WELL PRESERVED LATE PRECAMBRIAN PALEOSOLS FROM NORTHWEST SCOTLAND

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ABSTRACT: Two paleosol profiles were developed side by side on gneiss (Sheigra clay paleosol) and amphibolite (Staca clay paleosol) along the same unconformable contact of Lewisian metamorphic rocks (2.7-1.7 Ga) overlain by Torridonian sandstones (0.8 Ga) in northwestern Scotland. Although altered during burial by compaction and illitization, these profiles are well preserved compared with most other Precambrian paleosols that are metamorphosed to at least greenschist facies. In addition, deformation of pegmatite veins in a manner similar to soil creep can be taken as an indication that upper horizons of these paleosols have been preserved.

Soil development in both profiles was characterized by the oxidation of iron in biotite and magnetite, and by hydrolysis of plagioclase, orthoclase, and hornblende. Clay skins, soil-like microfabrics, and pedogenic smectite have been preserved in the profiles. Alteration of silicate grains other than quartz, microcline, and biotite has been intense, but chemical analyses failed to show dramatic removal of alkalis and alkaline earths from the profiles. Microscopic concretions similar to pedogenic carbonate were found in subsurface horizons of the Staca clay. Rare-earth elements are more abundant within the paleosols than in their parent materials.

These paleosols developed on a well drained landscape that was buried by Torridonian alluvial fans. Atmospheric oxidation was greater than revealed by studies of Archean paleosols, but probably still short of modern levels. Paleoclimate was subhumid and probably seasonal. Organic carbon and its light isotopic composition in the paleosol is evidence for photosynthetic microbial crusts. Paleosols of such great antiquity do not fit comfortably within modern soil taxonomies, and are for the moment assigned to the informal category of "Green Clays".

INTRODUCTION

Two problems particularly impede the study of Precambrian paleosols as evidence of ancient environments. First is how to recognize Precambrian paleosols, which lack fossil root traces that are diagnostic of paleosols younger than Silurian (Retallack 1992a). Discovery of Precambrian paleosols thus relies upon recognition of other evidence of biological activity, of soil horizons, and of soil structures. The emphasis of many previous studies of Precambrian paleosols has been geochemical (Holland and Beukes 1991; Holland et al. 1988; Holland and Zbinden 1988), but diagnostic soil features also are beginning to receive attention (Retallack 1986a, 1992a; Grandstaff et al. 1986).

A second problem is alteration after burial, metamorphism, and exposure, which has often obscured the original chemical and mineralogical composition of Precambrian paleosols (Palmer et al. 1989). Some alterations are obvious, such as the growth of porphyroblastic crystals during high-grade metamorphism (Barrientos and Selverstone 1987; Dash et al. 1987; Golani, 1989; Bol et al. 1989). Other kinds of alteration at low temperature and pressure, such as reddening of ferric hydroxides, are difficult to distinguish from soil formation (Schau and Henderson 1983; Retallack 1991a; Nesbitt 1992).

For these problems of recognition and alteration, there is a need for detailed studies of soil features of geologically young and little-altered Precambrian paleosols. Here we present such a study of two paleosols on different parent materials side by side along the same major unconformity between Lewisian (2.7 Ga) gneisses and Torridonian (0.8 Ga) sandstones in sea-cliff exposures near the hamlet of Sheigra in northwest Scotland. These paleosols were previously studied by Williams (1966, 1968, 1969) and Russell and Allison (1985).

Although we are mainly concerned here with building a data base of criteria for disentangling the effects of ancient soil formation and burial alteration, understanding paleoenvironments of the paleosols remains a fundamental motivation for our studies. The work reported here is relevant to current interest in the identification of Precambrian paleosols within classifications of modern soils (Retallack 1990) and the increased oxygen content of the atmosphere over the long sweep of Precambrian time (Holland and Beukes 1991; Holland and Zbinden 1988).

GEOLOGICAL BACKGROUND

In sea cliffs of northwestern Scotland from Cape Wrath south to Sheigra are exposed a variety of paleosols along the unconformity between Lewisian biotite gneiss, amphibolite, and microcline pegmatite and the overlying sandstones and conglomerates of the Applecross Formation of the Torridon Group (Fig. 1). The Lewisian gneisses are largely 2700 million years old (Scourian), but amphibolite-facies metamorphism and injection of pegmatites continued until some 1800 to 1700 million years ago (Laxfordian), especially in the area near Sheigra examined here (Watson 1983). Amphibolite near Sheigra forms a broad, nearly vertical dike some 14 m wide within the gneiss, striking northeast into the cliff but folded along a nearly vertical axis to an easterly strike on the rock platform. It was probably a tholeiitic diabase dike (post-Scourian), subsequently metamorphosed with the enclosing gneisses. The overlying Torridon Group has yielded a Rb-Sr isochron age of 810 ± 17 Ma, but deeper parts of the sequence within valleys within the unconformity to the south were filled with the Stoer Group, dated in the same way at 999 ± 24 Ma (Moorbath 1969). This Late Proterozoic (Riphean) age is consistent with evidence of microfossils found in gray shales and in metasedimentary pebbles of the Torridon Group (Peach et al. 1907; Naumova and Pavlovsky 1961; Sutton 1962; Downie 1962; Muir and Sutton 1970).

Both paleosol profiles studied are in the sea cliff east of the small hamlet of Sheigra (Fig. 1; British National Grid NC183607). Here Applecross conglomerate and sandstone dips 10°N and strikes 214°. The clayey paleosols pass gradationally down into crystalline parent materials but are abruptly truncated by the overlying conglomerates and sandstones (Fig. 2). The paleosols also show a variety of soil features in the field, including local variation in weathering similar to spheroidal weathering and bending of pegmatite veins similar to soil creep (Williams 1968).

The paleosols formed on a subdued landscape, with local hills up to 150 m high, buried by alluvial-fan deposits of the Applecross Formation (Williams 1966, 1969; Stewart 1972). To the south near Stoer and Aultbea, deeper valleys of an ancient landscape with topographic relief of up to 1000 m were filled with marine and fluvial deposits of the Stoer and Sleats Groups (Selley 1965, 1969, 1970). The alluvial-fan deposits of the Applecross Formation include clasts of Lewisian gneiss, amphibolite, and pegmatite, as well as exotic clasts compatible with derivation from mountains to the west, now presumably drifted off as the Greenland part of the Laurentian Shield (Muir and Sutton 1970; Allen et al. 1974). Paleomagnetic studies indicate that Torridonian alluvial fans accumulated at a paleolatitude of about 30°, probably south of the equator (Stewart and Irving 1974).

METHODS

Field observations and oriented samples of the paleosols were collected (by GJR) in 1984. Oriented petrographic thin sections were point counted

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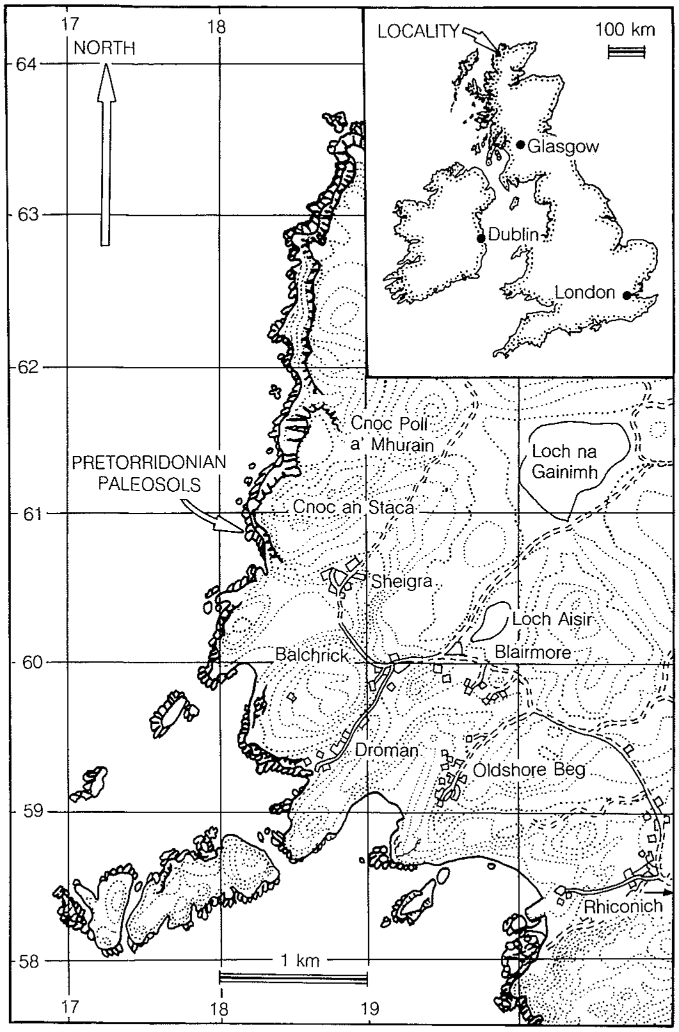

FIG. 1.—Location of pre-Torridonian paleosols studied in northwest Scotland.

using a Swift Automatic Point Counter. Chemical analysis for major elements and a variety of trace elements was done by atomic absorption, with ferrous ammonium sulfate titration for ferric iron and loss on ignition to 700°C for water and 1000°C for samples containing carbonate. Analyses of trace elements and of REE were from Neutron Activation, normalized to C1 carbonaceous chondrite abundances (after Anders and Ebihara 1982). Bulk density was determined from paraffin-coated clods weighed in and out of water. Clay minerals from the bulk rock as well as separates made by wet sieving, centrifuging, and hand-picking were identified from air-dried, ethylene-glycolated, and heated treatments using a computerized Rigaku Miniflex X-ray diffractometer (Cu radiation, Ni filter, step size 0.03 °2θ, count length 5 seconds). A CAMECA microprobe and JEOL scanning electron microscope was used to establish the chemical composition and micromorphology of mineral grains and their fine-grained weathering products. Total organic carbon and its isotopic composition was determined on selected samples by the Precambrian Paleobiology Research Group using methods outlined by Schopf (1983) and Schopf and Klein (1992).

TWO KINDS OF PALEOSOLS

The pre-Torridonian unconformity in the sea cliffs near Sheigra (Figs. 1, 2; NC183607) is remarkable for including a paleocatena of two different

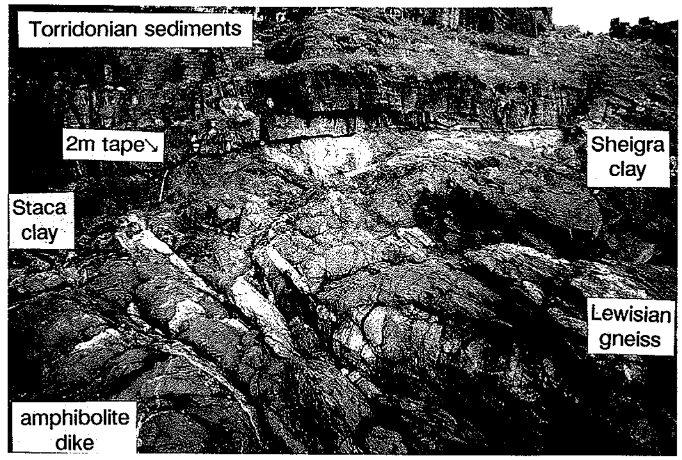

FIG. 2.—The type Sheigra clay (to right) and Staca clay (to left) paleosols in the sea cliff west of Sheigra, northwest Scotland. Tape to left of center on Staca paleosol is 6 ft (2 m).

paleosols: the Sheigra clay paleosol developed on biotite gneiss and the Staca clay paleosol developed on a broad amphibolite dike. The Sheigra clay has been named previously (Retallack 1986b, 1990) but this is its first detailed description (Fig. 3; Table 1). The Staca clay paleosol is 10 m to the north (Figs. 2, 4; Table 2). The name of this paleosol is from nearby Cnoc an Staca (NC184611), a Gaelic place name that can be translated "knoll of steep rocks" (MacLennan 1979).

OPTICAL PETROGRAPHY

An outstanding attraction of these paleosols is the coarse grain size of their parent materials. The alteration of large crystals is easy to see in optical thin section (Tables 3, 4), and both the weathering products and parent grains make large targets for analysis by electron microprobe (Fig. 5; Tables 5, 6). The amphibolite and gneiss share many minerals, the former being much richer in green hornblende (Figs. 3, 4). Other minerals of the parent material are quartz, K-feldspar (including microcline and perthite), plagioclase (andesine and oligoclase), biotite, apatite, zircon, ilmenite, and magnetite. Phases formed by alteration in cracks and around these large grains include albite, quartz, calcite, hematite, and fine-grained phyllosilicates such as chlorite, illite, and interlayered illite/smectite. Not all these alteration products are necessarily pedogenic or diagenetic, a subject to which we will return. In this section our purpose is merely to outline the sequence of alteration of primary grains observed in thin section.

Microcline and quartz were remarkably resistant to alteration, showing little change beyond mechanical disintegration near the tops of the profile, where they are the main recognizable grains in a clayey matrix (Fig. 5A, C). The accessory minerals, apatite and zircon, also show little change from bottom to top of the profile.

Ilmenite is present both as subhedral and anhedral opaque grains, in some cases mantled with anatase or rutile, both in the parent material and paleosols. Conversion of ilmenite and magnetite to hematite is most pronounced in the paleosol but is evident to some extent also in little-weathered parent rock. Concentrically layered collomorphous oxides are intimately intergrown with microlayers and "blebs" of silica (Fig. 6A). The outer surface of one of the concentrically layered collomorphous oxide grains is surrounded by radiating elongate crystals that appear to have grown on it (Fig. 6C). Microprobe analysis of large crystals in the halo gave a composition of silica (65.00%), alumina (18.52%), lime (0.02%), iron (0.44%), soda (0.46%), and potash (16.02%). Other interleaved acicular crystals include silica (58.70%), alumina (19.74%), iron (5.54%), lime

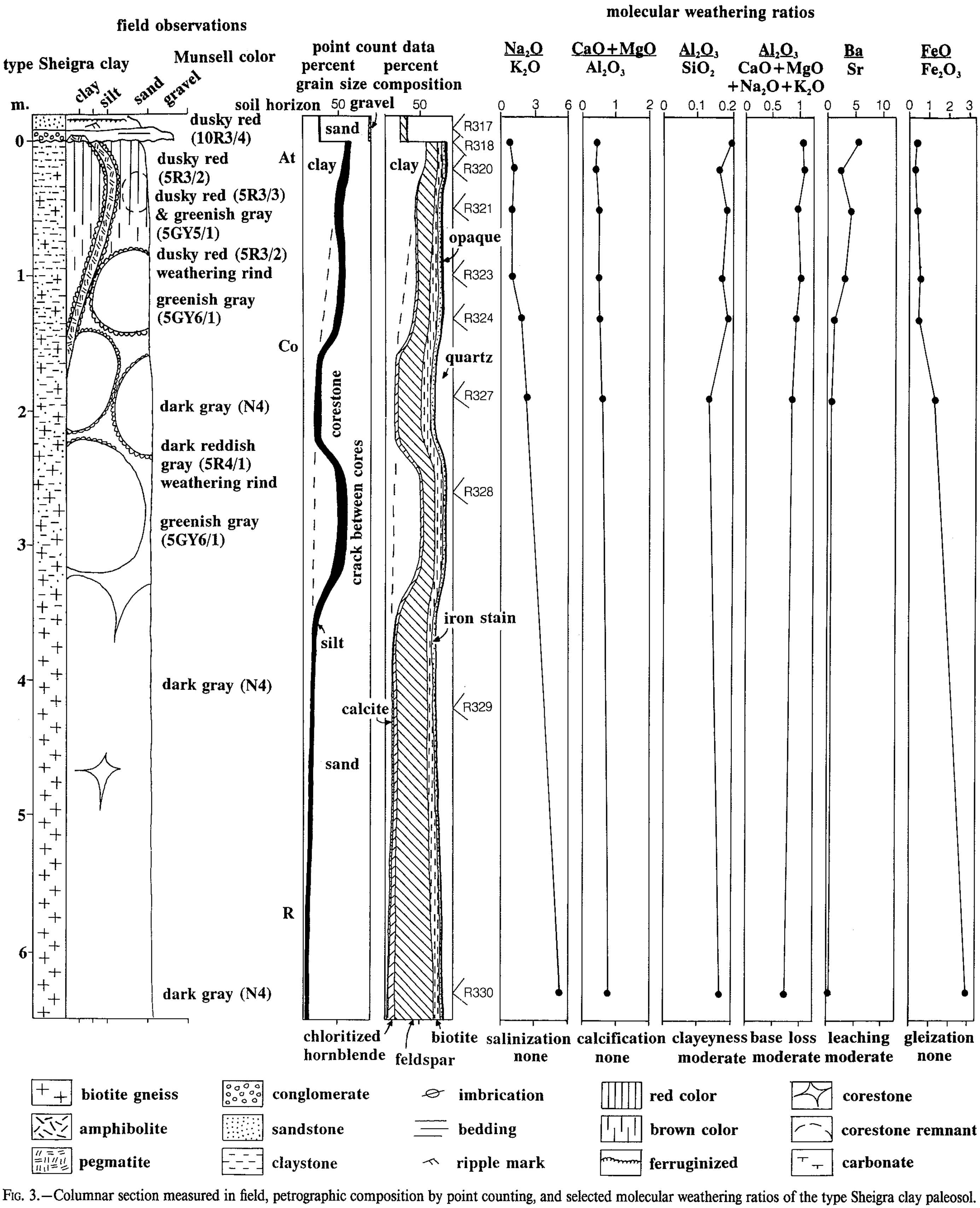

FIG. 3.—Columnar section measured in field, petrographic composition by point counting, and selected molecular weathering ratios of the type Sheigra clay paleosol.

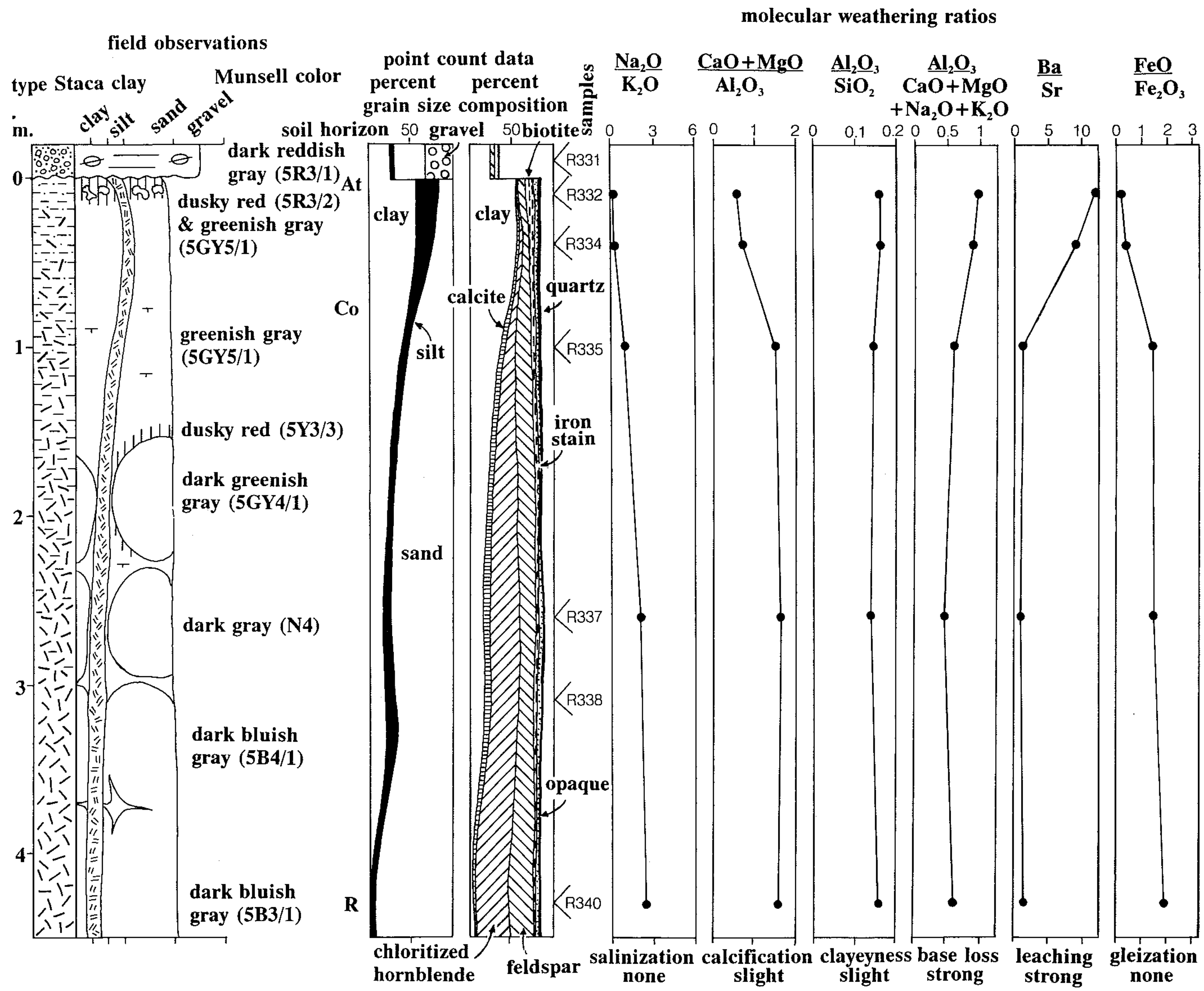

FIG. 4.—Columnar section measured in field, petrographic composition by point counting, and selected molecular weathering ratios of the type Staca clay paleosol.

(3.89%), soda (1.01%), and potash (10.57%). We interpret this as microperthite rind, probably formed during Laxfordian (1.7 Ga) metasomatism of the parent material, like other mineral changes described by Beach (1973), and then oxidized during Precambrian weathering of the ilmenite.

Biotite, plagioclase, and orthoclase within the paleosols are mantled and embayed by clay minerals, usually with hematite stain. These common products of hydrolysis also extend into fractures. Biotite sheets are splayed and bulged by interleaved clay and hematite. In the case of plagioclase, the oxides and clays are accompanied by small amounts of calcite. Within some biotites, both weathered and fresh, are small blebs and spindle-shaped grains that proved under the microprobe to be quartz and albite, and may be products of Laxfordian metasomatism of the parent material (see Beach 1973).

Little completely fresh hornblende was seen. Even samples from deep within the amphibolite profile or within corestones show pervasive replacement with a turbid green 14 Å chlorite with a ratio Fe/(Fe+Mg) of 0.48. This may be a product of Laxfordian metasomatism or subsequent retrograde metamorphism of the parent material. In more clayey parts of the paleosols, chloritic remnants of hornblende grains are enclosed by a more iron-rich chlorite (Fe/(Fe+Mg) of 0.54). Cleavage planes of replaced hornblende grains are lined by an even more iron-rich phyllosilicate identified as smectite (Fe/(Fe+Mg) of 0.66). In the final stage of amphibole alteration this iron-magnesium phyllosilicate is displaced by yellow smectite, and then transparent illite invading the remaining open space along the cleavage planes. Cleavage planes of hornblende also include small iron-stained grains of calcite (Fig. 7). Williams (1969) found that carbonate in the Sheigra clay stained as if it were ferroan dolomite, with minor amounts of ankerite, a determination supported by one of our microprobe analyses (Table 6).

CLAY MINERALOGY

Four different clayey phases were recognized through a combination of optical microscopy, scanning electron microscopy, microprobe analyses, and X-ray diffraction (Fig. 8).

Green chlorites formed mainly from hornblende are most abundant in the lower part of the Staca paleosol but are detectable throughout that profile. This chlorite has a distinct 7 Å peak and a small 14 Å peak that

TABLE 1.—*Description of the type Sheigra clay paleosol**

Top (cm)	Horizon	Rock Type	Prominent Colors	Other	Microfabric	Bottom contact
+14	—	Coarse-grained sandstone	Dusky red (10YR3/4), with some bands of olive (5Y5/3)	Grains of reddish black (10R2.5/1), red (10R5/6), and reddish brown (5YR4/3); bedded with local ripple-drift cross-lamination	Granular silasepic, with abundant quartz	Clear, smooth
+7	—	Conglomeratic sandstone	Mottled greenish gray (5GY6/1) and dusky red (10R3/4)	Most grains well rounded, vein quartz, 2–30 mm diameter, light gray (5YR7/1), in sandy matrix or forming short trains: near base, large (20 cm) flat angular clasts lie parallel to bedding near pegmatite veins from which they were derived	Granular silasepic with abundant vein quartz granules	Abrupt wavy
0	At	Claystone	Dusky red (5R3/2), with numerous crystal pseudomorphs of greenish gray (5GY5/1)	Coarse subangular blocky peds defined by slickensided clay skins of dusky red (5R3/2): quartz-microcline pegmatite veins protrude beyond top of profile, form ridges in outcrop, and are red (10YR4/6) with quartz of light gray (5Y7/1): non-calcareous	Agglomeroplasmic skelmosepic: microcline, quartz and oxidized biotite grains in illite-smectite matrix	Diffuse wavy
−16	AC	Sandy claystone	Dusky red (5R3/3), with crystals reddish black (5R2.5/1), light gray (5Y7/2), greenish gray (5GY5/1)	Pegmatite veins are red (10R4/6) and weak red (10R4/4) with quartz of light gray (5YR7/1): non-calcareous	Intertextic to agglomeroplasmic skelinsepic, with common quartz and microcline, and local illite-smectite matrix	Diffuse irregular
−65	Co	Sandy claystone	Greenish gray (5GY6/1), with relict crystals dusky red (5R3/2), and pinkish gray (5YR7/2)	Relict foliation of gneiss dipping 58°S along strike 279° is disrupted by slickensided joints forming a coset at low angle to the unconformity surface; relict corestones outlined by diffuse zones of dusky red (5R3/2) with some relict crystal outlines of pale olive (5Y7/3); pegmatites are weak red (10R4/3) and light gray to gray (5Y5/1): non-calcareous	Intertextic to agglomeroplasmic skelsepic: common quartz, microcline and biotite, but plagioclase and hornblende altered	Diffuse irregular
−170	Co	Sandy claystone	Greenish gray (5GY6/1), with crystals of reddish brown (5YR4/4) and black (N1)	Common large (two measured were 120 × 50 cm, and 150 × 45 cm) hard gneiss corestones of dark gray (N4), with crystals of dark gray (N4), black (N1), reddish brown (5YR5/3), and dusky red (10R3/2), and thick (7 cm) weathering rinds (diffusion sesquans) of dark reddish gray (5GY6/1), with crystals of dusky red (5R3/3) and greenish gray (5GY6/1): pegmatite veins are red (10R4/6) and light gray (5YR7/1): non-calcareous	Granular silasepic in corestones to intertextic skelsepic in intercore areas, which show more clayey matrix than corestones	Diffuse irregular
−460	R	Biotite gneiss	Gray (N5) and dark gray (N4), with crystals gray (N5), black (N1), and dusky red (5R3/2)	Foliation of quartz-rich and mafic-rich bands 6–8 mm wide: pegmatite veins are red (10R4/6) and light gray (5YR7/1): non-calcareous	Granular silasepic; with quartz, biotite, chloritized hornblende, microcline, and oligoclase	

* Microfabric terminology is from Brewer (1976), soil horizon terminology from Soil Survey Staff (1951, 1962), and color from Munsell Color (1975).

remains on shifting of the larger 14–15 Å peak from more abundant smectite. The chlorite is very iron rich. Chlorite could be clearly detected only from hand-picked patches of altered hornblende.

The most abundant clay is illite with perhaps minor interlayering of smectite, forming the light-gray-green soil matrix within which disintegrated silicate grains are floating. The basal X-ray reflection of this clay is broad, at 10.2–10.4 Å, and it shifted slightly on glycolation as in dioctahedral illite. Under the SEM this clay is stacks of curling flakes, sometimes with thin fingerlike protrusions (Fig. 8A).

Also seen were small patches of transparent illite a few hundreds of micrometers across, mainly within the clayey matrix of the uppermost parts of the paleosols. This clay has a sharper peak more nearly at 10 Å and is also richer in potassium, though on both counts would still fail to qualify as muscovite. It is of slightly larger grain size than the other clays, and some of the fine platelets are organized into vermiform units visible optically. Under SEM, fingerlike, elongate crystals and curling platelets are arranged into irregular books (Fig. 8B).

Smectite is a yellowish clay that invades the cleavage planes, and intragranular fissures, and coats grains, pore spaces, and intergranular cracks. The 14 Å peak of the smectite expands to 17 Å on glycolation and collapses to 10 Å on heating. Under SEM, this clay has a characteristic honeycomb structure (Fig. 8C).

BULK GEOCHEMISTRY

A variety of techniques have been proposed for massaging geochemical data on soils and paleosols in order to make them more readily understandable (Retallack 1990). Four of these have been used on our data (Tables 7–9).

Molecular weathering ratios are calculated by dividing the elemental oxide percents by their molecular weight as a proxy for a variety of common weathering reactions (Figs. 3, 4). This approach has been widely used (for example, Marbut 1935; Retallack 1983, 1991b) because it is relatively free from assumptions except that the material in question is indeed a paleosol and that changing molar proportions of elements are related to common weathering reactions found in soils today.

TABLE 2.—*Description of the type Staca clay paleosol**

Top (cm)	Horizon	Rock Type	Prominent Colors	Other	Microfabric	Bottom Contact
+20	–	Conglomerate	Dark reddish gray (5R3/1)	Abundant well-rounded pebbles mostly 4 cm in diameter, but some up to 5 cm; red (10R4/6), weak red (10R4/3), and pinkish gray (5YR6/2): this is the same conglomeratic layer, somewhat thicker than to the south where it overlies the type Sheigra clay: non-calcareous	Granular silasepic with abundant vein quartz granules	Abrupt wavy
0	At	Claystone	Dusky red (5R3/2), with prominent crystal relicts of greenish gray (5GY5/1)	Pegmatite dikes are dusky red (10R3/4) with crystals of olive (5Y5/4); moderately calcareous	Agglomeroplasmic skelmosepic, with quartz and microcline in clayey matrix, some ferruginized calcite veins	Diffuse wavy
−7	Co	Sandy claystone	Greenish gray (5GY5/1), with crystal relicts of dusky red (5R3/2) and dark red (5R3/6)	Pegmatite dikes are dusky red (10R3/3) and red (10R4/6), with crystals of light gray (5YR7/1): weakly calcareous	Agglomeroplsasmic skelmosepic to intertextic skelsepic, with quartz, microcline and oxidized biotite, some calcite veins	Diffuse irregular
−174	Co	Sandy claystone	Dark greenish gray (5GY4/1) with relict crystals of dusky red (5R3/3)	Common large corestones of amphibolite, dark gray (N4), with crystals of black (N2), dusky red (5R3/3), and dark greenish gray (5G4/1): weakly calcareous	Granular silasepic, with hornblende, quartz, microcline and plagioclase, the latter replaced by micritic concretions	Diffuse irregular
−270	R	Amphibolite	Dark bluish gray (5B4/1), with some crystals of black (N2) and red (5B3/1)	This may be a separate intrusion from that forming the upper part of the outcrop and is foliated at 56°S with strike of 96°: pegmatite dike of red (10R4/6) with light gray (5YR7/1) quartz: non-calcareous	Granular silasepic, with chloritized hornblende, as well as quartz, microcline, biotite, and plagioclase.	

* Microfabric terminology is from Brewer (1976), soil horizon terminology from Soil Survey Staff (1951, 1962), and color from Munsell Color (1975).

The molecular weathering ratios (Figs. 3, 4) demonstrate substantial chemical weathering of the Staca clay to a depth of 1 m, but the Sheigra clay is comparably weathered to 2 m, as would be expected considering its coarser grain size. Clay formation (Al_2O_3/SiO_2), base loss (Al_2O_3/$CaO+MgO+Na_2O+K_2O$) and leaching (Ba/Sr) were moderate to high in both profiles. Calcification is mild in the Staca clay but not detectable in the Sheigra clay, considering that the ratio $(CaO+MgO)/Al_2O_3$ reaches 10 or more in calcareous soils. Both ratios Na_2O and FeO/Fe_2O_3 have the form of a simple leaching function for the numerator, and so indicate no evident gleization or salinization of the profiles.

Our second approach is a mass-balance interpretation. Such approaches to understanding soil formation have been attempted since the last century (Thaer 1857; Brewer 1976). Here we have used the formulation of Brimhall et al. (1991). The procedure uses a particular constituent as a stable reference in order to calculate relative gains and losses of other constituents (also called the "transported mass fraction" or τ) as well as the change in volume of the material caused by soil formation and burial (also called "strain" or ϵ). It is assumed in this analysis that the parent material was uniform and represented by fresh rock lower in the profile, and that a particular constituent has been stable during weathering. These are reasonable assumptions for these Scottish paleosols. We chose TiO_2 as the stable constituent, because unaltered grains of ilmenite are common throughout the profiles (Fig. 6B), although a few partly altered grains were also seen (Fig. 6A, C). Hydrolysis and calcification evident from molecular weathering ratios reflect a geochemical environment favorable for persistence of ilmenite during soil formation. Relict crystal structure high in the profiles is evidence for a parent material like that below the paleosols. Volume change calculated by this means includes a component due to soil formation as well as one due to compaction after burial of the paleosols. Considering abundant framework quartz and little deformed pegmatite dikes, compaction of these Scottish paleosols could have been no more than 20%.

The enrichment of titania in the surface of each paleosol is evidence of overall contraction of the profiles, which can be calculated to be well in excess of the greatest likely compaction (20%) during burial (Fig. 9). Not surprisingly, then, most major elements were lost from the profiles, with the exception of Fe^{3+} in hematite and K in illite, and for the Staca clay only, some Ca+Mg in carbonate. There is variation due to corestone weathering, but the surface horizons of both paleosols have behaved similarly for most elements.

A third approach was to compare trace elements from the surface and parent material of both profiles normalized to nonvolatile composition of carbonaceous chondrites given by Anders and Ebihara (1982). Most trace elements, including rare-earth elements (La-Lu of Fig. 10), were enriched

TABLE 3.—*Textures (volume percent) from point counting of petrographic thin sections**

	Hz	cm	No.	Clay	Silt	Sand	Gravel	Texture
Sandstone	–	+5	R317	21.0	1.8	75.4	1.8	scl
Sheigra clay	At	5	R318	61.8	11.4	26.8	0	c
	At	20	R320	53.0	12.8	34.2	0	c
	Cr	50	R321	44.2	8.8	47.0	0	sc
	Cr	100	R323	52.6	10.2	37.2	0	c
	Cr	130	R324	43.4	11.4	45.2	0	sc
Corestone	Cr	190	R327	15.4	11.0	73.6	0	sl
	Cr	260	R328	50.8	15.8	33.4	0	c
	R	420	R329	14.2	2.6	83.2	0	sl
	R	630	R330	9.4	1.8	88.8	0	ls
Conglomerate	–	+5	R331	26.8	4.4	33.4	35.4	sc
Staca clay	At	5	R332	55.2	26.8	18.0	0	c
	Cr	40	R334	54.0	20.8	25.2	0	c
	Cr	100	R335	39.0	11.8	49.2	0	sc
	R	260	R337	16.8	8.2	75.0	0	sl
	R	320	R338	20.4	15.2	64.4	0	scl
Vein	R	430	R340	2.8	3.2	94.0	0	s

* Standard error (± 1 σ) of these 500-point counts is about 2 volume % (Van der Plas and Tobi 1965; Murphy 1983). Counts were made with a Swift automatic point counter by G. J. Retallack and G. S. Smith (R317 only). Abbreviations for textural classes are sand or sandy (s), clay or clayey (c), and loam or loamy (l).

TABLE 4.—*Mineral composition (volume percent) by pointing counting of Sheigra and Staca paleosols**

Paleosol	Horizon	Depth cm	Spem. No.	Dull Clay	Clear Clay	Total Clay	Chlorite	Carbonate	Feldspar	Amphibole	Biotite	Iron Stain	Opaque	Other	Quartz
Sandstone	–	+5	R317	21.2	0	21.2	0	0	9.6	0	0.4	0.2	1.0	1.0	66.6
Sheigra clay	At	5	R318	21.0	40.4	61.4	0	0	13.0	0	13.6	3.4	0.8	0.2	7.6
	At	20	R320	34.8	20.4	55.2	0.2	2.2	16.2	0	10.6	8.4	0.4	0.4	6.4
	Cr	50	R321	20.0	23.2	43.2	0.2	1.6	20.4	0	16.2	5.0	0	0.6	12.8
Pegmatite	Cr	50	R322	46.2	0.6	46.8	0	0	14.0	0	13.0	1.8	0.4	0	24.0
	Cr	100	R323	33.6	18.0	51.6	0.2	0.4	13.2	0	16.2	3.8	0.2	0.2	14.2
	Cr	130	R324	32.8	12.4	45.2	1.4	0	24.6	0	2.6	10.8	0.4	0.2	15.8
Corestone	Cr	190	R327	10.4	4.0	14.4	2.0	0.8	41.0	0	12.0	2.2	0.2	0.6	26.8
	Cr	260	R328	39.4	12.8	52.2	1.4	0	19.8	0	14.8	1.8	0	0.2	9.8
	Cr	420	R329	13.0	2.2	15.2	1.2	0.2	49.4	0	4.2	2.2	0.2	0.4	27.0
	R	630	R330	7.6	1.4	9.0	7.4	1.8	53.2	0	11.4	2.8	0.8	0.8	12.8
Conglomerate	–	+5	R331	25.2	0	25.2	0	0.2	6.8	0	0.6	0.4	2.2	0	64.6
Staca clay	At	5	R332	38.2	18.6	56.8	0.2	0	4.4	0	10.4	10.4	4.4	18.6	13.0
	Cr	40	R334	30.2	26.0	56.2	1.2	2.2	4.6	0	2.6	8.6	6.2	0.6	17.8
	Cr	100	R335	33.2	5.6	38.8	7.8	4.2	22.4	0	2.4	3.4	5.0	1.4	14.6
	R	260	R337	13.8	0.6	14.4	12.2	4.2	20.2	26.8	2.4	1.6	7.8	0.4	10.0
	R	320	R338	19.0	0	19.0	33.0	3.4	17.2	0	0	6.0	9.4	0.4	11.6
Vein	R	430	R340	4.0	0	4.0	36.6	1.4	27.4	0	1.6	3.6	7.4	0.6	17.4

* Error for 500 point counts as for Table 3. Counts by A. Mindszenty and G. J. Retallack (R322, 331 only). Dull clay includes turbid illite as well as yellow smectite, and clear clay includes low-birefringence illite presumably formed during late diagenesis. Feldspar mainly microcline, with minor plagioclase.

in paleosol surfaces compared with parent material, and there was little fractionation of light versus heavy rare-earth elements. In this respect, the paleosols are more like young soils of dry climates (Cullers et al. 1987; Price et al. 1991), where dust and carbonates accumulate, than like old, deeply weathered soils of humid climates (Nesbitt 1979; Duddy 1980; Braun et al. 1990). The paleosols showed no anomalous behavior of Ce, found in some lateritic soils (Banfield and Eggleton 1989; Braun et al. 1990). Both total concentrations and degree of enrichment of rare-earth elements were higher in Staca than in Sheigra paleosols, as in soils on basaltic rocks (Price et al. 1991) compared with those on granitic rocks (Nesbitt 1979).

A fourth approach is to plot weight percent of Al_2O_3, K_2O, and $CaO+Na_2O$ of noncarbonate phases on a ternary diagram to estimate the stoichiometric excess of potash, and thus relative effectiveness of illiti-

TABLE 5.—*Microprobe analyses (weight percent) of selected mineral phases in Sheigra and Staca paleosols**

Mineral	Spem No.	SiO_2	TiO_2	Al_2O_3	MnO	MgO	CaO	FeO	Na_2O	K_2O	H_2O	Total
Illite	R334	49.05	0.12	20.04	0	5.96	0.35	9.10	0.10	7.44	11.65	103.81
Smectite	R340	29.00	0.12	16.54	1.54	16.36	0.08	23.10	0	0.26	11.37	98.36
	R337	28.91	0.11	15.29	0.18	17.46	0.07	25.22	0.01	0.04	11.34	98.64
	R340	28.59	0.04	17.10	1.46	15.57	0.14	24.71	0	0.15	11.38	99.17
	R340	28.45	0.06	16.62	1.63	16.30	0.12	24.93	0.05	0.14	11.41	99.74
	R340	28.36	0.02	16.24	2.59	15.40	0.02	25.70	0.02	0.09	11.33	99.79
	R340	27.16	0.09	16.88	2.16	14.82	0.05	26.15	0.004	0.02	11.15	98.78
Chlorite	R337	37.51	0.10	14.07	0.55	12.20	0.13	17.16	0.05	3.15	10.71	99.72
	R335	36.35	0	18.73	0	10.54	0.30	19.47	0.28	2.88	10.51	99.07
	R335	34.82	0.0002	15.40	0	17.89	0.21	18.09	0.05	1.37	10.48	98.30
	R335	34.79	0	15.08	0.31	18.37	0.20	18.79	0.06	1.24	12.05	100.90
	R335	32.56	0.05	18.74	0	12.19	0.21	23.20	0.03	1.79	10.29	99.05
	R337	32.35	0.43	14.07	0.55	21.58	0.14	18.38	0.06	0.09	11.86	99.51
	R340	32.34	0.04	13.99	0.12	21.53	0.20	19.91	0.06	0.03	11.87	100.11
	R340	32.13	0	13.89	0	23.64	0.20	16.80	0.07	0.02	11.86	98.71
	R335	31.41	0.05	13.47	0	20.68	0.12	22.20	0.04	0.01	10.21	98.23
Hornblende	R337	41.07	1.35	11.84	0.29	8.25	11.32	19.74	1.45	1.63	1.94	98.87
	R337	41.07	1.49	12.12	0.42	8.21	11.23	20.48	1.74	1.66	1.96	100.07
	R337	40.89	1.55	11.89	0.43	8.05	11.32	20.20	1.39	1.61	1.94	99.26
	R337	40.70	1.28	11.88	0.30	8.17	11.52	21.08	1.44	1.65	1.94	99.95
	R337	40.47	1.28	12.02	0.34	8.38	11.34	20.32	1.56	1.65	1.94	99.36
Plagioclase	R337	61.20	0	25.09	0	0	6.25	0.11	8.09	0.11	0	100.94
	R337	60.78	0	24.64	0	0	6.06	0.13	8.24	0.20	0	100.05
	R337	60.74	0	25.09	0	0	6.51	0.12	7.96	0.12	0	100.54
	R337	60.46	0	25.16	0	0	6.63	0.07	7.99	0.12	0	100.43
K feldspar	R330	65.00	0	18.52	0	0	0.02	0.44	0.46	16.02	0	100.46
	R330	63.54	0	18.89	0	0	0	0.06	0.76	15.58	0	98.84
	R330	62.59	0	19.48	0	0	0.60	1.41	1.94	12.99	0	99.00
Albite	R330	69.14	0	19.36	0	0	0	0.76	12.16	0	0	101.43
	R340	68.63	0	20.04	0	0	0.003	0.41	11.90	0.08	0	101.06
	R330	68.61	0	19.71	0	0	0.09	0.54	11.83	0.03	0	100.80
	R340	67.73	0	19.67	0	0	0.16	0.18	10.72	1.66	0	100.12
Biotite	R337	35.93	2.53	15.32	0	11.47	0.05	19.81	0.12	7.87	10.62	103.72
	R337	35.33	2.49	15.20	0	11.27	0.08	20.17	0.07	7.63	10.49	102.73
Magnetite	R337	0.06	0	0.38	0.14	0.04	0	67.11	30.32	0	0	98.14
	R337	0.03	0.08	0.31	0.10	0	0	68.64	31.01	0	0	100.09
% Error	All	4.29	19.14	3.79	6.74	3.29	14.60	9.49	14.35	11.18	–	–

* Percent error is the average percentage error from the counting statistics of all the analyses listed.

zation during burial (Nesbitt 1992). Strongly illitized paleosols like those studied here deviate directly toward illite, rather than toward alumina-rich clays formed during weathering. It is perhaps unfair to plot minerals rich in Mg and Fe on this diagram, but it does reveal the compositional path of paleosol development compared with parent material.

BURIAL ALTERATION OF THE PALEOSOLS

A variety of alterations of mineral grains have been observed in these Scottish paleosols, due to metamorphism of the parent material, Precambrian soil formation, alteration after burial, and soil formation in outcrop. In this section we attempt to identify and assess the overprinting alteration due to burial of these paleosols.

Very little organic matter has been detected in these Precambrian paleosols (Table 10), just as in formerly well drained Quaternary paleosols, which are known to have had an original content of organic matter as much as an order of magnitude higher (Stevenson 1969; Retallack 1991a).

The paleosols have mottles of purple to scarlet red (5YR to 5R of Munsell Color 1975) from hematite of relatively coarse grain size (Morris et al. 1985; Torrent and Schwertmann 1987), as seen from the sharpness of the hematite peaks on X-ray diffractometer traces. Given the likely climate and duration of formation of these paleosols this color could equally have been attained by dehydration of brown to yellow iron hydroxides during soil formation or burial (Retallack 1991a). The original oxidation of these minerals, however, is almost certainly a Precambrian phenomenon, considering the petrographic (oxidation along veins and within biotite and hornblende grains), geochemical (low ferrous/ferric and high barium/strontium ratios), and physical evidence (clay skins, corestones) for good drainage.

TABLE 6.—*Microprobe analyses (weight percent) of carbonates from the Staca paleosol*

No.	CO_2	$MgCO_3$	$CaCO_3$	$FeCO_3$	Total
		High-magnesium calcite			
R335	1.35	4.67	91.32	2.85	100.18
		Low-magnesium calcite			
R335	1.35	0.72	96.73	0.89	99.70
R335	1.33	1.67	97.11	0.89	101.00
R340	1.36	1.41	95.39	0.96	99.12
R340	1.38	0.67	95.62	0.47	98.13
R340	1.32	2.39	97.95	0.13	101.58
% Error	–	25.9	1.3	5.7	–

* Error is calculated as the mean of the percent standard deviation from counting statistics for all these analyses.

The paleosols also have been compacted by overlying rocks. Above the paleosols are 450 m of Torridonian sediments below Lower Cambrian quartzites (Williams 1969), which in turn support about 1500 m of Cambrian and Ordovician limestones and dolostones (Walton 1983). To the south, however, the Torridonian sequence reaches a thickness of 7 km (Johnson 1983), and the southern facies are truncated in such a way that some of these Precambrian alluvial fans were eroded before deposition of the early Paleozoic platform sequence. In addition, the Moine Thrust, emplaced some 400–450 Ma ago (Caledonian), is only 20 km to the east

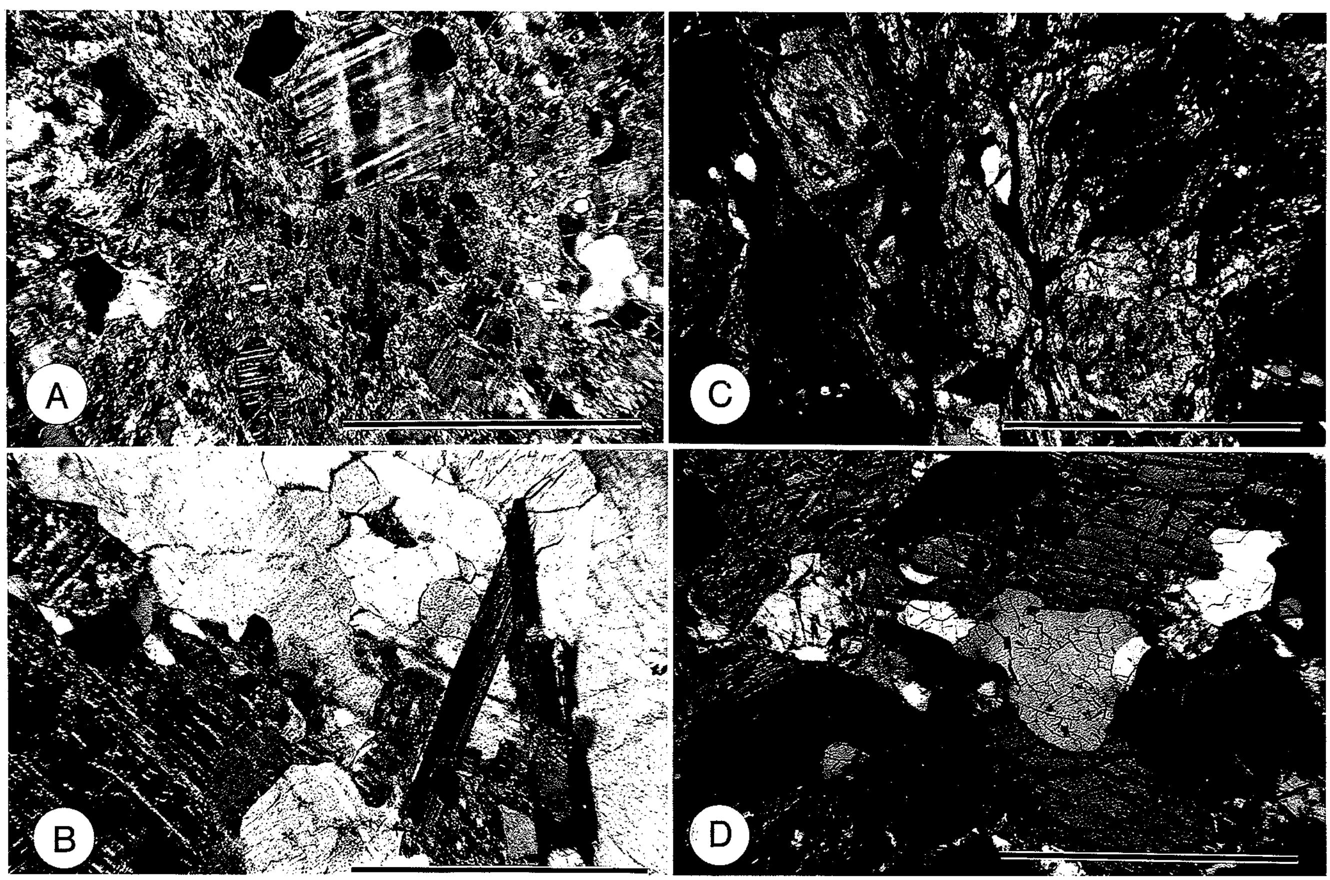

FIG. 5.—Photomicrographic appearance of weathering in the two profiles: **A)** clayey surface (R318) of the Sheigra clay paleosol and **B)** fresh gneiss (specimen R330); **C)** clayey surface (R323) of the Staca clay paleosol and **D)** fresh amphibolite (R337). All scale bars are 1 mm.

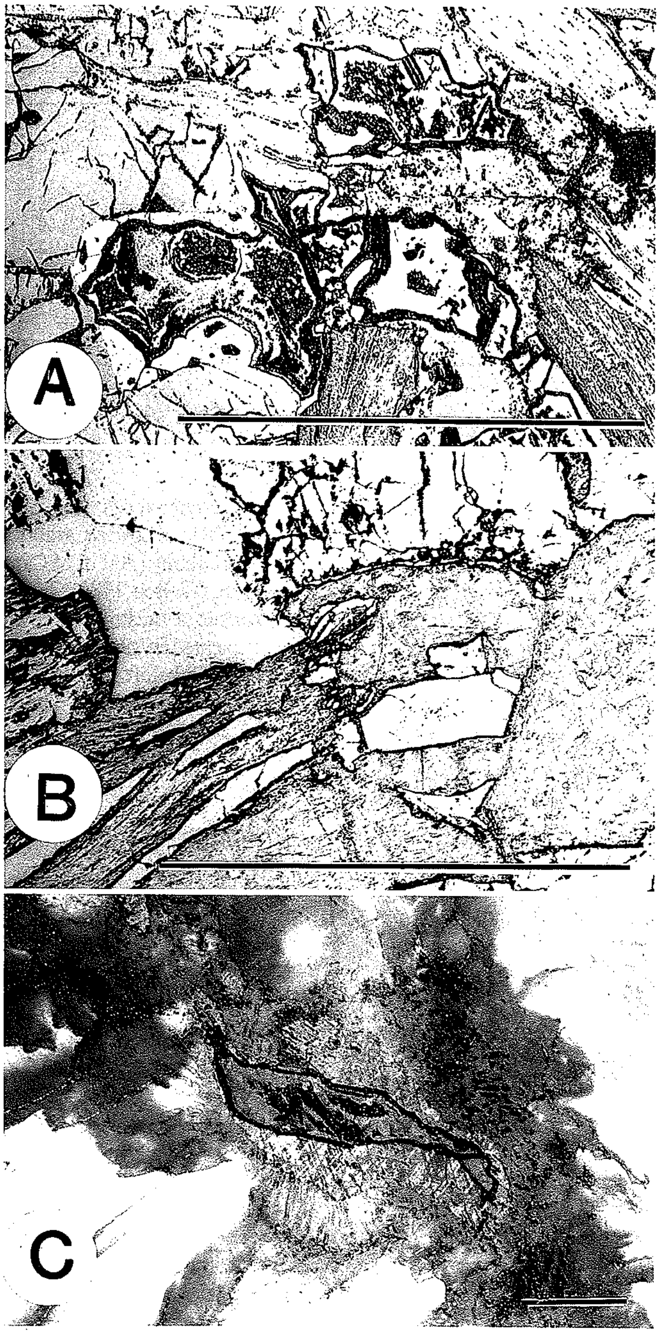

FIG. 6.—Opaque grains in reflected light: **A**) internally disrupted grains with collomorphous iron oxide, **B**) euhedral grain with growth lamellae, and **C**) grain with alteration rind of microperthite, all from parent material of type Sheigra clay paleosol (R330). Scale bars 1 mm for **A** and **B**, but 100 μm for **C**.

(Johnson 1983) and may have loaded these paleosols with metamorphic rocks overthrust from that direction.

Despite this potential for significant compaction, there are several lines of evidence that compaction was slight. First, there is no evidence of ptygmatic deformation of pegmatite dikes, even within the clayey upper part of the profile, where there is a well spaced, slickensided system of conjugate shears at a low angle to the ancient land surface. Ptygmatic deformation of dikes becomes obvious with compaction beyond about 0.8 times former thickness (Retallack 1986a).

FIG. 7.—Back-scattered electron micrograph of hornblende (A) weathered to carbonate (Ca) and three kinds of chlorite (Cl), light (Fe-rich), medium, and dark (Mg-rich), from the type Staca clay (R337). Scale bar 50 μm.

Second, the bulk density of the clayey portions of the paleosol (about 2.5 g/cm^3) has remained light compared with gneiss (about 2.7 g/cm^3) and amphibolite (about 2.8 g/cm^3). This is not as light as usual for surface soils (less than 2 g/cm^3; Retallack 1990, fig. 4.1), but within the range for saprolites on weathered bedrock. Compaction to 80% of the former thickness of the surface horizons from a more realistic density of 2 g/cm^3 for such a clayey soil with relict crystalline texture would result in the present density.

Third, most illite in the paleosol is relatively poorly crystallized. Specimens R318 and R324 in the Sheigra clay had the following indices of crystallinity, respectively, as measured from X-ray diffractograms of bulk samples: Weaver index of 2.45 and 2.01, Kubler index of 1.15 and 0.95° 2θ, and Weber index of 348 and 775. All these are values common under a burial regime of mild diagenesis, with temperatures below 120°C, overburden no more than 1500 m, and coalification less than bituminous grade (Frey 1987).

Fourth, there were no observed indicator minerals of low-grade metamorphism such as laumontite, prehnite, chloritoid, or stilpnomelane. Although this argument is based on negative evidence, these minerals were sought and could not have escaped detection by the varied analytical methods used.

Fifth, the overlying conglomerate and sandstone does not show crushed or deformed quartz grains. These grains show numerous tangential and long contacts, and some concavo-convex contacts, but few sutured contacts. These observations are compatible with burial to depths of about 1.5–2 km (Taylor 1950). At 2 km burial depth, a marine shale would be compacted to about 58% of its former thickness and a marine sandstone to about 70% (Baldwin and Butler 1985). Even this latter estimate is probably greater than would be expected in a coarse-grained paleosol on bedrock. From these lines of evidence, the pre-Torridonian paleosols were neither located below a Torridonian depocenter, nor completely onlapped by Paleozoic platform sediments, nor overthrust by the Caledonide crystalline rock. Instead, they were remarkably well preserved.

Another likely burial alteration of these paleosols is illitization of smectite (Retallack 1991a), which should have been significant at the depths of burial already outlined. Toward the top of both profiles there is marked enrichment in potash (Fig. 11) and a dramatic increase in the clear generation of relatively pure illite (Table 4; Fig. 8B). This clear illite postdates the yellow smectitic clay skins and the greenish-gray mixed-layer illite-smectite of the matrix. The overlying conglomerate also shows an authigenic growth of radially oriented illite around grains, as well as an even later patchy replacement of illitic matrix with green chlorite. These overlying sediments and their pore waters may have been an important source of potassium for illitization.

ALTERATION OF THE PALEOSOLS IN MODERN OUTCROP

Exposures in the sea cliff and rock platform near Sheigra are clear of soil (Fig. 2). The upper horizons of Sheigra and Staca paleosols have weathered back beneath an overhang of massive conglomerate of the basal Applecross Formation about 5 m above sea level. Thus a level some 50–150 cm below the tops of the profiles is the most likely to have been altered in outcrop, and a thin (10 cm) grass-covered soil (Orthent) has developed on colluvial debris at this level in some parts of the outcrop. This soil is probably much younger than 6000 years, when postglacial (Flandrian) transgression was completed and the current cycle of littoral erosion of the sea cliff was initiated. This is insufficient time and development to explain the clayeyness and rich red hues of the paleosol according to general models of soil development (Birkeland 1984). The brown hues of the loamy surface soils (10YR) and of ferruginized joints (7.5YR) beneath them are not found more than a few centimeters below the contact with bedrock, even within the 50–150 cm interval most likely to have been affected by weathering of the outcrop. All samples were taken more than 20 cm below the present surface.

Soils on more ancient postglacial (less than 13 Ka) geomorphic surfaces in the area around Sheigra are largely peaty (Histosols) or clay-poor, sandy, and ferruginized (Spodosols). These are both regimes of soil formation distinct from the organic-lean and clay-rich regime (lessivage) evident from the Precambrian paleosols. Clays in the surface soils were found by Williams (1968) to be mainly vermiculite and illite, the latter with basal spacing of 9.9–10.0 Å. Like Williams, we found no vermiculite in the paleosols, and that illite in them had basal spacings of 10.0–10.4 Å. Williams found no smectite in the modern soils, and we agree with him that this was produced by Precambrian rather than modern weathering, for the following reasons. First, clay skins of yellow smectite are cut by clear illite produced during burial diagenesis. Second, the larger-than-usual basal spacing of illite in the paleosols appears to be due to a small amount of interlayering with smectite.

In summary, the paleosols have been little affected by modern soil formation beyond local loosening of grains and orange-brown (7.5YR) ferruginization of joints, both of which were avoided during sampling.

PRECAMBRIAN PEDOGENIC FEATURES OF THE PALEOSOLS

Although some features of these paleosols are due to alteration during burial and exposure, there remains a residue of features produced by Precambrian weathering. Compared with other Precambrian paleosols these pre-Torridonian examples are remarkably well preserved and can serve as examples of the kinds of pedogenic features that can be observed in such ancient soils.

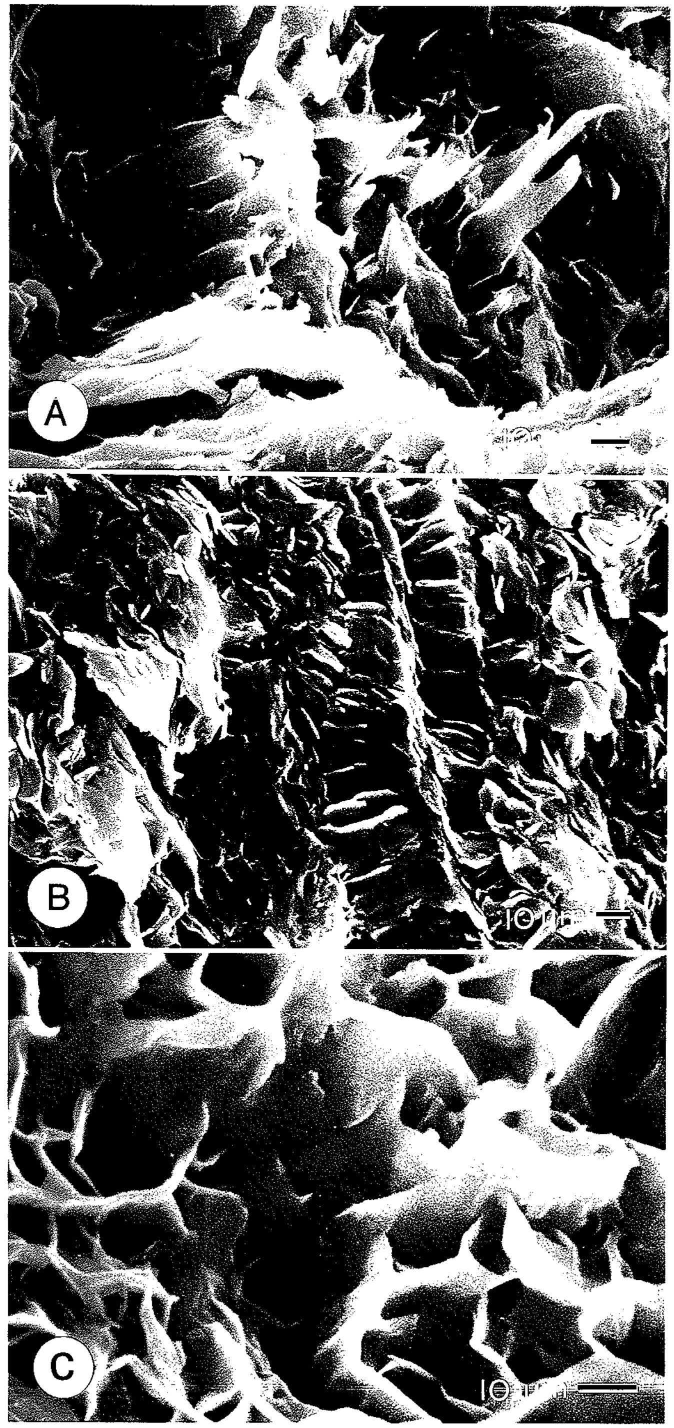

FIG. 8.—Three distinct kinds of clay in the Scottish paleosols as seen under scanning electron microscope: **A)** stacks and wisps of illite (R318), **B)** long books of illite (R337), **C)** honeycomb coat of smectite (R323).

Soil Horizons

The paleosols show the overall appearance of soil horizons, with an abrupt change to overlying Torridonian sediments contrasting with gradual transitions from a clayey surface through a thick horizon of corestones to unweathered gneiss and amphibolite (Fig. 2). This step-and-fade pattern can be seen also in chemical and petrographic depth functions of the profiles, if allowance is made for irregularities in the horizon of corestones (Figs. 3, 4).

In the horizon shorthand of soil science (Soil Survey Staff 1951, 1962; Birkeland 1984), the unweathered parent rock is an R horizon and the thick zone of relict crystal structure and corestones can be regarded as a Co horizon. Above that is a horizon of clay skins and clay enrichment, which we regard as the top of the soil, mainly because of the features interpreted below as soil creep (Fig. 12). No other soil-like materials have been found in the overlying conglomerates, or for that matter above other Precambrian clayey paleosols (Retallack 1990). Because we interpret the

TABLE 7.—*Major-element analyses (weight percent) by AA, loss on ignition and bulk density (g/cm³) of Sheigra and Staca paleosols**

Paleosol	Hz	−cm	No.	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	Total	g/cc
Type Sheigra clay	At	5	R318	57.06	0.47	19.27	5.81	0.89	0.04	2.54	0.45	2.20	6.14	0.36	4.94	100.29	2.60
	At	20	R320	64.88	0.08	17.67	2.54	0.28	0.03	1.86	0.76	3.01	4.44	0.11	4.77	100.38	2.67
	Cr	50	R321	57.07	0.49	18.39	6.49	0.98	0.05	2.53	0.84	2.44	6.12	0.30	4.43	100.27	2.66
(Pegmatite)	Cr	50	R322	63.01	0.58	16.53	4.74	1.19	0.06	2.23	1.05	3.25	4.19	0.25	3.27	100.37	2.60
	Cr	100	R323	61.58	0.44	17.89	3.83	1.03	0.05	2.49	0.58	2.26	5.56	0.28	4.41	100.41	2.57
	Cr	130	R324	57.89	0.38	19.11	4.67	0.98	0.06	2.56	1.40	3.62	4.69	0.36	4.43	100.25	2.55
(Corestone)	Cr	190	R327	67.52	0.37	15.70	2.03	1.27	0.06	1.52	2.70	4.28	2.49	0.22	2.21	100.06	2.66
	R	630	R330	61.38	0.28	17.49	2.27	3.06	0.13	3.37	3.06	5.05	1.38	0.25	2.63	100.30	2.69
Type Staca clay	At	5	R332	51.97	0.74	14.46	12.32	0.82	0.17	2.80	1.05	0.16	5.40	0.81	6.14	97.26	2.70
	Cr	40	R334	51.58	0.54	14.23	12.81	1.46	0.08	3.07	1.21	0.16	5.47	0.58	6.53	97.92	2.69
	Cr	100	R335	50.48	0.63	12.93	8.91	5.72	0.18	5.37	3.44	1.47	2.37	0.60	6.69	97.84	2.72
	R	260	R337	48.62	0.96	12.45	9.79	6.49	0.24	5.62	4.52	2.05	1.63	0.46	5.15	97.27	2.87
(Intercore vein)	R	430	R340	50.35	0.45	13.47	8.04	6.74	0.25	6.62	2.17	2.35	1.42	0.61	5.11	98.26	2.81
Error (σ)	—		All	0.25	0.02	0.17	0.15	0.08	0.05	0.10	0.05	0.01	.005	.004	—	—	0.05

* Analyses were from atomic absorption at the University of Oregon, Eugene, by Christine McBirney. Bulk density was calculated by weighing paraffin-coated clods in and out of water at the University of Oregon, Eugene, by Gregory J. Retallack. Errors were estimated from 10 replicate analyses of standard rock W2 for atomic absorption and from 10 replicates of rock R602 for bulk density.

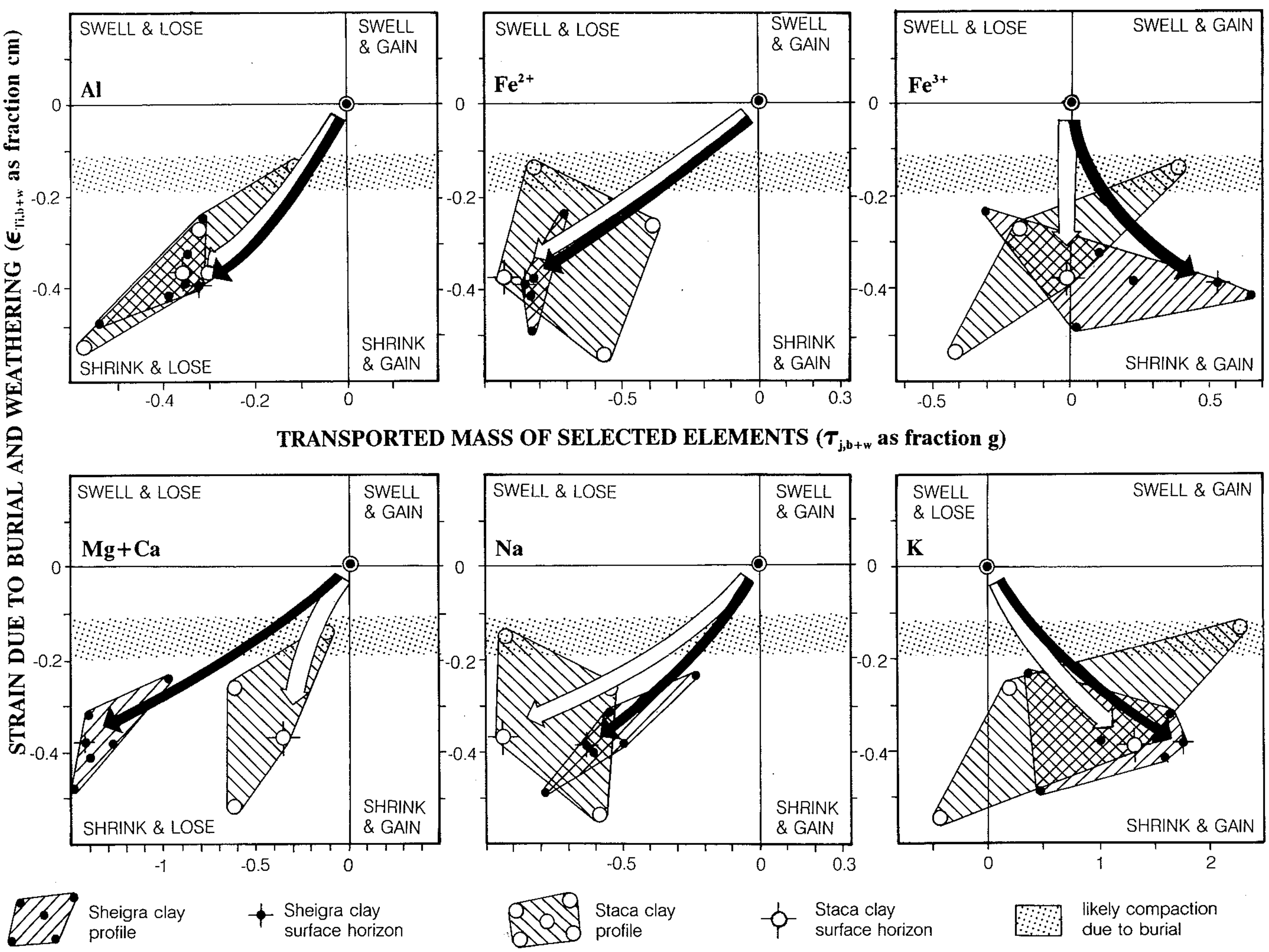

FIG. 9.—Titania-normalized changes in volume ϵ versus elemental abundance τ for Sheigra paleosol (closed symbols) and Staca paleosol (open symbols). Arrows indicate weathering trend from parent material plotted at the origin. Both paleosols are consistent in behavior of elements other than Fe^{3+} and Ca+Mg, and both show chemical weathering in excess of that explainable by compaction after burial (at intersection of trend and stippled band).

TABLE 8.—*Trace-element analyses (ppm) by AA of Sheigra and Staca paleosols**

Paleosol	Hz	−cm	No.	Ba	Co	Cr	Cu	Li	Ni	Rb	Sr	Zn
Type Sheigra clay	At	5	R318	363	33	16	8	9	21	193	43	84
	At	20	R320	262	35	7	4	6	11	124	101	42
	Cr	50	R321	991	37	14	7	11	17	211	144	91
(Pegmatite)	Cr	50	R322	471	55	9	5	14	13	170	268	93
	Cr	100	R323	440	40	11	5	11	13	185	102	79
	Cr	130	R324	308	35	14	57	13	13	179	221	81
(Corestone)	Cr	190	R327	626	58	11	6	12	9	90	647	61
	R	630	R330	269	42	12	10	37	13	45	614	83
Type Staca clay	At	5	R332	696	48	30	22	21	33	184	37	117
	Cr	40	R334	510	50	17	14	25	27	196	38	119
	Cr	100	R335	176	57	26	8	59	29	91	146	123
	R	260	R337	231	74	25	66	43	41	62	171	166
(Intercore vein)	R	430	R340	440	54	20	49	67	25	41	213	204
Error (σ)	–	–	All	12	1	2	2	1	1	2	4	3

* These analyses also by Christine McBirney, using atomic absorption, with errors from 10 replicates of standard rock W2.

clayey horizon as the top of the soil, it is designated At, rather than Bt characteristic of forested soils of Devonian and younger age.

Corestones

These are areas of relatively little-weathered rock within saprolitic lower parts of deep soils. As pointed out by Williams (1968, 1969), both corestones and broad exfoliation joints in these paleosols are locally cut by the pre-Torridonian unconformity, and this is evidence that they were original parts of the paleosol. Corestones are an obvious field feature of Precambrian paleosols, especially those developed on bedrock as opposed to sedimentary parent materials (Retallack 1986a).

Soil Creep

Following William's (1968, 1969) original suggestion of soil creep in these paleosols, one of us (GJR) confirmed that weather-resistant pegmatite dikes are indeed bent into the unconformity, which is locally littered with angular slabs of pegmatite distinct from the rounded quartz pebbles of the overlying conglomerate (Fig. 12). At the northern end of the outcrop near Sheigra, in contrast, where the unconformity has eroded down into the zone of corestones, pegmatite veins are abruptly truncated and the basal Torridonian conglomerates contain boulders of gneiss. To the south, surface litter of tottering and fallen slabs is preserved. Thus the land surface sloped to the north and the profile is relatively complete.

Clay Skins

Irregularly oriented slickensided planes of red clay are prominent in the upper part of the Sheigra clay but less common in the Staca clay. In thin section, these are internally laminated and cut other clay skins in a complex fashion (Fig 5C). In soils, such features are called illuviation argillans (Brewer 1976), and form by the washing down of clay into cracks. Their contorted irregular form is distinct from boxworks of hydrothermal or igneous veining. As evidence of low-temperature opening and filling of cracks, they are diagnostic features of paleosols.

Clay Microfabric

In thin sections viewed under crossed nicols, much of the clayey matrix of the paleosols presents a random arrangement of highly birefringent clay streaks within a matrix of darker clay (Fig. 5A, C). This kind of microfabric is called sepic plasmic fabric, in these profiles skelinsepic and skelmosepic (of Brewer 1976), and is diagnostic of soils and paleosols (Retallack and Wright 1991). It forms by shrinking and swelling, cracking and filling, in well drained and moderately to well developed soils (Brewer and Sleeman 1969). It requires a highly deviatoric system of small-scale stresses that are difficult to imagine outside of soil environments.

Destruction of Silicate Grains

In the tops of both paleosols almost all the minerals have been altered to clay, except for quartz and microcline, and even these are shattered or

TABLE 9.—*Chemical analyses by INAA of Sheigra and Staca paleosols**

	Sheigra Clay		Staca Clay		Standard Deviation (%)
	At R318	R R330	At R332	R R337	
	weight percent				
FeO	6.22 ± 0.02	5.06 ± 0.02	14.2 ± 0.03	15.3 ± 0.05	5
Na_2O	2.34 ± 0.02	5.02 ± 0.02	0.19 ± 0.01	2.18 ± 0.02	3
K_2O		< 7.5	< 14	< 14	
	parts per million				
Sc	11.2 ± 0.02	11.7 ± 0.02	35.2 ± 0.03	31.7 ± 0.03	3
Cr	14.2 ± 1.1	10.1 ± 1.0	33.3 ± 1.8	24.3 ± 1.7	10
Co	24.7 ± 0.15	41.2 ± 0.21	48.1 ± 0.19	80.4 ± 0.32	5
Ni	< 160	< 150	< 240	< 270	12
Zn	89 ± 3	89 ± 3	163 ± 4	126 ± 5	15
As	3.6 ± 0.7	< 6.3	12.0 ± 1.8	< 20	5
Sb	0.84 ± 0.07	0.20 ± 0.05	0.64 ± 0.10	< 0.99	5
Se	< 4.2	2.4 ± 0.4	< 6.0	4.1 ± 0.7	5
Rb	183 ± 5	48 ± 5	166 ± 6	58 ± 7	10
Cs	3.82 ± 0.11	1.15 ± 0.10	4.75 ± 0.16	2.37 ± 0.17	5
Sr	< 420	649 ± 44	< 630	< 690	12
Ba	350 ± 50	302 ± 30	814 ± 69	301 ± 62	10
La	30.8 ± 0.3	24.9 ± 0.2	45.7 ± 0.4	21.9 ± 0.3	3
Ce	57.1 ± 0.6	53.5	96.5	51.9	7
Nd	28.7	23.9 ± 3.5	53.6	26.8	12
Sm	5.89 ± 0.03	4.74 ± 0.02	11.7 ± 0.05	5.82 ± 0.04	5
Eu	1.49 ± 0.03	1.29 ± 0.02	3.39 ± 0.03	1.84 ± 0.03	5
Tb	0.75 ± 0.05	0.43 ± 0.04	1.97 ± 0.05	0.90 ± 0.06	5
Yb	1.52 ± 0.11	1.15 ± 0.08	5.58 ± 0.17	3.24 ± 0.16	5
Lu	0.23 ± 0.01	0.14 ± 0.01	0.84 ± 0.03	0.46 ± 0.03	5
Hf	4.74 ± 0.11	4.18 ± 0.10	6.88 ± 0.15	4.26 ± 0.15	5
Ta	0.63 ± 0.05	1.49 ± 0.05	1.94 ± 0.06	1.93 ± 0.08	5
W	78.3 ± 1.3	200.0 ± 1.2	123.0 ± 2.8	216.0 ± 3.2	12
Hg	< 0.15	< 0.14	< 0.25	< 0.33	5
Th	6.6 ± 0.1	9.7 ± 0.1	5.8 ± 0.1	3.7 ± 0.1	5
U	< 4.8	0.99 ± 0.23	5.90 ± 0.61	2.36 ± 0.46	7

* These analyses from Instrumental Neutron Activation Analysis are by A. G. Johnston and are Oregon State University Radiation Center Project Number 799. The figures beside the elements and oxides are percent standard deviation based on repeated counts of appropriate standards and may be a better indication of uncertainty than standard deviations given in table as ± from counting statistics. Total iron is given as FeO. The symbol < indicates maximum likely value from low peaks, from which statistically meaningful values could not be obtained.

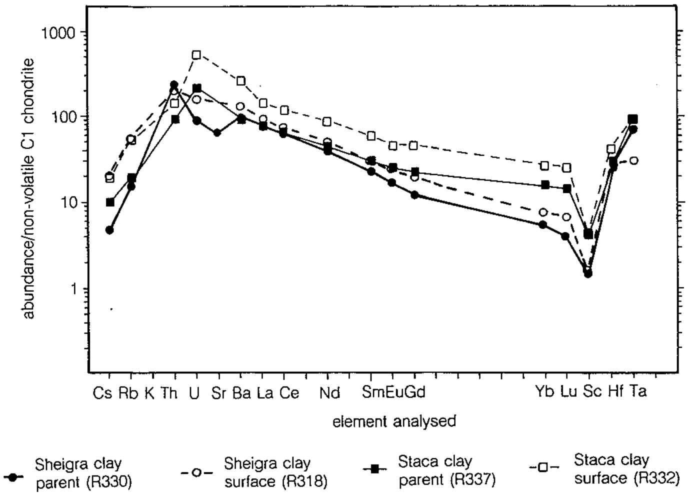

FIG. 10.—Abundance of trace elements, including REE from INAA analyses of Precambrian paleosols normalized to nonvolatile C1 carbonaceous chondrite composition (of Anders and Ebihara 1982).

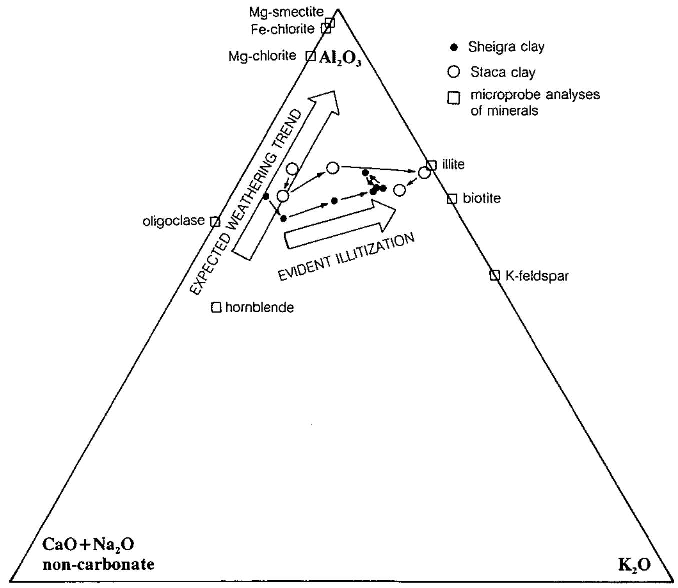

FIG. 11.—A-CN-K ternary diagram for chemical composition of Sheigra and Staca paleosols, showing strong deviation from normal weathering trends by illitization largely during burial.

embayed (Fig. 5A, C). Mineral grains could conceivably be altered this way hydrothermally. In the paleosols, however, the sharply truncated depositional contact on top of the zone of alteration makes that explanation unlikely. Nor were any distinctive hydrothermal minerals noted. Instead the sequence of mineral alteration follows the generally accepted weathering sequence established by Goldich (1938) using Cretaceous paleosols on granite in Minnesota.

Opaque Oxides

Irregular iron stain around opaque grains and trellislike grains formed by progressive oxidation of biotite (Williams 1968) are best explained as products of ancient weathering. Oxidized mineral grains are abundant in the profiles, and occasional oxidized opaque grains were found in little-weathered paleosol parent materials (Fig. 6A–C). These opaque oxides may have been dehydrated from iron hydoxides during burial (Retallack 1991a), but the oxidation was probably within the former soil.

Calcareous Micronodules

In the Staca paleosol, ferruginized carbonate veins cut across clay skins and across soil-like clayey microfabric (Fig. 13A). Such features are found in soils, but are not as characteristic of soils as are small micritic nodules. These engulf preexisting grains and may contain concentric ferruginized bands (Fig. 13B). Very similar replacive calcareous-ferruginous concretions have been recorded in Indian soils (Sehgal and Stoops 1972; Courty and Féderoff 1985). Such patchy, colloidal, episodic replacement is also characteristic of soils.

RECONSTRUCTED PALEOENVIRONMENT

Paleosols are products of a variety of environmental influences, whose role in forming soils at present is now becoming increasingly well understood (Birkeland 1984). Each of these is considered in the following paragraphs, along with an attempt to understand these Precambrian profiles in the context of classifications of soils (Soil Survey Staff 1975).

Atmospheric Composition

The Sheigra and Staca paleosols formed on felsic and mafic parent materials, respectively, side by side on the same ancient landscape some 0.8 Ga ago, like the Pronto and Denison paleosols that have been so useful as indicators of low atmospheric oxygenation 2.3 Ga ago (Holland 1984; Mossman and Farrow 1992). Unlike the pre-Huronian Pronto and Denison profiles, however, both Staca and Sheigra paleosols have retained ferric iron (Fig. 9) and so indicate increased oxygenation of the atmosphere since very early Proterozoic time. On the other hand, they are not as red and oxidized as many Phanerozoic Oxisols, lateritic and bauxitic soils, which commonly show cerium anomalies (Braun et al. 1990), not seen in these Scottish paleosols (Fig. 10).

How oxidizing the atmosphere was some 0.8 Ga can be estimated from the compositional reducing power of the parent rock (R value of Holland 1984): of the gneiss (R330) below the Sheigra paleosol, which is 0.024, and of the amphibolite (R337) below the Staca profile, which is 0.045. To oxidize such a mafic rock as the amphibolite requires at least 4.6×10^{-4} atm of O_2 (0.002 PAL) assuming modern CO_2 levels, or 4.6×10^{-2} atm of O_2 (0.22 PAL) assuming an equally unrealistic high CO_2 level of 3.4×10^{-2} atm (100 PAL). How much richer in oxygen cannot be determined from these paleosols, but indications from elsewhere are on the oxygenated side of this range. Comparable analysis of 1.9 Ga paleosols (reported by Wiggering and Beukes 1990; Holland and Beukes 1991) constrains the redox threshold value of parent-rock compositional reducing power to lie well above 0.049 for basalt, between 11.43 for iron-depleted siderite-facies banded iron formation, and 20.30 for greenalite-siderite banded iron formation. This translates to oxygenation between the limits of $3.8–6.9 \times 10^{-3}$ atm (0.02–0.03 PAL) of O_2 for the case of 1 PAL CO_2 and $3.8–6.9 \times 10^{-1}$ atm (1.8–3.3 PAL) of O_2 for the case of 100 PAL CO_2 at 1.9 Ga.

TABLE 10.—*Total organic carbon and its isotopic composition in and above the Sheigra clay paleosol**

Spem No.	PPRG No.	Depth (cm)	TOC mgC/g	$\delta^{13}C_{org}$‰
R317	1874	+10	0.10	−27.0
R320	1876	−20	0.10	−25.6
R321	1877	−50	0.04	−25.6

* Isotopic analyses are versus PDB, from J. W. Schopf (personal communication, 1992); PPRG is the research group of Schopf and Klein (1992).

FIG. 12.—Soil creep of pegmatite vein in upper part of Sheigra clay paleosol. Depth to zone of corestones from Torridonian sediments is 1.7 m.

The values most favored by Holland and Beukes (1991) from modeling of iron retention in the weathered siderite-facies parent material, which is mineralogically the most simple, are 3–8 × 10^{-2} atm O_2 (0.15–0.40 PAL) for 1.9 Ga. These values are adequate to explain the degree of oxidation observed in Late Ordovician (0.45 Ga) paleosols (Feakes et al. 1989), as well as the 0.8 Ga Scottish paleosols reported here. Atmospheric oxygen was a substantial fraction of its present abundance by Late Precambrian time.

Paleoclimate

The presence of carbonate micronodules in the Staca paleosol but not the Sheigra paleosol is evidence of a subhumid climate, or in soil jargon, a climate near the pedocal-pedalfer boundary or ustic/udic boundary, which varies from 600 to 1000 mm mean annual rainfall (Yaalon 1983). Carbonate nodules reach their greatest abundance 90–280 cm from the surface of the Staca clay. Using an empirical equation relating depth of carbonate accumulation to mean annual rainfall in surface soils (Retallack 1993), this implies mean annual rainfall of 609 ± 141 mm. Rainfall would have been greater if the paleosol had been compacted or eroded from a greater thickness of soil above the carbonate horizon. For compaction by 0.8, the carbonate horizon would have been at a depth of 113 cm and mean annual rainfall some 693 ± 141 mm. The higher, or subhumid, value is more realistic considering the depth of weathering of this clayey and only slightly calcareous paleosol adjacent to a noncalcareous paleosol. Subhumid climate also is compatible with mild enrichment of rare-earth elements in the profiles (Cullers et al. 1987; Price et al. 1991), without marked fractionation of heavy versus light rare earths (Nesbitt 1979; Duddy 1980; Braun et al. 1990).

No large clastic dikes or vertic structures were found as evidence of highly seasonal rainfall, unlike those seen in other Precambrian paleosols (Retallack 1986a). Nevertheless, some carbonate nodules do have concentric ferruginized rims, and the carbonate nodules are small and dispersed through a large part of the Staca clay. These are both features of surface soils on Indogangetic alluvium in India and Pakistan today, and are evidence of seasonality of rainfall (Retallack 1991b).

Paleotemperature is not well constrained by paleosols, other than those obviously affected by permafrost features (Retallack 1990, 1991b). The depth and degree of weathering observed in these paleosols are compatible with a temperate to tropical paleoclimate.

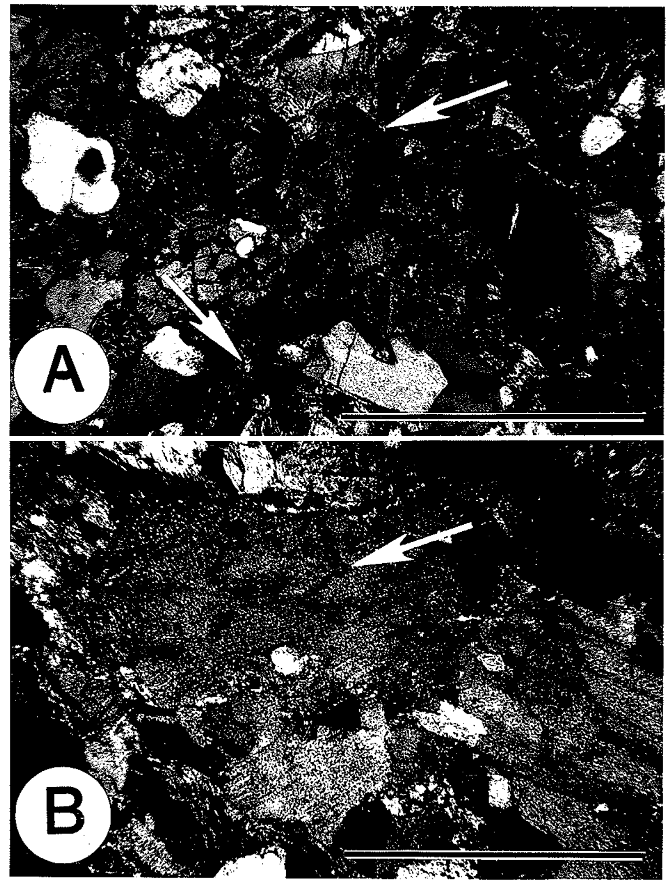

FIG. 13.—Photomicrographs under crossed nicols: **A)** iron-stained calcite vein (R335) and **B)** complexly zoned micritic micronodules (R338), both from the Staca paleosol. The carbonate micronodule replacing plagioclase has a ferruginized growth zone at arrow. Scale bars are both 1 mm.

Biota

Some kind of microbial life in these paleosols is likely. The most compelling evidence for it is the distribution and isotopic composition of organic carbon in the Sheigra clay (Table 10). The carbon is isotopically light, as is characteristic of carbon produced by bacteria, algae, and plants using the Calvin cycle (C_3) photosynthetic pathway (Schopf 1983; Cerling et al. 1991; Retallack 1992b). Although it is not abundant, this is a typical level for organic carbon in formerly well drained Cenozoic paleosols, in which organic carbon is about an order of magnitude less abundant than in comparable surface soils (Stevenson 1969; Retallack 1991a, 1991b). In addition, the organic carbon in the Sheigra clay paleosol declines in abundance downward from the surface of the paleosol. This distribution also makes it unlikely that the carbon is a modern contaminant. The uppermost sample of paleosol and overlying conglomerate were from a deep recess and overhanging cliff, respectively, whereas the sample from 50 cm down into the paleosol was broken out from a sloping rock face that nearby supports grass and soil. This is the first report of both isotopic and depth functions of organic carbon indicative of life in a Precambrian paleosol, although others have mentioned comparable unpublished results (Mossman and Farrow 1992).

Microbes and their mucopolysaccharides and other slimy products would

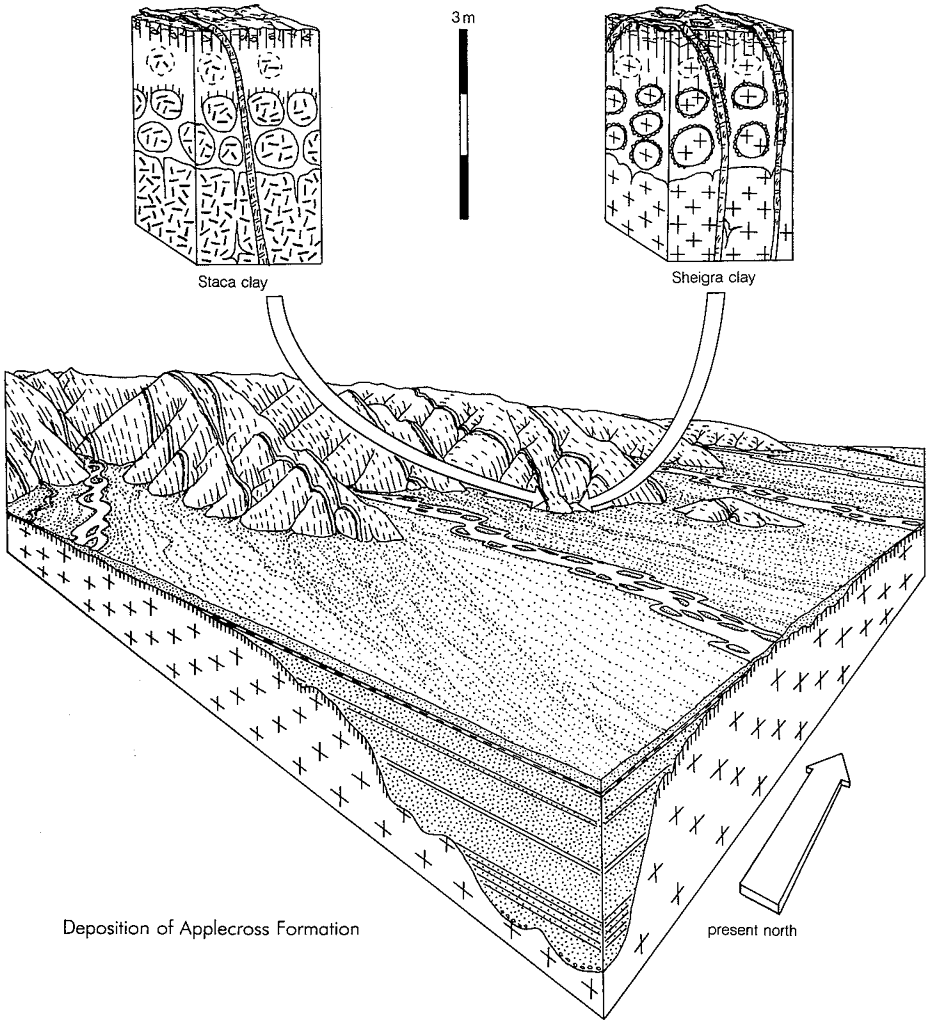

FIG. 14.—A reconstruction of pre-Torridonian landscape and soils in northwestern Scotland a billion years ago.

have encouraged development of clay skins around large clods observed in the upper part of these paleosols as well as the preservation of such thick clayey paleosols in an area of demonstrated paleotopographic relief. Without such organic stabilization of these clayey soils one would expect to find evidence of rill erosion, thin clay skins, and popcornlike clay clods found today in clayey desert badlands (Engelen 1973). Pre-Torridonian paleosols have been examined in many large outcrops (Williams 1968, 1969; Stewart 1972), yet nothing like these features of desert badlands have yet been reported.

Paleotopography

The paleotopographic setting of these paleosols established from regional mapping of the pre-Torridonian unconformity is a low-relief plain north of a region of deep valleys to the south (Fig. 14; Williams 1969; Johnson 1983). In the extensive (200 m long) exposure near Sheigra, paleotopographic relief is 2.3 m, with the paleosols eroded down to corestones in a broad gully to the north. There is no significant paleotopographic difference between Staca and Sheigra paleosols. An elevated terrain for both paleosols is supported by evidence of good drainage in the form of very low ferrous/ferric iron ratios (Figs. 3, 4), corestones of fresh rock, and soil creep of pegmatite dikes (Fig. 12). The abundant pegmatite dikes withstood weathering better than the enclosing gneiss and amphibolite, and the soil was probably crisscrossed by low walls and locally littered with slabs of quartz-rich pegmatite. A well drained landscape is inferred from observations at An Sochach, some 10 km east of these paleosols near Sheigra, where there is a rounded hill of Lewisian gneiss at least 150 m high. It lacks appreciable paleosol development and is mantled by breccia that can be interpreted as former scree around a rocky hill (Williams 1968).

Williams (1969) has interpreted the low-relief landscape of the paleosols as a piedmont with local hills at the foot of a mountain range now lost to the west and a source of alluvial fans covering the paleosols. Like those of the North American Basin and Range, this flat piedmont would have been a narrow zone at any one time. Williams argued that it was created by erosion, then covered by alluvial-fan deposits within a few tens to hundreds of thousands of years after formation. By this theory, the present width of this suballuvial bench was formed by westward retreat of the mountain front over a period of 8–50 million years. An alternative view not favored by Williams is to regard this landscape as an elevated peneplain, or etchplain in the sense of Büdel (1982), with local scree-mantled inselbergs, like those in peninsular India and southwestern Australia. By this second view, cover by alluvial fans would have been initiated by faulting and uplift to the west, similar to the Darling Scarp of Western Australia and the Eastern Ghats of India. We are inclined to follow the piedmont rather than the peneplain interpretation from our reassessment of the time for formation of the paleosols, which is more than expected in deserts but less than that of lateritic paleosols of India and southwestern Australia. Similarly thick and deeply weathered paleosols, much more impressive than the pre-Torridonian examples, are known from Precambrian rocks (Retallack 1990; Marmo 1992).

Parent Material

The parent material of the Sheigra and Staca paleosols was gneiss and amphibolite, respectively. Weather-resistant materials such as titania, microcline, and pegmatite dikes persist to the preserved surface of the paleosols, and so are evidence against soil formation incorporating significant sedimentary cover layers.

These pre-Torridonian paleosols signal the end of control of soil oxidation by parent-material reducing power, which is characteristic of pre-Huronian (2.3 Ga) and older Precambrian paleosols (Retallack 1990). Nevertheless, there are differences in the paleosols that can be related to the slightly coarser grain size and more felsic composition of the gneiss compared with amphibolite. For example, the visibly oxidized and clayey horizon is thinner in the Staca profile than in the Sheigra profile (Fig. 2), and so is the depth of chemical weathering (Figs. 3, 4). Only the Staca paleosol with its calcic hornblende has accumulated carbonate nodules, unlike the Sheigra paleosol.

Time for Formation

Both the Sheigra and the Staca paleosols are at a major regional unconformity, with ample geological time available for soil formation. Their metamorphic parent materials are no younger than 1800 Ma (Watson 1983). These were exhumed into a hilly terrain that began receiving sediment in its deepest parts some 999 Ma ago, but alluvial fans covered the formerly well drained soils reported here by 810 Ma (Johnson 1983). Not all these hundreds of millions of years would have been available for soil formation. Williams (1969) argues that this erosional landscape developed by slope replacement of a retreating mountain front in such a way that any part of it was exposed for only tens to hundreds of thousands of years after erosion down to bedrock.

The amount of clay formed in the paleosols compared with surface soils gives an rough idea of time involved in soil formation. Clay in the Sheigra paleosol can be estimated by a wedge ranging from 55 to 0 volume percent over 4 m, and the Staca clay from 57 to 0 volume percent over 3.8 m (Figs. 3, 4). The main components of this part of the paleosols are clay, quartz, and microcline, with bulk density comparable with that measured for the whole rock (2.6 and 2.7 g/cm^3, respectively), so this amounts to 193 and 167 g/cm^2 clay for each profile. In the coastal plain and piedmont of the southeastern United States, a region of 1040 mm mean annual rainfall and 10°C mean annual temperature, this amount of clay formation would take about 1.2 million years (Pavich et al. 1989; Markewich et al. 1990). Another way of assessing time for formation is by comparison with rates of saprolitization of 1 m per 100,000–500,000 years for similar soils in this same region (particularly profiles F11 and F18 near Fairfax, Virginia, of Pavich et al. 1989), which gives time for formation of the 1–2 m thick saprolitic Scottish paleosols as 100,000–1,000,000 years. These results are compatible with geochemical evidence for overall contraction of the mass of the paleosols during weathering (Fig. 9), which has been found in surface soils to follow a phase of net soil dilation after hundreds of thousands of years (Brimhall et al. 1991).

These estimates are compromised by presumably elevated CO_2 levels in the atmosphere of the Precambrian, when there were no forests like those of Virginia today. However, CO_2 produced by respiration in forest soils can climb as high as 110 times present atmospheric level (Brook et al. 1983), an unrealistically high level even for the Archean atmosphere (Kasting 1987). In the long term it appears that carbonic acid supplied by Archean rain has been supplanted by that generated by Phanerozoic forests (Retallack 1990). The mix of sources of weathering acid at 0.8 Ga when these Precambrian paleosols formed was probably more atmospheric than biogenic, and overall less than under forest soils today (Kasting 1987). The mild enrichment without fractionation by atomic weight of rare-earth elements in both the Staca and the Sheigra paleosols may reflect this, because carbonate ligands are suspected as the principal removers of rare-earth elements from deeply weathered soils today (Braun et al. 1990; Michard et al. 1987). In summary, then, these Scottish pre-Torridonian paleosols probably represent soil formation for several hundred thousand but less than a million years.

Classification of the Paleosols

In view of their presumed long time for formation, the Sheigra and Staca paleosols could be classified as Oxisols or Ultisols in the U.S. comprehensive classification of soils (Soil Survey Staff 1975, 1990). However, the common grains of microcline, and especially carbonate, rule out Oxisols or Ultisols, and both paleosols also lack the subsurface horizon of clay enrichment (argillic horizon) of Ultisols. In profile form these Precambrian paleosols are most like Inceptisols, particularly Eutrochepts or Eutropepts, but the amount of clay formed within them is unusual for such soils, which are usually weakly developed. Like many Precambrian paleosols, these Scottish profiles may be best ultimately placed within an extinct order of weakly oxidized well-drained soils, which have been informally termed "Green Clays" (by Retallack 1986b). More profiles of these enigmatic archaic soils will need to be examined before an extinct order of soils can be proposed seriously, and examples as well preserved as the Staca and Sheigra profiles will be important to this task.

Differences between modern and Precambrian paleosols frustrate attempts to use soil taxonomy to find comparable modern profiles. Nevertheless, in general thickness, overall degree of weathering, and distribution of carbonate, the Staca clay paleosol is similar to Kadirabad clay loam soil and the Sheigra clay to Patancheru sandy loam soil of India (Murthy et al. 1982). These soils, however, are classified as Typic Chromusterts and Udic Rhodustalfs, indicating that Kadirabad soils have a surface dark cracked clay-rich horizon and Patancheru soils have a subsurface red clay-rich horizon, not seen in the Precambrian paleosols or as clasts in overlying sediments. Both soils form on gently sloping pediments of the Medak district of Andhra Pradesh. Kadirabad soils formed from granodiorite and Patancheru soils formed from granite-gneiss in a monsoonal climate with mean annual rainfall of 730–760 mm and mean annual air temperature of 25.8°C, not very different from conditions envisaged for the Precambrian profiles. Although now heavily cultivated, their natural vegetation was probably grassy woodland of wattle (*Acacia* spp.) and neem (*Azadirachta indica*). Differences in classification between Precambrian and surface soils can be blamed largely on this very different ecosystem.

CONCLUSIONS

Our work confirms much of Williams' (1968, 1969) pioneering studies. The altered gneisses and amphibolites below basal Torridonian sediments were indeed Precambrian soils. Prominent soil features in the field include gradual conversion of rock to purple-red clay, with local corestones and soil creep phenomena, all truncated abruptly by overlying sediments. Under the microscope the paleosols show extensive conversion of silicate grains to clay, abundant clay skins, and scattered small micritic nodules, also indicative of soil formation. Finally, our petrographic and chemical analyses combined are evidence for soil-forming processes of hydrolysis and oxidation.

A remarkable aspect of these profiles, especially compared with other Precambrian paleosols, is their limited burial alteration. There has been significant illitization of the surface clayey horizons and some local albitization, but illite crystallinity among other indicators reveals burial by less than 1500 m and burial temperatures of less than 120°C. This is less altered than many Phanerozoic paleosols and accounts for the preservation of pedogenic smectite, which has not to our knowledge been reported yet from any other Precambrian paleosol.

Soil formation during Precambrian time presents special challenges, because little is known about the effects of purely microbial ecosystems on land in climates that now support tropical woodland or about an atmospheric mix of O_2 and CO_2 very different from today. For these reasons the identification of analogous surface soils and soil taxonomy is of limited usefulness for understanding Precambrian paleosols. Instead, we rely on general principles of soil formation and simplistic models based on these principles. Often these approaches are unsatisfactorily vague. For ground truth about soil formation under the extinct mix of environmental conditions that existed during Precambrian time, little-altered paleosols like the Staca and Sheigra paleosols are an especially valuable resource.

ACKNOWLEDGMENTS

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