Precambrian Research, 32 (1986) 195—232
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

REAPPRAISAL OF A 2200 MA-OLD PALEOSOL NEAR WATERVAL ONDER, SOUTH AFRICA

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(Received September 15, 1985; accepted November 25, 1985)

ABSTRACT

Retallack, G., 1986. Reappraisal of a 2200 Ma-old paleosol near Waterval Onder, South Africa. Precambrian Res., 32: 195—232.

A fossil soil (paleosol) about 2200 Ma-old in a deep roadcut between the towns of Waterval Boven and Waterval Onder (Transvaal), here called the Waterval Onder clay paleosol, was formerly reported to have formed on the Hekpoort Basalt. It is here thought to have developed on a thick (formerly at least 5.5 m), upward-fining and thinning sequence of sandstone, siltstone and shale derived from pre-existing soils on Hekpoort Basalt, within the basal portion of the fluvial Dwaal Heuvel Formation. In contrast to paleosols developed on Hekpoort Basalt, the Waterval Onder clay paleosol is little deformed, lacks corestones, and shows compositional variation and berthierine-rich laminae interpreted as relict bedding. The paleosol has a ridge and swale topography (linear or wavy gilgai) and a subsurface structure (mukkara) typical of modern Vertisols. Since there is considerable relict bedding, it probably formed over a short period of time (less than 7000 years) and was not 'self plowing' to the extent seen in modern Vertisols. Clastic dikes filled with sandstone from the overlying rocks, are evidence that the profile was periodically dry to depths of at least 2 m. More drab colored rocks below that may represent the level of permanent water table. Iron and manganese stain of layers and claystone granules within the surface horizon of the profile have some characteristics of modern rock varnish, and could be taken as evidence for life in the soil. The high ratio of ferric to ferrous iron near the surface of the paleosol, together with evidence from clastic dikes that this horizon was not waterlogged, is an indication of an oxidizing atmosphere, although probably one in which oxygen was much less abundant than it is today. The very low amounts of acid titratable bases in the shaly parent material of this paleosol and surficial loss of these bases within the paleosol, appear to have been produced by decarbonation reactions comparable to those found today in soils of vascular land plants. A likely source of weathering acid during the Precambrian is an elevated abundance of atmospheric carbon dioxide. Considering these likely atmospheric differences, together with the distribution of comparable modern Vertisols and other geological evidence, the climate under which this soil formed was probably semiarid to subhumid, seasonally dry and temperate.

INTRODUCTION

The term uniformitarianism was useful as a rallying cry against catastrophism and creationism in the early days of the growth of geological

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thought, but is not heard so much nowadays. Uniformitarianism is after all, merely the logical method of comparing the unknown (such as a deposit of the geological past) with the known (comparable modern deposits). This is essentially the method of all science (Gould, 1965). In this age of increasingly multidisciplinary research, it is not reasonable to imply that earth sciences have a special kind of scientific comparison. Moreover, simple modern comparison is not always appropriate, and has led to misleading inferences in the study of Precambrian fossil soils (Retallack, 1986).

A truly 'modern' soil, one which formed entirely during the present century, would show little development: only a little organic matter added to weakly altered parent material. It takes several hundreds to thousands of years for significant amounts of soil formation, which includes such processes as reddening, clay formation and accumulation of carbonates (Birkeland, 1984). Over millions of years these processes can produce thick residual materials, such as calcrete, laterite and bauxite. These exert influences of their own on further soil formation and may persist long after conditions (such as climate or vegetation) under which they formed have changed (Goudie, 1973).

Faced with this array of materials which can all be regarded as 'soils', with which should Precambrian paleosols be compared? Most Precambrian paleosols have been recognized at major geological unconformities, where the effects of weathering are often most intense and obvious, because of a long time of formation. Although easy to recognize, such unconformities are difficult to interpret. They are best compared with modern duricrusts (such as laterite) and other materials formed by deep weathering (such as saprolite). On the other hand, Precambrian paleosols formed over very short periods of time (only a hundred years or so) would not show enough soil alteration to be recognizable or to be useful indicators of past environments. Such weakly developed paleosols in Ordovician and younger rocks are recognized principally by the presence of fossil roots or trace fossils. Precambrian paleosols formed over intermediate periods of time (hundreds to several thousands of years) offer the most promise for interpreting paleoenvironments of the past, because they are developed enough to be recognizable and interpretable, but lacking complications of major unconformities. In this paper a start is made on the documentation and interpretation of Precambrian paleosols formed over these intermediate time spans, with the description of an example from South Africa.

The paleosol discussed here was discovered, described, chemically characterized, and interpreted by Button (1979). During my examination of his petrographic thin sections, and later during my own fieldwork in South Africa during March 1984, I could confirm many of his observations. However, I reached different conclusions as to the nature of this paleosol. In my view this paleosol formed not on basalt, but on a deposit of sandstone, siltstone and shale overlying another paleosol developed on basalt. This is not to deny the existence of paleosols at the basin-wide unconformity above the Hekpoort Basalt, and stratigraphically equivalent Ongeluk Volcanics (Button, 1975; Button and Tyler, 1981). In the roadcut which I examined, there is a coarse grained, red rock, with basaltic corestones, underlying the sandstone, siltstone and shale unit, but it is toward the outer part of the roadcut, within the zone of Cenozoic weathering. At most other localities, paleosols formed directly on Hekpoort Basalt are too deformed or metamorphosed for detailed study (Button, 1975, 1979). They remain undescribed.

GEOLOGICAL SETTING

Location

The Waterval Onder clay paleosol is in the roadcut 2.7 km west of Waterval Onder on national road 4, Transvaal, South Africa (Figs. 1—3). The

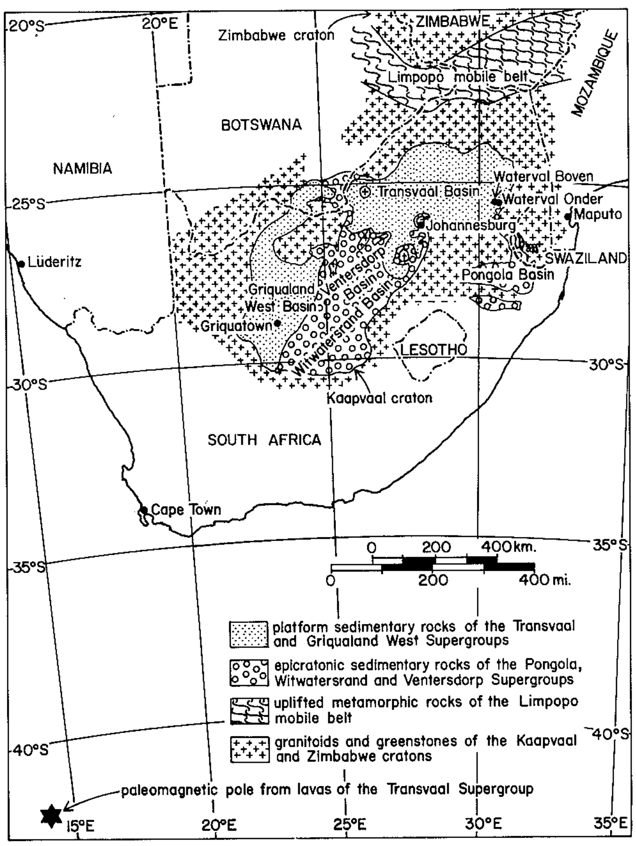

Fig. 1. Location of Waterval Onder and major tectonic units of southern Africa during the Archean/Proterozoic transition (2900—1800 Ma ago: modified from Tankard et al., 1982).

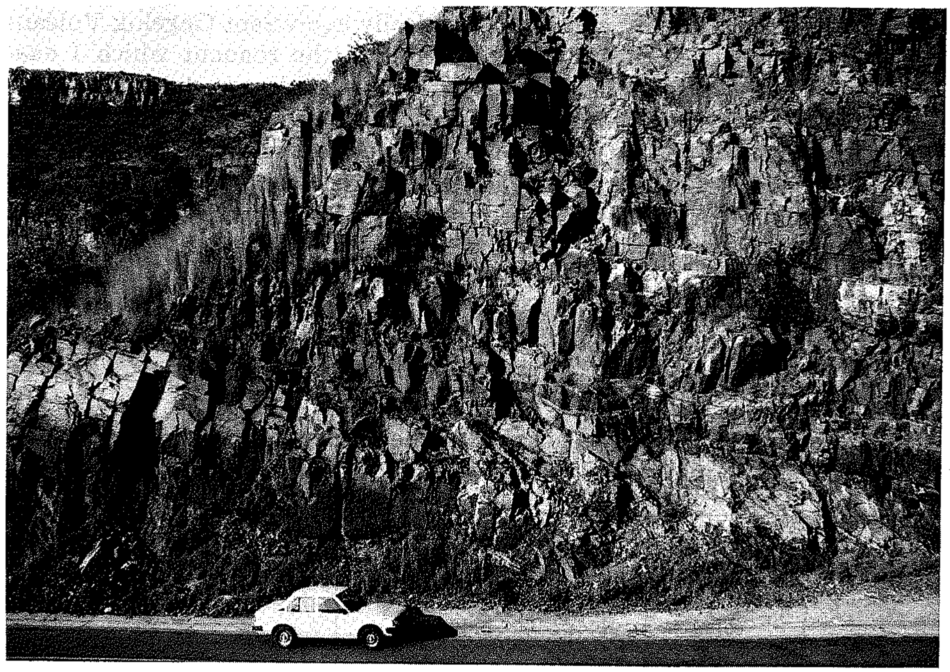

Fig. 2. Paleosol in the south face of the roadcut 2.7 km west of Waterval Onder, South Africa.

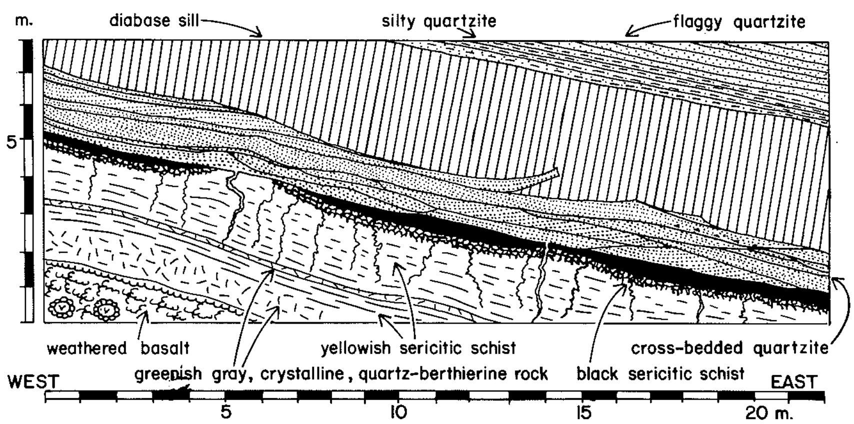

Fig. 3. Annotated field sketch of part of the south face of the roadcut near Waterval Onder, South Africa.

type profile was measured on the east side of the roadcut, on the south side of the road, just west of the bridge across the Eland River. The paleosol is also exposed just across the road, on the north side. This is at 25°38′65″ south and 30°21′45″ east on the 1:50 000 map 2530CB 'Waterval Boven'.

Stratigraphic position

This paleosol has been thought to have developed on top of the Hekpoort Basalt, but is now thought to have formed on a thin (4.5 m) basal shaly and sandy unit of the Dwaal Heuvel Formation. These are units of the Pretoria Group, on the eastern margin of the Transvaal Basin (Kent, 1980).

Metamorphic alteration

The paleosol was covered by at least 7 km of sedimentary rocks of the Pretoria Group, 3 km of Rooiberg Felsite, 5 km of intrusive Bushveld Complex and 1—2 km of mafic sills scattered within the sedimentary rocks. Even with a low geothermal gradient (20°C km^{-1}), burial temperature would have been at least 300°C. This is compatible with the presence of small amounts of chloritoid, stilpnomelane and epidote, which are typical of lower greenschist facies regional metamorphism (Button, 1979).

Geological age

Using the Rb/Sr method, shales of the Timeball Hill Formation, which underlies the Hekpoort Basalt, have been radiometrically dated at 2263 ± 84 Ma (Hamilton, 1977) and the basalt itself at 2224 ± 21 Ma (personal communication of Crampton in Button, 1979). The stratigraphically equivalent Ongeluk Volcanics have been dated by the Pb/Pb method as 2240 ± 57 Ma-old (Walraven et al., 1982). The basalts are no more than a few tens of millions of years older than the Waterval Onder paleosol, which formed after development of a clayey paleosol on basalt. The Waterval Onder paleosol was overlain by 7 km of mineralogically mature sediments of the Pretoria Group. These sediments were intruded by mafic rocks of the Bushveld Complex, radiometrically dated by the Rb/Sr method at 2095 ± 24 Ma (Hamilton, 1977).

DESCRIPTION OF THE WATERVAL ONDER CLAY PALEOSOL

Field and petrographic features

In the following description, the first number of each paragraph is the depth from the top of the paleosol to the top of the horizon (Fig. 4). Much of my terminology is from Brewer (1976), and colors of fresh rock in the field from Munsell Color (1975). Mineralogical composition was determined from point counting by Button (1979), but the mineral he called ‘chlorite’ and ‘chamosite’ is here called berthierine (after Brindley, 1982). I estimated percentages of clay, silt and sand (Fig. 4) by point counting the same thin sections using the grainsize scale of soil science (Soil Survey Staff, 1975),

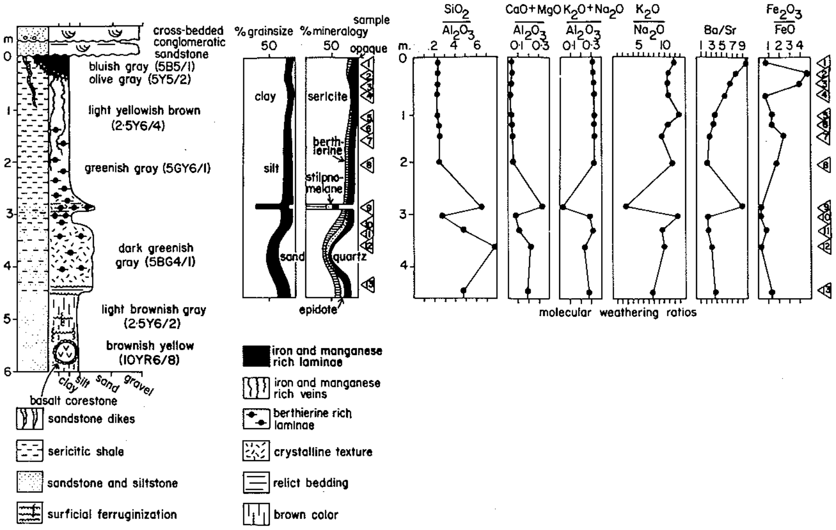

Fig. 4. Columnar section, percentage grainsize and mineralogy, and some molecular weathering ratios calculated from chemical composition of the Waterval Onder clay paleosol (given by Button, 1979; Holland, 1984).

rather than the Wentworth scale. Although sericite crystals are now silt size, they were counted as clay, from which they were presumably derived. Interpreted pre-metamorphic rock types (in parentheses) and the letter code for soil horizons (from the latest U.S. Department of Agriculture scheme; Birkeland, 1984) are justified in later sections of this paper.

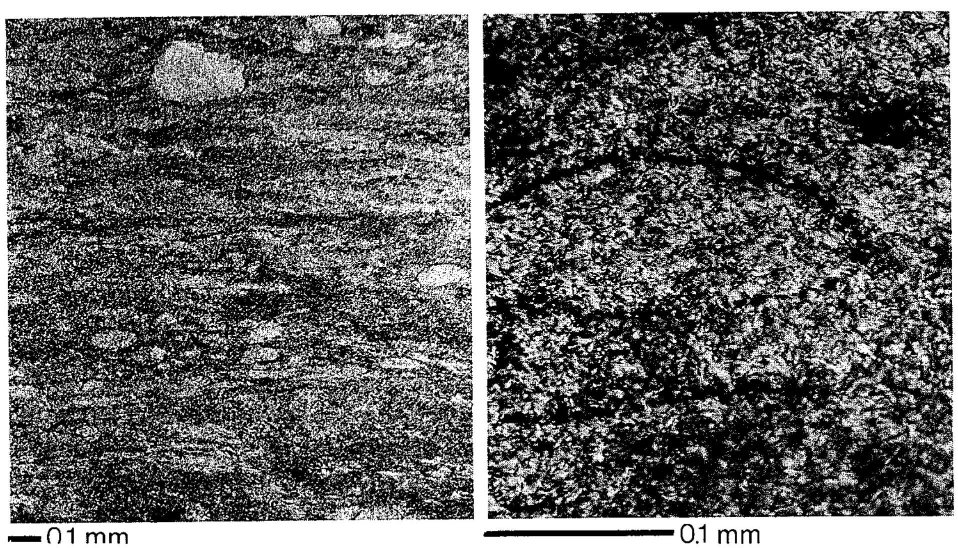

Fig. 5. Lamination, scattered crumb peds and possible rock varnish structure at two different magnifications in the lower A horizon (sample 2) of the Waterval Onder clay paleosol. Scale bars are 0.1 mm.

+0 cm; fluvial deposit; quartzite (coarse to medium grained, quartz sandstone); gray (5Y5/1) to dark bluish gray (5B4/1); trough and planar cross bedded, with sparse stringers of claystone clasts; individual grains and granules white (5B8/1), dark greenish gray (5G4/1), black (5B2/1) and yellow (5Y7/4); granular silasepic microtexture; mainly quartz, with common rock fragments, now altered to berthierine and sericite, and few pyrite crystals; abrupt wavy contact to

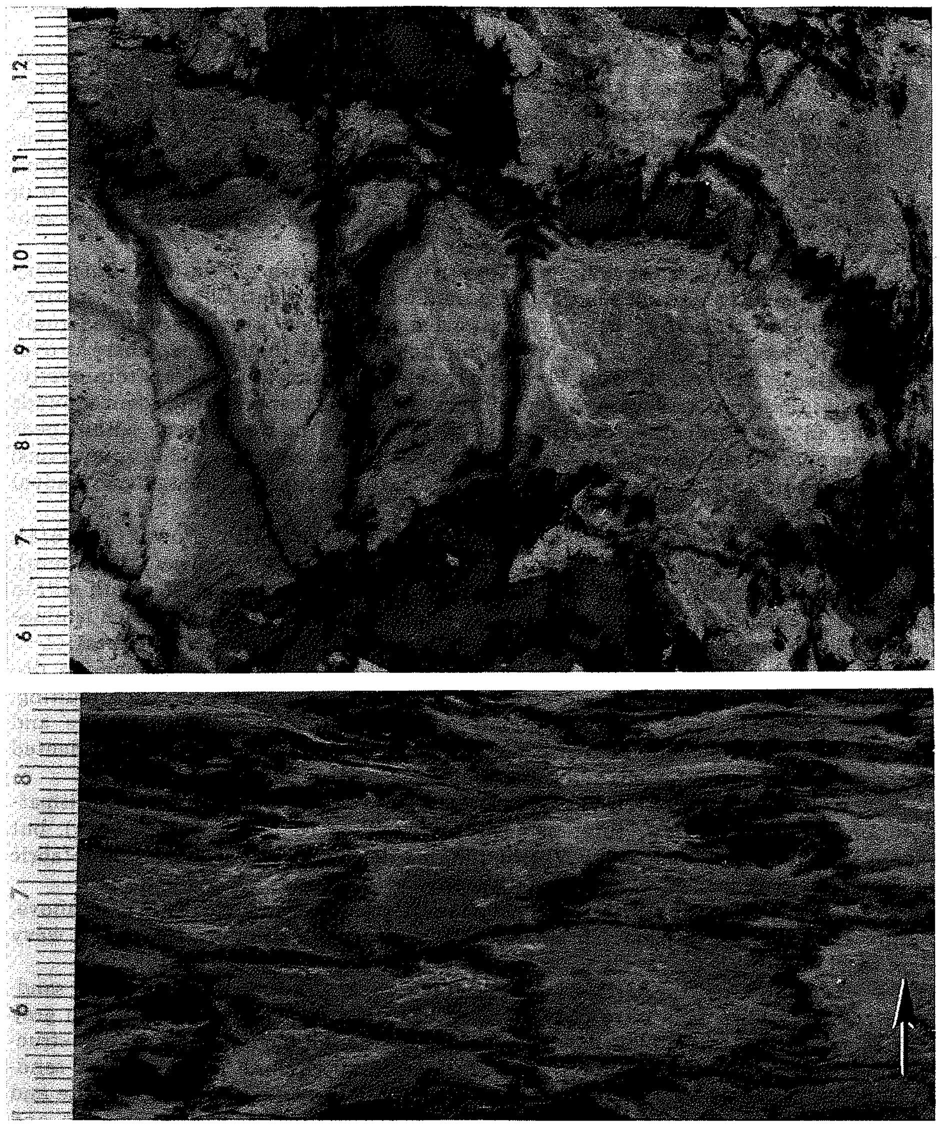

Fig. 6. Coarse, angular blocky structure in horizon AC (sample 3) in slabs sawn vertically (below) and horizontally (above) to the surface of the Waterval Onder clay paleosol. Scales are in cm and graduated in mm. Arrow indicates direction to top of paleosol.

—0 cm; A; sericitic schist (silty claystone); bluish gray (5B5/1); this horizon pinching out laterally at intervals of 10 m, where mounds of underlying light yellowish brown claystone of horizon C1 reach the surface (Figs. 2—3); prominent relict bedding accentuated by manganoferrans of very dark bluish gray (5B3/1); scattered very fine crumb peds, many of them broken rounds, defined by manganoferrans on top (less so on bottom) and laminar manganoferrans within overhangs (Fig. 5); indistinct, medium, angular blocky peds, more common near base of horizon; microfabric unistrial mosepic; mainly fine-grained sericite, with lesser quartz and berthierine, common leucoxene and ilmenite; gradual wavy contact to

—40 cm; AC; sericitic schist (silty claystone); olive gray (5Y5/2) to light yellowish brown (2.5Y6/4); this horizon also pinching out laterally with the horizon above (Figs. 2—3); prominent, coarse, angular blocky peds defined by bluish gray (5B5/1), illuviation, manganoferri-argillans (Fig. 6); few fine crumb peds; microfabric unistrial porphyroskelic omnisepic, with randomly oriented skeleton grains and some stress argillans (these apparently oriented with metamorphic fracture cleavage); mainly sericite, with sparse leucoxene and opaque minerals within peds, but abundant oriented opaque minerals in cutans; gradual wavy contact to

—55 cm; C1; sericitic schist (silty claystone); light yellowish brown (2.5Y6/4); very coarse blocky angular and prismatic peds, defined by ptygmatically folded, narrow (1—8 mm), bluish gray (5B5/1), manganoferri-argillans; few weak, very fine crumb peds; where horizons A and AC pinch out, this horizon is at the surface, where it contains ptygmatically deformed, wide (1—3 cm) skeletans (sandstone dikes) from overlying sandstone; these mounds were evidently linear ridges, as there are equivalent mounds along strike in the roadcut across the road to the north; microstructure within peds unistrial, porphyroskelic mosepic; mainly sericite, with randomly oriented opaque grains; opaques, chloritoid and leucoxene are concentrated within manganoferri-argillans; gradual irregular contact to

—120 cm; C2; sericitic schist (silty claystone); light yellowish brown (2.5Y6/4); very coarse prismatic peds, defined by ptygmatically deformed, narrow (1 to 5 mm), bluish gray (5B5/1) manganoferri-argillans, which become more rare with depth within this horizon; common prominent, subhorizontal, black (5B2/1), berthierine-rich laminae ('disc structures' of Button, 1979), of varying length (3—30 cm) and thickness (1—3 mm), with a surrounding neomanganoferran (Fig. 7); berthierine-rich laminae are displaced by skeleton grains, manganoferri-argillans, skeletans and brittle microfaulting; microfabric porphyroskelic mosepic; mainly sericite, with common berthierine and quartz (largely within berthierine-rich laminae or 'disc structure') and randomly oriented opaque minerals, leucoxene blebs and chloritoid laths; gradual wavy contact to

—140 cm; C2; sericitic schist (silty claystone); pale olive (5Y6/3); weak, coarse prismatic structure; few, subvertical, thin, manganoferri-argillans of bluish gray (5B4/1), but abundant, subhorizontal black (5B2/1) berthierine-rich laminae, as above; microfabric and mineralogy as above; wavy gradual contact to

—180 cm; C2; sericitic schist (silty claystone); greenish gray (5GY6/1); as above, but more abundant berthierine-rich laminae, with intergrown stilpnomelane; clear wavy contact to

—280 cm; Cg3; quartz—berthierine crystalline rock (medium-grained lithic sandstone); laminated with pale olive (5Y6/3), greenish gray (5GY5/1), dark greenish gray (5GY6/3) and black (5B2/1) layers (the latter are berthierine-rich laminae or 'disc structures'); sericitic portions with porphyroskelic insepic microfabric and abundant sand-sized skeleton grains of quartz, common ilmenite and leucoxene, and some epidote and magnetite; other layers with intertextic crystic fabric of intergrown berthierine and quartz (Fig. 8); clear wavy contact to

—288 cm; Cg3; sericitic schist (silty claystone); greenish gray (5GY6/1); common, dark bluish gray (5GY6/1) and black (5B2/1) berthierine-rich laminae; indistinct, medium, angular blocky peds in places; microfabric porphyroskelic mosepic; abundant

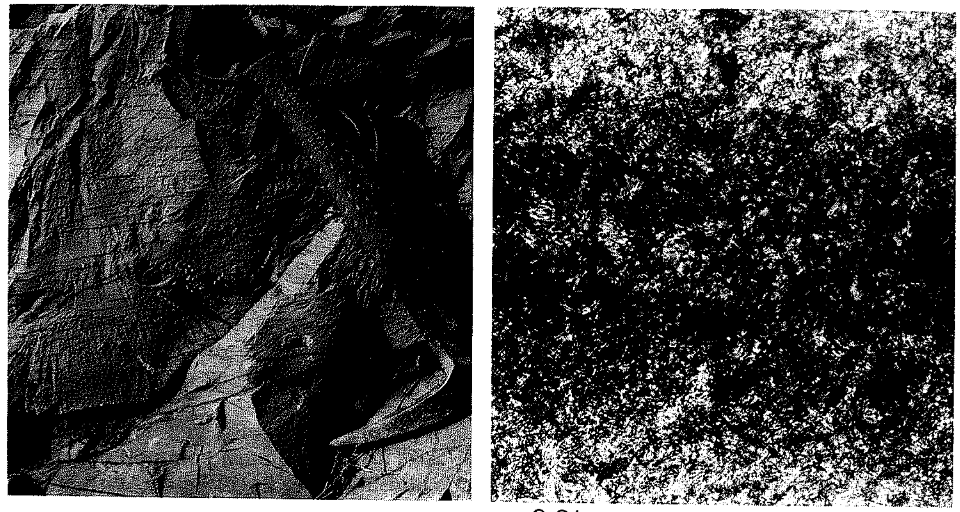

Fig. 7. Field appearance (left) and petrographic thin section (right) under plain light of berthierine-rich laminae from horizon C2 (sample 7) of the Waterval Onder clay paleosol. Scale bar for micrograph 0.01 mm.

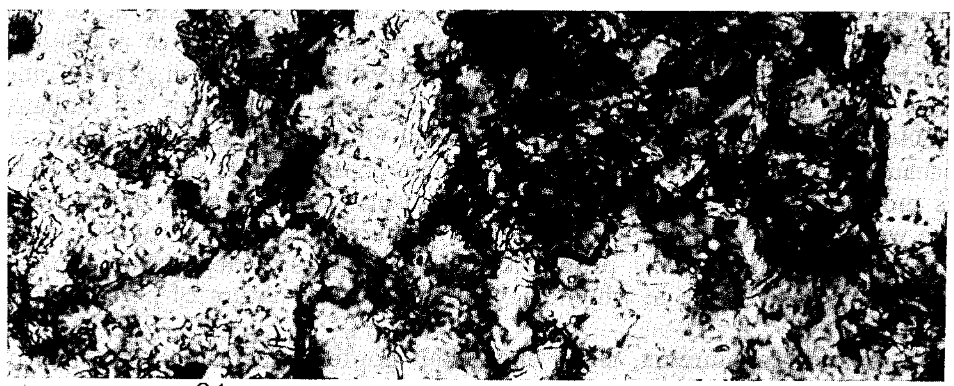

Fig. 8. Petrographic thin section under plain light of quartz (light) and berthierine (dark) of a crystalline rock interpreted here as a relict sandy bed of horizon Cg3 (sample 9) of the Waterval Onder clay paleosol. Scale bar is 0.1 mm.

rounded, sand-size grains of quartz; common ilmenite, leucoxene, epidote and magnetite; gradual wavy contact to

—320 cm; Cg3; quartz—berthierine crystalline rock (medium-grained lithic sandstone); dark bluish gray (5B5/1), few berthierine-rich laminae like those above; microfabric porphyroskelic insepic, to intertextic crystic near base of horizon; mineralogy as above; gradual wavy contact to

—340 cm; Cg3; quartz—berthierine crystalline rock (medium-grained lithic sandstone); dark bluish gray (5B4/1); with few berthierine-rich laminae and some other subhorizontal compositional banding; this is a massive, crystalline rock; microfabric intertextic crystic, with randomly arranged berthierine laths within a matrix of quartz; quartz may have a quasiferrimangan (leaving a clear margin wherever in contact with berthierine); also some stilpnomelane, leucoxene and magnetite; clear planar contact to

—400 cm; Cg3; quartz—berthierine crystalline rock (medium-grained lithic sandstone); dark greenish gray (5GY4/1) and black (5B2/1); otherwise similar to horizon above; clear planar contact to

—430 cm; Cg3; quartz—berthierine crystalline rock (siltstone); dark olive gray (5Y3/2), with some crystals weathered yellowish brown (5Y3/2), probably due to weathering in modern outcrop; otherwise similar to horizon above; abrupt moderately wavy contact to

—450 cm; 2Co; friable clayey sand (sandy siltstone); light brownish gray (2.5Y6/2) and light yellowish brown (2.5Y5/2) crystals almost equally abundant; prominent, yellowish red (5YR5/6) and red (2.5YR4/6), planar sesquans; this is the weathered top of a basalt flow and crops out closest to the eastern edge of the roadcut, where it has been altered to a friable, cave-forming rock by Cenozoic weathering; gradual wavy contact to

—530 cm; 2Co; friable clayey sand (sandy siltstone); brownish yellow (10YR6/8); very dark reddish brown (10YR3/2) and red (2.5YR4/6) sesquans; common corestones up to 40 cm diameter of little weathered, dark bluish gray (5B4/1) basalt, with a 2 cm wide, olive gray (5Y4/2), weathering rind.

Chemical characterization

Chemical analyses of 13 samples of the type Waterval Onder clay paleosol (from horizons A to Cg3), and of 4 samples of Hekpoort Basalt from elsewhere in South Africa, and of 1 sample of the basal zone of a paleosol on Hekpoort Basalt near Pretoria, were reported by Button (1979), and have been reprinted in a more readily available publication by Holland (1984). The basalt underlying the Waterval Onder clay was not analyzed, because it was highly weathered here. These chemical data are plotted here as molecular weathering ratios (Fig. 4). In these manipulations of the data, the weight percent of each major oxide in the ratio is divided by its molecular weight and so gives proportions of molecules, independent of their weight, of the volume in which they exist or of the nature of parent material.

Compared to alumina, silica and alkali content decreases very slightly toward the surface of the upper part of the profile (above 2 m). Alkaline earths increase slightly toward the surface compared to alumina, and so does potash compared to soda. Layers low in the profile, here interpreted to have been sandy and silty beds, are enriched in silica and alkaline earths, and depleted in alkalies compared to alumina. Since Na^+ is similar to and more soluble than K^+, like Sr^{++} compared to Ba^{++}, ratios of these elements reflect leaching. This is quite marked at the surface, but erratic at depth. The ratio of ferric to ferrous iron was calculated as a measure of oxidation and aeration of the profile. This is greatest in the blocky pedal part of the paleosol, between 30 and 50 cm, and declines below that. The lowest sample may have been oxidized in the present outcrop, as it was just above the strongly oxidized basalt near the outer portion of the roadcut. It also contains some friable deeply weathered grains, unlike the indurated massive clayey rocks of the rest of the profile.

ALTERATION AFTER BURIAL

Deformation

The Waterval Onder clay paleosol is surprisingly little deformed for its antiquity. It dips gently (12°) to the west (toward 256° magnetic azimuth), and is part of a thick, unfolded sedimentary sequence (Pretoria Group). There has been no discernible plastic deformation or shearing. The ridge and swale topography (gilgai microrelief and underlying mukkara structure of A horizon: to use terms of Paton, 1974) remains symmetrical. The blocky soil structure (peds of horizon AC) is crushed, but no more sheared than in comparable modern soils. Contorted black veinlets (illuviation manganoferri-argillans of horizons C1 and C2) remain perpendicular to the old land surface. The most prominent deformation is a well-spaced fracture cleavage at a high angle to bedding (Fig. 9). This locally disrupts all features of the paleosol by a few millimeters at most. It appears to have been produced by brittle deformation late in the burial history of the paleosol.

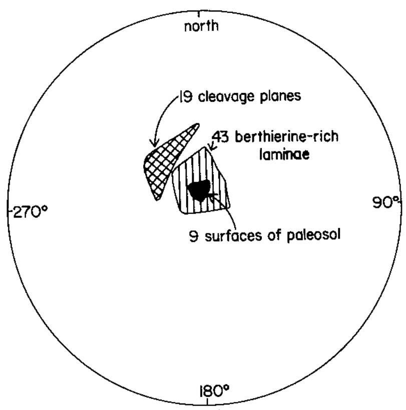

Fig. 9. Areas occupied by poles to the surface of the Waterval Onder clay paleosol, its berthierine-rich laminae and fracture cleavage on a Wolff net.

This mild deformation of the Waterval Onder clay paleosol contrasts with rodding and strong foliation in the clayey top of the Hekpoort Basalt elsewhere in the Transvaal Basin (Button, 1975). This stronger deformation is associated with a higher metamorphic grade, adjacent to large intrusions of the Bushveld Complex (Button, 1976, 1979). In contrast, this particular outcrop may have been protected from deformation by the thin diabase sill above the paleosol, or by formation in a local paleocanyon.

Compaction

Clastic dikes of sandstone (skeletans) and of opaque minerals (manganoferri-argillans) which were formerly more or less straight in the soil, are now ptygmatically folded by compaction. This is a fortunate circumstance, as it allows reconstruction of the amount of compaction, original densities and thicknesses of paleosol horizons.

Two such structures were carefully measured in the field: a sandstone dike now between 0 and 40 cm from the surface and compacted to 0.67 times its former length, and a contorted dark veinlet (manganoferri-argillan) now between 110 and 200 cm from the surface and compacted to 0.73 times its former length. The sandstone dike is in a gilgai ridge of the paleosol, so its compaction factor was used to estimate compaction of a similar material (sample 4) in the area of the gilgai swale, where rock density measurements have been made (by Button, 1979). The other estimate of compaction was in the area of the swale, and was used for the sample between 110 and 200 cm (number 7). Compaction factors (x) for the other samples at different depths (y) were extrapolated and interpolated between these two points using a linear formula ($y = 0.0074x + 0.62$). If these various compaction factors are multiplied by the present density of each sample, they show a fairly regular density gradient (see Fig. 12). If the compaction above each sample is estimated from the compaction factor for each sample, and these are added cumulatively down the profile, the depth of the lowest sample (13) now at 425 cm below the surface, was originally more like 540 cm.

While some of the simplifying assumptions used in these calculations, particularly linear variation of compaction factors down the profile, may be unrealistic, the densities and the density gradient gained are quite consistent with those of comparable modern soils (for example, the densities of Gray, Brown and Red Clays given by Stace et al., 1968).

Color change

Metamorphic alteration of lower greenschist grade (Button, 1979) has probably made these rocks more drab and green in color than they were originally. How much greener is uncertain. Considering the ratios used to approximate leaching (Ba/Sr) and oxidation (Fe_2O_3/FeO), and surficial increase in manganese in what appears to be rock varnish, color is likely to have been purple brown at the surface (horizons A and AC) and yellow in gilgai ridges and below the dark zone (horizon C1), but was probably not much more yellow than it is at present in subsurface horizons (C2 and Cg3).

Formation of berthierine-rich laminae

A curious and prominent feature of this paleosol attributed to metamorphism by Button (1979) are the dark, horizontal bands which he called 'disc structures'. These are laminar aggregates of berthierine (formerly known as chamosite), with a rim of iron and manganese decreasing in intensity into the surrounding sericite matrix (Fig. 7). They are much thinner (1—3 mm) than wide (3—30 cm), and are irregular in shape, rather than circular, as implied by the name 'disc structure'.

Button's metamorphic interpretation was based on his observation that berthierine-rich laminae 'cut cleanly through the subvertical contorted veinlets' and 'across millimeter-sized mottles of sericite in a leucoxene-rich sericitic groundmass' (Button, 1979, p. 15). My own inspection of oriented and sawn slabs did not completely confirm these observations. The berthierine-rich laminae are mostly disrupted by contorted subvertical veinlets and by sandstone clastic dikes (Fig. 10). A few subvertical veinlets were seen to be partly or completely cut by berthierine-rich laminae, and provide evidence of limited later remobilization of this mineral. I did not observe disc structure cutting across small 'mottles' of sericite. However, I did find berthierine-rich clasts in thin sections of sandstones of the overlying Dwaal Heuvel Formation. The inference from my own observations that the laminae are metamorphosed remnants of a very early feature of this paleosol, can be supported by other lines of evidence. As measured in the field with a Brunton compass, the berthierine-rich laminae are more variable in orientation than bedding, but are more consistently aligned with bedding than with fracture cleavage (Fig. 9). The berthierine-rich laminae are most abundant and laterally extensive in the lower part of the profile, where sandstone dikes and black veinlets are lacking. Berthierine-rich laminae are entirely absent in the blocky structured part of the profile

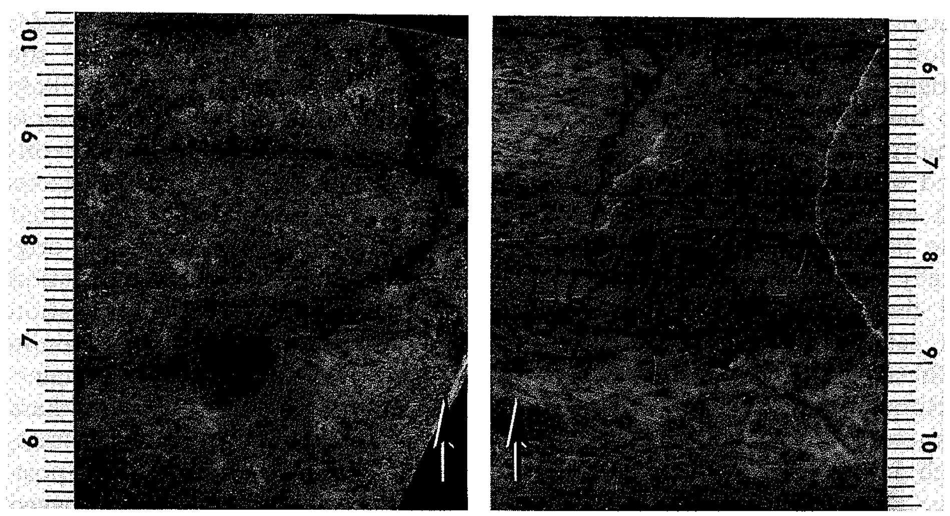

Fig. 10. Polished slabs from horizon Cg3 (sample 9), sawn vertically to the surface of the Waterval Onder clay paleosol showing berthierine-rich laminae cut by contorted black veinlets (illuviation manganoferri-argillans; right) and by a sandstone clastic dike (illuviation skeleton; left). Scales are in cm and graduated in mm. Arrows indicate direction to top of paleosol.

(horizon AC) and dark laminated upper part of the profile (horizon A). This is the reverse of the situation which would be expected if berthierine-rich laminae were produced as tension gashes by shearing of the clayey upper portion of the profile against more competent overlying quartzites. This proposed mechanism of Button (1975, 1979) may be appropriate for other outcrops of clayey rocks above the Hekpoort Basalt, but is unlikely for this one in which mukkara structure (of horizon A) and blocky structure (horizon AC) near the surface are little sheared. A final difficulty with a metamorphic origin of the berthierine-rich laminae is their halo of iron and manganese stain (neomanganoferran). This is evidence of mild oxidative alteration, unlikely to have occurred in the present deep roadcut or in the reducing chemical environment of diagenesis or greenschist grade metamorphism.

Considering the arguments already advanced for regarding the berthierine-rich laminae as remnants of original features of the soil, three likely explanations for them remain to be considered. They could be crushed remains of mineralized vesicles in the top of the basalt flow, or compacted stratiform nodules in a waterlogged part of the soil of either iron-manganese (placic horizons in the sense of Soil Survey Staff, 1975) or sideritic composition, or relict sedimentary laminae of minerals more mafic than the surrounding matrix. Interpretation as crushed vesicles must be rejected considering the lateral persistence of chamositic laminae. Their vertical spacing at intervals of a centimeter or more is also unusual for vesicles, which would be more densely packed if crushed nearly flat. Placic horizons of modern soils are usually concentrations of amorphous, opaque, iron and manganese oxides (Soil Survey Staff, 1975), quite different from berthierine. Even if the original mineralogy of these laminae was not berthierine, they are likely to have been mostly silicate minerals of comparable composition. Opaque amorphous minerals are found at the surface of the paleosol (A horizon) and within contorted black veinlets extending into the subsurface (horizons C1 and C2). Since these amorphous minerals are preserved it is likely that the berthierine-rich laminae also would have remained amorphous, if originally so. There are problems also with an original sideritic composition of the berthierine-rich beds and laminae, in that the diagenetic alteration of siderite into berthierine requires the presence of abundant kaolinite (Bhattacharya, 1983; Curtis, 1985). No trace of either mineral remains. Although sericitization obscures original clay mineralogy, the mukkara structure of the upper part and low alumina content of the berthierine-rich lower part of the paleosol are evidence against a kaolinitic composition. Nodules, layers or spherulites of siderite, abundant in waterlogged paleosols as old as Late Carboniferous (Retallack, 1986) do not resemble the berthierine-rich laminae of the Waterval Onder clay paleosol.

Interpretation as relict, mafic sandy or silty laminae is the most likely explanation for berthierine-rich laminae, by elimination, but this interpretation also accounts for other observations. In some cases, berthierine-rich laminae mark changes in grainsize, as if forming miniature placers at the base of sandy beds. In crystalline (formerly sandy or silty) parts of the profile, they are at angles to each other, like miniature placers within the toesets of ripple marks (Fig. 11).

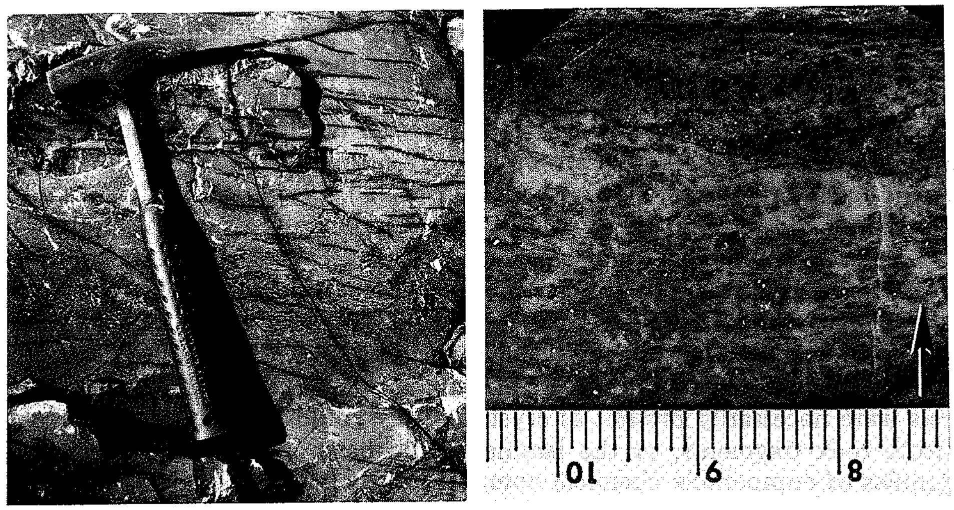

Fig. 11. Field appearance (left) and polished slab, sawn vertically to the surface of the Waterval Onder clay paleosol (right), both from horizon Cg3 (sample 10), showing forking of and angular discordance of berthierine-rich laminae and their association with areas of coarser grainsize suggestive of ripple marks. Arrow and scale as for Fig. 10.

This interpretation of the berthierine-rich laminae is my main reason for proposing that the Waterval Onder clay paleosol developed on an upward fining and thinning unit of sand, silt and clay overlying the Hekpoort Basalt, rather than on the basalt. Other observations supporting this view are the erratic, subhorizontal banding of the lower part of the profile (Fig. 4), the sharp contact of shaly rocks with the underlying basalt, and the very different appearance (banded) of the shaly rocks, compared to the basalt and its weathered top (massive with large corestones).

Berthierine is known in modern soils (Dixon, 1977), but for a number of reasons this is not thought to have been the original mineral of the berthierine-rich laminae. The berthierine crystals appear fresh and interlocking, and are at a high angle to the laminae containing them, aligned with the fracture cleavage (Fig. 7). Chloritoid and stilpnomelane associated preferentially with the berthierine probably formed during greenschist metamorphism. Only the well rounded quartz grains, and some of the opaque minerals, such as ilmenite and magnetite, appear to have been original minerals of the berthierine-rich laminae. Matrix quartz, berthierine, chloritoid and stilpnomelane, probably all formed during diagenesis and metamorphism.

The original mineralogy of these supposed sandy or silty, placer-like laminae, can only be guessed. A C.I.P.W. norm calculated (using A.R. McBirney's 3-10-85 version of program GPPII for Apple II) for a sample (number 9) which is unusually rich in laminae and thin beds of berthierine, gave mainly quartz (55 wt.%), with magnetite (18 wt.%) and hypersthene (16 wt.%), and lesser amounts of orthoclase (3 wt.%), ilmenite (2 wt.%) and albite (1 wt.%). Normative algorithms for petrographic purposes are not designed for assessing metasediments, but the mineralogy indicated is quite reasonable considering the normative composition of Hekpoort Basalt likely to have been upstream from the paleosol (sample 2 of Button, 1979 and Holland, 1984). This basalt was mostly normative anorthite (33 wt.%) with quartz (11 wt.%), hypersthene (17 wt.%), diopside (9 wt.%), albite (17 wt.%) and orthoclase (6 wt.%), and minor amounts of magnetite (1 wt.%), ilmenite (1 wt.%) and apatite (0.18 wt.%). Preferential destruction of feldspars according to Goldich's (1938) weathering series for sand-size minerals, combined with physical settling of heavy minerals such as magnetite, hypersthene and ilmenite, could account for the distinctive composition of the berthierine-rich laminae. This is not to imply that these minerals were necessarily present as crystals. They could also have been present within rock fragments, or the laminae could have consisted of clayey soil granules of comparable chemical composition. It is even possible that these rock fragments were largely kaolinite and siderite, derived from gleyed horizons of very strongly developed soils formed on Hekpoort Basalt. Unlike the Waterval Onder clay, some of these paleosols lack berthierine-rich laminae, but have a thick berthierine-rich horizon with a gradational contact down into little weathered basalt (Button and Tyler, 1981; H.D. Holland, personal communication, 1985). This view has the virtue of simplicity in relating the origin of all berthierine in these rocks to precursor siderite and kaolinite (a well known metamorphic reaction; Curtis, 1985), but there is as yet no positive evidence for this, or against a more complex original mineralogy.

Sericitization

Reconstructing the original nature of the fine-grained material of this paleosol is a more difficult task. It is now sericite of largely silt size. The way in which it has cracked (mukkara structure) and formed surficial ridges and swales (gilgai microrelief) is an indication that it was once largely clay. The persistence of what is here interpreted as relict bedding (berthierine-rich laminae throughout and compositional banding in lower part), despite the considerable likely time available for soil formation (about 7000 years as outlined in a later section), provides some physical constraints for the nature of the clay. It could not have included appreciable quantities of strongly swelling clays, such as smectite, which are responsible for the 'self-plowing' behavior of comparable modern soils (Buol et al., 1980). The paleosol lacks 'lentil' ped structure (described by Krishna and Perumal, 1948; Worrall, 1957; Johnson et al., 1962) and more complex microrelief features (such as melonholes and nuram gilgai, among others of Hallsworth et al., 1955; Paton, 1974) found in modern Vertisols with high smectite content. The paleosol is not so well developed, aluminous, or developed on ashy, easily weathered materials, that its clays are likely to have been appreciably kaolinitic. Thus the paleosol is most likely to have been illitic, with minor amounts of smectite and interstratified clays.

If the soil is assumed for the moment to have had a bulk composition comparable with that of the present paleosol, then its abundant K_2O (9 wt.%) is compatible with an illitic composition. To investigate this possibility further, a normative mineral composition was calculated for the shale on which the paleosol is thought to have formed (specimen 8), using the method outlined by Garrels and MacKenzie (1971, appendix C). In these calculations, minor chemical components (CaO, MnO and P_2O_5 each less than 0.05 wt.%) were disregarded. All the MgO was used to make hypersthene (using the formula $Mg_{0.5}Fe_{0.5}SiO_3$), all the TiO_2 for ilmenite ($FeTiO_3$), and all Na_2O for albite ($NaAlSi_3O_8$). The remaining iron was made into magnetite (Fe_3O_4). At this point all remaining Al_2O_3 (0.3249 moles/100 g rock) and K_2O (0.0998) were used to make illite [$K_{1.75}Al_4(Si_{6.25}Al_{1.75}O_{20}(OH)_4$] and orthoclase ($KAlSi_3O_8$). The proportions of each were calculated from the moles/100 g of K_2O in illite (Ki) derived from the following four simultaneous equations involving also moles/100 g of K_2O in orthoclase (Ko), and moles/100 g of Al_2O_3 in illite (Ali) and orthoclase (Alo): Ko = Alo: 3.286 Ki = Ali: Ki + Ko = 0.0998: Ali + Alo = 0.3249. The silica remaining after making illite and orthoclase was assigned to quartz. As a result of these calculations the normative mineralogy of the shale is presumed to have been largely illite (88 wt.%), with common albite (4 wt.%), hypersthene (3 wt.%), ilmenite (3 wt.%), and magnetite (2 wt.%), and minor orthoclase (1 wt.%) and quartz (1 wt.%). This is a reasonable result, since ilmenite, magnetite and quartz are present in the rock, in about the proportions indicated, and are also thought to be original minerals. Berthierine-rich laminae are sparse in this specimen, and as already argued, are thought to have been sandy layers of hypersthene and feldspar, or of rock fragments of comparable chemical composition.

While these calculations may seem reasonable, they are not entirely satisfactory, and highlight possible problems with the assumption that the bulk composition of the original soil was comparable with that of the present paleosol. Only with the very potash-rich formula used for illite, which is between that of muscovite [$K_2Al_4(Si_6Al_2)O_{20}(OH)_4$] and the most potash rich of the usual modern compositions of illite ($K_{1.5}Al_4(Si_{6.5}Al_{1.5})O_{20}(OH)_4$: Deer et al., 1963], could the normative calculations be completed with silica to spare for quartz. In other words, there is too much K_2O in the rock to allow for a normal modern illite composition and still have the quartz known to be present in the rock. Two possible explanations come to mind. It could be that Precambrian soils were less deeply weathered of potash (compared to other weatherable bases) than those modern soils forming under vascular land plants of humid climates, thus accounting for the high potash content of Precambrian and early Paleozoic shales and paleosols (Weaver, 1967; Retallack, 1986). Another possibility is that the original clays have gained potash during burial. In boreholes in the United States Gulf Coast, the proportion of illite in marine shales has been shown to increase from 20 to 80% at the expense of smectite, by addition of K^+ and Al^{3+} during acidic dissolution (from CO_2 of carbonates or organic matter) of orthoclase or muscovite, and loss of excess Si^{4+} from the system in pore water (Hower et al., 1976). Isotopic studies indicate that this mostly occurred at depths of 2100—2500 m (much shallower than the present depth of burial of the shales). It may have been triggered by freshwater flushing during a low stand of sea level, and terminated by dewatering on deeper burial (Morton, 1985). It could be that the illitic to sericitic composition of many Precambrian and early Paleozoic shales and paleosols reflects potassium metasomatism of this kind (Holland, 1984; Curtis, 1985).

These explanations for potash-rich shales and paleosols are not mutually exclusive, and each situation should be considered on its own merits. For the Waterval Onder clay paleosol, both seem likely. The main source of potash and alumina for illitization of clay would have been orthoclase grains in the sandy laminae (now berthierine-rich). Little potassium is likely to have been gained from overlying sandstones. A point-counted petrographic thin section of a specimen from above the paleosol contained mainly quartz (87 vol.%) with common sericitic and berthierine-rich rock fragments (10 vol.%). Two other sandstones from above the diabase sill were similar: 84% and 82% quartz, and 12% and 14% rock fragments. All three specimens were stained with sodium cobaltnitrate (using method of Houghton, 1980) and no potassium feldspar was detected. These are also an unlikely source of potassium because both reconstructed (Fig. 12) and absolute amounts of K_2O (Holland, 1984) decrease slightly toward the top of the paleosol. Some potassium could have been gained from the underlying basalt, which has very low amounts of K_2O (0.91 wt.%) and normative orthoclase (6 wt.%). Since the soil structure of the paleosol is compatible with a largely illitić composition, potassium metasomatism could have altered a more potash-poor illite than that used for normative mineralogical calculations, or a mixture of illite with minor amounts of smectite or interstratified clays, into the present sericite.

RECONSTRUCTION OF THE PALEOSOL

General features

Considering the various alterations to this paleosol after burial, the Waterval Onder clay is now seen as a weakly developed soil developed on shaly alluvium, largely illitic in composition, with sandy and silty beds of basaltic rock fragments, mafic minerals (perhaps hypersthene), feldspars and quartz (Fig. 1). The surface of the soil had moderate gilgai relief of about 30—50 cm. Elongate subparallel ridges of yellow clay were spaced at intervals of about 10 m. Ridges of the corresponding subsurface mukkara structure were deeply cracked. Swales between the ridges were filled with purple brown shale (horizon A). There were scattered small granules, washed in from the surrounding ridges, but most of the swale surfaces were smooth with varnish of iron and manganese. A transitional layer (horizon AC) between the upper dark material (A) and yellow clay within the mounds and below the swales (C1), had crude blocky structure (peds), defined by cracks filled with purple brown clay washed in from above (illuviation manganoferri-argillans). Scattered iron and manganese stained veins below that horizon defined a crude prismatic structure in the yellow claystone (horizon C1). This horizon had disrupted, relict sandy laminae of volcanic rock fragments, mafic minerals (such as hypersthene), feldspar and quartz. These sandy laminae were more abundant and laterally persistent lower in the profile (horizon C2), where the clay was more drab in color (probably greenish gray). Deeper within the profile (horizon Cg3), there were prominent, sandy, relict beds which were dark with mafic mineral grains and rock fragments (dark greenish gray). This horizon may also have been enriched in manganese or siderite as a gleyed part of the profile below the permanent water table.

The Waterval Onder clay was a base-rich soil, with exchangeable cations dominated by K^+. It was probably neutral to mildly alkaline in chemical reaction (considering the slight and irregular surficial increase in CaO). The water table was quite deep within the profile (mostly below an original depth of 2 m, and permanently below about 3.4 m). During dry periods, the soil cracked deeply (to former depths of about 2 m).

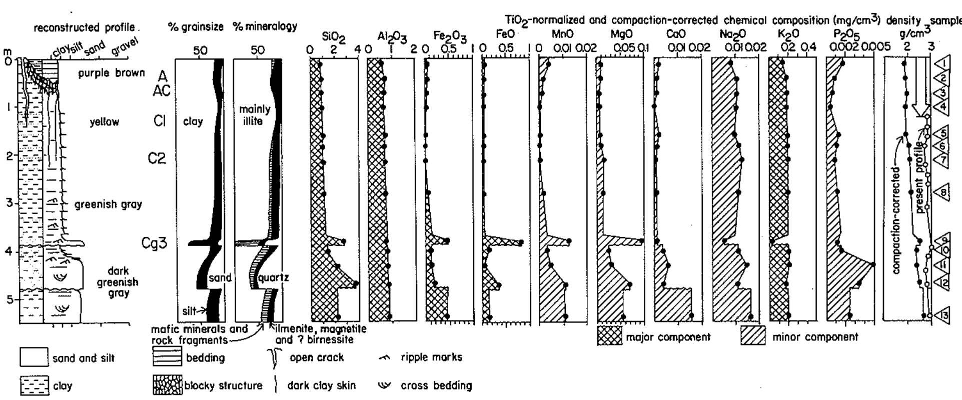

Fig. 12. Reconstructed columnar section, grainsize, mineralogy and chemical composition, as they may have been during formation of the Waterval Onder clay paleosol.

Chemical differentiation

Some of the alterations after burial already discussed can also be unravelled to reconstruct the former chemical composition of the original soil, from the present composition and density of the paleosol (given by Button, 1979, and Holland, 1984). For these calculations, titanium was assumed to have been stable throughout the profile, as it is in modern weathering provided pH is not extremely acidic or alkaline (Brewer, 1976). The erratic variation in chemical composition in the lower part of the profile (Figs. 4, 12) has been argued to reflect interbedded sandy and clayey beds in the parent material, so the samples were all normalized to the content of TiO_2 in grams per cubic centimeter of the lowest sample (number 8) which is likely to have had an original composition like that of the upper part of the profile. Gains and losses of major oxides relative to this TiO_2 datum are assumed to be due to ancient weathering, compaction, diagenesis and metamorphism. Diagenetic and metamorphic effects are not thought to have substantially altered the bulk composition of these rocks. The K_2O content of the sericite may be greater than that of the original clay, but is thought to have been derived locally and largely from the dissolution of potassium feldspar. The effects of compaction can be disentangled from the measured contortion of clastic dikes, as already discussed. Each of the major oxides for each sample have been multiplied by both the ratio of parent material to sample TiO_2 content and by a compaction factor, to reconstruct the chemical composition of the uncompacted soil (Fig. 12).

Reconstructed in this way, ancient weathering appears to have caused moderate loss of Al_2O_3 and K_2O, significant loss of SiO_2 and slight gain in Fe_2O_3 at the surface. There was very little MnO, MgO, CaO, Na_2O and P_2O_5 in the soil, and these oxides show either steady depletion more marked toward the surface (Na_2O) or depletion at depth and a slight gain at the surface (MnO, MgO, CaO, P_2O_5). Compared to the shaly upper part of the profile, beds thought to have been sandy in the lower part of the profile have much more SiO_2, Fe_2O_3, FeO, MnO, MgO, CaO and P_2O_5, about the same amount of Al_2O_3, and comparable to reduced amounts of K_2O and Na_2O.

The chemical analyses manipulated in this way are not greatly different from the raw analyses, although a number of irregularities in the upper portion of the profile have been smoothed out. These new results are, however, different from those presented by Button (1979), who attributed concentration of TiO_2 to compaction alone (disregarding volume changes during weathering) and who normalized major oxides to the TiO_2 content of the Hekpoort Basalt elsewhere in the Transvaal Basin (rather than the shale assumed to be parent material here).

CLASSIFICATION OF THE PALEOSOL

Because of its mukkara structure and gilgai microrelief (to use terms of Paton, 1974), this paleosol can be identified as a Vertisol, both of the U.S. Department of Agriculture (Soil Survey Staff, 1975) and of the F.A.O. classification (Fitzpatrick, 1980). The light yellowish color of the paleosol in ridges compared to the dark (likely purple brown) color in swales is most like that of Chromic Vertisols in the F.A.O. classification. Finer identification of the paleosol in the U.S.D.A. classification requires information concerning paleoclimate and the length of time that cracks were open during the year. If paleoclimate was as reconstructed in a later section of this work, then the most similar modern soils would be Chromuderts. In the classification of the Australian C.S.I.R.O. (Stace et al., 1968) the paleosol was a Brown Clay. It keys out to Ug5.22 in Northcote's (1974) descriptive classification. This means it was a uniform, clayey, cracking profile, developed on shale, and with a subsurface horizon redder than 2.5Y.

These modern kinds of soil form on freely drained, low rolling country in warm regions (more than 22°C mean annual soil temperature), with seasonally dry (deep cracks persisting up to 90 days per year) and humid to semiarid rainfall regimes. They support savanna woodland and hardwood forests, and are largely found on calcareous sedimentary rocks, basaltic igneous rocks, or alluvium derived from them. There is some controversy concerning the time it takes to form such soils. On clayey parent materials they form quickly (much less than 13 000 years), but it takes longer to generate such clayey soils from silty, sandy, or crystalline materials (Fitzpatrick, 1980; Buol et al., 1980).

A modern soil especially similar to the Waterval Onder paleosol has been described in the Darling Downs region of Queensland (Stace et al., 1968, p. 98, profile G; Paton, 1974). Like the Precambrian paleosol, the modern soil is unusual for soils with gilgai microrelief because it has either illite (40—50%) or randomly interstratified clay (40—50%), with kaolinite (30—40%), and quartz (10—20%), rather than smectite. It formed on a moderately well drained, alluvial fan, flanking low hills of sandstone and basalt. The modern soil has sparse ferromanganiferous concretions, but these are not developed nearly to the extent of the iron and manganese-rich laminae in the surface of the paleosol. The soil lacks calcium carbonate within the surface 30 cm, but does have calcareous concretions below that, unlike the paleosol. This may be a consequence of a climate drier (635 mm mean annual precipitation, which is arid in Holdridge's 1967, scheme) than envisaged for the paleosol. A seasonal, warm temperate climate is enjoyed by the modern soil, and is only broadly comparable to that envisaged for the paleosol. The vegetation of woodland and open grassy forest of eucalypts on the modern soil, is very different from conceivable vegetation of the paleosol.

RECONSTRUCTED PALEOENVIRONMENT

Much research on modern soils is concerned with establishing the variation of soil properties under different conditions of climate, organisms, topography, parent material and time of formation (Birkeland, 1984). Let us now look at some of these findings backwards, taking features of the Waterval Onder paleosol also found in modern soils as evidence for the environmental conditions under which it formed. Such an approach is most reasonable for geologically young paleosols (Retallack, 1983), and must be tempered with other lines of evidence and theoretical constraints, for interpreting very ancient paleosols formed under conditions quite different than encountered today.

Paleogeographic setting

The Hekpoort Basalt below the shaly sequence including the Waterval Onder clay paleosol is a regionally extensive (covering some 100 000 km^2 in the Transvaal Basin) continental plateau basalt (Button, 1975). The Hekpoort Basalt and its stratigraphic equivalent in northern Cape Province, the Ongeluk Volcanics, were extruded within a broad midcontinental basin developed on the Kaapvaal Craton and inundated by epeiric seas. This extensive epicratonic lowland of 2200 Ma ago was bordered to the north by hills of granulite and gneiss of the Limpopo fold belt, which was deformed and uplifted some 2700—2600 Ma ago (Fig. 1).

The Ongeluk Volcanics to the southwest were extruded into an epeiric sea (Grobler and Botha, 1976), but the Hekpoort Basalt was closer to the hills of the Limpopo belt and was extruded on land. Its massive flows with amygdaloidal tops probably formed a topography similar to the modern Columbia Plateau and Snake River Plain of the northwestern United States: a modern landscape well characterized by McKee (1972). This stark volcanic topography was softened somewhat by thick clayey soils developed on the basalts. These paleosols have not yet been studied in detail, because too metamorphosed or weathered in outcrops examined. These paleosols represent a major basin-wide hiatus in deposition (Button, 1975, 1979; Button and Tyler, 1981).

Sediments of the Dwaal Heuvel Formation, overlying paleosols on the basalt, formed in a variety of depositional environments. There were sandy, bedload streams to the northeast around the Limpopo source terrain, and shallow marine shales and oolitic ironstones to the southwest around Johannesburg. In the area around Waterval Onder, paleocurrent studies have revealed a major stream system draining west-southwest, parallel to the trend of the rejuvenated Limpopo fold belt (Button, 1975).

The sequence of sandstone, siltstone and shale on which the Waterval Onder clay formed, was probably also an alluvial deposit. It is an upward fining and thinning sequence. In places relict laminae of mafic composition (berthierine-rich laminae) are at low angles to each other, and may have been miniature placers in cross beds ranging in size from ripple marks to small trough cross beds. In these respects this sequence is similar to crevasse splay deposits of modern rivers (as characterized by Davis, 1983). This sequence is unlikely to have been a marine or intertidal deposit for the following reasons. Such deposits elsewhere in the Pretoria Group are more finely bedded, mineralogically mature, calcareous and stromatolite-bearing (Button and Vos, 1977). This locality is well inland (150 km) from marine rocks of the same age to the south (Button, 1975). This sequence is also unlikely to have been a lacustrine deposit, considering its inferred coarse grainsize, thick bedding, and cross bedding.

The Waterval Onder clay was a well drained soil, moderately oxidized and seasonally cracking to original depths of about 2 m. It formed a low terrace or a part of the flood basin away from the stream, where only the finest fractions of flood alluvium were deposited. Like comparable modern soils (Stace et al., 1968, p. 98; Paton, 1974), it probably sloped toward the stream at a gentle angle from adjacent scarps of basalt. Comparable modern linear or wavy gilgai ridges form perpendicular to contours on land surfaces sloping at angles of 15″ to 3° (Hallsworth et al., 1955).

Parent material

For a variety of reasons having to do with berthierine-rich laminae and paleogeography already discussed, the Waterval Onder clay is thought to have formed on the shaly upper portion of a crevasse splay deposit of a large stream. By this view, only physical and chemical modifications above a reconstructed depth of about 2 m can be regarded as produced by soil formation. The erratic variation in petrographic and chemical composition below that level is now seen as variation in original clayey and sandy beds of the parent material. Thus the original parent material of the upper portion of the profile is most closely approximated by the sample at a reconstructed depth of about 3 m (sample 8). This was probably a gray, illitic shale, with scattered sandy laminae rich in quartz, feldspar, mafic minerals (such as hypersthene), basaltic rock fragments, magnetite and ilmenite.

This is a very different concept of the parent material of this paleosol than earlier envisioned by Button (1979) or by me (Retallack and Button, manuscript cited by Holland, 1984), who thought that it formed on basalt. Paleosols formed on basalt may be widespread on top of the Hekpoort Basalt (Button and Tyler, 1981), but they have not been described in detail. If we consider the basaltic composition of relict sandy and silty beds, these basaltic soils were presumably also the main source of the shale on which the Waterval Onder clay developed. There may have been a minor felsic contribution to this alluvium from the Limpopo hills. This felsic material was more important higher within the Dwaal Heuvel Formation.

Time for formation

One consequence of my reappraisal of the parent material of this paleosol is that it now appears to have been a weakly developed soil (in the qualitative scheme of Retallack, 1984). There has been some development of mukkara structure, but much relict bedding (now berthierine-rich laminae) persists. This buried Vertisol had not yet turned over in the usual 'self plowing' behavior of such soils (Buol et al., 1980). Better developed mukkara structure has been found in post-glacial (that is younger than about 13 000 years) land surfaces in South Dakota (White and Agnew, 1968), and in Sudan (El Abedine et al., 1971). These modern soils, however, have smectite clays, more strongly swelling than illite envisaged for the paleosol.

The dark surface horizon of the paleosol, interpreted as an A horizon, extended to depths of 50 cm. In modern Vertisols, the radiocarbon age of organic matter at this depth is about 1600 years (Sharpenseel, 1973). Organic matter in modern Vertisols stabilizes soil structure and retards and localizes deep cracking of otherwise very unstable smectitic clays (Duchafour, 1982). If the lower amount of organic matter likely for the paleosol is balanced against its more stable illitic clay, then the short estimates of time for formation of A horizons and gilgai microrelief of modern Vertisols may be reasonable also for the paleosol.

Fortunately, there is another possible indicator of the time over which the paleosol formed. The iron—manganese laminae in the surface horizon (A) is here interpreted as rock varnish, like that most often seen in modern deserts. In thin section (Fig. 5), 36 of these thin layers were counted in 2 mm of vertical height. At this density, the present 40 cm thick dark surface horizon in the gilgai swale contains approximately 7200 layers. This must be approximated rather than counted because the lamination is disrupted by cracking in the lower part of this surface horizon. The rate at which a complete coat of rock varnish forms in modern environments varies enormously: from less than 10 000 years in deserts, less than 100 years in riverine settings, less than 40 years in periglacial environments and less than 6 months in the laboratory. In considering these rates and likely factors controlling them, Dorn and Oberlander (1982) proposed that the optimal conditions for varnish formation were high humidity and lack of biological competition against iron and manganese fixing bacteria. The level of biological competition for such terrestrial bacteria during the Precambrian is not known with certainty, but is likely to have been much lower than at present. If so then Precambrian conditions for varnish formation would have been most like those in the laboratory. If we also consider other features of the paleosol, such as evidence of clastic dikes for periodic drying, the varnished laminae could have formed on annual flood deposits. The extreme values for varnish formation only constrain the age of the paleosol from 70 Ma to 3000 years, but an estimate of about 7000 years based on the assumption of annual varnishing seems most likely, and is compatible with the degree of development of soil structure.

Paleoclimate

Clastic dikes penetrating the Waterval Onder clay could be taken as evidence for long dry periods, which were perhaps seasonal. The total amount of rainfall through the year was probably quite high, however, considering the very low amounts of CaO throughout the profile. This is much less than found near the base of the sedimentary sequence on which the paleosol developed and in the underlying basalt (Button, 1979; Holland, 1984). It is also low compared to calcareous modern aridland soils (compare analyses given by Marbut, 1935). Non-calcareous soils in the mid-continental United States are now restricted to climates wetter than about 500 mm mean annual precipitation in the cool north of Minnesota and 600 mm in the warm south of Texas (Birkeland, 1984). The Waterval Onder clay was a Vertisol, and modern soils of this kind form in climates with mean annual rainfall ranging from 180 to 1520 mm (Hallsworth et al., 1955).

Paleotemperature is more difficult to constrain. Modern Vertisols form in temperate to tropical climates, within 45° of latitude either side of the equator (Buol et al., 1980). The illitic composition of the paleosol could be taken as an indication of temperate climate, since illitic Vertisols are only known from such climates in Australia (Stace et al., 1968). Although Vertisols occur in cool climatic regions of North America, all are smectitic (Buol et al., 1980), presumably because of widespread andesitic ash in their parent materials. In subtropical Egypt and Sudan, in regions with non-ashy, sedimentary and basaltic source areas, Vertisols may contain abundant illite, but have comparable and greater amounts of smectite (El Abedine et al., 1971; El Attar and Jackson, 1973).

These modern comparisons constrain the paleoclimate of the Waterval Onder clay paleosol to the semiarid, temperate, moist to wet forest life zones of Holdridge (1967). Caution must be urged in using paleoclimatic inferences based on modern Vertisols, because these are restricted to climates where base saturation of the soil can remain high. Vertisols are currently rare in hot and wet regions because copious vegetation encourages deep weathering of clays, stable soil structure and non-fluctuating soil moisture. Under much less abundant vegetation during Precambrian time, they may have been more widespread in humid regions. Their spread to such regions would have been checked to some extent if carbon dioxide levels in the Precambrian atmosphere were higher than at present, resulting in warmer, more acidic and more corrosive rain.

Because of these uncertainties it is useful to supplement these interpretations of the paleosol with other information from rocks at comparable stratigraphic levels within South Africa. Button (1975) pointed out that the mineralogically mature quartz arenites of the Dwaal Heuvel Formation, thought to have formed in tidal shelves and braided streams, were produced by acidic, humid weathering in a tectonically stable environment. He argued that the landscape was effectively leached of calcium, iron and manganese, which contributed to thick deposits of ironstone, iron formation, manganese ore and limestone in sediment starved parts of the shallow sea to the south. In Cape Province to the south, glaciofluvial and glaciomarine deposits underlie the Ongeluk Volcanics (Visser, 1971), which is a stratigraphic equivalent of the Hekpoort Basalt. There is only limited evidence for glaciation in the Transvaal Basin (Tankard et al., 1982; J.N.J. Visser, personal communication, 1985). Presumably it was farther from the glaciers, which may have been centered on high land in southern Botswana. South Africa was at a high paleolatitude at this time (Piper, 1976). From this perspective, it is worth considering whether the microrelief interpreted here as gilgai may instead have been permafrosted patterned ground. I think this unlikely because compared to the sandstone dikes of the paleosol, ice wedge casts are wider, more strongly tapering, and complexly nested. Ice wedge casts preserved in the rock record, some of them as old as 2300 Ma (Young and Long, 1976; Seddon and Holyoak, 1985), have a wide, strongly tapering morphology and are associated with other evidence of glaciation. All these features are lacking in the Dwaal Heuvel Formation, which overlies a deeply weathered clayey paleosol developed on the Hekpoort Basalt (Button and Tyler, 1981) and is laterally equivalent to thick stromatolitic dolomites to the southwest (Button, 1975). Such soils and carbonates are not found in modern frigid climates, so it is likely that climate was more or less temperate by the time the Dwaal Heuvel Formation began to accumulate.

Atmospheric composition

The weakly oxidized surface horizons (high Fe_2O_3/FeO ratios in A, AC and upper C1) of the Waterval Onder clay paleosol are compatible with current dogma, in part based on paleosols (Holland, 1984), that about 2000 Ma ago the atmosphere was poorly oxygenated. This view has been opposed by those (Dimroth and Kimberley, 1976; Clemmey and Badham, 1982) who feel that the atmosphere was always oxygenic, and that interpretations to the contrary from paleosols are compromised by other factors, such as former waterlogging, impermeable clayey soil materials, soil organic matter, and metamorphism. These may be serious complications for the interpretation of some other Precambrian paleosols (Retallack, 1986), but they are well constrained in the case of the Waterval Onder clay paleosol. For a start, the upper part of this paleosol was exposed to the atmosphere, not waterlogged. Clastic dikes extend to depths of 2 m in the reconstructed soil. Below that it may have been permanently waterlogged. Some of the iron and manganese at depth could represent a gleyed zone (placic or sideritic horizon). These differences in oxidation within the profile are reflected in Fe_2O_3/FeO ratios, which are highest in the horizon (lowest A and AC) with the most open permeable structure. The ratio is very low at the surface, perhaps because of the reducing effect of organic matter once there, or because of burial by alluvial sands. It is anomalously high near the base of the profile, but this can be attributed to Cenozoic oxidation, since this part of the profile is exposed further toward the reddish outer edge of the roadcut. A second consideration is the impermeability of such a clayey soil to atmospheric gases. The preservation of soil structure and the correlative degree of oxidation (Fe_2O_3/FeO ratio) are indications that this texture did not prevent aeration of this paleosol. A third consideration is the reducing effect of soil organic matter, which is known to form stable dark coatings on soil structure (peds) of some modern Vertisols (Pellic Vertisols of F.A.O.: Duchafour, 1982). The Waterval Onder paleosol was not of this kind, and shows much coarser soil structure than most modern Vertisols. This was probably because of very limited influence of organic matter in subsurface (C1 and below) horizons and in gilgai ridges, which in any case have a surprisingly warm hue (2.5Y now, and perhaps more yellow originally). A final consideration is the reducing effect of burial metamorphism (Curtis, 1985), which in this case was within the lower greenschist facies (Button, 1979). Ordovician and Silurian paleosols in Pennsylvania metamorphosed to well within greenschist facies, have remained strongly oxidized and are dark red (Retallack, 1985). In the case of the Waterval Onder paleosol, it is difficult to understand why the degree of oxidation (Fe_2O_3/FeO ratio) would correlate with the openness of soil structure crushed early during burial, if the soil were reduced during metamorphism. Although soil and sedimentary fabrics have been somewhat obscured by metamorphism, their persistence is evidence against pervasive alteration.

Within the perspective provided by other Precambrian paleosols, the Waterval Onder paleosol is also compatible with the view that the hypothetical warm, acidic, CO_2 rich atmosphere of the early Earth was beginning to change to one more like the present atmosphere at the time it formed. Like all paleosols of its antiquity and older (Retallack, 1986) the Waterval Onder paleosol was very weakly calcareous. It was not, however, acidic in reaction as this would be incompatible with the slight surficial increase in CaO (Birkeland, 1984), gilgai microrelief (Buol et al., 1980) and rock varnish (Dorn and Oberlander, 1981). By contrast, older Precambrian paleosols are often deeply weathered (Holland, 1984; Retallack, 1986). Calcareous paleosols, which would have been quite alkaline (pH 8 or more) are known only as old as 1900 Ma (Campbell and Cecile, 1981; age from Easton, 1981), and silcretes of the kind now formed in deserts only as old as 1800 Ma (Ross and Chiarenzelli, 1985). Very high levels of CO_2 would also induce a warmer environment than envisaged for the Waterval Onder clay paleosol. Perhaps it is significant that the oldest evidence of glaciation is bracketed in age between about 2500 and 2000 Ma (Walker et al., 1983), and includes glacial rocks underlying volcanics laterally equivalent to the Hekpoort Basalt.

Precise quantification of these insights remains difficult, and perhaps dangerous. Holland (1984) has devised an ingenious simplified model of weathering which can be used to approximate the ratio of O_2/CO_2 in the primitive atmosphere by calculating the demand for these gases from the chemical composition of the parent material of paleosols and comparing the degree of oxidation caused by ancient weathering. After Button's (1979) and also my own (Retallack and Button manuscript cited by Holland, 1984) earlier opinion, Holland assumed in his calculations that the Waterval Onder paleosol formed on Hekpoort Basalt. Since the oxygen to carbon dioxide demand of the basalt would have been high ($R \doteqdot 4.5 \times 10^{-2}$) and the surface of the paleosol lost a good deal of its iron (33%) compared to fresh basalt, he concluded that the paleosol formed in an atmosphere in which the ratio of oxygen to carbon dioxide was very low ($O_2/CO_2 \leqslant 1.3 \pm 5$), many hundreds of times less than in the present atmosphere (where $O_2/CO_2 \doteqdot 600$). Since the Waterval Onder paleosol is now thought to have formed on sandstone, siltstone and shale rather than basalt, this result can only be considered reliable to the extent that the shale reflects the composition of the soils developed on basalt of its source terrain. This may be confirmed when a suitable example of a paleosol on Hekpoort Basalt is located for detailed study. Nevertheless, weathering of the Waterval Onder clay must be reconsidered. The sample (number 8) above the compositional banding interpreted as relict sandy beds, can be taken as closest to the likely parent material of the upper part of the profile. As reconstructed here the paleosol gained iron (22%) at the surface, but the oxygen demand of the parent material was low ($R \doteqdot 2.96 \times 10^{-2}$). It may have been somewhat higher than this if K_2O was added during diagenesis from elsewhere, rather than just locally exchanged from orthoclase to sericite, as already argued. This value is just below that determined by Holland ($R \doteqdot 3.0 \times 10^{-2}$) from other paleosols as a threshold value for the transition between iron loss and iron gain in paleosols some 3000 to 2000 Ma ago. Since the Waterval Onder clay paleosol was moderately well drained and open to atmospheric resupply (the less well constrained weathering model 2 of Holland, 1984), and in addition may have been colonized by iron and manganese fixing bacteria (as outlined in the next section), its gain in iron is not compelling evidence for a particular atmospheric ratio of O_2/CO_2. Nevertheless, its reinterpreted weathering behavior is consistent with Holland's (1984) general conclusions, based on other paleosols and the survival of detrital uraninite, that at about this time the partial pressure of oxygen in the atmosphere was at least lower than 1×10^{-2} atm (0.06 times present atmospheric level) and is likely to have been more than about 1×10^{-3} atm (0.006 present).

The level of atmospheric CO_2 is even less well constrained. The shaly parent material of the Waterval Onder paleosol has lost 1.9 mol kg^{-1} of acid titratable bases (MgO, CaO, K_2O, and Na_2O corrected by the method of Holland, 1984, p. 146) compared to its presumed source terrain of Hekpoort Basalt. This figure would have been different if there was appreciable diagenetic gain of K_2O from elsewhere, or a contribution of felsic material from the Limpopo hills: both effects already argued to be minor. Within the Waterval Onder paleosol, the top of the profile compared to its presumed parent material (sample 8) has lost only 0.06 mol kg^{-1} of titratable bases. This milder degree of acidic weathering reflects its less mafic parent material, second cycle weathering and likely shorter time of formation. The result for the shale, however, is close to the average for both Precambrian and Phanerozoic sediments, including those derived from such mafic materials (Garrels and MacKenzie, 1971; Holland, 1984). Since weathering through several cycles of erosion and high soil carbon dioxide levels produced by vascular plant roots and microbial decomposition of organic matter were much more significant to Phanerozoic than Precambrian weathering, a major source of weathering acid was needed during the Precambrian. The minerals and mineral weathering sequence in this and other Precambrian paleosols are not so unusual that strong organic, nitric, hydrochloric or sulphuric acids are likely to have been abundant. The acid was probably carbonic, as it is today, produced primarily by dissolution of atmospheric carbon dioxide. Thus, the CO_2 content of the atmosphere during the Precambrian was probably much greater than the present level (3.5×10^{-4} atm). It was also less than the current maximum mean growing season soil CO_2 content (3.7×10^{-2} atm or 110 times present atmospheric level). If atmospheric CO_2 were this abundant, then average base depletion of Precambrian sedimentary rocks would be greater than for modern sediments, since this high value of CO_2 is only found now in limited areas of the world under humid forests (Brook et al., 1983). Even at this maximum value, the greenhouse effect of this gas, countered by decreased solar luminosity 2000 Ma ago, would not have produced much higher than present mean global temperatures (Owen et al., 1979), thus allowing temperate climate envisaged for the Waterval Onder clay paleosol and glaciation in underlying rocks. At this maximum likely concentration, carbon dioxide would still have been a minor component of the atmosphere, much less abundant than oxygen in the present atmosphere.

Oxygen and carbon dioxide have been of concern because they are important to both chemical and biological weathering. However, even if both were present at the unlikely maximum values considered here, then this leaves unexplained most of the primeval atmosphere (0.95 atm). Holland (1962) proposed on theoretical grounds that most of the atmosphere 2200 Ma ago was nitrogen, as it is at present. This is a relatively unreactive gas during terrestrial weathering, and its presence, along with other inert gases, would not be expressed in the mineralogical or chemical composition of paleosols. The existence of atmospheric nitrogen during Precambrian times (at least as far back as 2500 Ma) can be inferred from isotopic evidence of nitrogen fixation in marine organic matter (Schidlowski et al., 1983). Such studies should also be made of paleosols, but they would have to be much more rich in organic matter than the Waterval Onder clay.

Was there life in the soil?

There are three lines of indirect evidence that the Waterval Onder clay supported some kind of microbial life: stable soil structure, features similar to rock varnish and surficial gains in trace elements usually associated with organic matter. A fourth possible line of evidence, analysis for organic carbon (generously performed by R.J. Dorn, University of California, Los Angeles) proved negative. This does not mean that it was not once there. Formerly well drained Quaternary paleosols contain much less organic matter than equivalent modern soils (Stevenson, 1969). Negligible amounts would be expected in well-drained, moderately oxidized Precambrian paleosols (Retallack, 1984) in which little was originally present.

A striking feature of the Waterval Onder clay is the very different soil structure of its surface (A and AC) compared to subsurface (C1) horizons. The blocky structure of horizon AC and the platy structure with scattered crumb peds of horizon A are much less well developed than the granular surface horizon (mollic epipedon) of modern grassland soils (Mollisols of U.S.D.A., or Chernozems of F.A.O. and C.S.I.R.O. classifications). On the other hand, the observed structure is more complex than that associated with microbial mats in modern deserts (discussed by Campbell, 1979; Krumbein and Giele, 1979; Wright, 1985). Soil structure of the paleosol is of the kind associated with organic parts of soils: called 'tilth' in the common parlance of agriculture and horticulture. Modern soil structure, in general, is stabilized by coatings (cutans) containing polysaccharides, iron oxides and other materials (Griffiths, 1965; Foster, 1981). By contrast, lower parts of the paleosol have a crude blocky or prismatic structure, of the kind seen in desert badlands and brick pits.

A characteristic feature of properly oriented petrographic thin sections of the upper (A) horizon of this paleosol is the way in which the upper surfaces of small clay granules (crumb peds) are more heavily encrusted with botryoidal to lamellar, opaque oxides of iron and manganese, than the lower surfaces of the clasts (Fig. 5). This is a characteristic feature of modern rock varnish, best known in deserts, and thought to be produced by the activity of bacteria, algae and fungi. While it is generally agreed that rock varnish forms biologically, opinions differ on the nature of organisms involved. Dorn and Oberlander (1981) thought that iron and manganese fixing bacteria such as *Pedomicrobium* and *Metallogenium* were mainly responsible. *Metallogenium* may have a fossil record, since it is identical to microfossils called *Eoastrion* from the 2000 Ma old Gunflint Chert of northwestern Ontario (Barghoorn and Tyler, 1965). Unfortunately, there is some question whether these curious modern stellate objects are crystals, organisms, colonies of organisms, or merely traces of their activity (Nealson, 1983; Margulis et al., 1983). A variety of microbes, including cyanobacteria, fungi and bacteria, currently live on and may be responsible for rock varnish (Dorn and Oberlander, 1981; Krumbein and Jens, 1981; Staley et al., 1982).

A final line of evidence for life in the Waterval Onder clay is the surficial enrichment in trace elements commonly complexed by organic matter: such as Ba, Cr, Cu, Ni, and Zn (Fig. 13: analyses cited by Button, 1979, and Holland, 1984). Most of these elements are also enriched in clayey parts of soils (Aubert and Pinta, 1977), but this is not a consideration for the upper part of the Waterval Onder paleosol, which is uniformly clayey. These enrichments are especially striking by comparison with the immobility or depletion of other elements analyzed, such as Zr, Rb and Sr, which are also distributed as they would be in modern soils (Aubert and Pinta, 1977). Also compatible with life in the paleosol is the surficial enrichment of phosphorus (Fig. 12). In modern soils this element is concentrated very near the surface, and to a lesser degree in a subsurface horizon, and most strongly depleted between these two levels (Smeck, 1973) just as in the Waterval Onder paleosol. Some modern forested soils provide an exception that proves the rule. Their phosphorus-depleted surface mineral horizons are below (or used to be) standing biomass rich in phosphorus (Smeck, 1973). The pattern of redistribution of phosphorus in the Waterval Onder clay was presumably more like that of modern grassland soils, in which most of the biomass was within the ground.

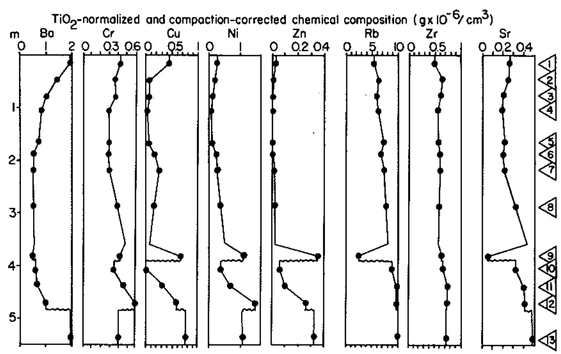

Fig. 13. Reconstructed trace element composition of the Waterval Onder clay paleosol.

The likelihood of life on land further clouds the already murky problems of assessing Precambrian biological productivity, terrestrial weathering, paleoclimate and atmospheric composition. In the Waterval Onder clay paleosol, microbial productivity and biomass was probably very low, because the rock varnish observed is very thin, the soil structure very crude, and its inhabitants were frequently (perhaps annually) either desiccated into inactivity or covered with sediment. At present, the average global productivity of ecosystems of extreme desert is only about 1.5 g m^{-2} $year^{-1}$ of organic carbon compared to 315 g m^{-2} $year^{-1}$ for savanna (forming on modern Inceptisols and Vertisols) and 900 g m^{-2} $year^{-1}$ for tropical rain forest (Whittaker and Likens, 1973). The low value is an unlikely maximum for the paleosol, since modern extreme deserts include some organisms larger than envisaged for the paleosol. Considering this low value, microbial contributions of carbon dioxide and other acidifying materials are likely to have been negligible compared to the atmospheric contribution. Microbes may, on the other hand, have been important in oxidation of iron and manganese, and in maintaining a certain amount of stability, in what would otherwise have been deeply eroded, clayey badlands. For other paleosols, particularly deeper, more sandy profiles formed over longer periods of time at major unconformities, such as the undescribed profile on top of the Hekpoort and Ongeluk volcanics, microbes may have been much more important chemically and physically.

The presence of life in the soil 2200 Ma ago may seem surprising from the perspective of Precambrian paleobiology, which has dealt largely with evidence from marine to intertidal rocks (Schopf, 1983; Knoll, 1985). There is, however, a growing body of indirect evidence for life in other Precambrian paleosols, and plausible theoretical arguments that life originated within soils in a 'primeval sludge' rather than a 'primeval soup' of primitive oceans (Retallack et al., 1984; Wright, 1985; Retallack, 1986).

CONCLUSIONS

I have been excited about this paleosol from the moment Andrew Button first brought it to my attention because of all the Precambrian paleosols known to me, this one is most like a modern soil. It is the oldest paleosol currently recognized away from a major unconformity, known to have its uppermost horizons preserved and identifiable within a modern soil taxonomy (Retallack, 1986). Here at last is a paleosol to which our hard won knowledge of modern soils could be applied with some justice. Here is a paleosol which can be compared in some detail with a modern soil. Indeed, there are modern soils similar in many respects (Stace et al., 1968, p. 98, profile G).

It has been a disappointment then, that many of my interpretations of this paleosol are of necessity so tentative. This is best appreciated by considering the genesis of one of the most obvious features of the paleosol, its mukkara structure. In modern soils, this structure is limited to soils of dry climates (less than 1520 mm mean annual precipitation), in which cation-rich clays can persist at neutral to alkaline pH. In modern soils also, it is tempered and localized by fluctuations in moisture, by soil organisms and by organic matter. Before evolution of abundant vascular land plants now characteristic of humid climates, mukkara structure may have formed in climates much wetter than at present. On the other hand, some of the modern weathering effects of vascular land plants could have been achieved if carbon dioxide were more abundant in the atmosphere, giving rainwater appreciably more acidic and corrosive characteristics than now. The likelihood of such homeostatic effects in the history of weathering on Earth is discomfiting, because it implies that similar results could be obtained by a balance of factors very different than exists now. For these reasons, I have only been able to place the broadest of constraints on dynamic aspects of Precambrian soil formation, such as atmospheric composition, climate and effects of soil microbes. Interpretation of these constraints must penetrate more deeply than simple modern comparison, into theoretical understanding of modern weathering, as in the simplified model used by Holland (1984). It is likely that future progress will rely increasingly on experimental and computer simulation of weathering under different variables.

For the moment, however, the following paleoenvironmental conditions appear likely for the Waterval Onder clay. Climate was probably semiarid to subhumid (500—1500 mm mean annual rainfall, but with high evaporation), frequently dry (perhaps seasonally) and temperate in general temperature range. There were low concentrations of both oxygen ($10^{-3} < pO_2 < 10^{-2}$ atm, or $0.06 < pO_2 < 0.006$ times present atmospheric level) and carbon dioxide ($10^{-3.5} < pCO_2 < 10^{-1.4}$ atm, or $1 < pCO_2 < 110$ times

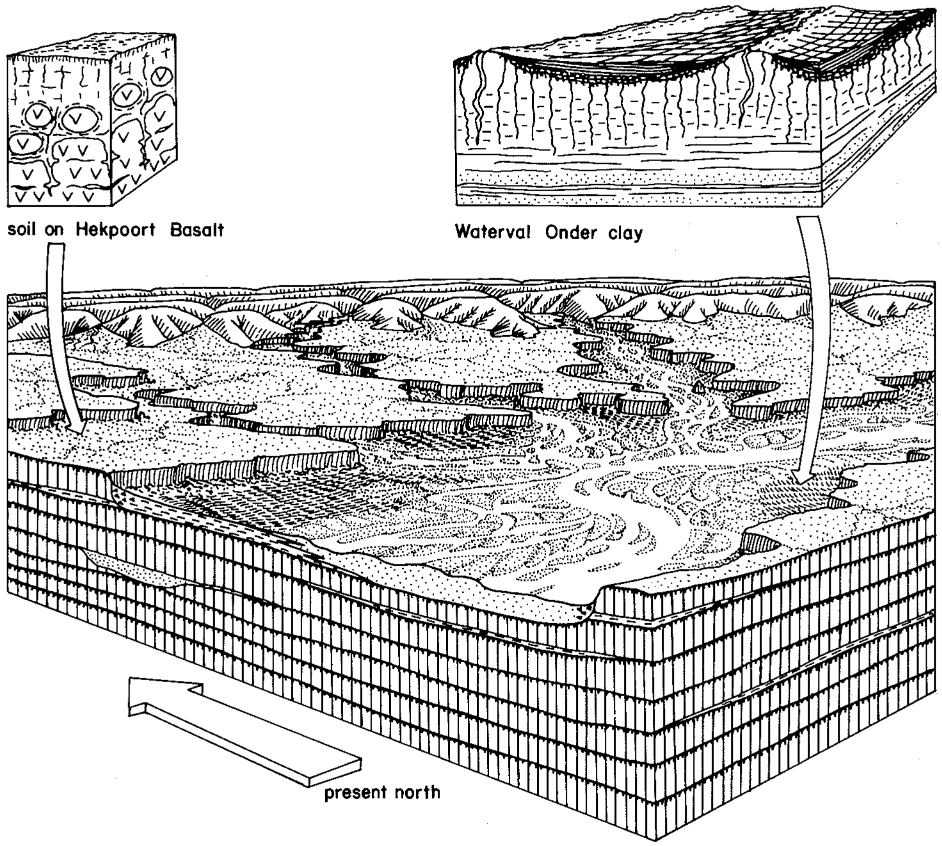

Fig. 14. Reconstructed paleoenvironment of the Waterval Onder clay paleosol (right) and the undescribed paleosol formed on Hekpoort Basalt (left) in southern Africa, some 2200 Ma ago.

present level). The soil was probably colonized by microbial communities, including iron and manganese fixing bacteria. These were effective in mitigating erosion of the soil and in accumulating phosphorus and a variety of trace elements, but their productivity (less than 1.5 g m^{-2} $year^{-1}$ organic carbon) was much less than that of vegetation in comparable modern soils (about 315 g m^{-2} $year^{-1}$). These tentative conclusions will have served their purpose well if they stimulate the additional studies of Precambrian paleosols and modeling of Precambrian environments that are so badly needed.

Dynamic aspects of Precambrian environments remain poorly constrained, but this study does provide a reasonable impression of at least one Precambrian terrestrial environment (Fig. 14). It was not entirely a barren, rocky, volcanic landscape, nor a flat sandy braid plain of ephemeral streams. The harsh outlines of extensive plateau basalts were softened by the development of clayey soils. On gently sloping young terraces of floodplains flanking active stream courses, the clayey soil was patterned into linear and wavy gilgai. The light colored, yellowish ridges contrasted with purple-brown swales. In the distance (toward present north) were low hills of granite and gneiss, and in the other direction (presently southwest) broad tidal flats of a shallow epicontinental sea.

ACKNOWLEDGMENTS

Andrew Button's generosity in lending rock specimens and thin sections, and in discussion, were instrumental in kindling my interest in this paleosol. I am also grateful for discussion with H.D. Holland, N. Tyler, D.E. Grandstaff, and M.M. Kimberley, and to the stimulating and friendly environment provided by two meetings of I.G.C.P. Project 157 'Early organic evolution and mineral and energy resources'. B. Tegge carefully checked some of my calculations, A.R. McBirney generously allowed use of his C.I.P.W. norm program, and R.J. Dorn attempted to analyze for organic carbon from the profile. The manuscript was greatly improved following careful reviews by S.W. Buol, N. Tyler, A.H. Knoll and H.D. Holland. My trip to South Africa was made easier by funding from the Wenner Gren Foundation for Anthropological Research to visit Kenya. Manuscript production has been assisted by N.S.F. Grant EAR 8503232.

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