weathering contrasts which impressed the early glacialists, and which have subsequently become known as 'weathering zones'.

Weathering zones are defined by Dyke (1977) as "units of the land surface which are distinguishable from each other on the basis of distinct weathering features that record different lengths of time through which they have formed" (p. 40). In restricting the explanation of weathering differences to exposure time, this definition prejudges their significance to chronology. Even where the other contributory factors of lithology and process can be held constant, some would disagree that the contrasts are valuable to chronology. The following broader definition is thus preferred for this discussion: 'a mappable terrain unit over which bedrock and/or surficial materials display a degree of weathering defined according to one or more *weathering criteria*, which distinguishes it from terrain bordering it along a *weathering break*. Its weathering characteristics may be interpreted as an effect of structure, process or time, or combinations thereof'.

The following sections discuss the origin of selected weathering criteria and their utility as measures of age. Emphasis is placed first on positive aspects in order that selected examples of their application may be dealt with, before discussion of objections and alternative hypotheses.

Chemical weathering

Crystal weathering. In contact with acidic solutions at weathering (vs. hydrothermal) temperatures, silicate minerals may lose ions in four ways: 1) diffusion through amorphous surface precipitates, 2) diffusion through crystalline precipitates, 3) diffusion through a leached layer beneath the crystal surface, and 4) surface diffusion into the solution (Petrovič 1976). Petrovič concluded that precipitates were less likely to form at weathering than at hydrothermal temperatures, and that, if a precipitate developed, ionic diffusion through it "may be unable to control the rate of feldspar dissolution for years or even thousands of years" (p. 1519). Berner (1978) and Holdren & Berner (1979) found no evidence of precipitation on feldspar crystal surfaces or of cation depletion beneath them, during experimental weathering with HF/HCl solutions at various temperatures. Rather, weathering proceeded by ionic diffusion concentrated at weak points in the molecular structure, leading to progressively denser and deeper etch pits, identical to those seen in feldspars from soils (Berner & Holdren 1979). Similar features were experimentally produced in pyroxenes and hornblende by Berner *et al.* (1980).

Studies of crystal etching as a chronological tool have proven fruitful in distinguishing: 1) Wisconsinan from pre-Wisconsinan tills on the basis of hypersthene etching (J.G. LaFleur, unpub., quoted in Birkeland, 1974, p. 160), and 2) glacial deposits in southern Baffin Island, ranging in age from 3.5 kabp to greater than 130 kabp (kabp = thousands years before present), on the basis of "mean maximum etching depth" (MMED) of hornblende crystals (Locke, 1979). Locke found that, at all depths where hornblende etching was microscopically detectable (to 1.45 m), MMED increased with age through an all-inclusive range of less than one to five micrometers. In the same general area Andrews and Miller (1972) reported variations in the density of etch pits on quartz grain surfaces with age of deposit. Deposits assigned to the Foxe Glaciation (< 130 kabp) showed no pitting of quartz surfaces; those of Pre-Foxe age had pits developed on glacially abraded microfractures; and those believed to be much older, but whose glacial origin or modification is in doubt, showed "considerable pitting and wasting" (p. 7) of quartz surfaces.

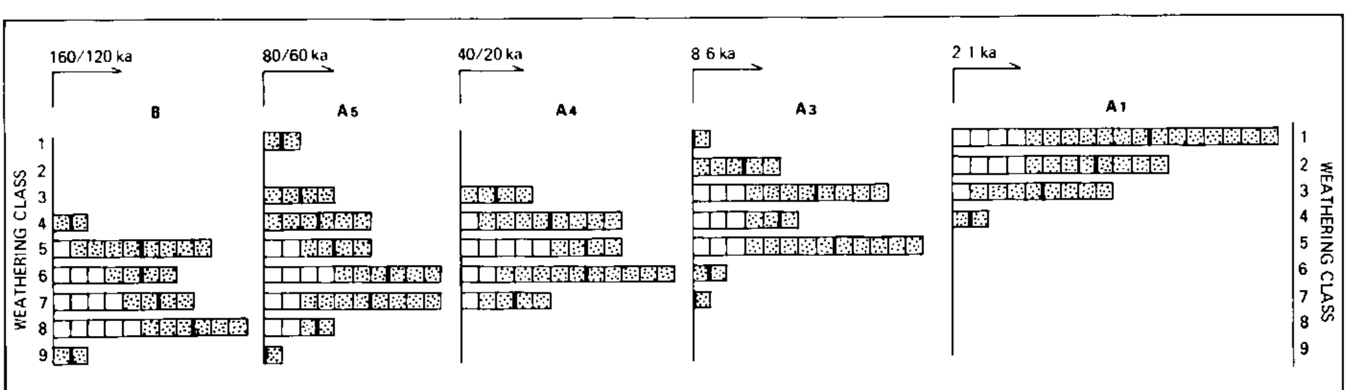

Figure 1 *Standard histograms of states of weathering of 25 boulders for the weathering zones of southern Cumberland Peninsula. Histograms are based on several samples of 25 boulders each. Each block represents one boulder; heavy bars mark the means and the dotted blocks are within 1δ of the mean. Total of blocks below the mean is 25. Duration of weathering in 1000's of years is shown above each zone. Modified from A. Dyke (1979).*

Boulder weathering. A variety of boulder weathering features has been the most widely used of geomorphic criteria of relative age. These features include: boulder frequency, ratio of fresh to weathered boulders, state of surface weathering, weathering rind thickness, and boulder angularity. Some of these criteria are applicable to bedrock surfaces also (e.g., state of weathering and angularity), and some are applicable to specific sizes of boulder (e.g., boulder frequency and surface pitting are restricted to larger boulders).

Surface boulder frequency and ratio of fresh to weathered boulders were used to good effect by Sharp (1969) to distinguish Tahoe (Early Wisconsinan) from Tioga (Late Wisconsinan) moraines near Convict Lake, Sierra Nevada, California. In this and similar studies in this area (Sharp and Birman, 1963; Birman, 1964) frequency counts were made at sites where neither burial nor erosion were likely to have distorted the influence of weathering. Lithological uniformity among moraines was safely assumed from the similarity of ice-flow routes through granitic terrane. Also, since these advances were of approximately the same extent, original boulder frequency was not likely to have varied due to distance of transport. This last circumstance does not prevail in the highlands of northeastern North America, where, at each stadial maximum, glaciers took many forms, from cirque glacier to ice cap (Dyke *et al.*, in press) and travelled different distances. Also, glacier extent within any one topographic setting differed markedly from stade to stade, so that original boulder frequency would likely have been greater in moraines nearer to source. Accordingly, these criteria have not been employed in these areas.

In any attempt to use criteria measuring the state of weathering, care must be taken to exclude non-weathering factors, such as progressive exposure of boulders and outcrops by erosion, and micro-environmental factors, such as variations in snow and vegetation cover. Salt weathering gradients away from coastlines should also be avoided. Several authors have produced classifications designed to measure the degree of boulder weathering. The most internally logical scale of weathering was devised by A. Dyke (1977, 1979) in a study of glacial deposits on southwest Cumberland Peninsula, Baffin Island. He has nine classes of weathering: 1) completely fresh, 2) surface stained by oxides, 3) surface rough due to crystal relief, but crystals not removable by hand, 4) crystals removable with thumbnail, 5) crystals removable with fingers by rubbing, 6) micro-pitted (< 1 cm), with exfoliation shells or weathering relief > 1 cm, 7) macro-pitted (> 1 cm) or inclusions protruding, 8) surface disintegrated, 9) deeply or completely disintegrated. Histograms of boulder frequency in each class (Fig. 1), show that modal class increases with age of terrain. Also, the distribution of classes is less peaked and less skewed with increasing age, since, as the state of weathering advances, boulder surfaces acquire more weathering features. Thus, as A. Dyke (1979) notes, a class 8 boulder also possesses characteristics of classes 6, 5 and 3. "The cumulative nature of the system reflects an orderly progression from less to more weathered, but does not imply that the intervals between classes are equal in terms of process or time" (p. 183).

Boyer (1972) devised a 6-class boulder weathering index for use with other criteria to differentiate weathering zones in the Maktak Fiord area of northern Cumberland Peninsula. This and similar classifications used by Dugdale (1972) more locally, and by Miller (1973) more broadly, on the peninsula, expressed degree of weathering in classes ranging from fresh with a smooth surface to deeply or completely disintegrated. Miller (1973) was able to differentiate deposits of Neoglacial age (< 3.5 kabp, classes 1 and 2) from Wisconsinan (8-120 kabp, class 3), pre-Wisconsinan (> 120 kabp, classes 4 and 5), and deposits much older (>> 120 kabp, class 6), but the system did not differentiate amongst Wisconsinan deposits, in contrast to their five-part subdivision to the south by A. Dyke.

Boyer (1972) and Pheasant (1971) included several other boulder weathering criteria in their differentiation of weathering zones on northern Cumberland Peninsula. These were corner angularity, degree of micro-pitting (< 1 cm), occurrence of large pits or macro-weathering features, height of inclusions, and presence/absence of, or percent with, a weathering rind. Rather than organize them into classes, Boyer and Pheasant (1974) subjected weathering data to statistical analysis which 1) supported the proposition that three weathering zones exist, and 2) revealed highest discriminatory power for the criteria of number of fresh rocks, number commonly micro-pitted, number with inclusions projecting more than 1 mm, corner angularity, and number macro-weathered. Nelson (1980) used Boyer's 6-class granular disintegration scale in an unsuccessful attempt to differentiate statistically amongst moraines older than the Late Foxe stade on Qivitu Peninsula, northern Cumberland Peninsula. Statistical comparison of these data with those compiled from known Late Foxe and Early Foxe moraines in the area showed, however, that the Qivitu deposits are

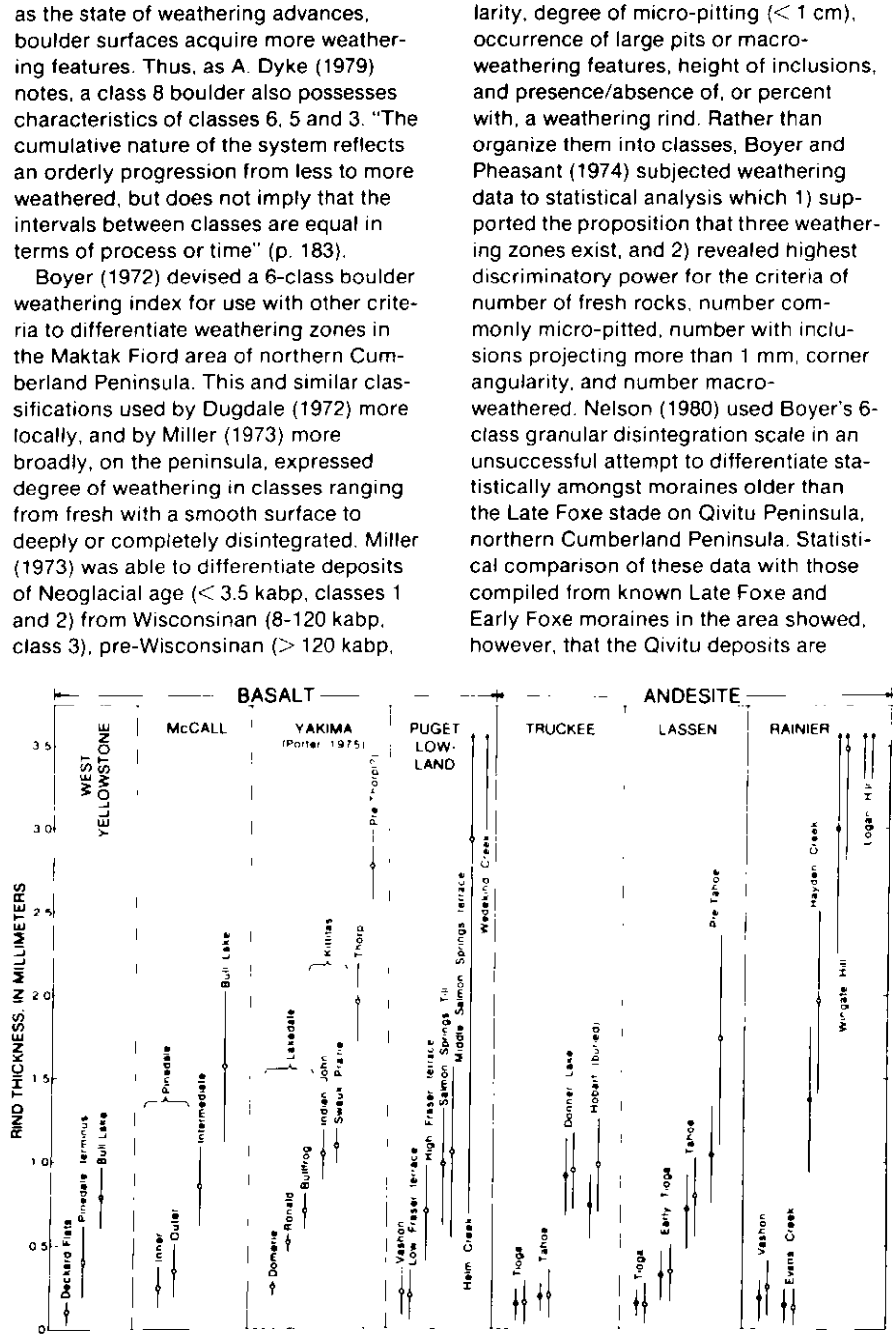

Figure 2 *Average weathering rind thickness for the sequence of deposits in each sampling area in the western U.S. Each point and bar represents the mean and 1σ range, respectively, of all rind measurements for each age of deposit in each area. In andesitic areas (right), fine-grained types are closed circles, coarse-grained are open circles. Arrows indicate data off the scale. From Colman and Pierce (1981).*

more closely related to the former. This is an unexpected and unexplained result, particularly puzzling in light of the expectation that salt weathering on this coastal foreland would have intensified weathering.

The thickness of weathering rinds on boulders in glacial deposits has proved an effective discriminator of weathering intervals, particularly in mountainous terrain in the western United States (Birkeland, 1973; Porter, 1975; Colman, 1981; and Colman and Pierce, 1981). Colman and Pierce amassed a set of 7355 rind thickness measurements from 150 sites in 17 different smaller areas of the American Cordillera, in which mean annual precipitation varies from 23 cm (Yakima Valley, Wash.) to 130 cm (Mt. Rainier, Wash.). Measurements were made on andesite and basalt clasts at 20 to 50 cm depth. With such environmental differences, time was not the only factor explaining rind thickness variation. Differences in rind thickness were noted between andesite and basalt; between coarse and fine grained andesites; between till and outwash; between steep and gently sloping sites; between forested and grass/sage bush cover types; and with mean annual precipitation.

In spite of these variations, plots of rind thickness versus stratigraphic age (assigned by these and earlier workers, partly on evidence independent of weathering criteria) showed close groupings (Fig. 2), which suggest that duration of weathering is an important variable. Tests of normality, analysis of variance, and determination of the significance of differences led Colman and Pierce to conclude that "age of deposit sampled is the most important source of variation in rind thickness" (p. 23).

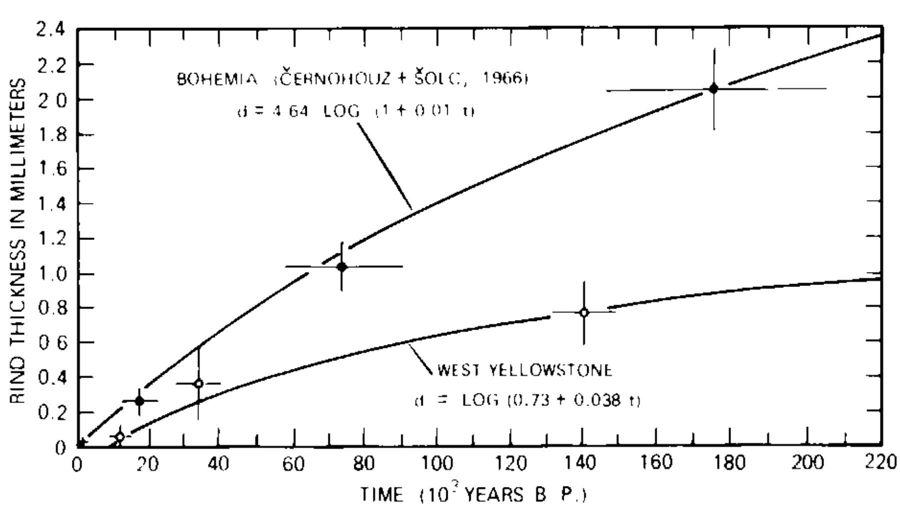

Figure 3 *Weathering-rind thickness vs. time curves for Bohemia and West Yellowstone. From Colman and Pierce (1981).*

In this and some other studies (e.g., Černohouz and Solč, 1966; Chinn, 1981), rates of rind development decelerate with time. Colman and Pierce (1981) preferred a logarithmic expression over linear and power functions to describe this. The general form, $d = A \log(1 + Bt)$ is the same as that used by Černohouz and Solč, where *A* and *B* are constants, and *t* is time (Fig. 3). In order to eliminate the rate constant from age calculations, Colman and Pierce proposed the use of rind thickness ratios. With no ages known, at least the magnitude of age difference between two deposits can be gauged. With one deposit dated, a ratio at least fixes the limits of age for undated deposits.

Decrease of weathering rate with time has been attributed to: 1) retardation of weathering by progressive precipitation of weathering products within the rind; 2) buffering of the weathering solution by release of liberated ions; and 3) weathering of smaller crystals at a faster rate than larger ones. The work of Berner and colleagues, and Petrovič, mentioned above, showed that protective precipitates do not form on silicate mineral grains in natural and laboratory environments. Progressive buffering would seem doubtful, since weathering solutions have residence times three or four orders of magnitude briefer than the millenia required to reveal a decrease of weathering rate. While it might be expected that smaller crystals will weather faster than larger, Colman and Pierce (1981) found that coarse grained andesites weather faster than fine grained ones. The effect is perhaps explained by rapid early weathering of ferro-magnesian minerals and slower effects on feldspars. However, Denner and Anderson (1962) found approximately the same element assemblage in rinds and fresh rock, indicating no preferred order of elemental loss during rind formation.

Further, surface clasts lose weathered material, as indicated by progressive roughening, loss of granules, and the formation of exfoliation shells. Subsurface clasts, however, are noticeably smoother and preserve angularity longer. Those measured by Colman and Pierce (1981) would seem to require at least 0.5 Ma, and possibly as long as 1.0 Ma, for rind thickness to attain constancy under the combined influences of penetration of the weathering front and possible surface loss (Fig. 3).

Due to paucity of suitable fine grained lithologies, weathering rind studies have generally not been feasible in eastern Canada. Late Precambrian diabases penetrating granite-gneisses in the Grenville inlier of Newfoundland's Northern Peninsula have been used by the writer in an attempt to differentiate three weathering zones, (C, B, and A), assigned by Grant (1977a) to pre-Wisconsinan, Early, and Late Wisconsinan glaciations, respectively. Modal rind thickness on till and felsenmeer was 1 mm in both of the older zones at 750-800 m. This is possibly due to loss of weathered material from the surface, which has reached equilibrium with rind penetration since Early Wisconsinan deglaciation of zone B. The youngest zone could only be sampled at 150 m elevation, where modal rind thickness was three times greater than on the older zones. This is probably an effect of the more energetic weathering environment influenced by higher temperature and denser vegetation.

Hydration rinds on fractured obsidian fragments are a special case of a weathering rind. They have been used successfully to date archaeological and geological contexts 10^2 to 10^5 years old (see Friedman and Long 1976, for a review and refinement of the method, Pierce *et al.* 1976, for an appliction to glacial chronology, and Michels 1967, for an archaeological application). Hydration rate is determined from:

$$k = Ae^{-E/RT} \text{ (Friedman and Long 1976)}$$

where k = hydration rate in $\mu m\ a^{-1}.10^3$, *A* is a constant, *E* is activation energy of the hydration process (cal. $mole^{-1}$), *R*, is the gas constant (cal. deg. K^{-1}. $mole^{-1}$), and *T*, is temperature (°K). Hydration rate is related to temperature at which it occurs, but while the form of a family of curves for different samples is the same, the slopes vary and are related to the chemistry of the samples. Increased silica content and refractive index increase the hydration rate and increased CaO and

MgO reduce it. Thus a *chemical index* was introduced and hydration rate plotted against temperature for obsidian samples of different chemical index. For example, at 10° C, obsidians with indices of 0, 20, and 40 have hydration rates of 0.2, 0.8, and 2.3 $\mu m\ a^{-1}.10^3$, respectively (Friedman and Long 1976, fig. 8, p. 351).

Preservation of glacial erosion microforms. This is an inverse weathering criterion inasmuch as preservation rather than erasure of initial morphology is measured, and it is often used with "state of weathering" criteria. On southwest Cumberland Peninsula, A. Dyke (1977, 1979) found striae and polish "very frequent" on bedrock within his weathering zones A1 (deglaciated 2.1 kabp) and A2 (deglaciated either 3.2 or 4.8 kabp), but "infrequent" in zone A3, deglaciated 8.6 kabp. By comparison, the modal class of boulder weathering in zone A1 is 1 (surface oxide staining), and that of zone 3 is class 5 (crystals removable by rubbing with fingers). It appears that in this region little erasure is achieved in the first 3-5 ka of exposure, but that after 8.6 ka polish and striae have disappeared. If striae are estimated to have been up to 3 mm deep, surface lowering proceeded at a maximum average rate of 0.35 mm per ka.

R. Dahl (1967) measured postglacial bedrock lowering on granites around Narvik, northern Norway, using glacially polished quartz and pegmatite veins as reference planes. His plot of lowering against elevation showed a wide scatter, but this is reduced when 72 anomalous surfaces are isolated. These include sites with wet moss, snow-covered sites, blockfield sites, and sites with marked glaciofluvial polishing. The remaining 168 values (each the mean of 50 measurements) shows an increase from zero at sea level to 15 mm at 110-120 m, and a general decline to 6 mm at 500 m. Above this, lowering remains at 6-7 mm, with increasing scatter (4-13 mm) up to 1300 m. Above the 90 m marine limit, dated at ca. 10 kabp, the general decline in surface lowering may be attributable to declining temperatures. Significantly, the lowering rate of 0.6 mm per ka at 20-80 m would remove striae up to 3 mm deep in 5 ka, compared with 8.6 ka in southern Baffin Island.

Preservation of glacial polish and striae can thus be viewed as evidence of less than 10 ka of weathering in maritime subarctic regions. However, care must be taken to exclude sites where "fresh" glacial surfaces may have only recently been exhumed from beneath a cover of much older till. The writer has recognised exhumed striae within weathering zone B (Grant 1977a) in west Newfoundland. England *et al.* (1981) report striae on carbonate bedrock emerging from beneath a pre-Late Wisconsinan till in eastern Ellesmere Island, N.W.T.

Mechanical Weathering

The periglacial morphoclimatic zone is distinguished by the importance of frost action in the weathering of rock and mass movement of detritus (French, 1976, pp. 2-4). The abundance in this zone of castellate cliffs, ridge-crest and plateau-top eminences (tors), and cryoplanation terraces (Reger and Péwé, 1976), as well as mantles of coarse, angular detritus (blockfields, felsenmeers: White, 1976a) over steep and gentle slopes has led to wide recognition that mechanical frost weathering is dominant. However, hydration and salt crystallization have also been recognized as concomitant chemical weathering processes with important mechanical effects under freeze-thaw regimes (Evans, 1970; White, 1976b).

The recognition of contrasts in the development of these bedrock and detrital forms across stadial ice limits of the last glaciation, and their recognition in morphoclimatic zones not presently periglacial has led to their use as criteria of the duration of postglacial periglacial weathering intervals and as indicators of the former extent of periglacial morphoclimatic conditions.

In the alpine periglacial zone, diverse accumulations of blocky detritus at the base of cliffs (see White, 1981 for their classification) are derived by freeze-thaw mechanical and chemical weathering. If the detritus forms an apron below a cliff, it may be further mobilized by creep and flow, particularly under permafrost conditions, and come to occupy distal, low-angle slopes as broad sheets (blockfields), or become channelized into block streams. Fossil features of this origin are known from beyond the limit of the last stadial glacial limit in Wisconsin (Smith, 1949) and Pennsylvania (Smith, 1953; Potter and Moss, 1968), and are attributed to periglacial conditions during at least the Late Wisconsinan.

Blockfields not associated with steep cliffs may mantle steep or gentle slopes. On steep slopes, rapid creep of blocks can possibly maintain the freeze-thaw weathering regime in bedrock below. Over gentle slopes, blocks are derived from the edges of tors (Dyke, 1976) and from the scarps of cryoplanation terraces (Reger and Péwé, 1976). A variety of features indicate that "rock heave" (L. Dyke, 1979, 1981) may explain some blockfields. These include frost-heaved blocks (Dilabio, 1982), frost-thrust blocks (Yardley, 1951), rubble-rimmed craters (Dilabio, 1978), rubble-rimmed polygonal crack networks (Kerr, 1977; L. Dyke, 1979, 1981), fissured rock mounds (Payette, 1978; Thom, 1978), and debris-covered rock mounds (Payette, 1978). Ice segregation was observed in association with the features recorded by L. Dyke (1979) and Thom (1978), and postulated by Dilabio (1978) and Kerr (1977). Elevated hydrostatic pressure was also postulated by L. Dyke (1979, 1981) as responsible for rock heave.

Observations of the rate of rock heave (L. Dyke, 1979, 1981) provide little evidence of the rate at which felsenmeer has developed over deglaciated surfaces in

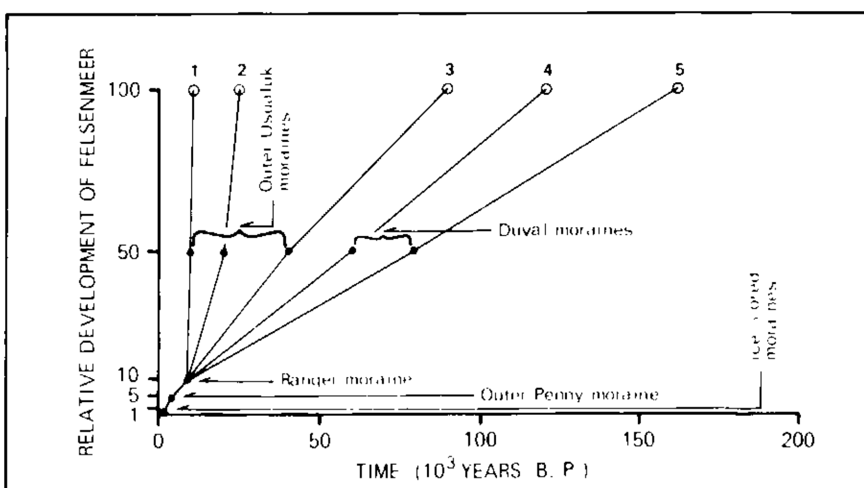

Figure 4 *Relative development of felsenmeer (y-axis) vs. age of weathering zone (x-axis) on southwest Cumberland Peninsula, from data of A. Dyke (1979). Solid circles show felsenmeer development over terrain behind labelled moraines. Note uncertainties of age of Outer Usualuk and Duval moraine sets. Open circles show five posible dates at which felsenmeer could become common (100). See text for discussion.*

the Quaternary period. More informative are records of the degree of bedrock disruption over terrains deglaciated at known times. In northern Canada, surfaces deglaciated in the Holocene have suffered only minor to moderate disruption by rock heave. Exceptions occur on some carbonate and flaggy sandstone terranes (e.g., on Somerset Island, N.W.T., Dyke, 1976; Kerr, 1977).

In southwest Cumberland Peninsula, A. Dyke (1977, 1979) found felsenmeer "common" over his weathering zone B, beyond the Duval moraines to which he tentatively assigned an age of 80-60 kabp, correlative with the Early Foxe (Alikdjuak) moraines of northern Cumberland Peninsula. However, beyond and above Early Foxe moraines, weathering zone II (> 120 kabp) is described by Boyer and Pheasant (1974) as exhibiting only "incipient felsenmeer." This poses a problem either of correlation or of weathering rates. An attempt is made here to extrapolate the varying degrees of felsenmeer development reported by A. Dyke (1979) in order to estimate the time required for it to become "common." Dyke's descriptive terms have been given values between 1 and 100 (with his concurrence, personal communication, 1982). Thus, "negligible" is assigned a value of 1, "very minor" 5, "minor" 10, "patchy" 50, and "common" 100. Regression analysis generated curves 1-5, Figure 4. Curves 1 and 2 are unreasonable, since felsenmeer becomes common within 25 ka. Curve 3 is not so unreasonable, since it arrives at "common" by 90 ka, but this means that common felsenmeer takes only 10-30 ka longer to form than "patchy." Curves 4 and 5 are considered most reasonable since they arrive at common felsenmeer at 120 ka and 160 ka, respectively. They imply that terrain in weathering zone B was deglaciated in the last interglacial or in the previous glaciation.

The chronological meaning of tors is more difficult to ascertain due to the probability that longer weathering is necessary to produce the relief of a tor than that of felsenmeer which typically surrounds it, and which is in part genetically related. Further, unless a one-cycle periglacial origin can be demonstrated, the relief of tors may pre-date periglacial conditions. Dyke (1976) demonstrated a clear genetic relationship amongst tors, cryoplanation terraces, and felsenmeer in gneissic terrane in northwest Somerset Island, N.W.T. These features occur in an area free of glacial erosion, bordered by terrain scoured by Late Wisconsinan ice. Erratics and patches of till were found within the weathered zone, but these were attributed to pre-Late Wisconsinan glaciation.

Tors have been widely recognised, with felsenmeer, as diagnostic landforms in the uppermost weathering zone in northern Labrador (Ives, 1957, 1958a,b), northern Cumberland Peninsula, Baffin Island (Pheasant & Andrews, 1972, 1973; Boyer and Pheasant, 1974), western Newfoundland (Grant, 1977a; Brookes, 1977), eastern Ellesmere Island (England *et al.*, 1981), and above the limit of Wisconsinan glaciation in Yukon Territory (Bostock, 1966).

Landform Modification

What Pierce (1979) called "muting of morphology" (Table I) referred specifically to lowering of slopes and summits, but the term could be generalized to include all erosional and depositional modifications of original glacial forms. Much of the modification is likely to occur immediately after deglaciation, when meltwater runoff is intense and vegetation cover incomplete. Price (1980) found "major changes in the landforms and drainage patterns . . . occurred in the first 50-100 years of their existence" (p. 91) in a recently deglaciated depositional terrain in Iceland. Welch (1970) also reported rapid changes of slope angle on moraines abandoned by the Athabasca Glacier, Alberta, up to 105 years ago, noting a reduction of maximum slope from 60° to ca. 30° in 35 years, but constancy in the following 70 years, possibly due to the persistence of ice cores.

Since colluvial accumulations can be expected to become bulkier and/or their morphology more modified with time, the bulk and form of talus in glaciated, high-relief terrain are potential indicators of the duration of the interval either in which they formed or in which they were modified. In west Newfoundland, the writer has found that inactive, compound talus forms occupy valleys which did not act as outlets for the Late Wisconsinan inland ice cap(s). They show lobate and ridged foot zones which resulted from permafrost aggradation in talus produced on ice-free valley sides. That permafrost existed here at low levels in Late Wisconsinan time is demonstrated by ice wedge casts which penetrate marine sediments of that age (Brookes, 1971). Sorted patterned ground forms also vary in their dimensions and degree of modern activity between weathering zones C and A of Grand (1977a) in west Newfoundland. In zone C both large diameter, presently inactive and smaller, active stone nets are found, whereas only the latter occur within zone A. Similarly, inactive gelifluction lobes are an order of magnitude larger in zone C than active ones in zones B and A.

Selected regional applications

Contrasts in both the type and degree of weathering across definite boundaries have been used as indicators of glacial limits, and accordingly as chronological tools since the early years of glacial studies in Canada. It is in the western and eastern mountains that these contrasts are best expressed, and where the existence of ice-free areas could most easily be predicted on theoretical grounds (e.g., projection into the mountains of theoretical ice-surface slopes from a distal ice margin). Thus, Bell (1884) noted that, on the Labrador coast:

> The mountains around Nachvak are steep, rough-sided, peaked and serrated, and have no appearance of having been glaciated, excepting close to sea level. The rocks are softened, eroded, and deeply decayed (p. 14 DD),

and Daly (1912) noted that, on the Okanagan Range of southern British Columbia:

> Above the limit of the ice the peaks are greatly disintegrated and widespread felsenmeers are usually present. At those levels the granites were sometimes seen to be deeply weathered, with the generation of many boulders of secular decay (p. 592).

In the eastern highlands of Canada, interpretations of these contrasts have swung between opposing concepts, summarized as "minimum" and "maximum" viewpoints by Ives (1978) in a comprehensive review. Briefly, "minimalists" interpreted weathering differences as evidence of several glaciations, decreasing in extent with age. "Maximalists" have seen them as originating during emergence of 'nunataks', following complete inundation by Late Wisconsinan ice. More recently, another group of "maximalists" see weathering zones as reflections of contrasts in thermal regime at the base of all-encompassing Late Wisconsinan ice.

The early minimalist position of workers such as Bell (1884) in northern Labrador, was strengthened by Coleman (1921), who extended it to Gaspé (Coleman, 1922) and to Newfoundland (Coleman, 1926). Following supersession of this view by a widely accepted maximalist position advocated by Flint (1943, 1947) and Demorest (1943), Ives (1957, 1958a, b) revived the minimum viewpoint in northern Labrador, noting the concordance of geomorphic and biologic evidence there with that from Scandinavia (Dahl, 1955). Later work by Ives, Andrews, and their co-workers applied weathering criteria, as well as rock-stratigraphic evidence, to establish a chronology of multiple glaciation in eastern Baffin Island, similar to that of northern Labrador, but improved with a burgeoning number of radiometric dates from proximal glaciomarine sediments, and amino-acid ratios in marine shells.

In the coastal mountains of northern Labrador weathering zones have been recognized by several workers since renewal of interest in their chronological significance by Ives (1957, 1958a, b). Reviews of these studies have been presented by intervals by Ives (1963, 1974, 1975, 1976, 1978). In his early work, Ives recognized three weathering zones in the central Torngat Mountains. The upper zone contained erratics in a distinctive mantle of felsenmeer. The lowest zone exhibited fresh features of glacial erosion and deposition, whereas the middle zone showed more weathered glacial landforms. Løken (1962) also recognized these zones in the northern Torngat Mountains, with boundaries at lower elevations than in Ives' area. He challenged Ives' view that the highest zone possessed erratics, believing the blocks to be inclusions weathered from local bedrock. Andrews (1963) recognized the lower two zones south of Ives' original area, where the boundary between them was marked by a prominent lateral moraine-kame terrace complex, which he named the Saglek Moraine. He also named the three Labrador weathering zones Saglek, Koroksoak and Torngat, from lowest (youngest) to highest (oldest). To the south of Andrews' area, Johnson (1969) identified these three zones, the Torngat zone containing indisputable erratics. Ives (1976) divided the upper weathering zone into two, reserving the name Torngat for the highest summits with felsenmeer, but no glacial erratics, and introduced the name Komaktorvik for a zone below this, with felsenmeer and erratics. Ives (1978) schematically depicted the sequence of weathering zones and glacial covers in the northern and central Torngats (Ives 1958a, 1976) and Nain-Okak areas (Andrews, 1963), as in Fig. 5.

Despite these congruent findings, no detailed weathering studies have so far been undertaken to determine the degree of difference between the zones in any area of northern Labrador, or the degree of similarity among representatives of each zone in different areas, (but see Gangloff, pers. comm., below). The only comparative criterion is the degree of felsenmeer development, which ranges from absent in the Saglek zone, to incipient in the Koroksoak zone, to mature in the Torngat and Komaktorvik zones. The Saglek zone was assigned to the last glaciation because of its fresh glacial forms, but the age of the bordering Saglek moraines is problematical. Ives (1976) recognized that the moraines may not be coeval throughout northern Labrador. Andrews (1977) distinguished them from a younger and more restricted set, called Tasiuyak, of probable Late Wisconsinan age. He tentatively assigned the Saglek moraines an Early Wisconsinan age, noting that this stade was separated from that of the Tasiuyak moraines by (? Middle Wisconsinan) marine submergence. Mayewski, *et al.* (1981) find the evidence for restricted Late Wisconsinan ice in these areas unconvincing. Moreover, Fillon and Harmes (1982) present evidence in favour of extensive Late Wisconsinan ice grounded on Saglek Bank, seaward of the Saglek moraines.

On northern Cumberland Peninsula, Baffin Island, weathering criteria were used by Pheasant (1971) and Boyer (1972) to distinguish weathering zones I, II, and III. Statistical analysis of their combined data from the Maktak-Narpaing Fiord area supported this discrimination (Boyer and Pheasant, 1974). As in Labrador, the boundaries between

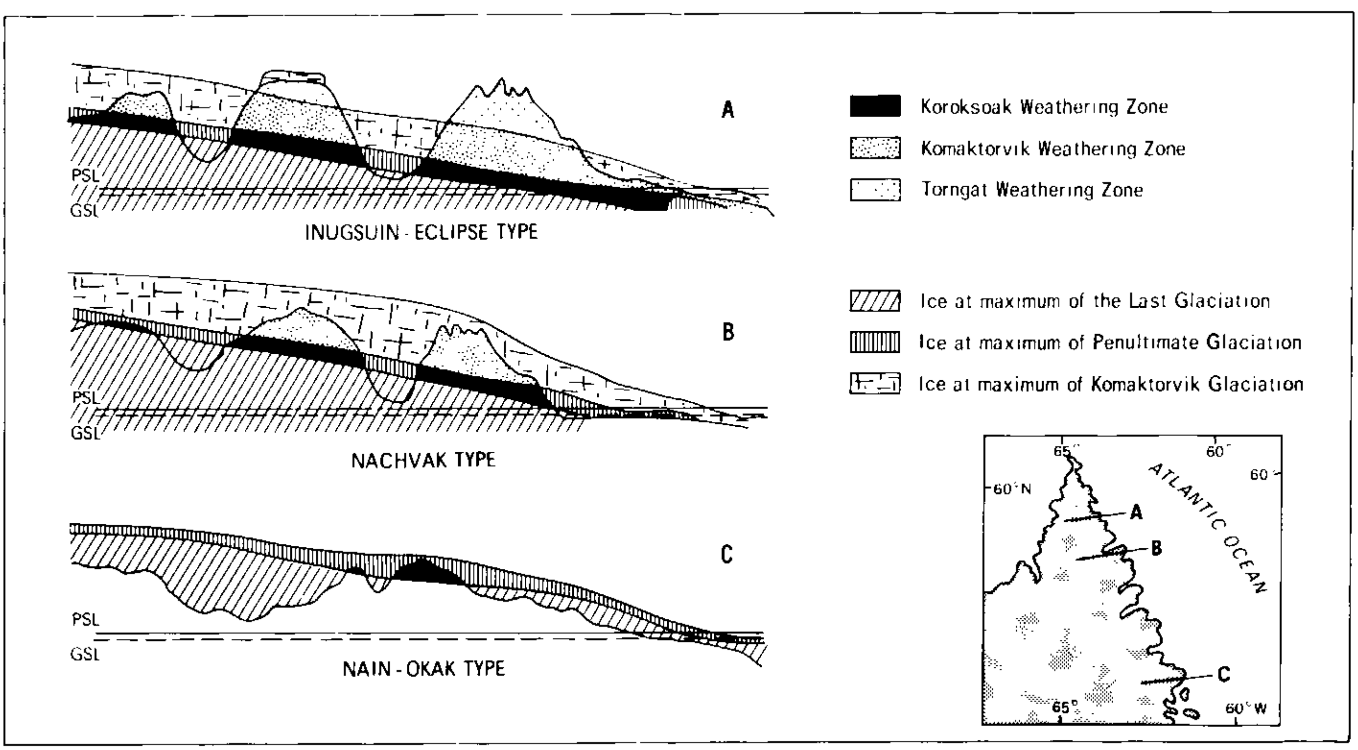

Figure 5 *Scheme for glacial history, glacial style, and disposition of weathering zones in the highlands of northern and central Labrador. Inset shows locations of A, B, C. PSL, GSL, present and glacial sea level. From Ives (1978).*

these zones slope seaward and were interpreted by Pheasant and Andrews (1973) as the upper limits of seaward-thinning glaciers. As in Labrador, the boundary between the lowest and middle zones is marked by moraine fragments which project seawards to end moraines on coastal forelands. Radiometric ages of 100-137 kabp on shells from glacio-marine deposits thought to be associated with the oldest of these moraines (Alikdjuak), led Pheasant and Andrews to assign the lowest weathering zone (III) to the Foxe (~ Wisconsinan) glaciation (ca. 120-8 kabp). The firmness of this correlation is, however, questioned by Mayewski *et al.* (1981). They feel a case can be made for the "Early Foxe" moraines on the coastal forelands being of pre-last interglacial age and for extending the Late Wisconsinan glacial limit seaward. Support for this comes from recent re-evaluation of amino-acid ratios in marine shells from sediments believed by Miller *et al.*, (1977) to be of last interglacial age. This shows that interstadial deposits formerly included in the Foxe Glaciation actually are at least 136 ka old (Szabo, *et al.* 1981). On the basis of ratios of clay mineral species and Fe_2O_3 content of soils from these weathering zones, Andrews (1974) tentatively assigned ages of 450 kabp to weathering zone II (and correlated it with the Koroksoak zone of Labrador), and 680 kabp to zone I (correlated with the Torngat zone of Labrador).

These uncertainities in the correlation of rock-stratigraphic units on Cumberland Peninsula are matched by the aforementioned discordances in weathering data from northern and southwestern parts of the peninsula (compare Pheasant and Andrews, 1973; Boyer and Pheasant, 1974, with Dyke, 1979). Dating control on Foxe glaciation stadial ice limits in the southwest, while not firm, is more reliable than in the north, and the orderly maturation of boulder weathering there with increasing age of surface lends confidence to it. With mechanical disruption of bedrock, A. Dyke's (1979) observations are not quantified, but felsenmeer is described as "common" beyond the Duval Moraines, over terrain which was deglaciated in either Early Foxe time (120-80 kabp) or before. Felsenmeer is described as "patchy" over terrain between the Duval and Ranger Moraines, which was deglaciated between 60 ka and 9 ka (although the duration of the stand at the early Holocene Ranger Moraine is unknown). These observations are in general accord with those from the mainland Shield zones beyond the tree-line, where felsenmeer is generally only incipient over terrain deglaciated in the Holocene. They contrast markedly with the observations on northern Cumberland Peninsula, where no felsenmeer is recorded from terrain deglaciated at various times within the last 120 ka. The difficulty is most easily resolved if the age of weathering zone III in that area is re-evaluated, as suggested by Mayewski *et al.* (1981).

In eastern Ellesmere Island, weathering zones with boundaries marked by moraines or erratics have been recognized by England *et al.* (1981). An uppermost zone (IIIb) is free of erratics or other evidence of glaciation, and is deeply oxidized and frost-shattered, with abundant tors and advanced solution of carbonates. At lower elevations, across a boundary marked only by the maximum extent of erratics, zone IIIa exhibits comparable weathering. This boundary declines to the southwest, and this, together with a Greenland provenance for the erratics, indicates that the oldest recognized glaciation here was from Greenland. Extensive areas remained ice-free during this glaciation, which, although undated, is believed to antedate the last interglacial considerably.

Weathering zone II on Ellesmere Island is characterized by advanced surface weathering, such as "highly frost-shattered and oxidized clasts" (England *et al.* 1981, p. 75) in moraines which mark its boundary with zone IIIa. Sloping eastward, this boundary marks the upper limit of east-moving Ellesmere Island ice. Recession appears to have been in progress before 35 kabp, possibly before 70 kabp (England, *et al.* 1978; 1981), so that the duration of Zone II weathering is comparable to that in A. Dyke's (1979) zone A5 (modal boulder weathering classes 6 and 7, infrequent macro-pits, patchy felsenmeer). Zone I on Ellesmere Island is bounded by the much more restricted end-Pleistocene/early Holocene Hazen Moraines (England, 1978), from which retreat was in progress ca. 8.0 kabp. Only slight weathering and morphological modification is reported from this zone, which is correlative with zone A3 of A. Dyke (1979).

In the Canadian Cordillera, while weathering differences were used early to map and gauge the status of glacial limits (e.g., Daly 1912), subsequent investigations have added little to characterising those differences or dating their development. Several factors account for this. First, in many of the more accessible areas of the southern Cordillera, Fraser (Wisconsinan) Glaciation ice cover was extensive. Any 'nunataks' were steep-sided peaks which would not likely retain earlier weathering products, and which have weathered rapidly since the Fraser maximum. Second, where glaciation was less extensive, in the drier northern interior ranges and basins, access is limited. Third, the few detailed studies have come from areas in which rock-stratigraphic methods and radiometric dating have provided acceptably reliable chronologies (Denton and Stuiver, 1967), so that it has not been necessary to rely upon other dating methods.

In the Yukon Territory, Bostock (1966) distinguished an older Nansen Drift from Klaza Drift by a contrast in till clast weathering and by modification of their morphology by solifluction. Above the Nansen Drift, in areas lacking evidence of glaciation, Bostock noted "castellated" outcrops, which, from their description and illustration (see Bostock 1966, Plate VI, p. 18) can be recognized as ridge-crest tors. Over a wider area of the Yukon, including that mapped by Bostock, Hughes *et al.* (1969) used "comparative morphology" to distinguish the same drifts and their equivalents. They found that deposits and associated morainic morphology were difficult to recognize and may not often have survived erosion on the steep sides of valleys which contained outlet glaciers from Wisconsinan mountain ice sheets. Alternatively, as Mercer (1956) noted in southern Baffin Island, moraines may never have formed if the firn limit on glaciers was greatly lowered.

In the Alaska Range, Wahrhaftig (1949) recognized five generations of "frost-moved rubbles" on and surrounding andesitic Jumbo Dome. He was unable to assign ages to each, but from stratigraphic and topographic relationships he related them to periods of periglacial frost-shattering and debris mobilization, interrupted by periods of stabilization and fluvial dissection.

Objections and alternative hypotheses

While weathering criteria have long been used successfully and without challenge to establish glacial chronologies in many areas, most notably the American Cordillera, their application to eastern Canadian uplands has occasionally been resisted. Much of the debate over the significance of weathering zones in these uplands stems from a difference in viewpoint (partly based on independent evidence) concerning the extent of the Late Wisconsinan Laurentide Ice Sheet and contiguous glacier complexes.

Objections arising from weathering studies have generally drawn attention to Holocene weathering rates which have been rapid enough to produce some features inferred by others to indicate a longer exposure. Watts (1979, 1981), for example, has noted rounding of outcrops, widening of joints, dense and deep pitting, and grussification on granitic terranes at a variety of elevations in the eastern Canadian Arctic. He draws attention to the influence of salt crystallization and high hornblende content in giving rise to many apparently advanced weathering features, and thus cautions against their interpretation as indicators of ice-free conditions during a glaciation. Many of these features are probably special cases of lithological or environmental enhancement of weathering intensity, and do not warrant fundamental reconsideration of weathering zones.

In northern Norway, R. Dahl (1966) denied the validity of weathering differences as evidence of 'nunataks'. He saw no regular trend in the lower limit of blockfields, which E. Dahl (1955) had interpreted as a glacial limit, finding instead that this was slope-controlled. R. Dahl preferred post-Weichselian frost-shattering over areas emerging early from a complete ice cover. He further saw macro-pitting as caused by intense "microgelivation" on near-horizontal lichen-covered surfaces of southerly aspect. Ives (1966) countered by asking how weathering pits could have developed on blocks which had been actively frost-churned in post-Weichselian time. He preferred to see them developing on stabilized blockfields formed in the Weichselian over areas which remained ice-free. This is affirmed by later work in Norway (Stromquist, 1973; Sollid and Sorbel, 1979).

Instances are reported in which an origin of felsenmeer by other than primary periglacial weathering can be proposed. In the Torngat Mountains of northern Labrador, P. Gangloff (pers. comm., 1982) finds the case for this origin weak for the felsenmeer which Ives (1957, 1958a, b) used to define the Torngat weathering zone. He finds no significant differences between the matrix of Wisconsinan till in the Saglek zone and felsenmeer fines (< 2 mm) with respect to texture, clay content and mineralogy, and quartz grain surface textures. Gangloff thus interprets the felsenmeer as a periglacially modified Wisconsinan till. In the Chic-Choc Mountains of Québec, Gray and Lafrenière (1981) also found no difference in these parameters amongst Late Wisconsinan glacial deposits on the upland flank, plateau-top blockfields, and polygonal terrain. They therefore suggested that the latter two units may represent tills transported only a short distance by a local plateau ice cap of possible pre-Late Wisconsinan age. In the southern Shield of Saskatchewan, Alley and Kupsch (1982) report allochthonous felsenmeer over glacial deposits and bedrock. This situation may be genetically related to one reported by L. Dredge (pers. comm., 1982) from northern Manitoba, where Late Wisconsinan ribbed moraine contains joint-bound blocks of felsenmeer moved over distances short enough for some blocks to be matched with bedrock hollows < 100 m up-glacier.

Glaciological counter-arguments to the chronological interpretation of weathering zones have arisen with the refinement of theory on thermodynamic regimes at glacier beds (Weertman, 1957, 1964; Boulton, 1972). Sugden (1974) developed an elegant model, incorporating glacier-bed thermodynamic regime, subglacial topography and bedrock porosity, to explain the contrasts between overdeepened troughs and "weathered" plateaus above them in Greenland. This hypothesis was later applied to northern Cumberland Peninsula, Baffin Island (Sugden and Watts, 1977) to account for the contrast between weathering zones I and II/III.

Along the eastern seaboard of Canada many plateaus at 500-1500 m elevation possess a surface relief of 100-200 m. This may lead to the suspicion that if, during a glaciation, ice was confined to shallow valleys, leaving some interfluve and summit areas ice-free, it would not be thick enough, nor would it slope sufficiently to raise basal shear stress to levels conducive to erosion. If so, morphological contrasts could not be explained by weathering over different intervals. This was tested by the writer in an area of west Newfoundland where Grant (1977a) identified three weathering zones. In one locality the boundary between the uppermost zone (C) and the middle zone (B) is marked by discontinuous moraine fragments, which permit the longitudinal ice surface slope to be determined at 0.007-0.014, and the maximum ice thickness in the adjacent valley to be calculated at 30 m. Basal shear stress would have been 185 - 370 kPa, which is in excess of values normally associated with flowing glacier ice (Paterson, 1981). Therefore, it is not necessary to project an ice surface over weathering zone C in order to produce sufficent basal shear stress for erosion in zone B valleys.

Whether or not it was "cold-based", the ice which emplaced erratics on west Newfoundland summits within weathering zone C must have been erosive. Plateaus covered with felsenmeer, contain not only far-travelled erratics, but also erratics of local lithologies which have been transported from the up-glacier to the down-glacier sides of the plateaus. This implies glacial erosion of the summits and formation of the felsenmeer after deglaciation. That this glaciation was not of Late Wisconsinan date is demonstrated by gradients on the upper boundary of the lowest weathering zone (A), which remains below these plateau surfaces, and which intersects end moraines and glaciomarine deposits ^{14}C-dated at 14.0-12.5 kabp near the present coast (Grant, 1977b).

The question of the survival of erratics emplaced on these plateaus during a pre-Wisconsinan glaciation while blockfields have formed around them is a thorny one. In its favour are the points that 1) the original density, size, and shape of erratics is not known, so the amount of erratic weathering is impossible to gauge, 2) erratics have occasionally suffered frost-shattering equally as severe as that visible in a surrounding blockfield, 3) decomposed erratics are usually granites in which decomposition progresses rapidly (Watts, 1981), and 4) in west Newfoundland, erratics of any lithology are not discernibly less weathered than blockfields of the same lithology in comparable topographic settings.

The glaciological counter-argument to the morphostratigraphic interpretation of weathering zones has led to a degree of polarization between the two camps. However, it is beginning to be realized that a case can be made for the existence of "cold-based" ice in specific cases where the field evidence is not as open to dual interpretation. For example, in northeast Somerset Island, N.W.T., A. Dyke (in press) has mapped an extensive terrain tract consisting of coarse gneissic and carbonate gruss near low interfluves, which stands upslope from finer-grained, transported residuum derived from the gruss by solifluction and slopewash. These slopes grade smoothly to fluvial valleys which are organized dendritically. Superimposed on this manifestly subaerial landscape are myriad side-hill meltwater channels which record successive ice marginal positions, occupied during retreat of a local ice cap. Preservation of delicate subaerial forms beneath this ice cover demands that the ice was "cold-based". Peripheral to this terrain, glacially scoured forms record erosion by "warm-based" ice below the equilibrium line. Marine sediments deposited over isostatically depressed coastal fringes on Somerset Island yielded shells with ^{14}C-dates no older than 9.3 kabp, indicating that the last ice to affect the island was of Late Wisconsinan age.

In the Canadian Appalachians, where for over a decade evidence has been accumulating which points to areally and vertically restricted Late Wisconsinan ice margins (Grant 1977b), certain features force consideration of the effects of "cold-based" ice. Whether this was of

Late Wisconsinan age is, however, presently considered doubtful in most cases.

In the highlands of northern New Brunswick, Gauthier (1980) reported granite weathered to depths of up to 60 m, and residual summit tors of the 'woolsack' type. The tors consist of rounded boulders up to 4 m in diameter, occasionally free-standing, more often attached to outcrops exhibiting exfoliation shells and large, compound weathering pits. Mineralogical analyses of the residuum by Wang *et al.* (1981) indicate the presence of kaolinite and gibbsite which indicate a pre-Wisconsinan and likely pre-Pleistocene age. Erratics are strewn around the tors and presumed basal till overlies the residuum. (Veillette and Nixon (1982), however, located metasedimentary inclusions in the granite, and caution against interpretation of "erratic" lithologies as foreign to the area.) Gauthier (1980) attributed the glacial evidence to *active* Late Wisconsinan ice which affected the highlands "to a lesser extent and during an earlier phase than the lowlands" (p. 281). Earlier, Gauthier (1978, 1979) had resorted to "cold-based" ice to preserve granite cliffs pre-weathered to castellate forms in an adjacent area.

In a somewhat comparable topographic setting in the highlands of Cape Breton Island, Nova Scotia, McKeague *et al.* (1982, submitted for publication) describe a saprolite, weathered from granite-gneiss and overlain by till. The till was emplaced by a regional ice sheet which crossed these highlands in pre-Late Wisconsinan time (Grant, 1977b). The preservation in the saprolite of gneissosity and pegmatite veins is thought to indicate minimal distrubance during glaciation, and, accordingly, to support the "cold-based" ice hypothesis. However, since it is not known if glacial erosion of the saprolite did in fact occur before the till was emplaced on it, this evidence is not conclusive.

On Îles de la Madeleine, Québec, Grant (1981) attributes deformation structures in Pennsylvanian sandstone to glaciotectonic stresses beneath "cold-based" ice, again within a regional pre-Late Wisconsinan ice sheet. Thus, in one case it is proposed that "cold-based" ice preserved delicate structures in a saprolite, while in another, such ice is supposed to have deformed more competent sandstone.

Conclusions

Weathering contrasts have been used to good effect, in some cases to define glacial limits, in others to erect relative chronologies for limits established from glacial evidence. More detailed studies of weathering differences are needed to permit the regional schemes so far erected to be correlated. It is perplexing to note that the two most detailed studies made in neighbouring areas of Baffin Island (Boyer and Pheasant, 1974; Dyke, 1979) differ in their descriptions of weathering zones which are believed from independent evidence to be correlative. As independent chronological control on weathering zone boundaries becomes available, the value of weathering differences to chronology will decline. However, complementary gains will be made in knowledge of the rates at which weathering processes operate and the products evolve (Andrews and Miller, 1980).

In spite of the longevity of weathering studies as aid to glacial chronology and wide acceptance of their value in the North American Cordillera, sceptics question the meaning of weathering zones in the highlands of eastern Canada. Opposing views are strongly held, since at stake is the much broader question of the extent and volume of the Laurentide Ice Sheet at the Late Wisconsinan stadial maximum. This is not the place for a detailed discussion of all the evidence, but that so far obtained from ice-marginal and marine deposits and landforms, radiometric and amino-acid dating, and the glacial limits and sea level histories inferred from it, complements rather than conflicts with that from weathering studies in favouring the interpretation of weathering zone boundaries as glacial limits. Opposing views are presently compelling more objectivity in the interpretation of field evidence.

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