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THE SABALOKA IGNEOUS COMPLEX, SUDAN

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[Pullout map]

CONTENTS

Sabaloka is one of the best exposed and most accessible of a large number of Younger Granite complexes in Sudan. These complexes have close affinities with the Younger Granites of western Africa and like them range widely in age. Sabaloka itself probably dates from the Proterozoic or early Palaeozoic. The paper includes a detailed map and description of the complex and presents the results of 20 new whole-rock chemical analyses.

Of the two main centres at Sabaloka, the large Cauldron Complex comprises a subsided block of basement overlain by up to 2 km of volcanic rocks and circumscribed by a polygonal zone of ring-fracturing. The fracture system was intruded by a ring-dyke of porphyritic microgranite after eruption of the volcanic rocks, and at about the same time a boss of mica granite with associated tin–tungsten mineralization was injected into the subsided block. There is also gravimetric evidence of subsurface granite intrusions in both the north and south of the cauldron, but no indications of any large mass of basic rock. Nearly all of the volcanic and intrusive rocks of the Cauldron Complex are thoroughly acidic, but a thin group of basaltic lavas lies at the base of the volcanic succession and a few minor intrusions are of basic and intermediate composition. The acidic rocks include metaluminous and subaluminous types, but peralkaline rocks are either absent or very minor in amount and altered beyond recognition. Lavas dominate the lower part of the volcanic succession whereas rhyolitic ignimbrites compose most of the upper part. Of the two main episodes of subsidence which formed the cauldron the first followed upon eruption of the lavas and produced a structural basin centred on the eastern margin of the present complex. Subsequent establishment of the ring-fracture system appears to have been consequent upon an extension of the magma chambers to the north, and was accompanied by voluminous ash-flow eruptions and the formation of a caldera. The second major subsidence post-dated all the volcanic rocks still preserved, and was probably followed by resurgent doming in the north, though the evidence on this point is not conclusive. The Cauldron Complex is classified as a 'Valles type' of caldera volcano.

The much smaller Tuleih Complex lies north of the Cauldron and includes a boss of quartz-syenite and subacid granite together with a plexus of smaller intrusions which include peralkaline intermediate and acidic rocks of comenditic character. The age of these intrusions relative to the Cauldron Complex is not known.

The chemistry of these various rock types reflects in many respects their close similarity to the Younger Granite association of western Africa, although the rocks of the Cauldron Complex are somewhat poorer in soda than most analysed acidic rocks from the Nigerian Younger Granites.

1. Introduction

Sabaloka is one of a large group of anorogenic igneous complexes in Sudan characterized by predominance of acidic rocks, shallow level of emplacement, the common presence of ring structures, and a tendency to include peralkaline variants. These complexes are conveniently referred to as 'Younger Granites' and have a great deal in common with the better known association of the same name in western Africa (Jacobson, Macleod & Black 1958; Black & Girod 1970; Bowden & Turner 1974). Their wide distribution in Sudan was recognized by Delany (1955, 1958) and has been discussed more recently by Vail (1972, 1973, 1976). Vail includes in his accounts of the Sudanese ring-complexes several which contain felsic rocks bearing feldspathoids. Such rocks are absent from Younger Granite complexes elsewhere in northern Africa and Arabia and thus belong to a different rock association. Excluding them, the Younger Granites are confined to the northern two-thirds of the country, where about 70 complexes have been located. Sabaloka is one of the most accessible since it lies astride the Nile at the Sixth Cataract, only 80 km north of Khartoum. The complex is noteworthy for the wide array of volcanic rocks preserved within a cauldron subsidence.

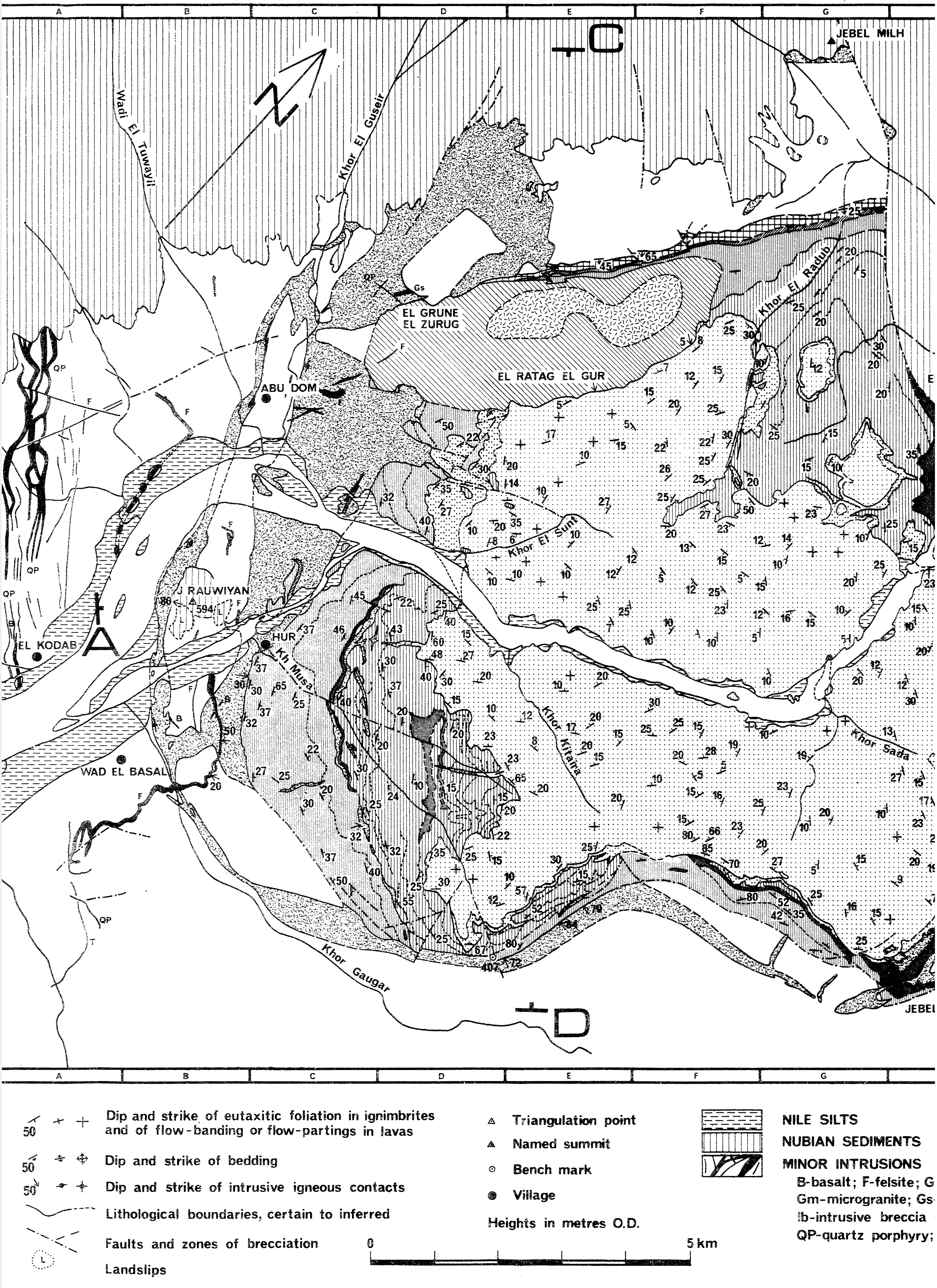
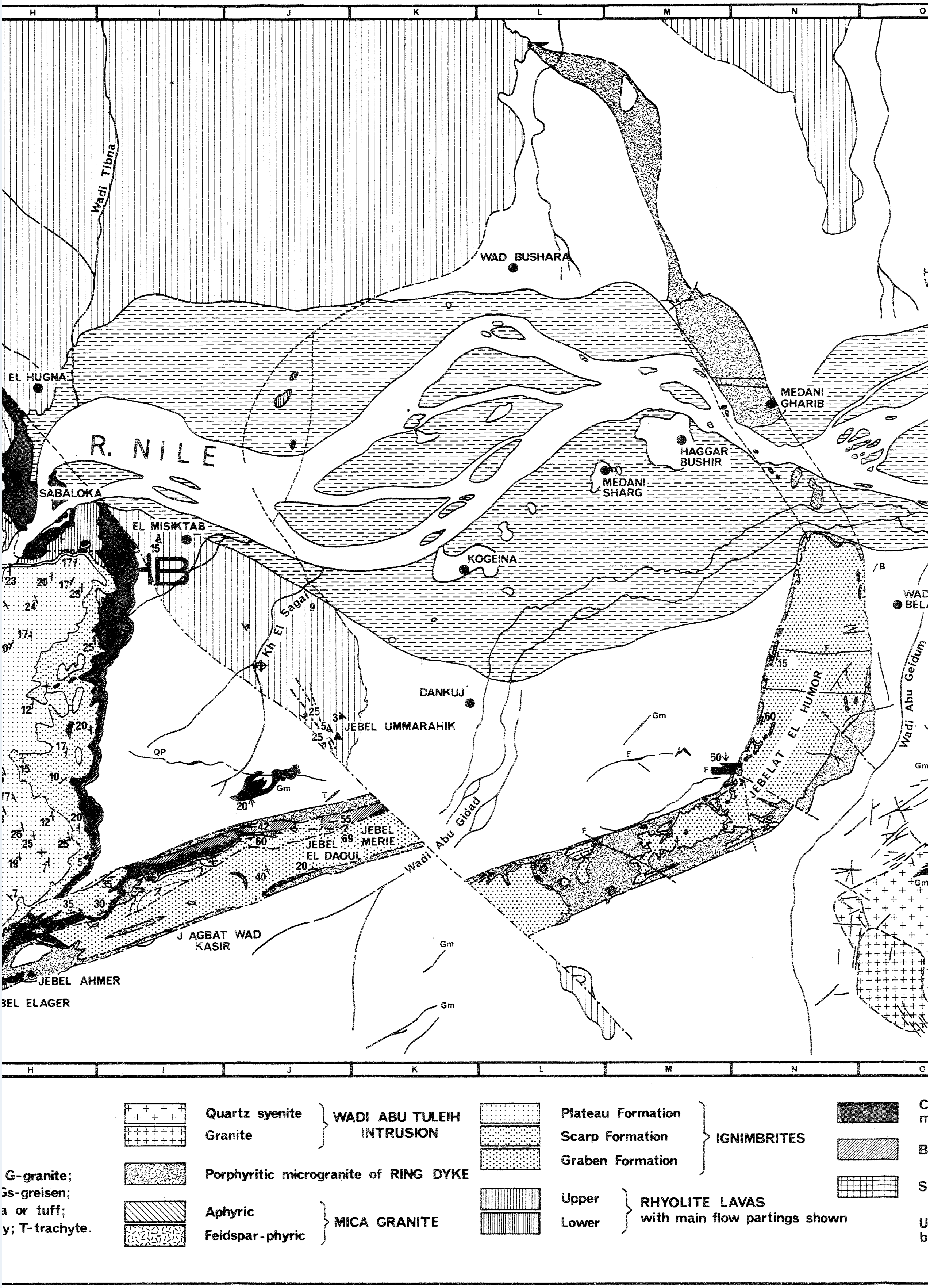
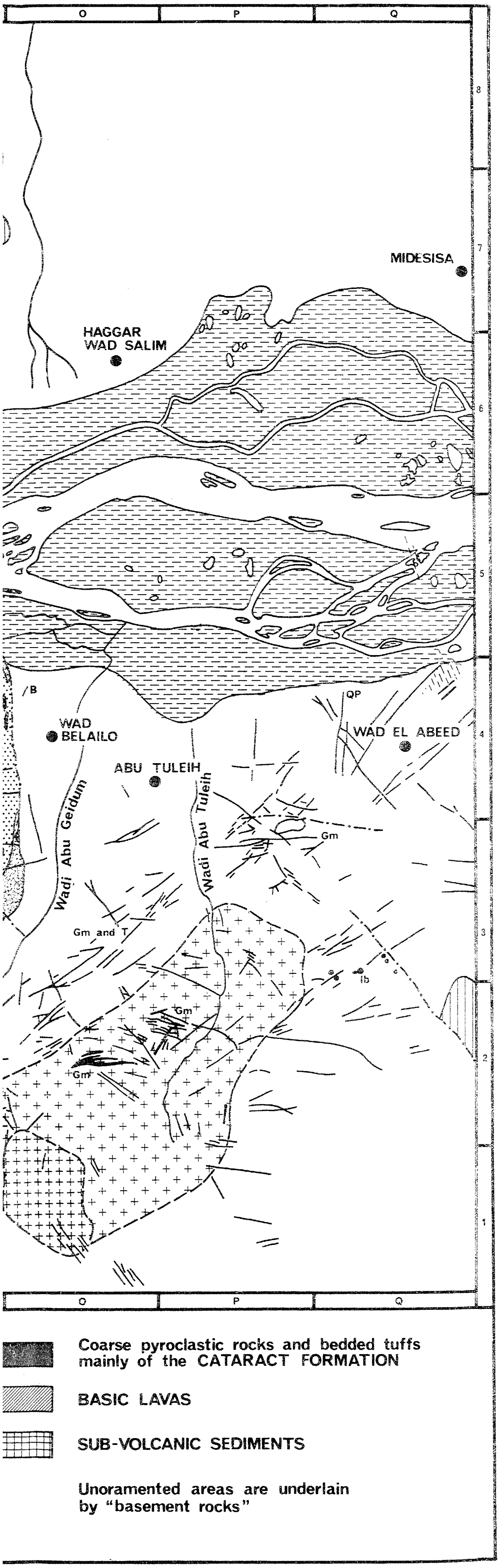


FIGURE 1. Geological



al map of the Sabaloka igneous complex. A–B and C–D mark the lines of section shown in figure 4.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ABU TULEIH
Wadi Abu Geidum
Wadi Abu Tuleih
Gm
Gm and T
Gm
Gm
Ib
Coarse pyroclastic rocks and bedded tuffs mainly of the CATARACT FORMATION
BASIC LAVAS
SUB-VOLCANIC SEDIMENTS
Unoramented areas are underlain by "basement rocks"

(*Facing p.* 597)

Despite a remarkable constancy of lithology and structure the Younger Granites of northern Africa and Arabia range widely in age, spanning Proterozoic to early Tertiary, or later, times. Dates are most firmly established in western Africa, where there is said to be an irregular but progressive southward decrease in age from over 500 Ma in the Aïr through 300 Ma in southern Niger to 150 Ma in southern Nigeria and perhaps to Lower Tertiary in Cameroun (Bowden &

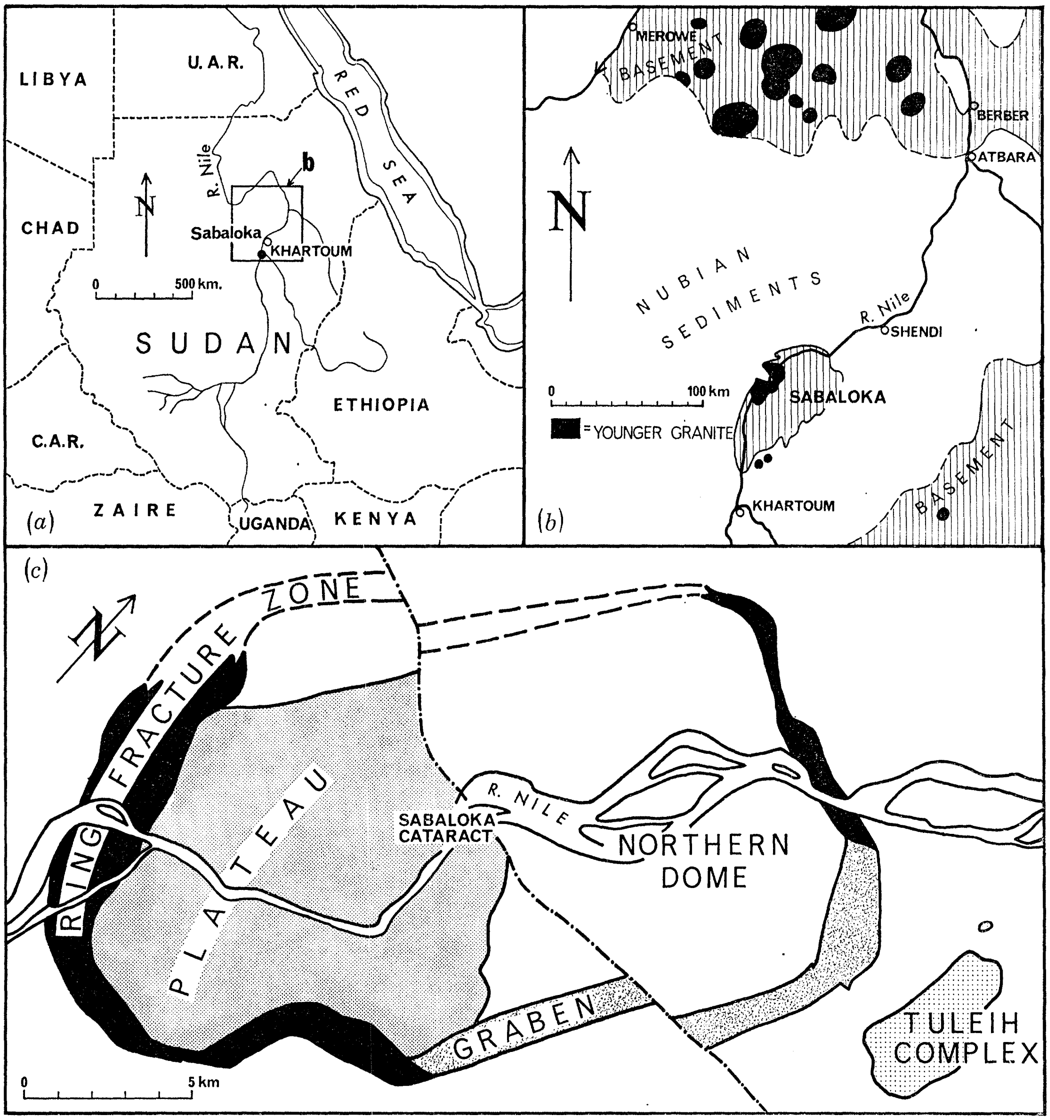

FIGURE 2. (*a*) Location of Sabaloka in the Republic of the Sudan. (*b*) Geological sketch of the region around Sabaloka based largely in Delany (1958) and Vail (1974). (*c*) Major structural units of the Sabaloka Complex. Nubian sediments and younger rocks omitted.

Turner 1974; Bowden, van Breemen, Hutchinson & Turner 1976). The age pattern is less clear in Sudan. At Jebel Uweinat, on the Libyan border, Burollet (1963) found granites intruding Lower Carboniferous sediments, and isotopic ages as young as the Eocene have been recorded (Vail 1974, 1976). On the other hand, at least some of the Sudanese Younger Granites are demonstrably older than Nubian sediments of Cretaceous age, among them Sabaloka where, moreover, two K–Ar determinations gave apparent ages near to 500 Ma for the granites

(Almond 1967). Briden (1973) favours an even greater age for Sabaloka since palaeomagnetic data on basic lavas fit best with a pole position about 700 Ma ago. The radiometric dates cannot be said to preclude this possibility since Vail & Rex (1970) found widespread indications of thermal overprinting during the Pan African Orogeny (550 ± 100 Ma) in many parts of Sudan, though there is no sign of a late metamorphism in the Sabaloka igneous rocks themselves. More evidence is needed to resolve these uncertainties; meanwhile the writer concludes that Sabaloka is likely to be of Proterozoic or Lower Palaeozoic age.

Hume (1937) was the first to mention the presence of granites at Sabaloka, but credit for the recognition of the ring structure and delineation of its main features belongs to Delany (1954, 1955, 1958). Later published work includes a general account by Whiteman (1971), descriptions of tin–tungsten mineralization and of ignimbrite feeder vents by Almond (1967, 1971), the comments on age by Vail & Rex (1970) and by Briden (1973) mentioned above, and an account of the gravity field associated with the complex by Sadig, Almond & Qureshi (1974). Khalil (1972, 1976) has provided geochemical data on the tin–tungsten mineralization.

This paper summarizes the results of a petrological and structural study made by the writer. The accompanying pullout map (figure 1) was compiled on a largely uncontrolled mosaic of aerial photographs supplied by the Sudan Survey Department. References to locations given in the text are related to the grid on the margin of this map. The chemical analyses were made by Dr E. L. P. Mercy in the Department of Geology, University of Edinburgh, and the author is very grateful for this provision. Thanks are also due to the University of Khartoum and Kingston Polytechnic for research support, and to the writer's former colleagues at Khartoum. Dr J. R. Vail and Mr D. G. A. Whitten read the original draft and made many useful comments, while the figures were redrawn with great care by Miss Linda Parry of Kingston Polytechnic.

2. General geology

In this account names of major units of the complex (figure 2) are capitalized since they will be referred to repeatedly throughout the paper.

The most prominent topographical feature at Sabaloka is a subcircular Plateau about 11 km across which rises 200 m above a surrounding pediment of gneisses and older granites. The Plateau is around 550 m a.s.l. and through it the Nile has cut a zigzag gorge which debouches at the Sabaloka Cataract itself. The river follows this course through ground higher than the surrounding pedeplain, which is at about 400 m a.s.l. to the southeast of the Plateau, a circumstance almost certainly due to superimposition of the Nile drainage from the Nubian sedimentary cover (Berry & Whiteman 1968). Nubian rocks still obscure the northwestern portion of the complex and cap Jebel Rauwiyan (594 m), a hill immediately southwest of the Plateau and the highest point in the area.

Acidic lavas and ignimbrites compose the Plateau and are deeply dissected by steep-sided valleys following master joints. To the south and west the Plateau is margined by boulder-covered hills marking the outcrop of a granitic ring-dyke intruded into a complicated Ring Fracture Zone. The ring structures pass northwestwards beneath the Nubian cover but to the northeast an arm of low hills trace out the continuation of the Ring Fracture Zone well beyond the Plateau, embracing a large area of gneissose rocks which form the core of the Northern Dome. Between the Plateau and the Nile at Haggar Bushir the Ring Fracture Zone includes a downfaulted strip of volcanic rocks and its therefore termed the Graben. The subsided block of volcanic rocks and basement within the Ring Fracture Zone forms the Sabaloka Cauldron. A large post-Nubian fault, named after Jebel Ummarahik, more or less bisects the Cauldron from east to west and causes a notable displacement of the Ring Fracture Zone.

Outside and immediately to the northeast of the Cauldron is a separate cluster of syenitic and granitic intrusions here called the Tuleih Complex. The age relations of this centre are in doubt, although the Cauldron is probably the older complex. The only peralkaline rocks found at Sabaloka are in the Tuleih area, though even here they occur in only small amount.

Table 1. Geological succession of the Sabaloka area

Superficial deposits	
Nile silts, alluvial fans, aeolian sands and lag gravels, sandy residual soils	*Cainozoic*
Unconformity	
Nubian formation	
Sandstones, grits, conglomerates, siltstones; often ferruginized or silicified	*Cretaceous*
Unconformity	
Sabaloka Igneous Complex	
Tuleih Complex	
(*m*) Minor intrusions of microgranite, microsyenite, trachyte and felsite; some peralkaline	
(*l*) Boss of microgranite	
(*k*) Quartz syenite of the main pluton	
Cauldron Complex	
(*j*) Minor intrusions of quartz-porphyry, felsite, microgranite and trachyte. Some may be of similar age to (*m*)	
(*i*) Local mineralization of (*g*) and (*h*) with tin and tungsten	
(*h*) Ring-dyke of porphyritic microgranite, partly overlapping in age with (*g*)	*Lower Palaeozoic or Proterozoic*
(*g*) Boss of mica granite	
(*f*) Plateau and Graben ignimbrite formations	
(*e*) Scarp ignimbrite formation	
(*d*) Upper rhyolite lavas	
(*c*) Cataract formation of volcaniclastic rocks	
(*b*) Lower rhyolite lavas	
(*a*) Basic lava formation	
Disconformity	
Subvolcanic sediments	
Conglomeratic sandstones, sandstones and shales	*Lower Palaeozoic or Proterozoic*
Unconformity	
Basement	
Quartzo-feldspathic gneisses, amphibolites, quartzites and batholithic intrusions of foliated and unfoliated granitic rocks	*Archaean or Proterozoic*

Table 1 summarizes the geological sequence. As in the Nigerian complexes most of the exposed intrusions are younger than the volcanic rocks so that it is possible to distinguish an earlier volcanic cycle (*a*–*f*) from a later intrusive cycle (*g*–*m*).

The term 'basement' is used in this account as a convenient designation for the plutonic metamorphic and igneous rocks into which the Sabaloka intrusions are emplaced. In Sudan the more formal term 'Basement Complex' is properly used to embrace all formations older than the Nubian (Andrew 1948) and so includes the rocks of the Sabaloka Complex itself.

The subvolcanic sediments of table 1 comprise a 40 m thick, fining-upward sequence of clastic sediments and are found only on the northwestern margin of the Plateau, where they rest unconformably on basement gneisses but are concordant in attitude with the overlying lavas. They have been strongly affected by the thermal aureole of the mica granite. The sediments may or may not have been deposited subaqueously but the volcanic succession above appears to be wholly subaerial. The base of the volcanic rocks is also well exposed along the northeastern scarp of the Plateau but here the basal lavas rest directly on gneisses. At neither of these two localities are there signs of significant relief on the subvolcanic surface. In contrast, there is considerable local relief on the sub-Nubian surface, and the underlying rocks are often weathered, as on Jebel Rauwiyan. The writer suspects that the present topography over the igneous complex is broadly as it was when first exhumed from beneath its Nubian cover, apart from limited dissection by the Nile and its tributaries.

3. Extrusive rocks

(a) *Basic lavas*

(i) *Field relations*

A thin formation of basic lavas is to be seen wherever the base of the volcanic succession is exposed, and identical material is widespread as fragments within ignimbrites and the ring-dyke. Apart from a few basaltic dykes intrusive into the basement these are the only basic rocks in the complex, but they have a wider distribution and more constant character than any other group.

The formation is well exposed in two areas. At the first, about 10–20 m of lavas rest on pre-volcanic sediments northwest of the Plateau (E6–G7). Feldspar-phyric and aphyric varieties are present. A comparable thickness of lavas is exposed along the northeastern scarp of the Plateau and here includes three flows, of which the central member is distinguished by relatively coarse grain and the presence of feldspar phenocrysts. At another locality, in the Graben (L2), the lavas are overlain by 15 m of well bedded tuffs.

It therefore seems that the centralized rhyolitic volcanicity which formed the main phase of activity at Sabaloka was superimposed on a thin but extensive field of basic lava flows with occasional pyroclastic cones.

(ii) *Petrography*

The lavas are platy-jointed rocks, black to dark-green in colour. Aphyric varieties contain flow-oriented plagioclase (An_{20-50}), about 30 % augite, opaque oxides and in some cases up to 10 % pseudomorphed olivine. The primary minerals have been more or less extensively altered to chlorite, carbonate and albite, but lavas metamorphosed by the mica granite are not affected in this way so it seems that alteration occurred at an early date. In porphyritic lavas the phenocrysts are of weakly zoned labradorite while the groundmass laths are more sodic (*ca.* An_{33}) than in most aphyric lavas, and augite is less abundant. Such rocks are transitional towards trachyte.

(b) *Rhyolite lavas*

(i) *Field relations*

Thoroughly acidic lavas make up over half the bulk of volcanic rocks preserved within the Cauldron and are particularly well exposed in the southwest of the Plateau, where they give rise to a broken topography of pale-coloured hills in distinct contrast with the less dissected ground and more sombre hues of the ignimbrite outcrops. Arcuate features reveal the regular dip of the flows towards the centre of the Plateau, orange-coloured features picking out massive lavas whereas autobrecciated flows show up grey. In the lower part of the succession some flows are brecciated throughout their thickness. The flows are separated in places by bedded tuffs and lenses of volcanic conglomerate, the most conspicuous of these horizons being up to 9 m thick and traceable for 4 km. The lava succession measures 1800 m on the southern flank of Khor Musa (C3), the base unseen and the top eroded. The lower half of this sequence is of weakly flow-banded, aphyric lavas in very thick flows and is topped by the main volcaniclastic horizon. The succeeding 450 m show stronger flow-banding and contain a few small phenocrysts of quartz and feldspar, while the uppermost 450 m is of very strongly flow-banded rocks with up to 10 % phenocrysts. Flow thickness is in the order of 130 m in the lower part of the succession and decreases to 40 m in the upper part, with about 25 flows in all.

On the north side of Khor Musa (D3) is an irregular diatreme of volcanic breccia and agglomerate which may mark a centre of lava eruption.

It is difficult to correlate the Khor Musa succession with the lavas which overlie the Cataract formation in the north of the Plateau. Mapping suggests that the main volcaniclastic horizon in Khor Musa may correlate with the Cataract formation, separating aphyric Lower rhyolites from Upper rhyolites with phenocrysts. If so, the distribution of these two formations implies that early acidic volcanicity centred in which is now the southeast of the Plateau whereas later activity enlarged the field in a northerly direction. At no time did rhyolite lavas reach into the northeast of the area.

(ii) *Petrography*

Pale-grey where least altered, the lavas are generally mottled with orange oxidation products which in autobrecciated rocks concentrate in the matrix. Thin sections reveal little, the most obvious variations being in the phenocrysts. These are entirely absent from many of the Lower rhyolites though a few sections contain rare quartz or kaolinized feldspar. All the Upper rhyolites contain sparse microphenocrysts, the quartz often conspicuously rounded, while feldspar may be euhedral or have rounded edges. Both potassium feldspar and oligoclase are normally present. Groundmass texture is often evenly cryptocrystalline, but micropoikilitic enclosure of feldspar needles in quartz often gives a patchy appearance and spherulitic and nodular variants also are found. Some rocks are further diversified by nests of relatively coarse quartz and fluorite. Chlorite, clay minerals and carbonate are common alteration products.

(*c*) *Volcaniclastic rocks of the Cataract formation*

(i) *Field relations and lithology*

These rocks are well exposed in a 6 km long section along the northeast scarp of the Plateau and are also prominent in the southern section of the Graben, though they die out northeastwards. The formation is absent from much of the Plateau west of the Nile but in the south may be represented by the volcaniclastic rocks which separate Upper and Lower rhyolite formations.

Variation along the northeastern scarp is illustrated in figure 3. On the banks of the Nile the greater part of the formation is composed of crudely bedded volcanic breccias in which a few blocks reach 2 m. Bedding is most evident near top and base, where there are also intercalations

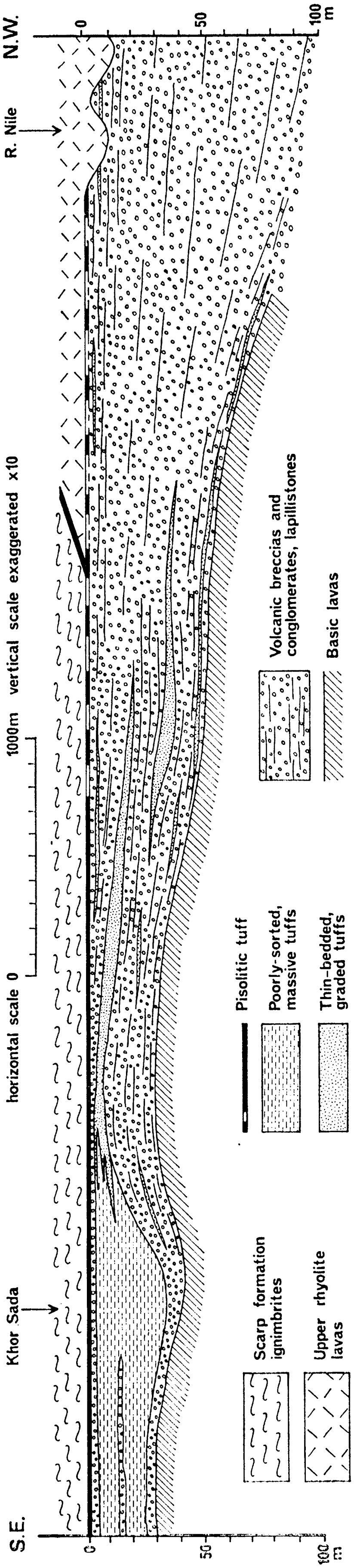

FIGURE 3. Lateral variation in the general lithology of the Cataract formation along the northeastern scarp of the Plateau.

of lapillistone and coarse tuff. Southwestwards crude bedding becomes characteristic of all the coarser rocks, while lenses of well bedded and often very well graded tuff appear and make up about one quarter of the succession in the central section. There is fair correlation between grain size and bed thickness but within-bed sorting is nevertheless poor, with lapilli and coarse ash making up about one third of the volume in even the coarsest beds. Sorting is much better in the thin-bedded tuffs, in which most beds are between 0.5 cm and 5 cm thick and grade from coarse to fine tuff but sometimes include lapilli near the base. Bedding is generally parallel but rare instances of cross-bedding and wash-out have been found near the base. A low degree of rounding is characteristic of all except the larger clasts in the volcanic conglomerates.

Fragment composition varies with grade and from bed to bed, but the large clasts (over 64 mm) are almost exclusively of rhyolite lava. Rare blocks of weakly welded ignimbrite from near the top of the formation 2 km northwest of Khor Sada testify to earlier ash-flow activity. Acidic lava clasts also dominate the lapilli grades but are often accompanied by porphyritic pumice and less commonly by basic lava. The coarse ash grades show more compositional variation; rhyolite predominates in breccia matrices and some graded tuffs whereas in most graded tuffs and lapillistones broken crystals are most abundant in this size range, and lithic clasts are pumice or perlite rather than massive rhyolite. In all these rocks feldspars are extensively decomposed and the fine ash grades are too altered to be determined.

The massive tuffs in the southeastern section of the scarp differ from all the rock types mentioned so far. These pale-green, platy-jointed rocks are well exposed in a gully 1 km southeast of Khor Sada, where two 8 m units of massive tuff are separated by 3 m of well sorted rhyolite breccia. Both units contain scattered chips of rhyolite and there is a rapid transition from the breccia into the upper tuff. Thin sections show the tuffs to be petrographically similar to ignimbrites higher in the succession, though welding is absent, and these rocks probably originated as low-temperature ash-flows.

A distinctive pisolitic tuff, 1.5 m thick, forms a useful marker for the boundary between the Cataract formation and overlying Scarp ignimbrites.

(ii) *Mode of emplacement*

In assessing the ways in which these rocks may have been deposited, three points seem particularly significant. First, apart from the massive tuffs the formation is composed of thin sheets showing marked lateral variation in grade and sorting but little sign of channelling. This feature, and proximity to the subvolcanic surface suggests deposition on low ground from wide-fronted flows. Secondly, the fine tuffs are too localized to have been deposited by air-fall eruptions, and since in places they pass by lateral or vertical gradation into the coarser rocks these, too, are unlikely candidates for such an origin. Rounding of some larger clasts does not in itself rule out an air-fall origin since rounding could have taken place by attrition in fluidized vent fillings. Thirdly, although fragments of rhyolite lava bulk large in the formation as a whole the high proportion of crystals and pumice in the finer grades is best explained by contemporary pyroclastic activity, and prohibits an explanation based wholly on epiclastic processes.

The breccias, conglomerates, lapillistones and graded tuffs are closely interrelated, and the presence of rudimentary bedding in even the coarsest units rules out deposition from subaerial ash-flows or from mud-flows of the more viscous sort (Fisher 1960; Crandell 1957). Their character is more in keeping with deposition from sheet floods or water-rich mud-flows debouching from volcanic uplands onto adjacent plains, with local sorting of the finer fractions in transient streams and small bodies of standing water. Explosive eruptions in the source area may have triggered the generative floods and contributed pyroclastic fragments, but the coarser material probably derived from the mantle of epiclastic debris. Very few fragments were picked up after the flows left the uplands, for the immediately underlying basic lavas are very sparsely represented among the clasts.

(d) *Ignimbrites*

Over much of the Plateau a thick sheet of ignimbrites rests unconformably on the rhyolite lavas and Cataract formation. There is marked angular discordance in the Plateau interior but almost none along the northeastern scarp. The lower, Scarp formation of thin, discontinuous and variable character has been distinguished from a much thicker and more uniform Plateau formation. In addition, downfaulted ignimbrites within the Graben have much in common with the upper part of the Plateau formation and may be its lateral equivalent, but since there is some doubt in making this correlation the Graben ignimbrites are treated separately.

(i) *Scarp formation*

Patchy distribution of these rocks is due to their confinement within pre-existing valleys and basins. Around Khor Sada (H3), four or five ash-flows are present and total 50–70 m in thickness, the bases of the flows often rich in rhyolite rubble. The ignimbrites are pale-grey where fresh, but alteration has often added tints of pink or green. Welding is present throughout much of their thickness. Along the southeastern margin of the Plateau the formation is generally only a few metres thick but it locally thickens into steep-sided depressions, in some of which the eutaxitic foliation is very strong and steeply inclined. These structures are believed to include both vents and valley infills (Almond 1971).

Little is seen of the Scarp ignimbrites in the Plateau interior but they reappear in force on the west bank of the river east of Abu Dom (D5) and there include welded tuffs, volcanic conglomerates and graded tuffs. Thickness is about 175 m close to the Nile but there is rapid thinning and decrease in overall grain size to the north. The even more abrupt reduction which must occur across the river to the south suggests a scarp boundary, possibly fault-controlled, while high relief within the basin is shown by the presence of rhyolite inliers within the main outcrop. The sequence includes five ash-flow units of which the lowest three have thick breccias at their bases.

The Scarp ignimbrites are distinguished petrographically by the relatively low content (*ca.* 15 %) and small size (*ca.* 1 mm) of primary crystals of quartz and feldspar. Primary ferromagnesian minerals are nearly always altered, but green biotite was found in two rocks. Pumice makes up 10–20 % of these rocks, mostly as small fiamme, while matrix shards average about 0.15 mm. Variation in degree of welding indicate that each flow cooled independently, and some were hot enough to allow strong welding near the base even though the overburden was only a few metres. Alteration is more advanced than in the other ignimbrite formations.

(ii) *Plateau formation*

These rocks rest conformably on the Scarp ignimbrites but in places overlap them to lie with angular discordance on rhyolite lavas. The base is often marked by a metre or two of lava rubble succeeded by 20 m or more of almost black tuff, above which comes a gradual change to a paler grey coloration. The black zone corresponds approximately to the zone of maximum welding and tends to show the best columnar jointing. Total thickness in the Plateau is in the order of 1000 m, but no signs of composite structure have been seen. The base of the sheet dips inwards at 25–75° but for some tens of metres above it the attitude of the eutaxitic foliation is often influenced by irregularities of the underlying surface. Accommodation structures of this kind are particularly clear where ignimbrites are banked against a rhyolite ridge in the north of the Plateau (F5–G5). In places exhumed summits of lava hills are exposed within the main outcrop (e.g. G3–G4).

Table 2. Variation (volume %) in the primary crystal contents of graben and plateau ignimbrites

	I	II	III	IV
quartz	9.3	12.7	17.2	16.8
K-feldspar	10.0	13.2	19.0	33.0
plagioclase	1.7	1.7	1.5	3.0
mafic minerals	0.6	0.6	0.5	0.8
Totals	21.6	28.2	38.2	53.6

I. Average of three specimens from near the base of the Plateau formation.
II. Average of three specimens from the top of the Plateau.
III. Average of two specimens from near the base of the Graben formation.
IV. Average of two specimens from the centre of the Graben.

Feeder vents are well exposed at two places in the south of the Plateau (D3–D4) and in both cases the base of the ignimbrites passes downwards into a northeast trending dyke about 50 m wide. At deeper levels the dykes were intruded by microgranites of the ring-dyke suite, but near surface they are filled by ignimbrite displaying parataxitic foliation parallel to the contacts and columnar joints normal to the foliation. These structures pass continuously into the shallow-dipping foliation and steep columnar joints of the overlying sheet. The ignimbrite filling of the dykes appears to have been compressed by inward collapse of the walls following eruption of the ash-flows (Almond 1971).

Table 2 shows that the content of primary crystals shows a notable increase upwards within the ignimbrite sheet, potassium feldspar and inverted high quartz predominating. The average size of these crystals (*ca.* 2 mm) is distinctly greater than in the Scarp formation. Potassium feldspar often contains clouded patches due to incipient development of microperthitic intergrowths, the value of $2V$ increasing from 45° in optically homogeneous feldspar to about 62° in clouded areas (cf. Smith 1974, p. 447). Plagioclase composition is typically near mid-andesine. Primary mafic minerals are sparse but include pale-green clinopyroxene, fayalitic olivine and brown biotite, with occasional iron oxide, zircon and allanite. Pumice forms up to 20 % of these rocks as fiamme 2–3 cm long in weakly welded rocks. Small clasts of rhyolite lava are a constant minor constituent while fragments of basic lava are sparse but widespread. Granodiorite xenoliths have a localized distribution in the north of the Plateau, particular nearly the western margin, where some exceed 50 cm. Vertical variations in devitrification textures are similar to those in the Graben ignimbrites, described below.

(iii) *Graben formation*

The outcrop of these rocks is marked by dark-coloured, rounded hills which extend from Jebel Ahmer (H1) in the south to near Wad Belailo (N4) in the north. North of Jebel Ahmer the base is seen to rest on volcanic conglomerates of the Cataract formation. Eutaxitic foliations are generally only obvious near the margins of the Graben, but enough have been measured to suggest that the structure is broadly synclinal.

The Graben ignimbrites are petrographically like those of the Plateau formation but contain even more primary crystals, the proportion of feldspar increasing upwards (table 2). The character of the devitrification textures also changes in this direction. In the lower, visibly eutaxitic horizons the groundmass is largely cryptocrystalline, though shard form is preserved by axiolitic devitrification and streaks of collapsed pumice are picked out by slightly coarser grain. A few micrographitic patches also occur, often within pumice. Above the basal zone devitrification becomes evenly microcrystalline, with indistinct shards and pumice no longer visible. At still higher levels the texture coarsens and shards disappear entirely. The groundmass here consists of alkali feldspar enclosing blebs of quartz, together with a little plagioclase, biotite and greenish-brown amphibole, also poikilitic. In the coarsest patches the place of amphibole is taken by pale-green clinopyroxene. Textures and mineralogy of these rocks are reminiscent of those in quartzo-feldspathic hornfels and must be the result of high temperature autometamorphism.

(iv) *Correlation of Plateau and Graben formations*

The Graben formation rests directly on Cataract volcanic conglomerates without intervening Scarp ignimbrites although their petrography invites correlation with the upper part of the Plateau formation. One possible explanation of this apparent anomaly in sequence is that before eruption of the ignimbrites the Graben region was uplifted by early caldera faulting or flexuring, so that the ash-flows overlapped onto high ground and only the uppermost part of the Plateau sheet covered it completely. Alternatively the Graben formation might be a quite separate, early formation confined within a proto-Graben depression, but this explanation does not take account of the close petrographic similarities between Plateau and Graben rock types.

4. Intrusive rocks of the Cauldron Complex

Intrusions linked with the evolution of the Cauldron include the ring-dyke, a small mass of mica granite in the northwest of the Plateau, and a number of minor intrusions. The cluster of bosses and dykes northeast of the Cauldron is described later (§ 5).

(*a*) *Mica granite*

(i) *Field relations*

This is the only major intrusion exposed within the subsided block of the Cauldron. Its oval outcrop measures 5 km by 2 km and forms low-lying country diversified by small tors. Age relations with the ring-dyke are not straightforward. Outcrop shape of the dyke and a chilled contact against the western end of the granite (C6) make it seem that the dyke is a wholly younger intrusion, but at El Grune El Zurug (D6) an early injection of ring-dyke was affected by mineralization associated with the mica granite before emplacement of a later pulse of similar magma. Evidently the two intrusions overlapped in time.

The granite comprises an older, fine-grained and porphyritic mass enclosed within a younger and generally coarser aphyric intrusion. Xenoliths of porphyritic type are widespread within the coarser granite and aplite veins cut both rock types. The outer margin of the coarser granite is moderately chilled and locally porphyritic. Pegmatitic patches, miarolitic cavities, quartz pods and veins occur sparingly in the vicinity of the contact, where they are associated with alteration and a weak tin–tungsten mineralization of the granite margin and adjacent country rocks. Veins of granite penetrate the adjacent volcanic rocks at several points and shallow-dipping contacts can be found along the southeastern margin. This feature and the apparent conformity of the northwestern contact with southeast dipping lavas initially suggest that the intrusion is lensoid and inclined towards the centre of the Plateau. Two objections to this concept are worth noting, the first being that gravimetric data fit best with a steep northwestern contact (Sadig *et al.* 1974) and the second, that pegmatites, veins and mineralization would be expected to concentrate along an outward dipping contact and yet are most in evidence along the northwestern margin.

The granite has a strong metamorphic aureole, in contrast to the ring-dyke, a feature doubtless due to a higher volatile content in the magma and in keeping with the presence of pegmatites and mineralization.

(ii) *Petrography*

The mica granites are buff-coloured, highly leucocratic rocks. The porphyritic variety contains approximately equal amounts of quartz, microperthite and albitic plagioclase, with some 5 % biotite and a little iron oxide and muscovite. Although the rock contains about 15 % phenocrysts only the large microperthites (up to 15 mm) are conspicuous in hand specimen, though close examination reveals smaller phenocrysts of quartz and plagioclase. The groundmass varies irregularly between micrgranitic and rudimentarily microgranophyric, with an average grain of about 0.5 mm. Plagioclase composition is usually near An_{10} but in one rock it is zoned from An_{22} to An_{10}. The brown to greenish-brown biotite has been partly replaced by chlorite, and muscovite is interstitial. Apart from a few grains of magnetite the most common accessory is colourless to purple-spotted fluorite. Minute grains of zircon and allanite are found within biotite.

The outer granite is typically hypidiomorphic granular with an average grain size over 1 mm, but some specimens are finer and tend towards granophyric intergrowth. Constituents are similar to those in the inner granite except that the chief mica is a pale-yellow, weakly pleochroic variety thought to be a leached biotite. It has n_β 1.595, $2V_\alpha$ 5° and is closely associated with aggregates of hydrous iron oxide. These properties are similar to those of a grey zinnwaldite found in greisens related to the tin–tungsten mineralization and also found sparsely in one granite specimen, accompanying the yellow mica. A little muscovite also occurs in these rocks.

The aplites are equigranular and contain quartz, subhedral plagioclase and only minor amounts of potassium feldspar and muscovite.

A number of authors have emphasized the rôle of late-stage albitization and related hydrothermal alteration in modifying the geochemistry and textures of some Nigerian Younger Granites (see, for example, Bowden & Turner 1974), and have suggested a relation between these processes and tin mineralization. It is therefore relevant to remark that the petrography of the outer granite contains indications of late-stage modification, but there is no clear evidence that this included albitization. The discrete and often zoned subhedral plagioclases are taken to be of magmatic crystallization, but it is possible that some crude perthitic intergrowths are partly of replacement origin. The limited chemical data (table 5) reveal little difference between the two types of granite.

Specimens of metamorphosed sediments and lavas collected within 200 m of the northwestern contact of the granite contain the following mineral assemblages:

Pelitic	Cordierite – quartz – magnetite – muscovite
Semi-pelitic	{Quartz – muscovite – biotite Quartz – muscovite – andalusite
Quartzo-feldspathic	Quartz – K-feldspar – Na-plagioclase – muscovite ± biotite ± pale-green garnet
Basic	Andesine (or labradorite) – actinolite – magnetite.

The most critical assemblage is the last. In it, the acicular habit, pale-green colour and small extinction angle (*ca.* 10°) of the amphibole indicate actinolite rather than the hornblende one might expect, but association of actinolite with calcic plagioclase is found in some other contact aureoles (Miyashiro 1973, p. 288). For instance, Seki (1957) found an actinolite-calcic plagioclase zone between greenschist (albite-epidote amphibolite) and amphibolite (hornblende hornfels) facies rocks in the Arisu area of Japan.

The outer limit of the aureole has not been defined, but in gneiss collected 400 m from the northwest contact decussate mats of biotite and chlorite replace original ferromagnesian constituents, plagioclase is altered to white mica and quartz recrystallized, though potassium feldspar is not visibly affected. In ignimbrite 300 m from the southeast contact metamorphic effects include the growth of poeciloblastic biotite and of chequered exsolution textures in potassium feldspar.

(*b*) *Ring-dyke microgranite*

(i) *Introduction*

Delany (1954, 1955, 1958) was the first to recognize the ring-dyke at Sabaloka and suggested that it had been intruded into the marginal fracture of a caldera subsidence.

Later mapping has shown that the intrusion is multiple and emplaced into a system of polygonal ring fractures of essentially vertical attitude. Subsidence of the central block partly or wholly preceded emplacement of the dyke but before this the fractures had been flushed through by gas streams carrying fluidized fragments. Explosion breccias, intrusive breccias and tuffisites consequently still mark out parts of the fracture system not intruded by ring-dyke microgranite and are locally found as remnants on the contacts of the dyke. The ring intrusion is almost complete although there is some uncertainty about the continuity in the section covered by Nubian sediments to the west of the Nile. Ring-dyke magma also penetrated the central block, exploiting flow-partings to form sills and following steep radial fractures (C3–C4) which in at least two cases had already provided passageway for erupting ignimbrites.

The ring-dyke suite is collectively referred to as 'porphyritic microgranite' though some fine-grained varieties are properly rhyolites. There is great variation in the size and abundance of phenocrysts and also in colour. The pale-grey of the freshest rocks has usually been modified to shades of red, brown or yellow, with green and black variants in the chilled margin.

(ii) *Field relations*

The external contact of the dyke follows the outer margin of the Ring Fracture Zone and before displacement by the Ummarahik Fault had a maximum diameter of 27 km and a minimum diameter of 14 km. In the north microgranite occupies the full width of the Fracture Zone but complexity enters with the appearance of volcanic rocks, and from Wad Belailo (O4) to Jebel Elager (H1) the intrusion is discontinuous and largely guided by the marginal faults of the Graben. In Jebelat El Humor (L2 and M2) magma has also exploited cross-fractures whereas south of the Ummarahik Fault several intrusions follow fractures within the Graben but parallel to its length. There are petrographic differences between these various bodies and clear evidence of multiple injection is displayed $1\frac{1}{2}$ km north of the Ummarahik Fault (L2), where a plexus of dykes intrudes volcanic rocks but is cross-cut by a mass of microgranite occupying the full width of the Graben. This is in turn cut by a dyke of porphyritic rhyolite which follows the northwestern boundary fault.

Table 3. Modal composition (volume %) of ring-dyke rocks

	I	II
felsic phenocrysts		
K-feldspar	5.9 (0–11.2)	23.2 (16.5–31.0)
Plagioclase		4.2 (2.5–7.6)
Quartz		10.9 (6.7–18.2)
mafic minerals†	11.0 (10.2–12.0)	3.1 (0.8–6.8)
groundmass‡		
Feldspars	62.5 (54.8–70.2)	58.6 (49.8–70.3)‡
Quartz	20.6 (19.1–22.0)	

I. Average percentages and ranges for three specimens of early, sparsely porphyritic microgranite.
II. Average percentages and ranges for seven porphyritic dyke rocks; their groundmasses vary from rhyolitic to microgranitic.
† This figure is for total mafic constituents and in some instances includes a few phenocrysts.
‡ Two relatively coarse-grained rocks of group II gave feldspar: quartz ratios near to 2:1.

From the southern termination of the Graben to Wad El Basal (B3) the ring intrusion skirts the southeast of the Plateau as a single dyke along the outer margin of the Ring Fracture Zone. Here again is evidence of multiple injection, especially in the north–south section in E2, where a central member of sparsely porphyritic rock is flanked by strongly porphyritic microgranites which chill hardly at all against the older intrusion, though showing marked reduction of grain size against the country rocks. This part of the dyke also provides good exposures of the inner contact dipping at around 70° towards the centre of the Plateau. Linear features on aerial photographs of the southeastern sector of the dyke give an impression of multiplicity which is partly deceptive since such features often mark differences between zones of differential cooling. On the ground, perceptible decrease in grain size may begin as much as 40 m within the contact and is often accompanied by a reduction in joint spacing, while for a metre or so within the contact the magma has been chilled to splintery, flow-banded rhyolite with fewer phenocrysts than the interior. Contact metamorphism of the country rocks is, however, very slight.

The ring-dyke attains its greatest width in the region of Jebel Rauwiyan (B4) where it encloses a large block of hornblendic basement gneisses. In this sector contacts are generally subvertical but an outward dip of 75° was recorded immediately southwest of Jebel Rauwiyan, and northeast of Wad El Basal the eastern contact of the inner branch of the dyke conforms to the inclination of the lava flows and dips at only 30° towards the Plateau.

West of the Nile the outcrop of the dyke is less obviously controlled by ring fractures, probably because its path was partly blocked by the mica granite. The magma was forced to exploit cross-fractures and the resulting outcrop shapes are lobate rather than linear.

(iii) *Petrography*

Modal analyses of ring-dyke rocks are averaged in table 3. The early, sparsely porphyritic type forms only a few *in situ* intrusions but is common as xenoliths. Only a few of these rocks contain phenocrysts of quartz and the texture of the groundmass is unlike that of later intrusions in being hypidiomorphic granular, both feldspars forming subhedral tablets towards which quartz is interstitial. The main ferromagnesian constituents are pale-green amphibole and brown to green biotite, while accessories include opaque oxides, zircon and zoned apatite, with occasional epidote and apatite.

In the commoner, more richly porphyritic rocks potassium feldspar occurs as tablets up to 1 cm which, judging from $2V_\alpha$, are probably sanidine cryptoperthites containing local microperthitic areas. Plagioclase (*ca.* An_{30}) is in small, euhedral phenocrysts often enclosed by potassium feldspar. Quartz phenocrysts are nearly always rounded, or corroded by caries. The groundmass varies from cryptocrystalline and flow-banded at the contacts through spherulitic and microgranophyric in less chilled rocks to evenly microgranitic in the middle of large intrusions. Amphibole and biotite rarely occur as phenocrysts but tend to cluster with each other and accessory constituents. The amphibole is typically anhedral and pleochroic, with γ olive green $> \beta$ green $> \alpha$ pale brown, $2V_\alpha$ about 20° and maximum extinction up to 20°, properties which suggest a hastingsite-rich composition. Fringes of a paler green, prismatic amphibole with rather larger $2V$ occur locally. Amphibole continued to grow into the later stages of crystallization and often encloses well-shaped grains of groundmass quartz and feldspar. In some rocks the mica is golden brown whereas in others it is pleochroic from deep green to pale straw-yellow. It sometimes forms subrounded clusters which may pseudomorph an earlier mineral. As accessories, opaque oxides and zircon are ubiquitous, while allanite, opidote and fluorite are very common and titanite and apatite less so. Products of alteration include hematite, chlorite, carbonates and clay minerals. A few rocks contain small vugs lined by chlorite, fluorite, quartz or epidote, sometimes with a filling of prehnite.

The ring-dyke suite has features in common with the Plateau and Graben ignimbrites and if the dyke represents late-emplaced de-gassed batches of the same magma one might expect it to show affinity with the upper parts of the ignimbrite sheets by containing a concentration of phenocrysts. Comparison of tables 2 and 3 shows, however, that ranges in phenocryst content and relative abundance do not allow clear distinctions between random selections of dyke rocks and ignimbrites, at least with the limited data available. A more detailed comparison would have to take into account such factors as possible flow differentiation within the dyke and fractionation of crystals from matrix during eruption of ignimbrites (Walker 1972).

The rocks of the ring-dyke also resemble some of the 'amphibole-fayalite granites' of the Nigerian complexes (Jacobson *et al.* 1958). In that region original fayalite and pyroxene are generally represented by soda-iron amphiboles and biotite. It is therefore worth recalling that clinopyroxene and fayalite are found unaltered in a number of specimens of the Plateau and Graben ignimbrites.

(*c*) *Minor intrusions*

There are surprisingly few minor intrusions within and around the Caudron, and the greatest concentration lies southwest of it where a swarm of quartz-porphyry dykes runs northwest through El Kodab (A4). The dykes are up to 100 m across but of inconstant width and direction, frequently anastomosing or wedging out. Rock types resembled those of the ring-dyke, though phenocrysts tend to be larger and quartz displays a bluish cast. Some other intrusions can be dated relative to the swarm, and among those which are older are several dykes of porphyritic olivine basalt and a few dykes and sheets of quartz trachyte. The trachytes are of particular interest in being the only intermediate rocks associated with the Cauldron Complex. They include both feldspar-phyric and aphyric types and their groundmass textures are similar to those of some peralkaline trachytes in the Wadi Abu Tuleih area. Unfortunately they are everywhere altered and their mafic constituents replaced by iron oxides. A group of late felsite minor intrusions are younger than the quartz-porphyry swarm, and similar rocks also form dykes which cut across the Graben in Jebelat El Humor. Many of these rocks are aphyric but some contain a few phenocrysts of quartz and feldspar. The abundant groundmass quartz is typically poikilitic towards small feldspar laths. Again there is a possibility that some of these rocks had peralkaline mineralogy, but alteration is advanced, with heavy replacement by clay minerals, sericite, silica and haematite.

5. Intrusive rocks of the Wadi Abu Tuleih Complex

Outside and to the north of the Cauldron, Delany (1958, p. 117) noted an intrusive mass which she described as being of 'soda granite' or 'riebeckite granite', with an associated network of dykes. The writer has found the 'soda granite' to be considerably larger than is shown on Delany's map, but being composed principally of hornblende syenite it does not qualify for the terminology used by her. On the other hand, peralkaline rock types do occur among the related minor intrusions.

(a) *The main intrusion*

The interior of this intrusion is well exposed on the north but the contact zone is less easily examined. The predominant rocks are buff-coloured, moderately coarse syenites with mafic clusters up to 5 mm and colour indices between 9 and 17. Enclosures of country rock are common, the largest over 200 m long, but the only good exposure of the contact is on the eastern side of the intrusion (P2), where the syenite sends veins into adjacent granitic gneisses and encloses blocks of it. Chilling is less marked than in the ring-dyke. Variations in grain size from about 1.25 to 2.5 mm occur in the centre of the pluton and may indicate multiple intrusion. More certain is the separate identity of the oval mass of microgranite at the southern end of the syenite, which from its shape appears to be younger than the main intrusion.

Modal composition of the syenite is shown in table 4. The feldspars include subhedral plagioclases zoned from andesine to mid-oligoclase and enclosed by large, poorly-shaped perthites. Some of these early plagioclases are antiperthitic. Plagioclase also occurs in rounded grains and polygonal mosaics enclosed within mafic clusters or surrounding them. The chief mafic constituent is anhedral hornblende of strongly absorbing type, generally pleochroic from deep reddish-brown to yellowish-brown but sometimes pleochroic in brownish-greens. Small values of $2V$ (*ca.* 25°) suggest a ferrohastingsite-rich composition. The hornblendes usually enclose opaque oxides and a grain or two of clinopyroxene, zircon or plagioclase, but some are strongly poikilitic towards granular plagioclase. The pale-green clinopyroxene, though also anhedral, is more evenly distributed. Accessories include interstitial quartz and well-formed zircon and apatite. A few rocks contain a small amount of biotite, and allanite is found rarely.

The southern microgranite is of finer and more even grain than the syenite and lacks mafic clusters. Subhedral perthite and antiperthite occur in roughly equal amounts with interstitial

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quartz and a few prismatic pseudomorphs after some primary ferromagnesian mineral, now replaced by iron oxides and quartz. This mode of alteration is like that of aegerine, so it is possible that the granite is of peralkaline affinity.

(*b*) *Minor intrusions*

The area contains a higher concentration of minor intrusions than any other part of Sabaloka. Over 200 dykes and several small plugs have been mapped and a wide variety of rocks found in them. Most dykes are small, but some of the largest exceed 10 m in width and form a plexus near the centre of the main pluton. These dykes are related in composition to the syenites but contain more quartz (*ca.* 10 %) and lack pyroxene and early plagioclase. They may be a late fraction of the syenite magma, but if so it is surprising to find that they often contain angular xenoliths of meta-dolerite which were presumably gained at depth.

Table 4. Modal composition (volume %) of syenite and microgranite from the Wadi Abu Tuleih complex

	I	II
Perthite and antiperthite	62.7	78.4
Plagioclase	22.3	—
Quartz	1.8	16.6
Hornblende	11.2	—
Clinopyroxene	1.3	—
Fe oxide/quartz pseudomorphs	—	5.0
Opaque iron oxides	0.5	—
Others	0.2	—

I. Average of three syenite analyses.
II. One analysis of microgranite.

Rocks with sodic ferromagnesian minerals are found only among the smaller intrusions, side by side with non-peralkaline felsites and quartz-porphyries. The peralkaline intrusions are younger than most other intrusions which they intersect. Good examples are found $1\frac{1}{2}$ km north of the syenite intrusion (P3, 4), where a small boss of feldspar-phyric microgranite is intruded on its northeastern side by a heterogeneous plug of peralkaline microgranites and micro-syenites. All these rocks have perthite phenocrysts and the more acidic have phenocrysts of quartz. Groundmass textures are hypidiomorphic in microsyenites but xenomorphic granular in microgranites. In one acidic type reddish-brown biotite is accompanied and partly enclosed by a blue sodic amphibole while in another the amphibole is prismatic and astrophyllite is present as well as biotite. In some of the quartz-microsyenites sodic amphibole is the only coloured mineral while others have a green amphibole and a small amount of red-brown aenigmatite.

Elsewhere, peralkaline dykes are mostly trachytic and contain feldspar phenocrysts which are either finely microperthitic with large $2V$ or optically homogeneous with small $2V$. Coarser mesoperthite predominates in the groundmass, with variable amounts of interstitial quartz and either a bluish-green amphibole or aegerine as the chief ferromagnesian constituent. Opaque oxides, rutile and zircon are common minor accessories while some contain abundant fluorite both in the groundmass and as veinlets with quartz and magnetite.

A very fresh sample of rhyolite collected from a small dyke west of the syenite intrusion proved to be the most strongly peralkaline rock so far found at Sabaloka. The rock is dark-green in hand specimen and contains numerous equidimentional microphenocrysts of quartz and microcline, with a smaller number of pale-grey to yellowish-grey clinopyroxene phenocrysts. The pyroxenes form clusters with haematite and are fringed by blue amphibole and aegerine needles. The matrix is mainly composed of microcline, quartz, sodic plagioclase and numerous prisms of aegerine, while zircon is an abundant accessory. A chemical analysis of this and two other rocks with peralkaline mineralogy shows their affinity to be comenditic rather than pantelleritic, according to criteria used by Macdonald (1975).

Finally, a group of small, breccia-filled diatremes pierce the gneisses immediately northeast of the main syenite (Q3). The constituent fragments were probably derived from the adjacent gneisses; quartz and feldspar figure largely, with smaller amounts of hornblende, apatite and zircon. Mosaic quartz, white mica and haematite are major components of the matrix, with carbonates, chlorite and fluorite locally abundant.

6. Structure

Dips on rhyolite lavas shown on figure 1 were measured on flow-featuring and flow-banding, avoiding rocks clearly affected by primary deformation. Fortunately, the ramp-structure which imparts a strong up-flow inclination to the banding of some acidic lavas seems to be poorly developed at Sabaloka, and the method has provided a consistent structural pattern. There is, of course, no way of determining the initial attitudes of the flows. Measurements on welded ignimbrites record the attitude of the eutaxitic foliation and are less reliable guides to sheet dip since compaction of ash-flow tuffs over a surface of strong relief produces marked supratenuous folding of the foliation. Measurements remote from the lower contact of an ash-flow cooling unit are likely to be the most useful, a point kept in mind in making the structural interpretation.

Some major subsurface features of the complex have been deduced from a combination of structural and gravimetric measurements (Sadig *et al.* 1974), although in the absence of a sufficient density contrast it is impossible to distinguish between intrusive and extrusive acidic rocks by geophysical means.

(*a*) *The Plateau*

The volcanic rocks attain their greatest stratigraphical thickness of about 2800 m in the south (figure 4), and extrapolations based on dip and stratigraphy suggest that the subvolcanic surface reaches its lowest point below the southern lobe of the Plateau. It is estimated that the vertical thickness of volcanic rocks is here about 2000 m, with the base at −1400 m b.s.l. This decreases to about 1300 m against the ring-fractures to the southeast, where downthrow on the ring is at a maximum, but there are no outcrops of volcanic rocks outside the ring from which total displacement can be estimated. North and west of their thickest point the volcanic rocks diminish rapidly to less than 1000 m, and the structure of the basal surface in the north of the Plateau can be described as broadly synclinal with a gentle plunge to the southwest.

Over much of the Plateau there is angular discordance between ignimbrites and underlying volcanic rocks, and although both groups take part in the overall centrepetal inclination it is evident that there were at least two main episodes of subsidence. To explore this point further a map was prepared on which dips on rocks older than the ash-flows are shown corrected to their pre-ignimbrite values on the assumption that the eutaxitic foliation was initially subhorizontal (figure 5*a*). Steep and confused dips in lavas near the southeastern margin of the Plateau have been omitted. The corrected dips on the older volcanic rocks define a

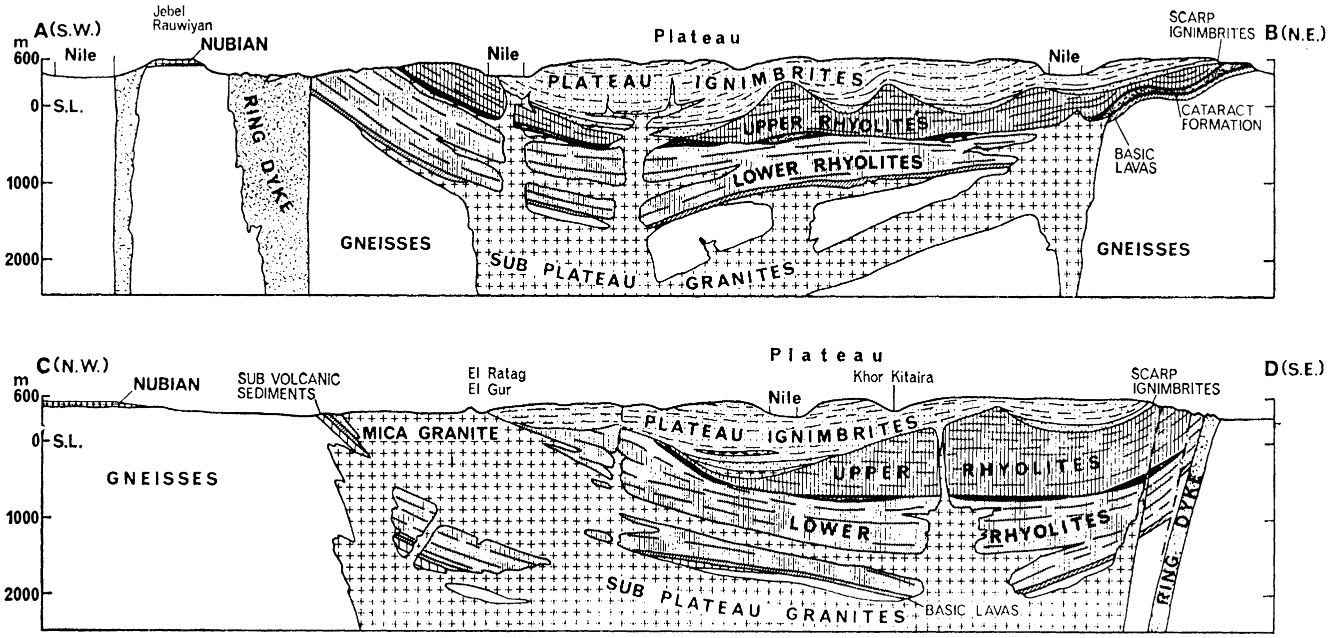

Figure 4. Cross-sections of the Plateau. For location of the section lines see figure 1. The detailed form of the sub-Plateau granites is conjectural, but the general mass distribution is based on geophysical evidence (Sadig *et al.* 1974) while the base of the volcanic succession is derived from extrapolated dips. Note that the inward dip of the ring-dyke at the southern end of section C–D is atypical insofar as the intrusion has an overall subvertical attitude.

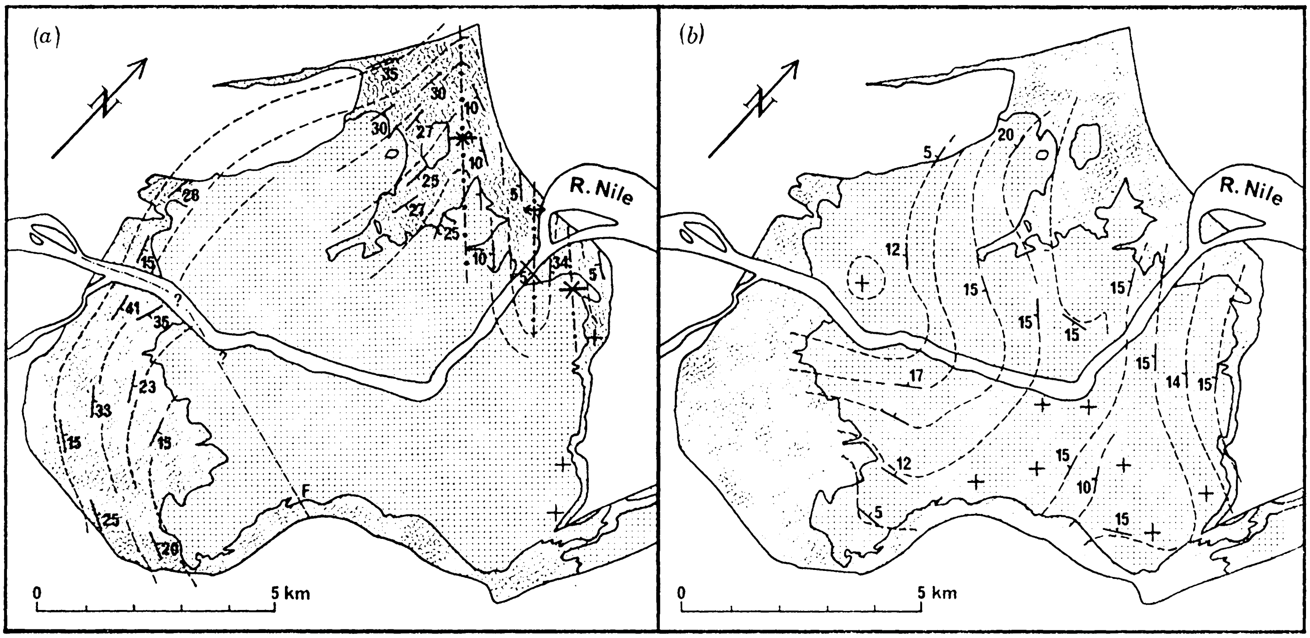

Figure 5. Structures produced by the two main periods of subsidence within the Plateau, shown by generalized dips. (*a*) Structure of the volcanic rocks prior to the onset of the ash-flow eruptions. Dips have been corrected for the effects of later subsidence on the assumption that eutaxitic foliation in the overlying ignimbrites was originally horizontal, except near the base. The outcrop of volcanic rocks older than the ignimbrites is shown by heavy stipple, the ignimbrite outcrop in light stipple. (*b*) Generalized attitudes of the eutaxitic foliation in the ignimbrites. Dips measured near the lower contact of the ignimbrite sheet have been omitted. Ornament as in (*a*).

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northwesterly elongated basin which is deepest in the east of the Plateau. Since the rhyolite lavas thin out and eventually disappear in this direction it is likely that the basin developed on the flank of the rhyolite lava field and formed a topographical depression. No satisfactory explanation can be offered for the northwest trending open folds which deform lavas and volcaniclastic rocks around El Hugna, but they are presumably connected in origin with this early period of subsidence.

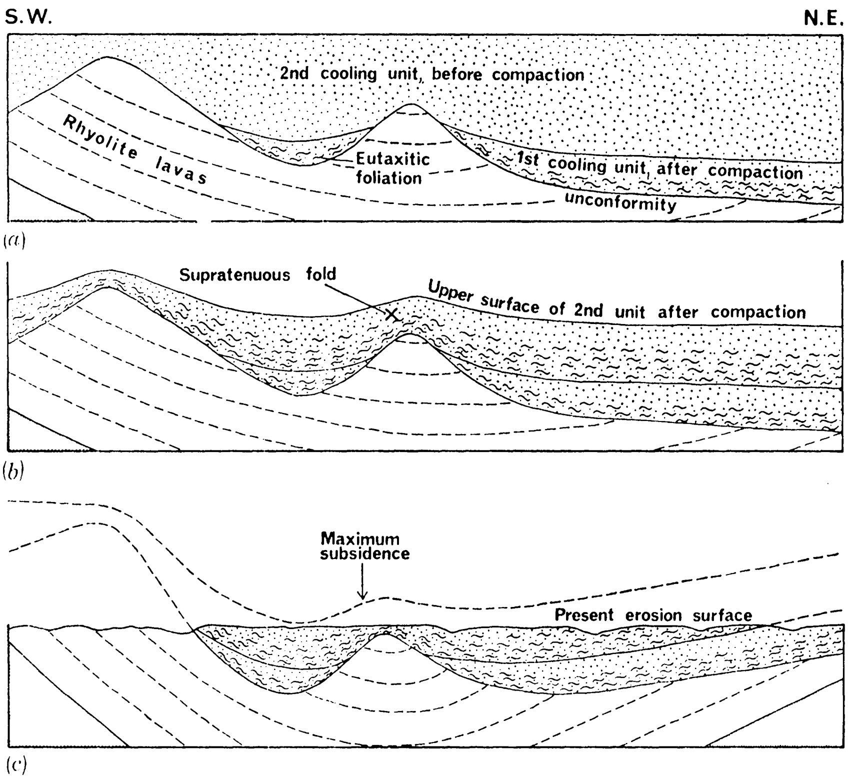

FIGURE 6. Structural relations between ignimbrites and rhyolite lavas in the Plateau, shown diagrammatically. (*a*) Successive ash-flows fill a basin bordered to the southwest by hills carved out of rhyolite lavas. (*b*) Compaction of the ignimbrites over the rhyolite hills produces supratenuous folds defined by the eutaxitic foliation. (*c*) Later subsidence completes the present structure. Foliations dip mainly towards the centre of subsidence but the pattern is modified by supratenuous folds over buried and exhumed hills of rhyolite lava.

A generalized interpretation of dips on the ignimbrite formations (figure 5*b*) defines a focus of late subsidence west of the Nile, in an area where there were high hills at the time the ash-flows were erupted. Consequently the ignimbrites are here very variable in thickness and lithology, but generally thinner than elsewhere in the Plateau, with exhumed summits of rhyolite hills emerging through them. Inferred structural relations are shown diagrammatically in figure 6, and it follows that neither period of subsidence focused upon the southern lobe of the Plateau, although the volcanic rocks here reach their greatest thickness.

(*b*) *The Ring Fracture Zone*

This structure everywhere comprises two or more concentric ring-fractures and associated cross-fractures, the average width of the whole Zone being about 1 km. Some fractures are clearly faults, others are marked only by brecciation. In plan view the Zone is an elongate polygon with a re-entrant in the southeast and a large section in the northwest concealed by

Nubian sediments. Ring-dyke microgranite picks out many features, both large and small, but in the south there are several which have escaped intrusion.

In the Graben sector the boundary faults are subvertical and the volcanic rocks disposed synclinally. Where it adjoins the Plateau the Graben is complicated by subsidiary downfolds and rotated blocks separated by internal faults (figure 7*b*) while north of the Ummarahik fault large masses of volcanic rock have been stoped away by the ring-dyke magma (figure 7*a*).

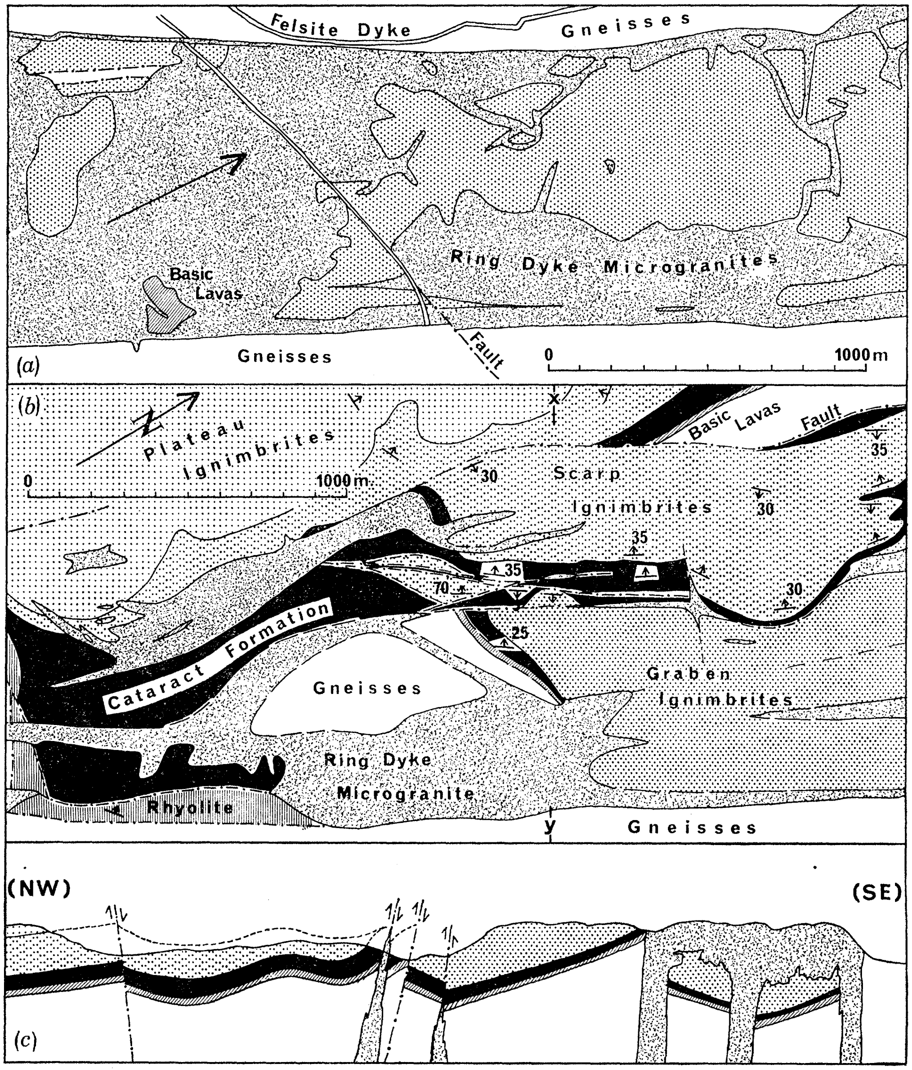

FIGURE 7. (*a*) Outcrop pattern produced by stoped emplacement of ring-dyke microgranite into the volcanic rocks of the Graben (located in M2 of figure 1). (*b*) Map of the southwestern termination of the Graben (H1 and H2 of figure 1). (*c*) Interpretive section across the southwestern sector of the Graben. It is assumed that the Plateau ignimbrites overlap the Scarp ignimbrites and are laterally equivalent to the Graben ignimbrites. The horizontal scale is twice that of the map (*b*) and the vertical scale is true.

On the other hand, the wide zone of fracturing around the southern and southwestern margins of the Plateau cannot be a simple graben because the innermost fractures are faults downthrowing towards the centre of the Plateau. This is particularly clear around Trig. Point 407 m (D1) where all the fractures are either inward-throwing faults or breccia zones. Around Rauwiyan Island (B4), however, the inward throw is much less and the branched ring-dyke encloses large blocks of basement rock. Gravity anomalies in this area (Sadig *et al.* 1974) are best satisfied by assuming that the basement on Rauwiyan Island itself is continuous downward whereas north of the Nile, near Abu Dom, it appears to be thin and has perhaps been eroded by stoping from below. The very wide inner branch of the ring-dyke in this area must have been emplaced by piecemeal stoping and/or local graben subsidence.

The Fracture Zone is probably very steep over a depth of several kilometres. Outward inclinations of 70–80° were measured on the outer fracture on Rauwiyan Island and inward dips of 72–84° near Trig. Point 407 m (D1). Inward and outward dips of a similar order are deduced from the gravity data on some other sectors, and the overall shape is thought to be cylindroidal with a subvertical axis.

Stoping may have provided much of the space now occupied by the ring-dyke but a further contribution arose from basinal deformation of the subsided block. This is probably the reason for the greater width of the fracture zone and ring-dyke on the longer major diameter of the complex. Estimates based on geological sections suggest lateral foreshortening could be as much as 5 %, amounting to 1.3 km on the longer diameter.

There are problems in deciding the age of ring-fracturing relative to the various volcanic formations. The fracture system was of necessity well-established before emplacement of the ring-dyke and some of the fractures had already acted as feeding channels to preceding ash-flow eruptions. On the other hand there is no evidence that ring-fracturing accompanied the early subsidence of the rhyolite lavas, and indeed existing fractures truncate the basin formed at this time (figure 5*a*). Moreover, north of the Plateau the ring encompasses a large area well beyond the rhyolite lava field but within the region covered by ash-flows. The ring-structure appears to be related to a late configuration of the magma chamber, and it is therefore likely that fracturing immediately preceded and accompanied the main phase of ash-flow activity and was finally completed during the subsidence which followed.

(*c*) *Major structure of the Cauldron Complex*

In the later stages of its active history the ring-complex must have been expressed at the surface by a caldera of similar size to that of the Ring Fracture Zone, that is, some 27 km by 14 km. These dimensions place Sabaloka into a class of large volcanic subsidence structures which includes such relatively young analogues as the Cainozoic calderas of Valles, New Mexico (Smith, Bailey & Ross 1961), Timber Mountain, Nevada (Carr & Quinlivan 1968), and some of the New Zealand calderas (Cole 1967). There are similarities, too, with even larger structures such as the Mogollen Plateau of New Mexico, which Elston, Coney & Rhodes (1968) regard as a giant ring-structure 125 km in diameter, with a marginal graben and centrepetally-dipping volcanic succession. Among subvolcanic ring-complexes the closest analogies are with some of the Younger Granites of Northern Nigeria (Jacobson *et al.* 1958).

In terms of Williams's (1941) original classification of calderas Sabaloka is of 'Glen Coe' type in that the depression at the surface was the result of underlying cauldron subsidence along a ring-fracture. McBirney & Williams's (1969) later revision places more emphasis on the nature of the associated volcanic rocks and relation of the caldera to antecedent volcanoes. By these criteria Sabaloka is of 'Valles' type, in which collapse follows the discharge of large volumes of acidic pyroclastic flows from fissures unrelated to pre-existing volcanoes. In a classification of ring-complexes by Sveshnikova (1968) Sabaloka is used as the type example of a class characterized by a single ring-dyke and only modest subsidence. However, the present study has shown that the subvolcanic surface is in places depressed about 2 km below the present surface, and the apparent simplicity of the intrusion pattern is almost certainly the consequence of a shallow level of erosion.

One of the most useful results of the gravity survey (Sadig *et al.* 1974) was to show that a low-density mass several kilometres in diameter underlies much of the Plateau. This body is presumably granitic in composition and extends to a depth in the order of 5.5 km in the area in which the volcanic rocks are also at their thickest. The mica granite may be an exposed portion of this sub-Plateau intrusion complex. There is no indication in the gravity picture of any large subsurface mass of basic rock. A second gravity low over basement rocks within the northern half of the cauldron is thought to mark the position of another, but completely unexposed granite lying at no great depth and of limited vertical thickness. The gravity contours suggest that this intrusion is about 7 km in diameter and separated by basement rocks from the sub-Plateau granite. The presence of a subsurface granite in the north would help to explain some puzzling features of the complex's overall structure. The granite could be a cupola related to the northerly extension of the magma chamber which, it is postulated, post-dated early subsidence. Its presence reduces the apparent asymmetry between intrusions, volcanic rocks and the Ring Fracture Zone. The present distribution of volcanic rocks within the Cauldron is partly the result of the concentration of early activity into the south, and partly the consequence of late- or post-volcanic deformation of the subsided block, which deepened the southern basin and created a complementary dome in the north. If there is an intrusion beneath the Northern Dome it is likely to have played a decisive rôle in dome formation, either as a passive mass which by strengthening part of the caldera floor led to differential subsidence of the sinking block, or as an active agent which by the force of its intrusion raised up a 'resurgent dome' within the already subsided caldera floor.

An explanation based on resurgent doming makes an attractive hypothesis, but beyond a general comparability of size and association with such well-established resurgent cauldrons as Valles and Timber Mountain (Smith & Bailey 1968) there is no firm evidence on the crucial matter of the relative ages of cauldron subsidence and doming. Both events mainly post-dated the ash-flow eruptions which brought extrusive activity to a close but some doubt attaches to their relationships with the ring-dyke microgranites. Elsewhere (Almond 1971) it has been proposed that microgranite magma was forced upwards into ignimbrite-filled vents during a deflation which followed eruption of the ash-flows, while the vent contents were still hot. Deflation could have preceded or accompanied cauldron subsidence but in any case must have occurred before any possible resurgence. On the other hand, microgranite has been intruded into both flanks of the Graben, which is an integral part of the Northern Dome. So if microgranite was intruded during a single, brief period one must conclude that this was more or less simultaneous with both cauldron formation and doming, and the structures are best ascribed to differential subsidence. But this may be an oversimplification. There is plenty of evidence to show that the ring-dyke is a multiple intrusion which includes a number of petrographic variants and displays an overlapping age relation with the mica granite. Magmas of the ring-dyke suite may well have been available for a considerable period and their continued intrusion into the Cauldron could have been the prime cause of a late resurgence. Among more circumstantial evidence it is relevant to remark that the ignimbrites of the Graben are much more altered than those of the Plateau, and pervasive alteration is apparently a common feature of volcanic rocks affected by resurgent doming (Smith & Bailey 1968, p. 632). On the other hand, in the resurgent cauldrons described by these authors domes are sited more or less centrally within the subsided block and there are characteristic fracture patterns which have not so far been recognized at Sabaloka. The writer is disposed to conclude that although the available data is insufficient to discriminate between differential collapse and positive uplift the analogies with better-understood cauldrons of Cainozoic age make the resurgency hypothesis distinctly the more credible. General characters common to nearly all resurgent cauldrons include the association with silicic volcanicity and voluminous ash-flows, a more or less coherent

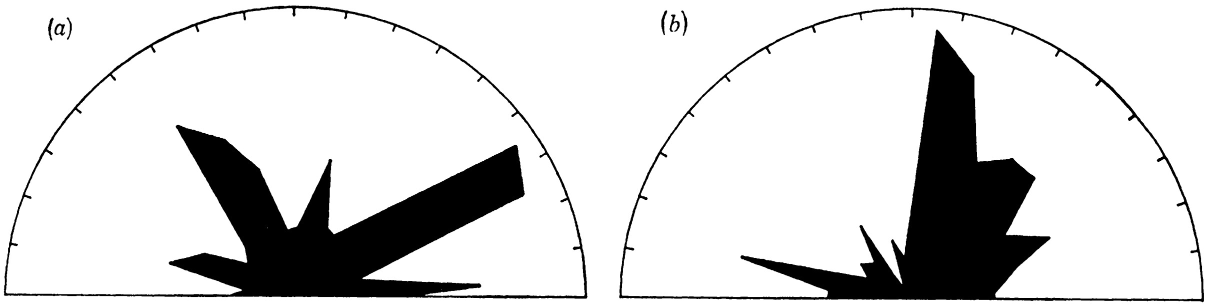

FIGURE 8. (*a*) Rose diagram of master joint azimuths in the volcanic rocks of the Plateau, based on 400 master joints measured from aerial photographs. North is towards the top of the page. (*b*) Rose diagram of dyke azimuths in the Tuleih Complex, using 270 mapped dykes.

subsided block, diameters of about 10 miles or more and a cratonic or post-orogenic environment. All these features are present at Sabaloka. Smith & Bailey themselves point to the Liruei ring-complex of northern Nigeria as one likely example of an eroded resurgent cauldron. As emphasized elsewhere, there are the closest parallels between the Younger Granites of Nigeria and those of the Sudan. Liruei itself (Jacobson *et al.* 1958) has much in common with Sabaloka, with a polygonal ring-dyke of granite porphyry surrounding a cauldron containing a thick succession of acidic volcanic rocks and intruded by biotite granite with which tin–tungsten mineralization is associated. In terms of a denudation sequence of cauldrons, Sabaloka is exposed at much the same structural level as Liruei, though the pronounced doming at Sabaloka exposes the subvolcanic basement more extensively.

(*d*) *The Wadi Abu Tuleih Complex*

The structure and age relations of this small cluster of intrusions remains largely unknown, and as yet there is no gravity data for the area in which they lie. The concentration of the Tuleih dykes into the NNE, with a subsidiary maximum in the WNW, indicates a stress pattern different from that which controlled structures in the Cauldron, where the master joint system in volcanic rocks shows maxima near NW and NE (figure 8). Nevertheless, comparative petrography does not rule out a possible link between the felsite minor intrusions which are the youngest rocks in the Cauldron and some of the peralkaline dyke rocks at Wadi Abu Tuleih, though an undoubted peralkaline rock has yet to be found within the Cauldron. But even if such a link were proved it would throw little light on the age of the main syenite-granite intrusion at Tuleih, which is older than most, at least, of the dykes in the area.

(*e*) *Post-Nubian faulting*

The prominent fault of Ummarahik roughly bisects the Cauldron Complex in a NNW direction. Gravity interpretation requires about 1000 m of Nubian sediments on the north side of the main fracture, though a parallel, antithetic fault reduces this thickness to 300 m or so 1 km further north. Vertical displacement decreases rapidly eastwards and beyond Jebel Ummarahik itself basement gneisses outcrop on both sides of the fault, and 7 km east of the Jebel the throw appears to have reversed. Between these two localities the fault crosses the Ring Fracture Zone with a dextral strike-slip displacement of 2 km defined by the offset of vertical structures. There is almost certainly a comparable displacement on the western side of the ring, but the field relations are obscured by overlying Nubian sediments.

At Jebel Ummarahik the Nubian sediments are deformed into an asymmetrical, synclinal drag-fold, with dips of up to 25° N adjacent to the fault. Ancillary faults trending at small angles clockwise from the main fracture are associated with silicified breccias and subhorizontally striated slickenside surfaces. Clearly both lateral and vertical components of displacement post-date the Nubian sediments and there is some evidence, in the attitude of the slickensides, that these components developed separately. The location of this major fault between two subsurface granite masses makes one suspect that the igneous complex strongly influenced the generative stress field, but it is also possible that the fault is a regeneration of an older structure which initially located the complex itself.

Several small faults displace the Nubian unconformity to the west of the Plateau, and others displacing rocks of the Ring Fracture Zone in Jebelat El Humor are likely to be of similar age. Moreover, throughout the area occasional narrow zones of brecciation and silicification traverse the basement gneisses or follow the courses of minor intrusions (see also Medani & Vail 1974).

7. Petrochemistry and petrogenesis

(a) *Analytical data*

Of the 20 rocks which have been chemically analysed, 15 are from the Cauldron and 5 from the Tuleih Complex (table 5). Of these, the single representative of the early basic lavas (388) is best discussed separately because of its compositional isolation from the other rocks and the probability that its composition has been significantly modified by alteration. The analysed rock is aphyric, lacks any trace of former olivine and has a modal plagioclase composition near An_{50}. It lies towards the basic end of the compositional range of the formation. The presence of normative quartz and hypersthene in the analysis might be taken to indicate tholeiitic affinity, and the major oxide composition is compatible with a rock near the basalt/trachybasalt borderline of a tholeiitic or transitional suite. On the other hand, alteration has almost certainly increased the oxidation ratio and so added to the degree of silica saturation in the norm. Recalculation of the norm with Fe_2O_3 reduced to an arbitrary 1.5 %, for instance, results in the disappearance of free silica from the norm and its replacement by over 3 % normative olivine.

A number of recent papers have discussed the use of minor and trace elements in the discrimination of basalt magma types (Pearce & Cann 1973; Floyd & Winchester 1975). The mobility of the relevant elements under conditions of low-grade metamorphism has also been assessed (Smith & Smith 1976; Wood, Gibson & Thompson 1976). Data is available for applying to the Sabaloka basalt those criteria said to be the most reliable for altered rocks, but when this is done the discriminants give an equivocal answer. Thus, ratios involving Y, Nb, Zr and P suggest that the Sabaloka basalt is marginally within the tholeiite field, whereas the high absolute amounts of Ti and P are more in keeping with an alkalic basalt. In view of the limited but significant scatter of points for such criteria found by Smith & Smith (1976) on specimens from a single, altered lava flow it would be clearly unwise to base any firm conclusions on the results of one analysis. The best that can be said is that the Sabaloka lavas are probably either transitional in character or not far removed from it.

The remaining analyses cover the major variations within the intermediate and acidic rocks of the Cauldron and Tuleih Complexes. Close similarities to the Younger Granites of Western Africa make many of the general comments of Jacobson *et al.* (1958), Bowden & Turner (1974) and others equally applicable to the Sabaloka rocks. One may instance the consistently low CaO and MgO of the association, the tendency of some rocks towards subaluminous and peralkaline compositions due to a combination of modestly high alkalies and Al_2O_3 deficiency,

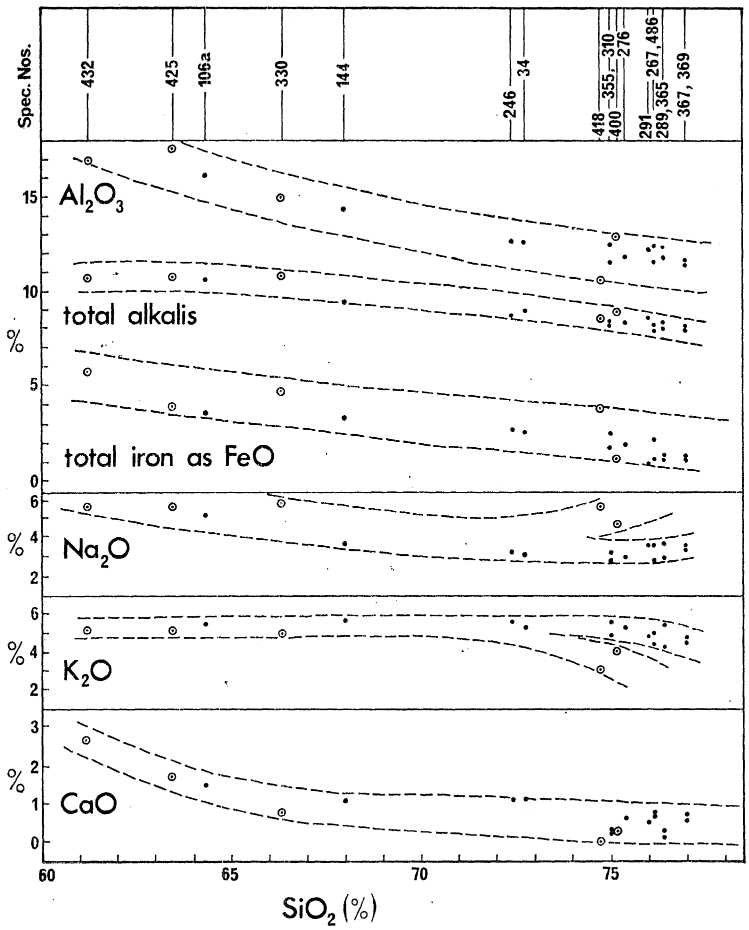

FIGURE 9. Variation diagram of major oxides plotted against SiO_2 for Sabaloka intermediate and acidic rocks. Rocks from the Tuleih Complex ⊙; rocks from the Cauldron Complex ●.

the abundance of fluorine among the minor constituents and the positive correlation between zirconium content and agpaitic coefficient (here expressed as molecular proportions

$$Na_2O + K_2O/Al_2O_3).$$

Though incomplete data for Nb, Y and Sn do not allow firm conclusions it may be that these elements also correlate positively with agpaitic coefficient.

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TABLE 5. CHEMICAL ANALYSES OF IGNEOUS ROCKS FROM SABALOKA

Analyst: E. L. P. Mercy

sample no.	Basic lava	Rhyolite lavas				Rhyolite ignimbrites				Ring-dyke microgranites			Mica granites		Tuleih quartz syenites		Quartz trachyte	Sodic micro-syenite	Sodic micro-granites	
	388	355	365	310	289	367	369	267	34	144	246	276	291	486	432	425	106a	330	400	418
SiO_2	47.16	75.02	76.35	74.98	76.30	76.95	76.89	76.01	72.73	68.00	72.39	75.38	75.99	76.16	61.12	63.40	64.30	66.29	75.16	74.72
TiO_2	3.12	0.14	0.17	0.30	0.15	0.09	0.09	0.21	0.30	0.54	0.33	0.21	0.085	0.07	0.74	0.34	0.18	0.25	0.08	0.18
Al_2O_3	15.81	12.62	12.45	11.61	11.96	11.46	11.63	11.65	12.68	14.32	12.77	11.94	12.38	12.48	16.90	17.70	16.14	15.09	13.04	10.76
Fe_2O_3	4.43	1.90	1.28	2.85	1.50	0.73	0.75	1.26	0.94	1.07	1.46	0.53	0.28	0.64	1.73	1.34	4.09	1.71	0.43	2.93
FeO	8.51	0.08	0.09	0.09	0.09	0.71	0.74	1.10	1.88	2.44	1.52	1.54	0.74	0.68	4.30	2.76	0.01	3.26	0.97	1.34
MnO	0.14	0.003	0.006	0.022	0.017	0.022	0.020	0.036	0.048	0.054	0.037	0.033	0.024	0.02	0.18	0.17	0.11	0.13	0.02	0.12
MgO	5.19	0.22	0.02	0.14	0.16	0.05	0.02	0.20	0.05	0.48	0.12	0.19	0.16	0.05	0.05	0.02	0.36	0.14	0.05	0.08
CaO	8.11	0.36	0.32	0.24	0.13	0.70	0.65	0.74	1.16	1.62	1.18	0.64	0.57	0.77	2.68	1.68	1.49	0.78	0.34	0.09
Na_2O	3.32	3.32	3.86	2.84	3.06	3.45	3.52	2.96	3.16	3.72	3.28	3.01	3.74	3.79	5.63	5.75	5.19	5.90	4.86	5.67
K_2O	0.43	4.96	4.34	5.66	5.52	4.78	4.70	5.14	5.90	5.71	5.70	5.37	4.99	4.55	5.06	5.13	5.57	5.02	4.17	3.09
P_2O_5	0.36	0.014	0.014	0.027	0.006	0.004	0.004	0.016	0.015	0.064	0.021	0.026	0.024	0.02	0.12	0.05	0.039	0.03	0.01	0.01
CO_2	0.73	0.03	0.09	0.05	0.03	0.05	0.01	0.02	0.02	0.12	0.05	0.08	0.08	0.10	0.02	0.02	0.40	0.12	0.03	0.09
F	0.05	0.21	0.04	0.02	0.10	0.24	0.21	0.02	0.08	0.10	0.10	0.13	0.30	0.27	0.08	0.11	0.07	0.03	0.26	0.07
S	tr.	0.010	0.005	0.035	0.17	0.005	0.004	0.002	0.092	0.004	0.005	0.003	0.004	tr.	tr.	tr.	tr.	tr.	tr.	tr.
H_2O-	0.29	0.38	0.15	0.33	0.20	0.09	0.09	0.38	0.10	0.11	0.18	0.06	0.06	0.13	0.16	0.25	0.61	0.51	0.19	0.07
H_2O+	2.58	0.80	0.62	0.72	0.50	0.56	0.42	0.43	0.45	1.05	0.59	0.48	0.55	0.30	0.30	0.32	1.69	0.72	0.37	0.12
total	100.23	100.07	99.80	99.91	99.89	99.89	99.75	100.17	99.60	99.40	99.73	99.62	99.98	100.03	99.07	99.04	100.25	99.98	99.98	99.34
Agpaitic coefficient	0.38	0.86	0.88	0.93	0.92	0.95	0.93	0.89	0.92	0.86	0.91	0.91	0.93	0.89	0.87	0.85	0.91	0.99	0.96	1.17
C.I.P.W. mass percent norms																				
Q	2.1	36.1	37.1	36.2	37.1	37.4	37.2	36.9	29.0	19.9	28.9	35.3	34.4	35.4	2.3	5.8	10.4	10.1	30.7	30.6
or	2.8	29.5	25.6	33.4	32.8	28.4	27.8	30.6	35.0	33.9	33.9	31.7	29.5	26.7	30.0	30.6	32.8	29.5	24.5	18.3
ab	28.3	28.3	32.5	24.1	25.7	29.3	29.9	25.2	26.7	31.4	27.8	25.7	31.4	32.0	47.7	48.2	44.0	49.8	40.9	38.2
an	26.7	0.3	0.8	1.1	0.3	1.4	1.9	3.1	2.8	5.3	3.1	1.7	0.6	1.9	5.8	7.5	4.2	—	—	—
C	—	1.6	1.1	0.4	0.8	—	—	—	—	—	—	0.5	0.6	0.6	—	—	—	—	0.6	—
ac	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	8.32
di	5.2	—	—	—	—	—	—	0.4	2.1	1.2	1.7	—	—	—	5.9	—	0.2	2.7	—	—
hy	17.4	0.5	—	0.4	0.4	0.8	0.7	0.9	1.0	3.4	0.5	2.5	1.3	0.8	2.6	3.6	0.8	3.1	1.3	2.71
mt	6.5	—	—	—	—	1.2	1.2	1.9	1.2	1.6	2.1	0.7	0.5	0.9	2.6	1.9	—	2.5	0.7	—
hm	—	1.9	1.3	2.9	1.4	—	—	—	—	—	—	—	—	—	—	—	4.2	—	—	—
il	5.9	0.2	0.2	0.2	0.2	0.2	0.2	0.5	0.6	1.1	0.6	0.5	0.1	—	1.4	0.8	0.3	0.5	0.2	0.30
ru	—	0.1	0.1	0.2	0.1	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
ap	1.0	—	—	—	—	—	—	—	—	—	—	—	—	—	0.3	—	—	0.1	—	—
fr	0.1	0.4	0.1	—	0.2	0.5	0.4	—	0.2	0.2	0.2	0.2	0.6	0.5	0.2	0.2	0.2	0.7	0.5	0.15
pr	—	—	—	—	—	—	—	—	0.2	0.3	0.1	0.2	—	—	—	—	0.9	0.3	—	—
cc	1.7	—	0.2	—	—	0.1	—	—	—	—	—	—	—	—	—	—	0.2	0.4	—	1.10
trace elements (parts per million)																				
Rb	< 10	280	195	145	275	210	230	150	130	140	140	150	400	400	50	100	150	120	260	460
Sr	410	20	10	50	10	15	15	30	55	100	60	30	20	20	265	130	45	30	10	< 10
Ba	300	0	0	435	0	10	0	70	245	580	245	55	0	20	3200	860	0	160	30	10
Zr	275	220	375	560	350	195	205	320	455	625	525	370	125	155	540	600	1020	1850	510	5100
Nb	25	—	—	—	—	—	—	—	—	—	—	—	—	20	35	85	—	115	265	325
Y	30	—	—	—	—	—	—	—	—	—	—	—	—	110	50	120	—	80	160	340
Sn	< 10	—	—	—	—	—	—	—	—	—	—	—	—	< 10	< 10	< 10	—	< 10	10	35

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Locations and lithological notes

388. N. sector of Graben, near Wad Belailo (N4). Block of aphyric basalt in ignimbrite. The flow-oriented laths of dusty-looking labradorite are subophitically related to a pale-brown augite. Abundant primary and secondary opaque oxides and a large amount of secondary chlorite.

355. S.W. margin of Plateau (C3). Delicately flow-banded, massive rhyolite. Wholly cryptocrystalline apart from rare kaolinized feldspar phenocrysts.

365. S.W. sector of Plateau (D3). Massive, coarsely flow-banded, aphyric rhyolite. Cryptocrystalline except for streaks of fine-grained quartz parallel to the flow-banding and some small patches of fluorite.

310. S.W. sector of Plateau (D2). Strongly flow-banded rhyolite with small, rounded phenocrysts of quartz (4%) and oligoclase (2%). The groundmass has fine-grained and patchy devitrification texture.

289. N.E. scarp of Plateau (I4). Strongly flow-banded, devitrified rhyolite with small (1–2 mm) phenocrysts of quartz (5%) and indeterminate, altered feldspar (4%).

367. S.W. sector of Plateau (D3). Strongly eutaxitic ignimbrite 10 m above the base of the Plateau formation. Phenocrysts of K-feldspar (9%), quartz (7%) and andesine (3%).

369. Southern lobe of Plateau (D2). Strongly eutaxitic ignimbrite 7 m above the base of the Plateau formation. Phenocrysts of quartz (9%), K-feldspar (9%) and a little plagioclase, biotite and allanite.

267. S. sector of Graben (I2). Lower part of the Graben Ignimbrite. Massive, crystal-rich with moderately strong eutaxitic foliation partly obscured by recrystallization. Phenocrysts mainly quartz (21%) and K-feldspar (20%) with small amounts of plagioclase and a variety of ferromagnesian minerals. The groundmass is patchily recrystallized to a fine microgranitic texture.

34. N. sector of Graben (N4). Massive and very crystal-rich variety of the Graben Ignimbrite lacking visible eutaxitic foliation. Phenocrysts of K-feldspar (32%), quartz (8%), oligoclase (6%) and ferromagnesian minerals (4%), mainly hedenburgite and fayalite.

144. S. margin of Plateau (E2). An early facies of the ring-dyke suite, often xenolithic, and with only a small and variable content of phenocrysts, nearly all K-feldspar. Groundmass of subhedral alkali feldspar, interstitial quartz, dark-green amphibole and brown mica.

246. S. margin of Plateau (G1). Richly porphyritic microgranite typical of much of the ring-dyke. Phenocrysts of K-feldspar (30%), plagioclase and quartz total over 40% of the rock. Green hornblende is the chief ferromagnesian constituents but amounts to less than 5%. Biotite, opaque oxides, zircon, fluorite and allanite are accessory.

276. S. sector of Graben (I2). A late, rhyolitic facies of the ring-dyke suite. Phenocrysts of K-feldspar (18%), quartz (18%), plagioclase (3%) and clinopyroxene part-altered to hornblende (2%) in a microcrystalline groundmass.

291. N.W. of Plateau (E6). Early, sparsely porphyritic variety of mica granite. Phenocrysts of perthite up to 1 cm accompanied by smaller phenocrysts of sodic plagioclase, quartz and biotite. The microgranitic groundmass contains similar minerals plus muscovite. Fluorite is a common accessory.

486. N.W. of Plateau (D6). Late phase of mica granite with only rare phenocrysts. Mineralogically similar to 291 except in that the chief mica is a pale yellowish-green replacement of biotite.

432. Tuleih Complex, from near centre of main pluton (O2). Relatively fine-grained syenite with mafic clusters up to 2 mm. Perthite (64%) and antiperthitic oligoclase are the main components, with clusters of greenish-brown hornblende enclosing clinopyroxene, opaque oxides, a little zircon and apatite. About 2% interstitial quartz is also present.

425. Tuleih Complex, near W. margin of main pluton (O2). Mineralogically similar to 432 but coarser-grained, with mafic clusters up to 5 mm.

106a. S.W. of Plateau (A1). Inclined sheet 2 m thick cutting gneisses. Sparse phenocrysts of K-feldspar in a matrix of fluxioned alkali feldspar laths and a little interstitial quartz. The rock may have originally contained about 5% mafic minerals but this is now largely replaced by haematite accompanied by late-grown fluorite and carbonate.

330. Tuleih Complex, near northern end of main pluton (P3). Sheet (1–2 m) cutting syenite. Feldspar-phyric microsyenite with antiperthitic albite and perthite as the chief constituents, accompanied by riebeckite-arfvedsonite and a little quartz, both interstitial to the feldspars.

400. Tuleih Complex, from small boss of riebeckite-biotite microgranite N. of the main pluton (P4). Phenocrysts of perthite (1 mm) and smaller quartz in a microgranite matrix with red-brown biotite locally enclosed by riebeckite-arfvedsonite. Fluorite and zircon are common accessories.

418. Tuleih Complex, from a small dyke of dark-green rock cutting gneisses W. of the main pluton (O3). Peralkaline microgranite with microcline, quartz and zoned sodic clinopyroxenes as the chief components, both as phenocrysts and in the groundmass. Some of the larger pyroxenes form clusters with haematite and are fringed by sodic amphibole. A little sodic plagioclase is enclosed with microcline, and zircon is an abundant accessory.

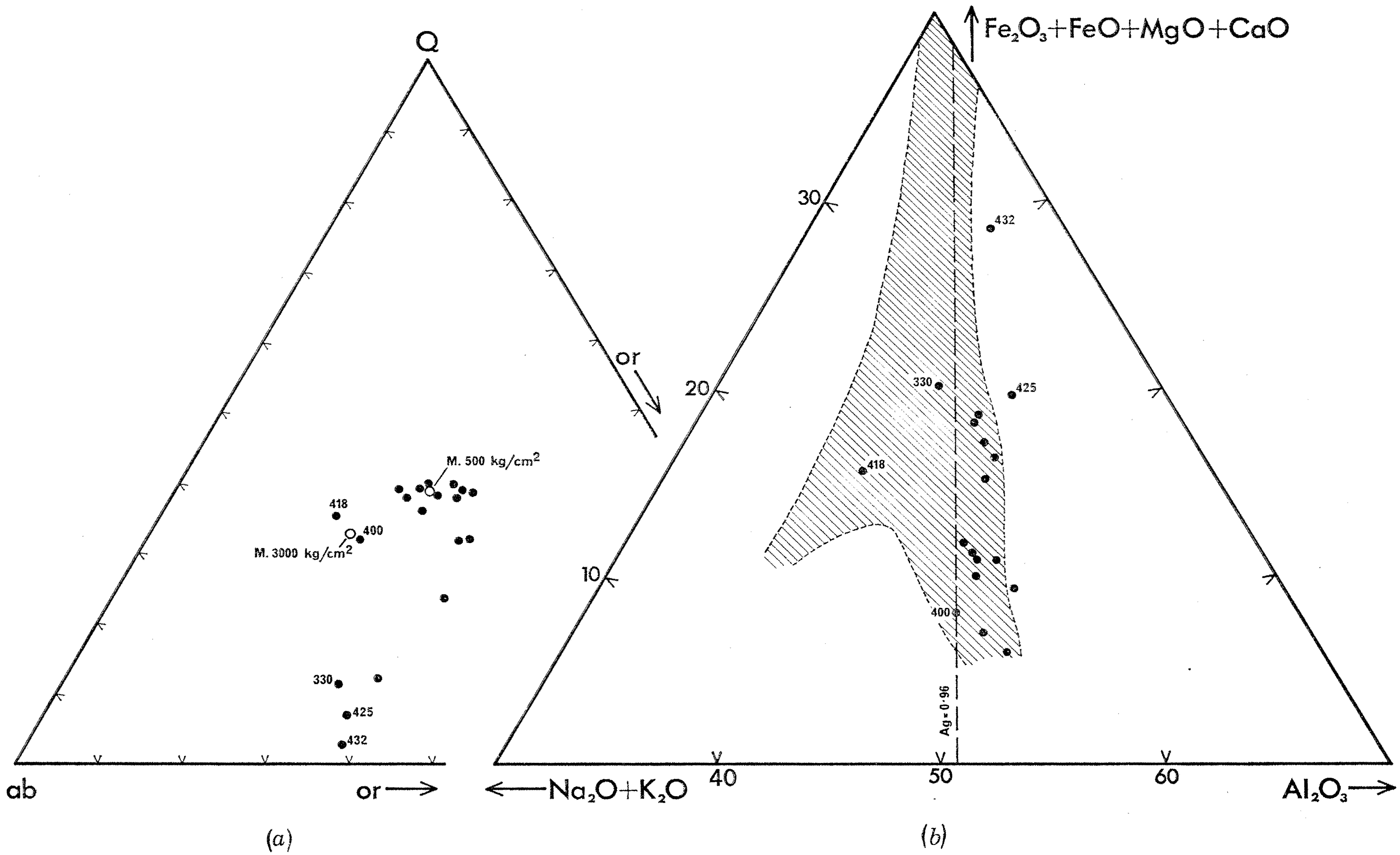

FIGURE 10. (*a*) Normative quartz, albite and orthoclase of Sabaloka intermediate and acidic rocks projected onto the anhydrous base of the system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2–H_2O. The minima for water vapour pressures of 500 kg/cm^2 and 3000 kg/cm^2 are from Tuttle & Bowen (1958). Specimens 330, 400 and 418 are rocks containing sodic ferromagnesian minerals as essential constituents. Points 425 and 432 are quartz syenites from the Tuleih Complex. (*b*) Sabaloka intermediate and acidic rocks plotted on part of the triangular diagram basic oxides/alkalies/alumina expressed in molecular proportions. The shaded area is the field of the Nigerian Younger Granite association as given by Jacobson *et al.* (1958). A line representing an agpaitic coefficient of 0.96 is also shown on the diagram.

On the other hand, the rocks of the Sabaloka Cauldron are poorer in Na_2O and have lower Na_2O/K_2O ratios than most analysed Nigerian granites. In these, Na_2O normally exceeds 4 % even in non-peralkaline rocks, whereas in all except one of the Cauldron rocks values are well below this figure. This distinction does not, however, extend to the rocks of the Tuleih Complex.

Major oxides of the Sabaloka suite show quite regular variations when plotted against silica in the range SiO_2 61–77 % (figure 9): Al_2O_3, total iron, CaO and (more surprisingly) total alkalies all decrease steadily with increasing SiO_2. However, if Na_2O and K_2O are examined separately it becomes evident that these oxides are at consistently different levels in the Cauldron and Tuleih Complexes, although in both Na_2O tends to rise and K_2O to fall in the most acidic rocks. The differences in alkali content coupled with the tendencies towards peralkaline acidic rocks at Tuleih and peraluminous acidic rocks in the Cauldron suggest that the magma batches from which the two Complexes crystallized were of separate, though doubtless related, origin.

A plot of basic oxides against alumina and total alkalis (figure 10*b*) illustrates the point that the Sabaloka suite is on the whole richer in alumina than the Nigerian Younger Granites. Moreover, the mineralogical indicators of alumina deficiency first appear in rocks with agpaitic coefficients less than unity; the line Ag = 0.96 in figure 10*b* approximates to the boundary between the two mineralogies and passes through a transitional rock (400) containing both riebeckite and biotite. Several of the Cauldron rocks plot close to this boundary. The appearance of sodic ferromagnesian minerals in rocks not strictly peralkaline in chemistry is possibly an indication that the late-stage residuals were fractionating in a peralkaline direction due to the earlier crystallization of an aluminous phase, probably a lime-bearing plagioclase or biotite. Such a process is only possible in a magma which is already on the borderline of alumina deficiency (Bailey & Schairer 1966). Alternatively it is relevant to note that Noble (1970) and Macdonald (1975) both conclude from analyses of peralkaline obsidians and equivalent crystalline and secondarily hydrated rocks that significant losses of Na_2O often occur during crystallization. Such losses would, of course, reduce the value of the agpaitic coefficient.

Jacobson *et al.* (1958) remark upon relatively high Rb contents in Nigerian Younger Granites while Mackay, Beer & Rockingham (1952) and Bowden & van Breemen (1972) give figures showing that Rb enrichment is particularly a feature of the more albitic granites, both peralkaline and non-peralkaline. Many of these rocks have K/Rb ratios of less than 100, compared with the normal range of 150–300 found by Ahrens, Pinson & Kearns (1952). Sabaloka shows similar Rb enrichment in the strongly peralkaline microgranite (418) and non-peralkaline mica granite (291, 486). However, when ratios for the whole suite are superimposed on the K/Rb plot of Ahrens *et al.* (1952) (figure 11) it is clear that apart from the modestly enriched mica granite most of the Cauldron rocks fall within the normal scatter limits whereas the Tuleih specimens display a much wider spread, with strong Rb depletion in syenite (432, 425) contrasting with the notable enrichment in aegerine microgranite. Ahrens *et al.* consider anomalously low K/Rb ratios to be indicative of highly fractionated residual magmas whereas Bowden & Turner (1974) speculate that hydrothermal activity believed responsible for late-stage albitization may have upgraded the Rb content of some Nigerian granites. Where there is a gradation from depletion in more basic rocks to enrichment in acidic rocks of the same complex, as at Tuleih, a process of fractionation seems likely. The situation is reminiscent of that found by Upton (1960) within layered syenites of Kûngnât (Gardar Province) and the sheets of soda granite which intrude them. At Kûngnât there is incontrovertible evidence of crystal fractionation within the syenites, and Upton relates Rb depletion in early cumulates to relatively high contents of Ca and Ba in alkali feldspars (Taylor & Heier 1958; Heier & Taylor 1959). A similar situation may have obtained during crystallization of the Tuleih syenites, which also have relatively high contents of Ca and Ba. Upton interprets the Rb-enriched peralkaline granites as low temperature residues of the layered syenites they inject. Here, again, the acidic peralkaline rocks show indications of post-consolidation recrystallization, including albitization.

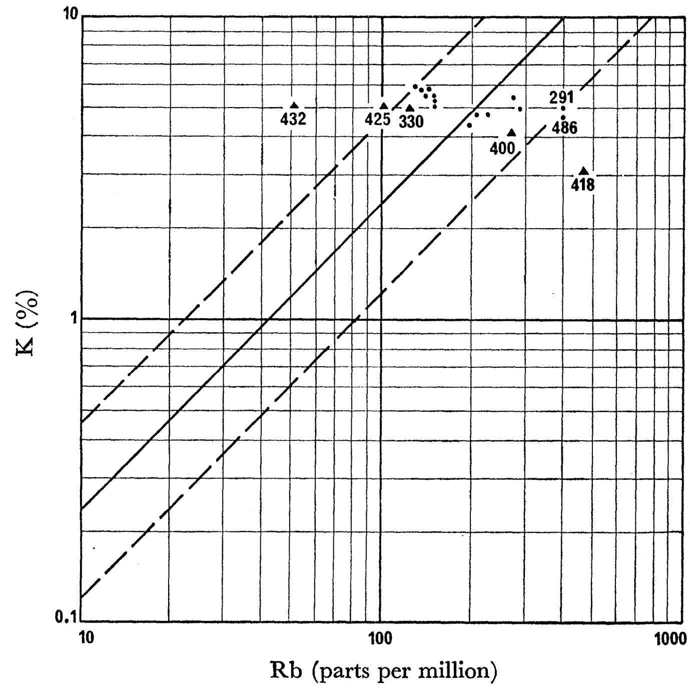

FIGURE 11. Plots of potassium against rubidium for Sabaloka rocks, superimposed on the K/Rb curve and scatter limits of Ahrens *et al.* (1952). ▲, Rocks from the Tuleih Complex; ●, rocks from the Cauldron Complex.

(*b*) *Petrogenesis*

Schemes which seek to account for the origin of the Younger Granite type of association include those which regard the acidic rocks as residuals of crystal fractionation and those supposing that the granite magmas are essentially partial melts of crustal or upper mantle rocks.

Jacobson *et al.* (1958) tentatively suggested that amphibole-fayalite granite magma or syenite magma may have been parental to the more acidic granite types, but they were unable to suggest any mechanism for the split into aluminous and alumina-deficient branches. Black (1965) proposed a basaltic parent magma of tholeiitic affinity; this may fractionate directly to give the non-peralkaline series, but on some occasions there is an initial separation of anorthositic cumulates, leaving an alumina-depleted magma which subsequently fractionates to give peralkaline acidic residua. This idea originated in suggestions by Upton (1960) on the petrogenesis of the Gardar rocks, coupled with the presence of early anorthosites in several of the Aïr complexes of Niger, but anorthosites are uncommon in Younger Granites as a whole.

There are several variants on the general thesis that the acidic rocks originated as partial melts. Bailey & Schairer (1966) proposed that primative granite magmas of non-orogenic regions are normally peralkaline and are partial melts of basic or intermediate rocks in the lower crust or upper mantle. Since most sialic rocks are peraluminous there will commonly be contamination of the primative melt to give the associated aluminous granites. Bowden (1970), on the other hand, initially preferred a theory deriving both granite series from upper crustal rocks, early melts being peralkaline and more extensive melts aluminous. However, experimental melting of supposed parental monzonite (Brown & Bowden 1973) seems to show that while this process may yield aluminous granites it is unlikely to generate peralkaline liquids. Moreover, strontium isotope data (van Breemen, Hutchinson & Bowden 1975) on the whole favour deeper origins, at least for the aluminous granites. In one of the most recent syntheses (Bowden & Turner 1974) it is proposed that melting of a high pressure equivalent of granodiorite in the lower crust generated peraluminous liquids which rose diapirically to the brittle zone of the upper crust, preceded by a wave of basaltic magmas from the mantle. Here, sufficient superheat remained to initiate localized partial fusion of basement granites to give peralkaline liquids.

While not claiming that the Sabaloka data throws new light on the petrogenesis of Younger Granite associations a few comments may be useful if only to reinforce observations made by previous authors.

(i) Widespread eruption of basic lava at the beginning of volcanicity must mean that the developing thermal anomaly first became an effective generator of magmas at upper mantle levels. The subsequent cut-off in supply of basic magma during the main phases of activity is less likely to have been due to cooling in the mantle than to the interposition of a physical barrier between mantle and surface, possibly in the form of a laterally extensive zone of melting in the crust. Gravity evidence makes it clear that no large volumes of dense material were ever emplaced high in the crust, so precluding any appeal to fractional crystallization of basic magma at low pressure as a source for the felsic rocks (see also Ajakaiye 1970).

(ii) The proposal of Jacobson *et al.* (1958) of parental magmas in the compositional range syenite to sub-acid granite may be a viable alternative to low pressure basic magma fractionation and is not ruled out by gravity considerations. The volume relations and relative ages of the rocks within the Tuleih Complex are compatible with a mechanism involving the limited fractionation of a large volume of parental quartz syenite magma to give small amounts of subaluminous and peralkaline residua. The problem is, as usual, to explain how a distinctly non-peralkaline parent could have fractionated in a peralkaline direction. In the Cauldron Complex, where the question of peralkalinity does not arise, only a few small intrusions of trachyte and the early, subsilicic phase of the ring-dyke indicate the former existence of anything other than thoroughly acidic magmas. If the chemistry of the analysed subacidic rocks (table 5, 106a, 144) is any guide a syenitic parent magma here might have been poorer in lime and soda than the syenite at Tuleih.

(iii) Plots of normative albite, orthoclase and quartz projected onto the anhydrous base of the 'granite tetrahedron' (figure 10*a*) show marked clustering of the high silica, non-peralkaline rocks of the Cauldron around the minimum for 500 kg/cm² water pressure (cf. Freeth 1972), while less acidic rocks plot in or near the thermal valley within the feldspar field. Such a distribution can be attributed to either fractional crystallization or equilibrium partial melting. The plotting of the two peralkaline acidic rocks (400, 418) near to the 3000 kg/cm² minimum is clearly anomalous, since these are hypersolvus rocks and their plotted positions cannot be explained by high water pressures. Bailey & Schairer (1966) have pointed out that the system as devised is a residua system only for liquids in which the alkali-alumina ratio is near unity and remark that discrepancies between plotted peralkaline compositions and the quartz-feldspar minima have no intrinsic significance. Even so it is interesting to find that the plotted positions of the two peralkaline rocks from Sabaloka are displaced from the low pressure minima in much the same sense as are the 'strongly albitized' peralkaline and non-peralkaline granites shown by Bowden & Turner (1974). These authors suggest that hydrothermal modification may be responsible for their anomalous compositions. On the other hand, the work of Carmichael & MacKenzie (1963) and Thompson & MacKenzie (1967), discussed recently in this context by Roux & Varet (1975), suggests that in peralkaline quartz-alkali feldspar systems the thermal valley leads from the binary minimum on the alkali feldspar join towards a ternary minimum nearer to the albite-quartz side line than in the non-peralkaline 'granite system'. The more strongly peralkaline the liquids the more the thermal valley is diverted towards a direction of soda enrichment. Plots of peralkaline acidic rocks on the sodic side of the thermal valley of the granite system may thus be explained by crystal-liquid equilibria as plausibly as by appeals to hydrothermal albitization. Such an explanation cannot, of course, be extended to anomalously sodic non-peralkaline granites.

8. Tin–tungsten mineralization

A detailed account of a small area of tin–tungsten mineralization immediately west of the Sabaloka Plateau has been published elsewhere (Almond 1967). In brief, quartz veins carrying cassiterite and wolfram have been found at El Grune El Zurug (D6), north of Abu Dom, mainly within an early phase of the ring-dyke microgranite but also in the margin of the mica granite and adjacent country rocks. The veins are individually very small but form a stockwork on El Grune El Zurug itself. As well as cassiterite and wolfram the quartz veins carry minor amounts of chalcedony, iron oxides, molybdenite, fluorite and traces of galena, malachite and various other secondary minerals. A lenticular body of greisen within the stockwork contains druses lined with quartz and cassiterite.

A few comments can be added here derived from later work carried out mainly by Khalil (1972, 1976 and personal communications). In 1967 (Almond 1967, p. 9) it could not be proved that mineralization was associated with the mica granite rather than the ring-dyke microgranite although the mineralogy of the mica granite made it seem the more likely parent. This was confirmed when a more careful search showed that the veins are not entirely restricted to the neighbourhood of El Grune El Zurug but occur at several other localities along the north-western and southeastern contacts of the mica granite. They are, however, very few in number, small and only rarely carry traces of cassiterite or wolfram. Along the northwestern contact greisenization surrounds areas where veins are found. Veins in the country rocks are rarely far from the mica granite and at El Grune El Zurug itself the mica granite is separated from the main stockwork only by a narrow intrusion of younger ring-dyke microgranite (and not also by basement rocks, as is incorrectly shown in Almond 1967, figure 2). At one locality, however, veinlets found as far as 1 km from the mica granite (E7) carry a little molybdenite, green fluorite and green mica. These veinlets are within the main intrusion of the ring-dyke and may be younger than those at El Grune El Zurug. It is the presence of molybdenite which is significant here, since fluorite-bearing veins are found widely, if sparsely, in both the Cauldron and Tuleih Complexes.

Khalil (1976) has made a detailed geochemical survey of the secondary dispersions of a number of elements in the residual soils of the El Grune El Zurug area. He has been able to demonstrate anomalous values in a number of related element groups and has shown that from north to south there is a spatial zonation in the order W – Sn – Mo – Zn – Pb – Ag. Correlation coefficients suggest three phases of mineralization:

(i) W, Sn, Mo.
(ii) (Pb, Ag, Bi, As), Zn, (Mo, Wo, Sn).
(iii) Pb, Ag, As, Bi.

Estimates of productivity based on these dispersion haloes suggest to Khalil that the amounts of metal are uneconomically small. A similar conclusion was reached after exploratory drilling by the Digna Company of Khartoum. In a wider context the discovery of this tin–tungsten mineralization may prove a useful pointer. Some of the Nigerian Younger Granites are important sources of tin, and there are many Younger Granite complexes scattered throughout northern Sudan among which a search should be made for richer mineralization than is found at Sabaloka.

9. Discussion of the igneous history of the Complex

One cannot be certain whether the igneous activity at Sabaloka preceded or followed the Pan-African orogeny, or how closely, but since the effects of this event in north-central Sudan were mainly limited to a widespread but relatively modest rise in temperature, with little or no accompanying tectonism, it would seem reasonable to label the complex 'anorogenic'. Such a conclusion apparently holds good for other north African Younger Granite associations of different ages and in various regions. In the terminology of plate tectonics, this was within-plate, continental activity.

The predominantly basic volcanicity with which activity began was probably the first surface expression of a developing thermal high in the upper mantle. The basic lavas which now form the lowest unit of the volcanic succession may be only remnants of a much more extensive, uncentralized volcanic field. Unfortunately, the altered condition of these rocks does not allow their unequivocal assignment to any of the currently recognized basaltic associations, though it is clear that the rocks were neither strongly alkalic nor notably silica oversaturated.

The almost exclusively acidic activity which persisted throughout the greater part of the volcanic history followed so abruptly on the initial basic phase that it is easy to believe that there was a sharp change in the magma source zone at this time, perhaps from the upper mantle into the lower crust. Moreover, activity simultaneously became centralized and began to build up a major volcanic superstructure. At first the erupted rocks were almost exclusively lava flows, though extrusion was intermittently punctuated by explosive outbursts. The virtual absence of phenocrysts from the oldest rhyolite flows and their rounded shapes in later flows suggests that the magmas were being generated in a relatively dry environment so that they remained near to, or even rose above, their liquidus temperatures as they moved upwards into lower pressure environments. The extensive nature of individual flows and the sparsity of pyroclastic rocks also argue that the viscosity and volatile content of the magmas were low at this stage. One important interval of explosive eruption did, however, interrupt the steady outpouring of lavas and produced the Cataract formation, with its mixture of low-temperature ash-flows, sparse air-fall deposits and water-lubricated debris flows.

The second phase of acidic volcanicity was characterized by voluminous eruptions of high-temperature ash-flow deposits, but during the interval preceding it several important events occurred which imply significant changes in the configuration of the underlying magma chambers. The phase of rhyolite lava extrusion ended with subsidence of the superstructure, forming a structural basin centred upon an area in the southeast which presumably lay above the apex of the early magma chamber. No ring-fracturing accompanied this subsidence, perhaps indicating that the chamber was either too deep-seated or not laterally extensive enough to permit annular fracturing of its roof rocks. Subsidence may have temporarily sealed off outlets to the surface or the supply of magma from depth may have waned, but in any case the formation of the lava basin was followed by a quiescent interval during which the early volcanic rocks were deeply eroded. This interval was long enough to allow the development of a local relief of several hundred metres in the rhyolite uplands, but its length must not be overestimated since in the low ground to the northeast there was little erosion of the rocks later covered by ignimbrites of the second phase. During the interval magma was probably accumulating at a high level in the crust and extending the old reservoirs towards the north. Consequently the roof became progressively weaker and was perhaps eventually domed by increasing pressure of magma below it. Conditions became right for initiation of the Ring Fracture Zone, and when ash-flow eruptions began they were fed partly by vents located on annular fracture lines. The earliest ash-flows, represented by the Scarp ignimbrites, were restricted to the southern half of the present Cauldron and may have been fed by vents which tapped the old magma chamber but the succeeding Plateau/Graben ignimbrites spread widely and must have extended beyond the present limits of the ring-structure. These younger ash-flows emerged from dyke-shaped vents situated both along the Ring Fracture Zone and within the circumscribed block. It is at about this time, too, that a caldera must have begun to develop, although it is difficult to discern its relationship to the individual ignimbrite formations. However, the correlation of the Graben ignimbrite with the upper part of the Plateau ignimbrite, proposed in this paper, carries with it the requirement that the Scarp and Plateau ignimbrites overlapped onto older rocks in the vicinity of the present Graben, and it is possible that these formations were partly confined in that direction by caldera walls. But the main increment of cauldron subsidence was certainly added after the youngest of the exposed ignimbrites were erupted, for all formations take part in the broadly centrepetal structure of the Plateau, defining a centre of subsidence west of the Nile.

Though chemically not very different from the acidic lavas the rhyolitic ignimbrites are notably dissimilar in containing a high proportion of primary crystals, mainly quartz and two feldspars but also small amounts of clinopyroxene, fayalitic olivine and biotite. This is understandable if the magma were stored for some time in shallow magma chambers, and the degree of crystallization attained may have been sufficient to raise the volatile content of the residuals high enough to ensure that when eruption finally occurred it was explosive in nature. Not, however, excessively explosive, for there is little in the way of air-fall deposits associated with the ignimbrites, apart from thin vitric tuffs which in places underlie the Plateau ignimbrite and no doubt record brief Plinian phases at the beginning of ash-flow eruptions. Moreover the way in which the ignimbrites of the Plateau pass continuously downwards into similar material filling dyke-shaped feeder vents (§ 3 (*d*) (ii) and Almond 1971) suggests that the mode of eruptions was by the 'boiling over' of highly frothed and disaggregating magma rather than by the avalanching of density currents from the overloaded base of a vertical eruption column. Upward increase in crystal content within the Plateau/Graben sheet is evidence for crystal settling within the magma chamber, so that the first erupted ashes were relatively impoverished in crystalline material. Inspection of chemical analyses of ignimbrites and ring-dyke rocks with different crystal contents (table 5) shows, however, a very poor correlation between total content of crystals and chemical composition, which may reflect sorting between the different crystal species but more like merely means that the difference in bulk composition between crystalline phases and residual liquids in these near-minimum temperature magma compositions was so small that there was little scope for effective chemical fractionation.

The major intrusions of the Cauldron Complex, that is, the ring-dyke and mica granite, also post-date all preserved volcanic rocks. Because of petrographic similarities it is believed that the ring-dyke magmas derived from the same chamber as did the Plateau/Graben ignimbrites. The mica granite is distinctly different and presumably occupied a different chamber or cupola, but field evidence shows that its intrusion overlapped with successive ring-dyke injections. It is therefore quite possible that sinking of the Cauldron block provided the force which drove both these magmas upwards into their present positions. At this point one encounters the question of whether the Northern Dome was also formed at this time, by differential subsidence, or later, due to a resurgence of magma pressure. Although it requires an additional pulse of magma the writer favours the second explanation, partly by analogy with younger cauldrons in which the evidence is much clearer, but also because it neatly accounts for the Graben structure of the northern sector of the Ring Fracture Zone and for the presence of a 'hidden granite' beneath the Dome, detected by gravimetric survey.

Finally, it is unfortunately impossible to say much of the rôle played by the Tuleih intrusions in the development of the Complex as a whole, for even the broad age relations of these rocks to those of the Cauldron is unknown. What is evident is that the origin or evolution, or both, of the magmas involved was in some way critically different from that in the Cauldron, so that syenites and minor amounts of peralkaline intermediate and acidic rocks are distinctive constituents.

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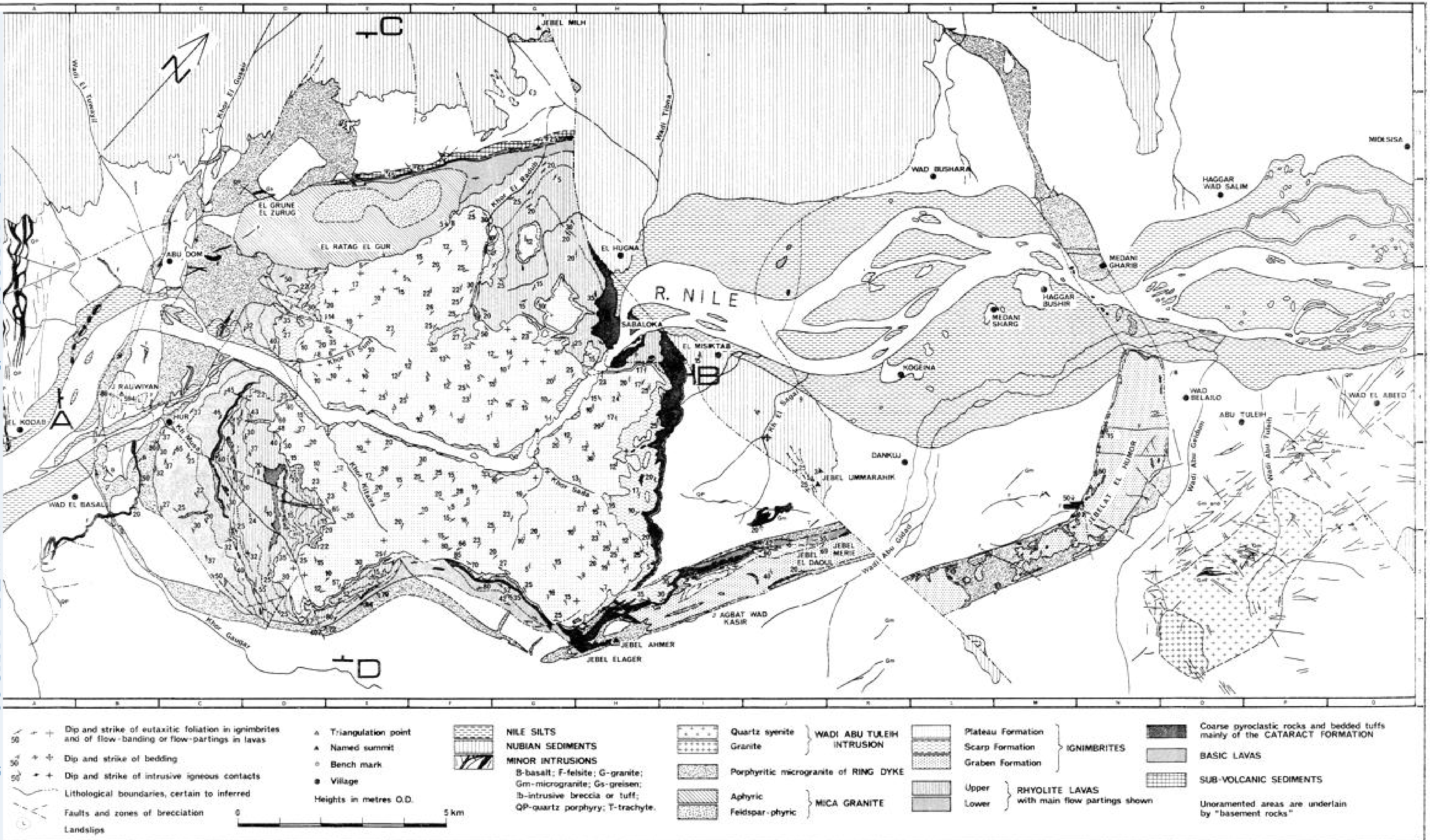

Figure 1. Geological map of the Sabaloka igneous complex. A–B and C–D mark the lines of section shown in figure 4.

(Facing p. 597)