Structure and Evolution of the Paresis Igneous Complex, South West Africa

by

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ABSTRACT

The Jurassic igneous province of northern Damaraland comprises a series of ring-complexes with which Paresis is correlated. Central igneous activity at Paresis consisted of an earlier, predominantly volcanic phase followed by the intrusion of felsic ring-dykes and stocks. The intrusive rocks were emplaced into the subsided volcanic superstructure.

Volcanic products were exclusively rhyolitic and three cycles are distinguished, characterised, respectively, by felspar rhyolite, quartz-felspar porphyry and comendite. Each of these commenced with the building of a high stratovolcano and ended with the intrusion of dykes of a similar composition. Cycles I and III were terminated by cauldron subsidences and a final collapse occurred along a ring-fault encompassing both of the earlier cauldrons. The regular, centripetal dips of the rhyolite are attributed to the control of subsidence by conical fractures. Distribution and textures of the rhyolites are consistent with their emplacement as incandescent ash flows which were subsequently compacted, welded and recrystallized.

Basalt flows—intercalated with rhyolites of Cycles I and II—and associated gabbroic dykes, are correlated with the regional Stormberg volcanism and are not directly linked to central igneous activity.

Dykes of syenite, bostonite, and microgranite are shown to have been intruded, in that order, under high magmatic pressure along conical fractures. The arcuate form of the syenite dyke of the Central Massif is ascribed to asymmetric collapse of the comendite caldera-block. The different focal depths of the syenite and bostonite conical dykes—about 17 Km. and 6 Km., respectively—are taken as evidence of a compositional stratification in the magma reservoir. Last in the intrusive sequence was a composite stock of felspathoidal rocks—located on the final ring-fault.

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I. INTRODUCTION

A. General Geology of Northern Damaraland

1. *Pre-Karroo Rocks*

The basement of this region consists of a coarse-grained, porphyritic granite, the age of which has been determined at 1,700±70 m.y. (Clifford *et al.*, 1962). Overlying the granite are strongly folded and metamorphosed sediments of the late Proterozoic Damara System comprising essentially quartzites, marbles and mica-schists, with a total thickness of probably more than 10,000 metres. The regional structure of the Damara is characterised by intense folding along NE-SW axes, but deviations from this trend are common. The metasediments are extensively intruded by concordant bodies of syntectonic Salem granite showing aplitic and pegmatitic phases locally. Numerous age determinations on pegmatite minerals from the Karibib area (Nicolaysen *in* Holmes and Cahen, 1957; Jamieson and Schreiner, 1957; Clifford *et al.*, op. cit.) fall within the interval 510±40 m.y.

2. *The Karroo System*

Resting unconformably on the irregularly eroded Damara formations is a succession of coarse terrestrial sediments—chiefly conglomerates, arkoses, sandstones and minor shales (Gevers and Frommurze, 1929)—for which a Karroo age (Permian to mid-Triassic) has been established by Reuning (1924a) and Gürich (1926) from fossil evidence. The sediments have been neither folded nor metamorphosed, except locally, and the present distribution of their outliers (Fig. 1) indicates an extensive area of deposition, considered by Korn and Martin (1954) to have been a shallow basin elongated in a northeasterly direction.

These sediments are followed by great thicknesses of volcanics which transgress the margins of the depositional basin in many places. Reuning (1924a, 1929) has described the extrusives—a succession of basalts, andesites, and rhyolites—and assigned to them a Stormberg (i.e. early Jurassic) age. Numerous dolerite dykes, with granophyric phases, traverse the Damara and Karroo deposits. Most of the dykes strike NNE but some, particularly near the coast, have a WNW strike.

3. *Post-Karroo igneous activity*

Following the Stormberg volcanism, a final phase of igneous activity in northern Damaraland became localised at several centres which are distributed along a belt trending roughly north-east (Fig. 1), within the Damara orogen.

Present knowledge concerning the post-Karroo igneous complexes has been reviewed by Martin, Mathias and Simpson (1960) and King and Sutherland (1960), who also give comprehensive bibliographies of the relevant literature.

Several of the complexes (Cape Cross, Messum, Brandberg, Okonjeje, Erongo) are intrusive into Karroo rocks but their upper age limits have not been established with certainty. The igneous rocks of Paresis penetrate the Damara metasediments but are nowhere in contact with later formations. However, whole-rock Rb/Sr determination on a suite of Paresis comendites has yielded an age of 130 m.y. (personal communication: Mr. W. Manton, Bernard Price Institute of Geophysical Research, Johannesburg) which—on the basis of regional and petrographic association—can be considered applicable to the other Damaraland post-Karroo ring-complexes.

The complexes exhibit considerable petrographic variation. The discordant granitic plutons of Brandberg, Gross and Klein Spitzkuppe, and Erongo, contrast

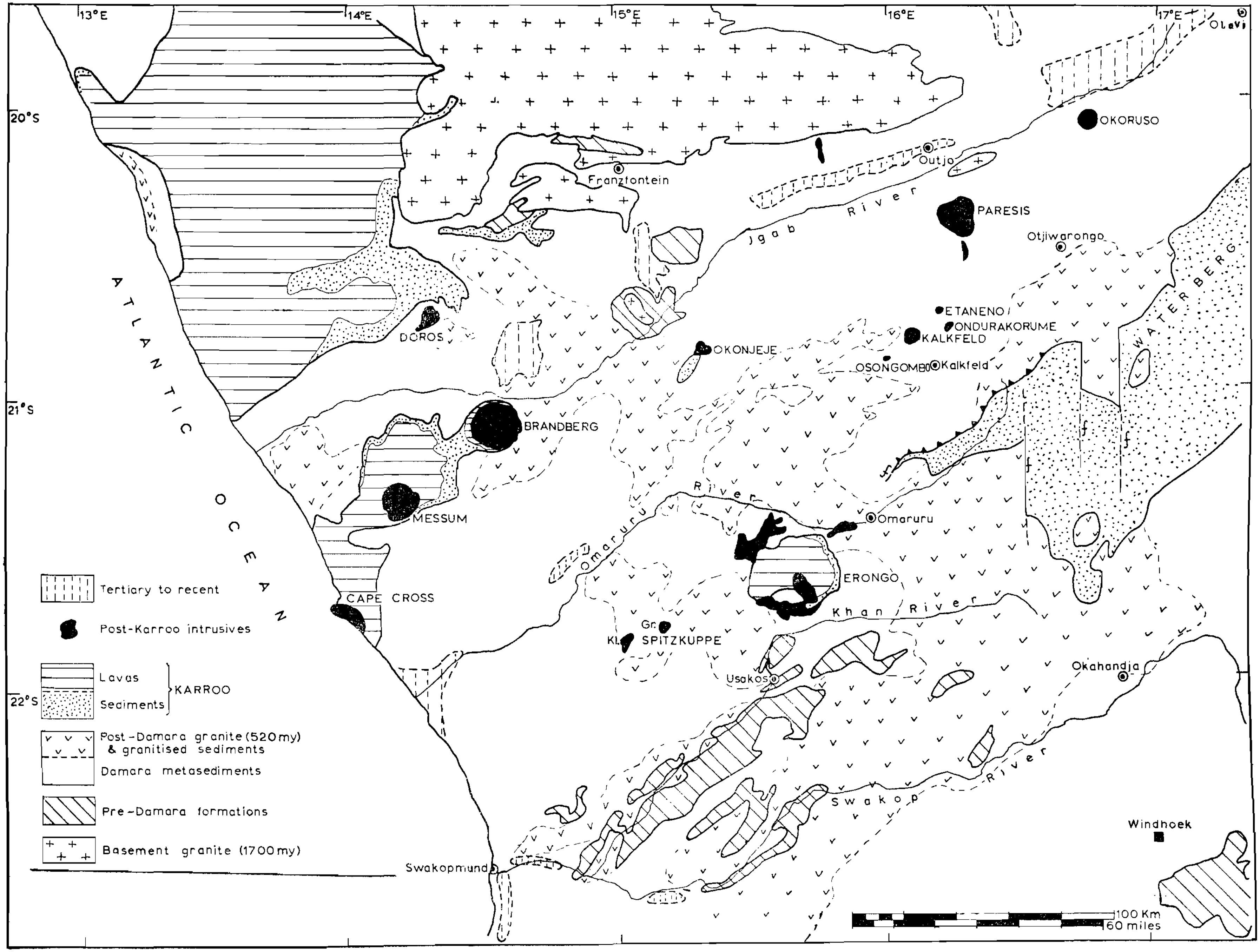

Fig. 1. The Damaraland igneous province (modified after H. Martin)

with the differentiated complexes of Cape Cross, Messum, Okonjeje, and Paresis which comprise a wide range of rock types with marked alkaline affinities. Doros is a volcanic neck consisting of olivine-dolerite which is cut by aegerine-bearing bostonite dykes (Reuning and Martin, 1957). The igneous centres of Kalkfeld, Ondurakorume, Osongombo, and Okoruso are characterised by prominent carbonatitic components, syenitic and foyaitic rocks. There is no apparent correspondence between petrographic character and location of the complexes, but the clustering of the carbonatitic centres at the northeastern end of the belt may be significant.

Structural studies by Cloos (1919), Korn and Martin (1954), Simpson (1954), and others, including the present author, leave no doubt as to the ubiquity of ring structures in the Damaraland igneous complexes. In some (Messum, Brandberg, Erongo, Okonjeje, Paresis) there is evidence of cauldron subsidence associated with the igneous activity. Korn and Martin (op. cit.) consider it likely that Kalkfeld and Okoruso are also caldera-like structures.

The suggestion, first made by Clough *et al* (1909), that at least some igneous ring-complexes are the subvolcanic expressions of calderas is applicable to the Damaraland igneous province. We may thus view the above-mentioned complexes as a denudation series on the lines envisaged by Reynolds (1956) and Buddington (1959). The preservation of the volcanic superstructure at Paresis emphasizes its significance as the high-level counterpart—on structural, if not erosional, grounds—of associated complexes showing evidence of central volcanic activity.

B. Physiography of Paresis

The Paresis Mountains are situated about 270 Km. northwest of Windhoek, capital of South West Africa, and the Gross Paresis-North survey beacon lies on the co-ordinates 20° 15′ south and 15° 42′ east. Provincial roads connecting the towns of Otjiwarongo, Outjo, and Kalkfeld form a rough triangle about the complex and farm roads provide access to its outer ridges at numerous points.

The several structural units constituting the Paresis Complex (Fig. 2) are also reflected in the topography and, for ease of reference, it will be convenient to enumerate them. (*a*) The "Central Massif" of almost perfect circular outline (diameter 9 Km.), comprising inward-dipping rhyolite flows marginally intruded by syenitic rocks (Pl. III, Fig. 2); (*b*) the low hills of the "North-West Margin", built up of conical dykes intercalated with steeply tilted, truncated rhyolite flows; (*c*) the "Paresis-North Ridge"—a rectangular unit, exposing the entire volcanic succession, which has been tilted and elevated so that the flows now dip towards the Central Massif (Pl. III, Fig. 1); and (*d*) the semicircular "Paresis-South Area", southeast of Gemsbok Valley, made up of rhyolite flows and agglomerate. A broken chain of low hills, chiefly microgranite, form a circle (diameter 18 Km.) about the subsided volcanics. This circle represents the outer ring-fault delineating the Greater Paresis Cauldron, bedded rocks outside it showing no signs of having participated in the subsidence.

II. FIELD RELATIONS AND STRUCTURE OF THE IGNEOUS UNITS

A. General Statement

The regional NE strike of the Damara basement is reflected by the prominent marble ridges to the northeast of the complex. North and south of the complex the strike diverges, so that at the Kameelfeld dyke it trends 10° north of west, whereas at Klein Paresis the basement strikes NNE. West of the complex the basement is thrown into relatively tight folds with a general southwesterly plunge. It is evident that Paresis is situated at a structural discontinuity. This poses the question of the extent to which the local deformation may be attributed to the effects of Paresis volcanism.

The outer ring-fault (RF-3, Fig. 2), delineating the Greater Paresis Cauldron, is a plane along which major tilting and subsidence took place. In contrast with the dislocated volcanics inside the ring-fault, Damara sediments exposed along its northern periphery show virtually no deviation from their general attitude—even up to within a few metres of the fault. This can only imply that the crustal displacements attendant on the Paresis volcanism were restricted to the interior of the outer ring-fault and that the structural discontinuity of the basement antedates it. The possibility of large-scale updoming of the country-rocks, as a precursor to the effusive phase, was considered but found to be unsupported by field evidence. Abrupt local deviation from the regional attitude is a common feature of the basement in Damaraland and is associated with intrusions of the late Precambrian Salem granite (Gevers and Frommurze, 1929). Outcrops of this syntectonic granite occur only about 25 Km. to the southeast of the complex and it is reasonable to attribute to it the local deformation of the Damara metasediments.

The Damara System in the immediate environs of Paresis is represented chiefly by fine- to medium-grained marbles, pelitic quartz-biotite schists, and felspathic quartzites, correlated with the Khomas Series (upper Damara). The complex is situated in the biotite hornfels zone fringing the northwestern margin of the central granitized belt of the Damara system. Alteration commensurate with low-grade, regional metamorphism may be observed in all basement rocks exposed. Contact-metamorphism related to the Paresis intrusives, however, is weakly developed and is confined to several marble outcrops on the northern border.

The pattern of igneous activity at Paresis may conveniently be divided into an earlier volcanic phase, characterised by the production of great quantities of rhyolite from several vents, and a later intrusive phase during which syenitic rocks were emplaced along pre-existing fractures. It should be noted, however, that throughout the volcanic cycle rhyolitic dykes were intruded intermittently, the most recent of which were synchronous with the first syenite ring-dyke. The volcanic rocks are confined to the area of subsidence, and of the intrusives only several linear dykes occur outside it.

B. The Rhyolites

In areal extent, rhyolitic extrusives occupy about 210 sq. Km., or nearly 90% of all igneous rocks exposed in the Paresis complex. Three volcanic cycles, each comprising a distinctive assemblage of rhyolites, have been recognised: 1. felspar rhyolite, 2. quartz-felspar porphyry, and 3. comendite, emplaced in that order. Each cycle commenced with effusive activity, building high cones, and ended with the emplacement of dykes of similar composition. Basalt flows are intercalated with the rhyolites of cycles 1 and 2.

1. *Felspar rhyolite*

This—the lowest member of the volcanic sequence—is typically a dense, hard rock with phenocrysts of alkali felspar set in a brown to purplish-grey, aphanitic matrix showing fluidal textures in places. Quartz phenocrysts are rare and iron oxides, probably pseudomorphous after pyroxene or amphibole, are the only dark minerals.

Igneous activity began with the effusion of alternating flows of streaky, vesicular felspar rhyolite and amygdaloidal basalt. Thin, irregular lenses of these rocks are interbedded and, together with later intrusives, form a discontinuous arc of low foot-hills on the outer flank of the Paresis-North Ridge and the North-West Margin. This succession is exposed only on the northern periphery of the complex where the tilted and truncated flows have centripetal dips, commonly in the range 30° to 50°.

The felspar rhyolite is most extensively developed in the Paresis-South Area. A traverse from the eastern margin, along the line of section D-E (Pl. IV), reveals a sequence of irregularly alternating, massive flows and densely welded pyroclastic deposits which are inclined towards centre F with dips ranging from 35° to 70°. Pyroclasts vary considerably, both in form and size, but in all occurrences consist of the same material as their matrix. Spherical to ovoid lapilli and bombs, up to 20 cm. in diameter, form crudely bedded deposits which show a characteristic knobbly weathered surface. Their occurrence and form, which commonly includes a flattened pedicle, speak for their ejection in a fluid or plastic condition. The bulk of the flow-breccia contains, in contrast, a jumbled assemblage of angular rhyolite fragments which are clearly lithic vent-ejecta. There is an increase in the size of the largest fragments, from about 50 cm. to rafts of more than 5 metres diameter near the vent. The rhyolite here occurs as large, low, domical outcrops in which the fragments are so densely fused into their matrix that it exfoliates in the manner of a completely homogeneous rock.

The converging dips of the rhyolite flows in the Paresis-South Area indicate centre F as the focal point for the collapse of the volcanic cone. Thinning of the felspar rhyolite deposits with increasing distance from centre F, further mark this area as the approximate location of the summit of a former volcanic cone. Thicknesses of the deposits constituting this cone (measured perpendicular to the base of the outermost flows) are as follows:- Paresis-South Area, section D-E—2,500 m.; below Gross Paresis-North—1,000 m.; North-West Margin—350 m. Welded deposits of rounded air-fall pyroclastics become progressively finer, more effectively sorted and thinly bedded with increasing distance from the vent (Pl. I, Fig. 1). On the Paresis North Ridge numerous beds of densely welded lapilli may be found which, upon weathering, produce knobbly aggregates in the upper levels of a flow grading into massive, homogeneous rock within a distance of 1 to 2 m.

Having adduced evidence to show that the distribution of felspar rhyolite material was radial from centre F, it is now possible to reconstruct the primordial volcano which these products represent. Section D-E (Pl. IV) passes through centre F, which may be taken as the approximate position of the summit of the hypothetical cone. The progressive decrease in the dip—from a maximum of about 70° in the peripheral flows, to about 35° near the centre—is consistent with the tilting of a wedge-shaped section thickening westward. Extrapolation of upper and lower flows, allowance being made for flexuring of the floor, produces a section of a formalized volcanic cone having a summit height of 5,000 m. and an angle of rest of 22°. The calculated height is a maximum value because widening of the summit crater by successive explosions will have considerably lowered its level. Nevertheless, it is remarkably close to the estimated height of 4-5,000 m. for the original Messum volcano (Korn and Martin, 1954). Using the height of 5,000 m. and the previously given thickness of felspar rhyolite deposits, extrapolation to zero thickness shows the following distances covered by the rhyolite in directions from centre F:— NNW—13 Km.; NNE—16 Km.; ESE—12.5 Km. These are great distances for rhyolite to be transported as lava, particularly when it is noted that the farthest flows are conspicuously free of brecciation. The bulbous forms, normally associated with silicic lavas, are also nowhere in evidence. If, however, these rocks were emplaced as mobile ash-flows, as the writer proposes, a mechanism for their transportation is immediately available and the above distances probably represent minimum values.

It is generally impossible to decide from outcrops or hand specimens what proportion of the angular fragments are of ejective origin and what the products of autobrecciation. The larger blocks, which diminish rapidly in size and abundance with distance from the vent, are probably entirely lithic vent-ejecta and rafts of earlier, consolidated rhyolite detached by the new flows. Even of the finer fragmental inclusions, only a small proportion could have resulted from autobrecciation. The picture which emerges then, is of a Vesuvian-type volcanism, characterised by the violent expulsion of both lithic and plastic ejecta. Modifying the Vesuvian pattern, is the simultaneous evolution of incandescent ash-flows, suggested by the densely welded matrix of the pyroclastics.

The Klein Paresis granophyre dyke is 6.5 Km. long and rises to a height of 1,882 m. The rock is remarkably uniform in composition and texture, characterised by zoned felspar in a micrographic ground-mass which is finely granular in places. Petrographically it resembles certain of the felspar rhyolite flows and chemically it is obviously a member of that group (Siedner, 1965). Accordingly, the granophyre has been interpreted as the hypabyssal representative of the felspar rhyolite, probably emplaced at the close of that cycle. Because of its massive fabric, the attitude of the Klein Paresis dyke at depth cannot be determined. However, the alignment of this dyke, the Paresis South micro-granite bodies, and the Rusthof bostonite dyke is markedly parallel to the Damara basement structures. Basic dykes, ranging from fine-grained dolerite to coarse olivine gabbro, intrude the granophyre at several localities and a dyke-like body of fine-grained quartz bostonite occurs on the western flank.

In the Paresis-South area, beds of coarse, unsorted air-fall agglomerate overlie the felspar rhyolite sequence. The compacted, but unwelded, agglomerate comprises fragments that vary in size from ash to blocks a metre or more in diameter, successive layers being separated by thin sheets of tuff. From a maximum thickness of 400 m. at Schwarzenfels, the agglomerate thins out northward to less than 30 m. below Monday Plateau. Felspar rhyolite material constitutes the base of the agglomerate which, at the Paresis-South Plateau and northward, becomes progressively enriched in quartz-felspar porphyry ejecta, giving way to solid flows above (Pl. I, Fig. 3). In the Schwarzenfels area this component is absent and the felspar rhyolite agglomerate is overlain directly by quartz-felspar porphyry flows. Apart from several small, scattered outcrops of agglomerate 10 to 20 m. thick in Kudu Valley, no other occurrences of unwelded air-fall pyroclastics have been found at Paresis.

2. *Quartz-felspar porphyry*

The rocks of this group are characterised by an abundance of quartz and felspar phenocrysts set in an aphanitic ground-mass whose colour varies between grey, purplish, and red-brown. Alkali felspar is the most prominent felspar but Na-rich plagioclase may also be present—a high proportion of these as angular fragments. Quartz occurs as shards and rounded grains commonly showing strong "corrosion". The fresh rock is hard and massive, and flow-tectures are confined to the extrusive rocks. Dykes generally have a well-developed granophyric texture which, however, is also found in many of the flows.

Quartz-felspar porphyry overlies the felspar rhyolite and intrudes it as fine-grained dykes. Two volcanic vents have been attributed to cycle 2: centre P1 in Gemsbok Valley and centre P2 on the Paresis-North Ridge. In the Paresis-South Area quartz-felspar porphyry flows constitute the dissected plateau-capping on the agglomerate but elsewhere in the complex they overlie the felspar rhyolite directly. Quartz-felspar porphyry flows are conformably overlain by comendite everywhere in the Paresis Complex. The comendite contains porphyry xenoliths in numerous localities and no evidence has been found indicating an overlapping of these two cycles. The occurrence of several bodies of quartz-felspar porphyry, apparently intrusive into the comendite is, therefore, puzzling.

In block B/4* several small outcrops of normal porphyry are surrounded by fine-grained, streaky comendite with irregular, disturbed attitudes but generally dipping away from the porphyry. The comendite is not veined by the porphyry but does, on the contrary, contain xenoliths of the latter. Conversely, no comendite inclusions have been seen in the quartz-felspar porphyry—in this area or elsewhere. These features, taken together with the absence of a hypabyssal texture in the porphyry, rule out the possibility of its having intruded the comendite in a magmatic condition. These relations are, however, adequately explained as a result of updoming produced by syenite rising into a thick sequence of rhyolitic flows along a zone of weakness. Presumably the great load of volcanics and the absence of a major fracture prevented the syenite from penetrating to the surface, which it was able to achieve only when it encountered the pre-existing ring-fault, a short distance to the northwest. This interpretation is illustrated in section E, Fig. 3. The two outcrops of quartz-felspar porphyry in block C/3 display the same general features as the B/4 outcrops and their present position is attributed to a similar mechanism. The only significant difference is that, being situated on a major fracture-zone (i.e. on ring-fault RF-2) they are in close proximity to the syenite which was able to reach surface here. As in the B/4 occurrences, the updomed porphyry is locally sheared but its generally massive character shows few other signs of dislocation, which contrasts with the fractured and generally disturbed comendite overlying it. The nonporphyritic, streaky comendite, intimately associated with the quartz-felspar porphyry in all but one of the updomed areas, belongs to the late intrusive phase of the comendite cycle which coincided, essentially, with the intrusion of syenite. In several localities the comendite appears to have been emplaced at higher levels in the fracture-zones than the syenite and to have invaded the domed, shattered quartz-felspar porphyry (section A-C1, Pl. IV).

The commencement of the quartz-felspar porphyry cycle followed closely on the final explosive phase of the felspar rhyolite, shown by the intimate association

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*B/4 refers to grid on geological and structural maps, Pl. IV and Fig. 2 respectively.

of these components in the upper levels of the agglomerate in the Paresis-South Area. The transition zone is, however, relatively abrupt—at most some 50 m. thick—suggesting that the focus of volcanic activity had by then shifted from centre F to a new centre. The only other occurrences of basal agglomerate belonging to the second eruptive cycle are several outcrops in Kudu Valley, but these are free of felspar rhyolite debris. Although the Paresis-South agglomerate may be expected to continue for some distance into the Central Massif, the known distribution and extent of the quartz-felspar porphyry agglomerate can only be adequately explained if it emanated from a single, central source at the northeastern end of Gemsbok Valley. Direct evidence for the location of this vent, P1, is sparse and consists of several low, exfoliating domes of vent-breccia and lava.

The interpretation of Gemsbok Valley as a major fracture-zone is supported by a strong lineament (on aerial photographs) where its extension traverses the crest and the flanks of Monday Plateau. A source for the quartz-felspar porphyry in a fissure-vent located on this lineament is consistent with its alignment parallel to the regional strike of the Damara basement and would, moreover, account for the absence of topographic evidence of a central vent. The porphyry flows attain their greatest thickness of 900 m. at Monday Plateau. Apparently unfaulted exposures in Kudu Valley have an average thickness of 400 m., on the western periphery of 500 m., and on the North-West Margin of 250 m. A source in the upper reaches of Gemsbok Valley is clearly required to explain this distribution. Calculations from extrapolated sections, using the above values, indicate a minimum thickness of 1,400 m. for the quartz-felspar porphyry products at centre P1. Using the same methods as for the felspar rhyolite, minimum distances travelled by the porphyry flows from centre P1 are: NE—8 Km., N—11 Km., and W—12 Km

The porphyry is, in many places, closely associated with basaltic flows and dykes. Basaltic xenoliths, causing local contamination of the porphyry, are a widespread feature.

The porphyry body occurring near centre F, has the normal texture of the flows, but there can be little doubt about its intrusive origin since it is surrounded on all sides by the earlier rhyolite and shows moderate to strong shearing at the contacts. Other intrusions of quartz-felspar porphyry include the coarse-grained marginal bodies at F/6 and the Kameelfeld dyke, north of the main complex. The latter dyke is a linear body, some 7 Km. long, whose attitude conforms closely to the surrounding Damara schists and marbles.

No evidence has been found indicating the relative ages of the quartz-felspar porphyry dykes. However, the position and attitude of the Kameelfeld dyke with respect to centre P2, and of the Paresis-South bodies with respect to centre P1, suggest a cone-sheet-type character for these dykes. It is probable that they were emplaced only after extrusive activity in the various centres had stopped and the conduits blocked with plugs of congealed rhyolite. Subsequent increase in pressure on porphyry magma remaining in the reservoir-conduit actuated its emplacement as dykes.

3. *Comendite*

Though showing a remarkable degree of chemical homogeneity, the comendites vary considerably in texture and mineralogy. The most abundant type contains riebeckite glomerocrysts, in places accompanied by hornblende or aegerine, set in a very fine, granular matrix of quartz and felspar. Short lenses of micropegmatite selvedged by riebeckite are widely developed in the unaltered rock giving it a streaky texture. The late intrusive comendite is chilled and non-porphyritic, commonly associated with syenite and bostonite dykes. Comendite of the Central Massif has been extensively subjected to deuteric alteration and the resulting rock is characterised by black or rusty-brown phenocrysts in a reddish-brown ground-mass. Felspar phenocrysts are scattered sparsely throughout the central comendite but in some places they constitute more than 10% (vol.) of the rock which also tends to have a coarser matrix. The above-mentioned textural varieties commonly grade into one another and are all hard, compact rocks with strong jointing, which is irregularly columnar in some localities and manifested as prominent vertical slabs in others.

Comenditic rhyolite represents the final extrusive episode of the volcanic phase at Paresis. Two volcanic centres have been correlated with this cycle: Vent C1 within the Central Massif and Vent C2 on the Paresis-North Ridge. Vent C2 truncates the earlier Vent P2 and is itself dislocated by later faulting. Its products, comprising a lower unit of fine-grained, streaky rhyolite and a thin upper flow of felsite, are overlain by comendite emanating from centre C1.

The dimensions of the reconstructed comendite cone are presented with somewhat less confidence than those for the earlier volcanoes; absence of an upper contact and the differential displacement of blocks are potential sources of error. Nevertheless, results based on the apparent thickness of comenditic products in three sections are in good agreement, indicating an average height of 2,200 m., from base to apex, and minimum distances travelled by the comendite from centre C1: NE—10 Km.; NNW—13 Km.; and NW—8 Km. Estimates of cone dimensions for the segment southwest to east from centre C1 were inconclusive because of few and irregular dips shown by the comendite in that area.

Small xenoliths of earlier volcanics, particularly of quartz-felspar porphyry and basalt, abound in many places but a basal agglomerate was not found. Such a unit may, of course, be buried under the thick pile of the Central Massif, but the absence of an indication, in even the thick tilted section exposed by fault Fd, creates doubt as to its existence. The apparent anomaly is largely resolved if the central comendite conduit is located on a strongly fractured zone, formed by the intersection of ring-fault RF-1 and the major northwest trending fracture splitting the central cauldron. The quantity of foreign fragments in the comendite is not commensurate with even the smallest of vents, but can be readily explained as incidental material detached from the walls of a well-developed fissure in the upward passage of the magma.

Dominating the Central Massif, somewhat southeast of its mid-point, is a group of volcanic domes, tightly clustered and, in some instances, deformed against one another. In plan they are oval, the long axes varying from 1 Km. to 2 Km. Of the thirteen domes mapped, four have axial ratios <2, seven have ratios from 2 to 3, and two domes have ratios >3. In the field, the domes appear as steep-sided, exfoliating ridges (Pl. I, Fig. 2) the crests of which generally have the form of whalebacks, but in some instances, are dissected plateaux. The essential structural features of the domes are, however, best studied on aerial photographs. These reveal that they consist of progressively decreasing, onion-like shells which generally show up as ridges or ledges parallel to the outline (Pl. II). Erosion has been particularly active around the perimeters of the domes which are commonly defined by deep canyons.

Jointing is strongly developed in the dome area. Within individual domes, the major joint-plane (exfoliation omitted) is generally parallel to the long axis, and a weaker one normal to it. The most persistent joints are very steep to vertical, but a horizontal set is prominent in many localities. The well-defined system of regional jointing is reflected in the NW-NNW joint pattern of the domes (Pl. II).

The regional joints have exercised a strong influence over topographic development, in particular the drainage pattern, and show up clearly as straight lineaments which can be traced for distances of up to 3 Km. The joints, which are superimposed on the domes and closely parallel to the several major north-west trending faults of the complex, provide no evidence—in the form of displacement, brecciation, or shearing—of differential movement. They are, therefore, best explained as a cooling feature of the dome-complex as a whole—originating under conditions of anisotropic pressure due to the collapse of the central volcanic cone.

At what stage of the structural history of the Central Massif were the domes themselves formed, and under what conditions? Their central position and internal features leave little doubt that the domes are a vent phenomenon and that their growth was entirely endogenous. If they were extruded from the active comendite volcano, subsequent collapse would have left unequivocal signs of fracturing or displacement in the solid state. Not only are the domes singularly free of brecciated material and internal dislocation, but their smooth, curved outlines and indentations against one another strongly suggest that they were emplaced, more or less simultaneously, in a plastic condition after cessation of the major deforming stresses. Evidence for their synchronous emplacement is provided by the continuity and regularity of the regional joints, which could have formed only in a uniformly cooling mass.

The domes are essentially massive comendite showing only a vague and irregular accumulation of amphibole glomerocrysts in planes parallel to the exfoliation shells. Where banding is observable, its dip is vertical or steeply outward. The surrounding flows show little sign of having been forcibly intruded and their dips commonly persist unchanged up to within a short distance of the domes. During the 1925 eruption at Santorini, Washington (1926, p. 6) observed the foot of the Foque Kameni dome to be " . . . surrounded by a thin ring of bright-red incandescence, evidently an encircling crevice. . . . At intervals there issued from the site of the crevice. . . a semicircular battery of narrow jets . . . which always exploded simultaneously and formed a crown around the dome." There is a strong suggestion here that the dome was partly emplaced by "burning" its path through the cone which, if the parallel may be drawn, would explain the paucity of intrusion features surrounding the Paresis domes. There is no evidence of the mushroom form, postulated by Williams (1932) for the Divide Peak domes in the Lassen area, which might conceal marginal intrusion features. Rather they resemble the rhyolite domes of the Marysville Buttes where Williams (1929) has noted that marginal flow-banding is either vertical or inclined steeply outwards. It is not unlikely, however, that the Paresis domes narrow gradually downwards and that the high talus slopes, commonly surrounding their base, cover a narrow zone of brecciation and dislocation. There is also a strong resemblance to the general features displayed by the trachytic dome at Ragged Hill, Ascension Island. Though Daly (1925) has portrayed that dome as a mushroom form resting in a shallow depression with a narrow central feeder, the vertical to steeply outward-dipping exfoliation planes at the margins are perhaps more consistent with the essentially cylindrical form envisaged for the Paresis domes.

The final comenditic product—the very fine, nonporphyritic variety—is closely associated with the early syenitic intrusions and also occurs to a minor extent as discrete dykes in the earlier rhyolites. There is abundant evidence that the final comendite dykes were contemporaneous with the syenites. Thus at the foot of the eastern flank of the Paresis-North Ridge, fine-grained comendite is intruded by porphyritic bostonite but higher on the slope encloses a large body of this rock in a composite dyke. In Kudu Valley, similar relations may be observed with the syenite. On the Northwest Margin, at C/3, a large outcrop of intimately mixed and mutually deformed fragments of syenite and fine, streaky comendite provide the most striking evidence. On quantitative grounds, the comendite might appear to be the host, but both rocks contain plastically deformed pieces of the other and it is quite obvious that their emplacement was simultaneous. Structural evidence shows that the intrusion of the syenites was roughly coeval with the final collapse of the volcanic complex. The linking of the intrusive comendite to the subsidence mechanism and thence to the volcanic domes, completes the pattern of this cycle.

C. Basalts and Gabbros

Basic rocks represent less than 1% of all igneous outcrops at Paresis and comprise essentially basalts and gabbroic rocks. The former occur as a series of thin, regular flows and are greatly predominant in areal extent. Outcrops of gabbro and related dolerite have been found only at Klein Paresis where they intrude the granophyre and the Damara rocks.

The basalts are all very fine-grained, nonporphyritic rocks and most commonly amygdaloidal. They occur predominantly in the vicinity of the outer ring-fault and generally show a high degree of alteration. Their concentric distribution and essentially conformable relations to the rhyolites with which they are interbedded, together with an absence of intrusive evidence, make it fairly certain that the great bulk of basalts exposed represent tilted and truncated flows. Some of the small basalt outcrops could, conceivably, be minor dykes but their mode of emplacement is generally difficult to establish in the field because of poorly exposed contacts and the lack of distinctive textures.

The relationship of the basalts to the three rhyolite groups shows some interesting features. Basalt was erupted intermittently throughout the felspar rhyolite cycle but the flows are intercalated only at the extreme margins of the complex. Contamination is absent and the rocks do not contain xenoliths of each other—facts which suggest that their source-vents were widely separated and that the flows overlapped in the area exposed by ring-faults on the northern margin. The presence of a basalt vent to the north is indicated also by the marked decrease in the abundance and thickness of basalt flows going southward along the periphery of the complex. The absence of xenoliths implies further that the felspar rhyolite was erupted from a single central source, because any later vent would have had to penetrate the overlying basalt and incorporate it in the resulting agglomerate or flows.

The emission of basalt flows continued until the end of the quartz-felspar porphyry cycle, but in diminishing volume and frequency—only two thin flows being associated with this rhyolite. As with the felspar rhyolite, the restriction of intercalated flows to the outer margins is significant in so far as it provides an indication of the prevailing topography as well as additional evidence for the external origin of the basalt. The distribution of the mobile basalt flows would have been determined largely by hydrostatic factors—terminating against the foot of the current volcanic cone. A noteworthy aspect of the association of basalt and quartz-felspar porphyry is the widespread distribution of basalt inclusions leading, in some places, to intense contamination of the host-rock.

The comendites are distinguished from the earlier rhyolites by the complete absence of basaltic flows, but xenoliths are widely distributed.

Any attempt to explain the origin of the basalt xenoliths must take into account the virtual absence of felspar rhyolite inclusions in the later rhyolites—which should be present if the xenoliths were derived from flows covering the vents of porphyry and comendite. Coupled with this observation is the exceptionally high concentration of basalt inclusions in the high-lying quartz-felspar porphyry of the Paresis-South Area. The consistently non-amygdaloidal texture of the xenoliths is another significant feature.

The above observations fall into a pattern if it is postulated that the xenoliths are derived from a dyke or sill emplaced along the Gemsbok Valley fissure before the beginning of the quartz-felspar porphyry cycle. Early effusive porphyry (and comendite) could thus be expected to have effectively cleared the vents of basaltic material, allowing later rhyolite to rise uncontaminated. The operation of some such mechanism is clearly demonstrated by the comendites of the Central Massif, in which the concentration of basalt (and porphyry) xenoliths is greatest in the lower flows but very sparse in the upper flows and endogenous domes. No direct evidence for the existence of a basalt dyke in Gemsbok Valley has been found, but its presence there is entirely feasible in the light of the nearby Klein Paresis gabbros and the small post-agglomerate dyke at D/6.

The basic rocks intruding the granophyre and the Damara metasediments at Klein Paresis range from coarse olivine gabbro to fine-grained dolerite—the latter occurring as narrow apophyses. The gabbro is virtually unaltered and no signs of contamination of or by the country-rocks were seen. Basic dykes in the granophyre are up to about 15 m. wide, but their continuity on the crest and the flanks of Klein Paresis could not be determined. Apart from their orientation, which is parallel to a prominent vertical joint-plane in the granophyre and to the strike of the Damara, no other data which might point to a particular source of the gabbro dykes are available. The presence of a large body of gabbro below the granophyre is, however, suggested by the coarse grain of some of the small outcrops, and by the alignment of the large exposure at D/9 with the outcrops on the eastern flank, and the strongly sheared crest between them. Chemical affinities between the gabbro and basalts (Siedner, 1965) leave little doubt that they are genetically related.

On the basis of present knowledge, we can only conjecture on the origin of the mafic rocks at Paresis. However, two points relevant in this respect are, firstly the evidence that post-Karroo igneous activity in northern Damaraland became centralized soon after the Stormberg volcanism (Korn and Martin, 1954) and, secondly, the indications that the Paresis basalts emanated from vents outside the caldera region. As a working hypothesis, it is therefore proposed to draw the Karroo and "post-Karroo" igneous cycles closer together chronologically and to correlate the mafic flows and the dykes at Paresis with the final Stormberg phase. This interpretation, though based on purely circumstantial evidence, does not conflict with the known facts. On the contrary, by postulating that the basalts are products of a regional phenomenon, their intercalation with rhyolites becomes comprehensible in terms of alternating pulses in discrete reservoirs. The basalt-rhyolite association at Messum may also be considered in this light. The writer suggests that the central complexes represent granitic magma—generated by palingenesis of basement-granite—which rose to epizonal levels before leaving the reservoir by way of effusive activity and intrusive emplacement. The Stormberg lavas and dykes, on the other hand, originate from considerably deeper levels, at which basaltic magma was tapped by the regional fracture system.

D. The Intrusive Phase

Closely following the culminating phase of rhyolite volcanism, a suite of syenitic rocks was emplaced as dykes and small plugs along fractures formed by the foundering volcano. The intrusives, which range in composition from microgranite through syenites to foyaites, represent about 10% of exposed rocks at Paresis.

1. *Syenites*

First in the sequence, and most abundant, is a medium- to coarse-grained, granular syenite comprising about 75% alkali felspar, up to 5% quartz, interstitial clinopyroxene, fayalitic olivine, and occasional hornblende in varying concentrations. In places the texture becomes markedly porphyritic, the minerals more euhedral, and quartz may increase up to 15%. This porphyritic syenite occurs as a marginal phase to the even-grained syenite in some outcrops—the transition being generally gradational.

The syenite is most extensively developed around the circumference of the Central Massif as a series of discontinuous outcrops. Its distribution there is markedly annular, occupying 220° of arc, of which 125° represents the Gemsbok Valley exposure and 35° the Kudu Valley outcrops—the latter being taken as continuous. The Gemsbok Valley body shows great regularity in its arcuate outline, particularly at its external boundary which follows closely the trace of the ring-fault (RF-2), departing from it only at the extremities. The syenite has moderately upwarped the comendite at the central protruberance but, apart from this, the country-rocks are surprisingly free of intrusion tectonics.

In contrast, the syenite bodies on the northwestern side of the Central Massif are discontinuous and irregular units showing evidence of forcible emplacement and are, moreover, not restricted to the immediate vicinity of the ring-fracture. As discussed previously, updoming of the quartz-felspar porphyry into the overlying comendite, in this region, is directly attributable to syenite intrusion.

The configuration of the Gemsbok Valley unit clearly indicates that it was emplaced within an arcuate fissure whose shape it closely reflects. In contrast, syenite on the North-West Margin created space for itself by doming, wedging, and generally dislocating the tilted rhyolite flows. It is important to note, in this respect, that nowhere in the complex does the syenite show evidence of stoping in the form of xenoliths or contamination.

The virtual restriction of the syenite (and, indeed, all succeeding intrusive rocks) to major fracture-zones and its intimate association with late intrusive comendite, leaves no doubt that its emplacement coincided essentially with the activation of the faults and fractures involved in the formation of the final caldera.

2. *Bostonites*

The rocks classified as bostonites occur typically as small plugs and narrow dykes with a pronounced concentric distribution, mainly within the peripheral fracture-zone of the northern half of the complex. The texture varies from trachytic to porphyritic; it may be extremely fine-grained, as in the Rusthof dyke, or pegmatitic, as in the bostonite mantling the foyaite stock.

The trachytic bostonite is younger than the syenite which it intrudes at B/3 and on the outer margin of the Paresis-North Ridge. Porphyritic bostonite, however, forms composite dykes (Fg and Fh) with the late intrusive comendite, which makes it essentially contemporaneous with the syenite.

On the western and northwestern periphery, narrow bostonitic dykes are emplaced concordantly within the tilted volcanics. Although low-angle transgressions of the country-rocks do occur, the dykes have essentially similar dips of between 40° and 60° towards the centre. They vary from 5 to 50 metres in width and commonly show signs of forcible intrusion. The post-subsidence age of the dykes shows that their present conical disposition is a primary feature related to their mode of emplacement, thus making them directly analogous to the cone-sheets typically developed in the British Tertiary ring-complexes.

3. *Microgranite*

Microgranite is exposed sporadically around the periphery of the complex as low hillocks which, though small in area and far apart, lie on a near-perfect circle with an average diameter of 17 Km. There can be no doubt that the emplacement of the microgranite was controlled by a well-developed, continuous ring-fracture (RF-3) which, it may be observed, defines the outer limit of the subsidence at Paresis—as far as this can be expressed in terms of inward-tilted volcanics.

The relative time of emplacement of the microgranite is difficult to establish because intrusive contacts with the syenitic rocks, apart from an ambiguous outcrop at C/2, have not been found in the field. Furthermore, the massive nature of the microgranite and its isolated outcrops give no idea of its configuration at depth. Although the ring-fault it occupies may, by analogy with others at Paresis, be assumed to have a vertical to steeply centripetal dip, it is improbable that the source of the microgranite was tapped by the base of the fault. On petrochemical evidence it is suggested that the microgranite represents the residual fraction of the Paresis magma remaining in the upper levels of the widened conduit. Structural arguments will be advanced (section IIId) to show that the microgranite represents a low-angle, conical dyke, spatially related to the bostonites.

4. *Felspathoidal rocks*

Occurring on the northwestern periphery of the complex is a composite stock of undersaturated alkalic rocks, roughly oval in outline with axes measuring 3 Km. and 1.3 Km. It comprises an outer ring of coarse-grained, leucocratic foyaite with the central portion occupied by tinguaite and dark, fine-grained phonolite.

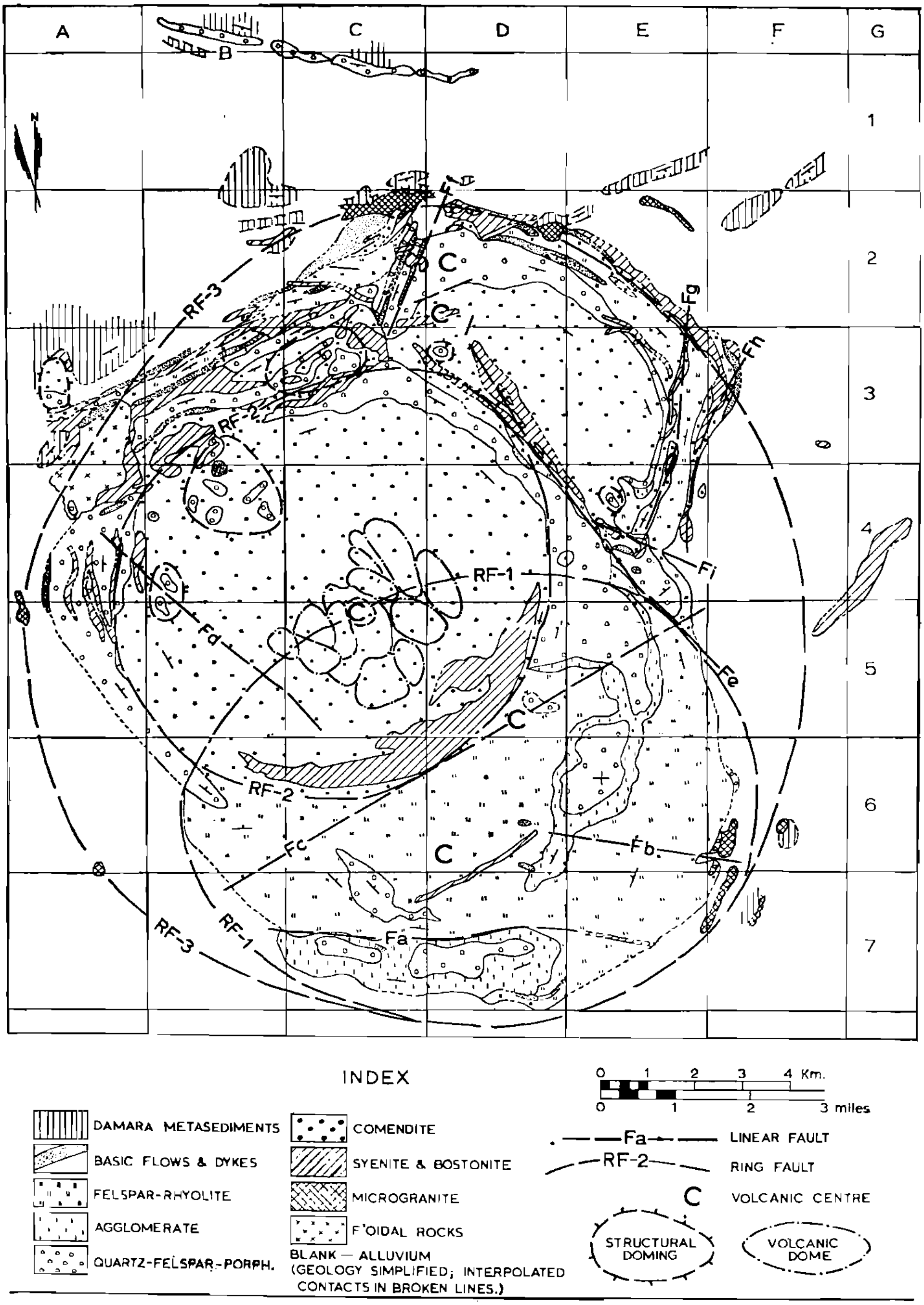
From the general distribution of their outcrops it would seem that the phonolite was emplaced as a cylindrical plug (conduit?) within the foyaite and was itself cut by later tinguaite. This relationship is indicated also by exposures in the centre where tinguaite veins contain fragments of phonolite, whereas the latter appears to intrude between the foyaite and the bostonite at the southern tip of the stock. Intrusive relationships between the foyaite and its bostonitic mantle have not been seen, the two rocks grading into each other by imperceptible degrees. The texture of the bostonite in this area varies considerably, but is invariably more coarse-grained and equigranular than elsewhere in the complex. In addition to a progressive diminution of their quartz content, the rocks immediately surrounding the foyaite bear the marks of intense metasomatism resulting in strong clouding of the felspars, the introduction of aegerine, magnetite and, to a lesser extent, riebeckite and fluorite. In most specimens, however, the dark minerals have been largely decomposed to iron and titanium oxides, leaving only patches of the original material.

III. STRUCTURAL EVOLUTION OF THE PARESIS COMPLEX

A. Volcanic Sequence

A brief recapitulation will serve to emphasize the changing character of the Paresis volcanism as related to its structural history. The felspar rhyolite cycle was characterised by the production of vast quantities of flow-breccia and massif rhyolite interbedded with air-fall agglomerate, erupted from a central vent which built up a high strato-volcano. This cycle closed with the paroxysmal eruption of coarse pyroclastics—probably representing the disintegrated summit of the volcanic cone. The quartz-felspar porphyry cycle began soon after with the production of a small quantity of basal agglomerate and thereafter is represented by a continuous sequence of unbrecciated, massive rhyolite, erupted mainly from centre P1. The final comendite suite is likewise massive and unbrecciated but differs from the earlier cycles in being, apparently, devoid of any basal agglomerate which might represent material derived from the drilling of a central vent. This marked decrease in the volume of lithic ejecta is attributed to earlier ring-faults that were opened by magmatic pressure lifting the block segments and, thus, provided effective conduits to the reservoir. The implied doming effect and its result is comparable with that postualted by Smith *et al* (1962) for the Valles caldera, New Mexico.

It is worth noting the remarkable similarity between the pattern of volcanic activity at Paresis and that related to the late Pre-cambrian ring-complexes of Nigeria. Jacobson *et al* (1958) have recognised two groups of rhyolites: the earlier group, comprising typical products of vent-extrusion from simple, central volcanoes, shows autobrecciation and interbedding with pyroclastic deposits; the late rhyolites, which are much more extensive, are usually porphyritic with a microcrystalline ground-mass suggestive of rapid chilling and contain no pyroclastic rocks. The authors consider that the origin of the later rhyolites is satisfactorily explained by magma welling up major ring-faults subsequent to cauldron subsidence. However, they make no attempt to explain how an acid lava, of whatever origin, can be widely distributed to form great plateaux and yet retain its micro-crystalline, unbrecciated texture.


A
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RF-1
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Fa
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INDEX
DAMARA METASEDIMENTS
BASIC FLOWS & DYKES
FELSPAR-RHYOLITE
AGGLOMERATE
QUARTZ-FELSPAR-PORPH.
COMENDITE
SYENITE & BOSTONITE
MICROGRANITE
F'OIDAL ROCKS
BLANK — ALLUVIUM
(GEOLOGY SIMPLIFIED; INTERPOLATED CONTACTS IN BROKEN LINES.)
0 1 2 3 4 Km.
0 1 2 3 miles
Fa LINEAR FAULT
RF-2 RING FAULT
C VOLCANIC CENTRE
STRUCTURAL DOMING
VOLCANIC DOME

Fig. 2

This is exactly the problem presented by the Paresis volcanics and their mode of emplacement must be elucidated before any suggestion can be made regarding the nature of the eruptive mechanism.

B. Mode of Rhyolite Emplacement

As mentioned previously, the bedded flows of all three groups are conspicuously lacking in autobrecciation and the bulbous forms normally associated with rhyolitic lavas. Indeed, peripherally exposed rocks are commonly more massive and contain less clastic material than those near the vents. Moreover, the probable minimum distances travelled by the rhyolites (up to 16 Km.) speak for a great mobility of erupted material.

The mobility of ash-flows is a well documented phenomenon and the interpretation of the Paresis rhyolites as the welded and extensively recrystallised products of such deposits explains numerous, otherwise puzzling, observations. From the lateral dimensions of the numerous ash-flows cited by Smith (1960a) and others, it will be apparent that the distances estimated for the Paresis units are readily accounted for. The dense, non-vesicular and devitrified character of most Paresis rhyolites differs from the majority of studied ignimbrites which, though variable in the extent and intensity of their welded character, generally display a much greater porosity and typically retain traces of the original glass shards as an integral part of their texture, (*see* Ross and Smith, 1961; and Smith, 1960b). It should be remembered, however, that almost all the deposits for which an ash-flow origin is well established are of relatively recent age (Tertiary and younger), form extensive sheets and plateaux, and have generally not experienced protracted periods of metamorphism. Considering the far greater age of the Paresis volcanics (130 m.y.) and the fact that all the presently exposed rocks have been subjected to the recrystallising effects of increased heat and emanations resulting from cauldron subsidence, it becomes possible to reconcile the textural differences with a similar mode of origin.

C. Structural Relationships

It is now possible to integrate the stratigraphic relations with the structural units constituting the Paresis Complex. The map (Fig. 2) and photographs (Pl. III) show the volcanic rocks to have pronounced centripetal dips everywhere and to be restricted to the area bounded by the outer ring-fault (RF-3). Thicknesses of the truncated flows indicate a much wider original distribution, which suggests tilting on ring-fault RF-3 and subsequent removal by erosion of volcanic deposits outside this fault.

The first cauldron, Paresis-South Area, is an unambiguous example of central collapse and involves a circular area of some 10 Km. diameter. Although the position of the bounding ring-fault (RF-1) is mostly ill-defined, its existence is well established by the concentric distribution of the felspar rhyolite with respect to centre F and the steep dips of the peripheral flows; shearing and brecciation mark its trace around Monday Plateau, while quartz-felspar porphyry and microgranite dykes are probably indicative of its presence on the east side. Displacement along RF-1 began with the collapse of the volcano, towards the end of the felspar rhyolite cycle, and continued at least up to the emplacement of the quartz-felspar porphyry flows, as shown by their attitude.

The Gemsbok Valley fault (Fc) is parallel to the regional strike of the Damara basement and clearly represents a plane of weakness inherited from it. Probably, it also started to form as a fracture during this period, opening into a fissure-vent during the quartz-felspar porphyry cycle and finally becoming active as a fault related to the collapse of the comendite cauldron. Evidence favouring fault Fc as a conduit has been discussed previously and it need only be added here that the small volume of basal agglomerate in this cycle is consistent with the presence of an effective opening to the magma reservoir—facilitating the exit of magmatic material without the production of large quantities of pyroclastics. The thicknesses of porphyry flows at various places in the complex, indicate that a central cone was built, but there is no evidence that this cycle included the formation of a cauldron.

Truncating the Paresis-South subsidence in the northwest, is the second cauldron represented by the Central Massif. Inward tilted rhyolite flows are concentrically exposed and their repetition on the North-West Margin and the Paresis North Ridge are unequivocal indications of central subsidence controlled by strikingly regular ring-faults. However, local deviations from an ideal centripetal inclination are noted. Areas showing broadly uniform attitudes may be considered as representing blocks which became detached and underwent differential subsidence and/or tilting as the volcanic cone collapsed. Small zones of brecciation and shearing have been found along apparent boundaries between adjacent blocks, but on the whole, direct evidence is sparse because the inferred boundaries mostly follow the deeply-incised beds of streams.

The thickness of flows is greatest northwest of vent C1, which is in accordance with its postulated position on the flank of the quartz-felspar porphyry cone—resulting in a directional bias to their distribution. The total absence of comendite southeast of the Gemsbok fault may thus be attributed largely to the shielding effect of the P1 cone. In attempting to determine the cause for this position of vent C1, one notes that it lies on the intersection of two prominent structural features. The first of these is the northwest trending fracture-zone bisecting the cauldron, one pole of which is represented by the bostonite-foyaite stock, the large syenite body, and the updomed quartz-felspar porphyry, the other pole by the local protruberance of syenite from the Gemsbok Valley dyke. Secondly, the the interpolated trace of ring-fault RF-1 exactly passes through centre C1. Although the northwest fracture-zone was not strongly developed, its influence served to localise igneous activity on the ring-fault in a manner analogous to the emplacement of the foyaite stock on ring-fault RF-3.

The situation of volcanic vents on ring-fractures is a prominent feature also at Slieve Gullion (Richey, 1932a) and several of the Nigerian ring-complexes (Jacobson *et al* 1958). Smith *et al* (1962) have recorded no less than ten rhyolite vents on the 8-mile diameter ring-fault delineating the Valles caldera, New Mexico. Closely parallel to the northwest fracture zone are fault Fd, the Kudu Valley fault (Fe) and several lineaments of minor importance. It is plain that two mutually perpendicular directions of weakness were operative in determining the form of, and relationships between, the various structural units at Paresis. This is a local reflection of the regional northeast and northwest structure-pattern of Damaraland.

The Paresis-North Ridge participated in the formation of the Greater Paresis Cauldron as a discrete block which was tilted along a northwest trending axis.

It is bounded on the southwest by the Kudu Valley fault (Fe) which is tangential to the Paresis-South and Central Cauldrons, and on the northeast by the outer ring-fault. Both faults are almost entirely occupied by intrusive rocks. The eastern side of the block is marked by two northward-trending composite dykes of comendite and porphyritic bostonite. The outer dyke (Fh) is interpreted as a fault truncating the marginal flows, but the inner dyke is not associated with any displacement and probably fills a fissure. The block is bounded on the west by a strong fault (Ff), manifested as a steep scarp, about 50 m. high, where it has sheared through vent P2. Small bodies of porphyritic bostonite are associated with this fault and with the fracture cutting vent C2.

D. Dykes, Ring-faults and Cauldron Subsidence

The evidence so far presented shows that the volcanic rocks occupy areas of subsidence bounded by circular faults. Any postulate regarding the collapse mechanism and the emplacement of the late intrusives clearly hinges on the configuration of the ring-faults at depth. It must be stated at the outset, however, that nowhere at Paresis has an actual fault surface been seen, intrusions and erosion having effectively destroyed all direct evidence such as mylonitization, slickensiding, etc. Interpretation is, therefore, necessarily dependent upon circumstantial evidence related to the subsidence.

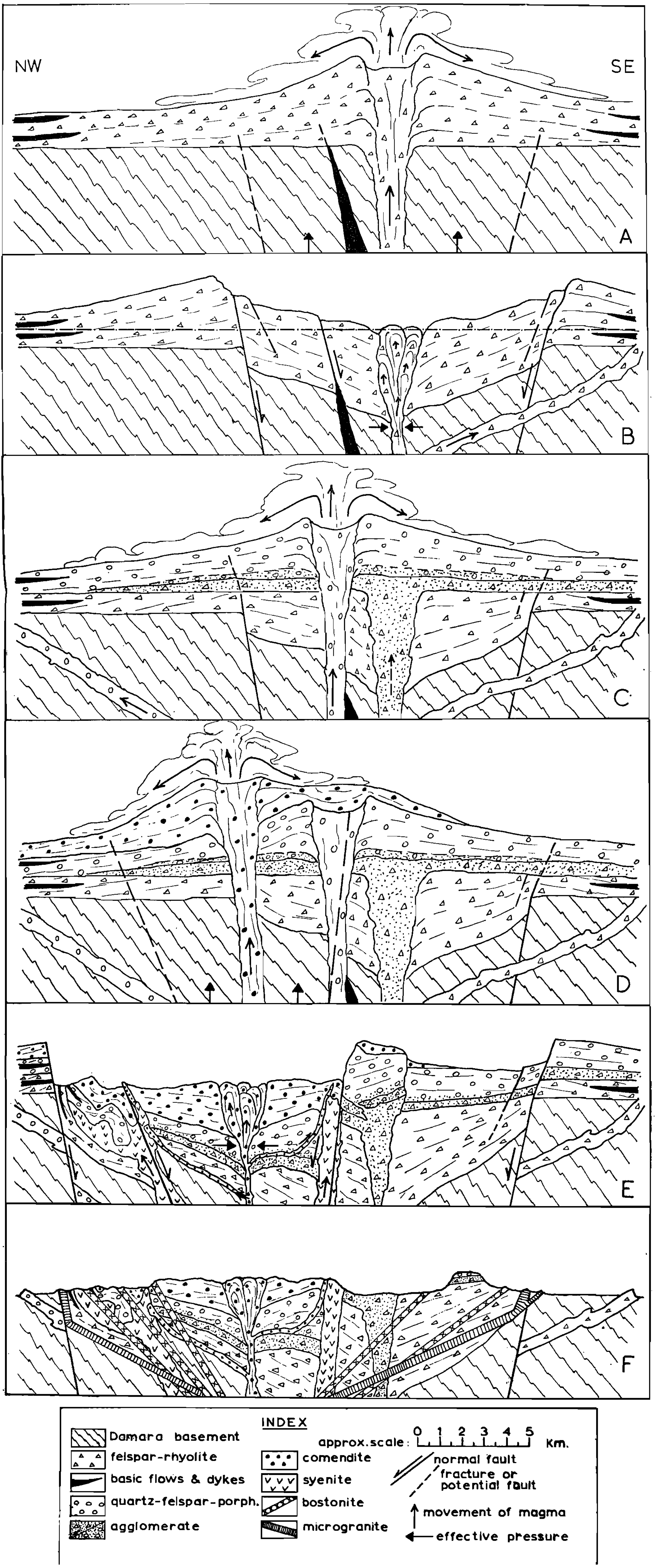
It is appropriate to examine first the periphery of the central cauldron along which syenite, representing the trace of the ring-fracture, is intermittently exposed. The intrusive comendite in Kudu Valley, coeval with the syenite, shows vertical to steeply centripetal dips at numerous localities. The thin tongue of comendite bordering the syenite in the northeastern corner of Gemsbok Valley also has vertical flow-lines. The western and northwestern sides of the ring-fault are less clearly defined, mainly because of the complex dislocation of the volcanics resulting from irregular, multiple intrusions of syenite. In the immediate vicinity of the inferred position of RF-2, the picture resembles that on the eastern side but the rhyolites dip at markedly lower angles—commonly in the range of 60° to 70°. It is inferred that the dips of the flows contiguous to the ring-fault reflect the general attitude of the fault to a depth of at least several hundred metres.

The comendite flows constituting the Central Massif also show great regularity in their dips—commonly in the range 40° to 60° in the outer belt, and 25° to 40° near the centre—allowance being made for uneven depositional surfaces and differential tilting of segments within the subsiding cone. Even so, the dislocation of the segments, relative to each other, is surprisingly slight, considering the diameter of the cauldron, and has not significantly distorted the overall collapse pattern. Similar features were observed by Cloos (1941) in the Schwabian diatremes, where large blocks of Jurassic country-rocks had subsided virtually without rotation. The flows do not become horizontal because the focus of subsidence is occupied by the endogenous domes, but their strongly developed inward tilt makes this cauldron and, indeed, the entire Paresis subsidence, structurally analogous to the cauldrons of Glen Coe (Clough *et al*, 1909), Ossipee (Kingsley, 1931), and Glitrevann (Oftedahl, 1953).

Let us first consider the collapse mechanism in terms of the tilted volcanics, for their markedly regular attitude clearly holds the key to the problem. The regularity of dip and the consistency of outcrop carry several important implications: firstly, that the subsidence and disintegration of the comendite cone was a single, continuous episode affecting all parts of it simultaneously and roughly to the same extent—rotation and differential subsidence of the segments being insignificant; secondly, that a powerful guiding influence—essentially concentric in its effects—was operative in reversing the originally quaquaversal dips of the rhyolites. It may also be assumed that a wide conduit, probably filled with unconsolidated material, provided an effective space for the adjustment of inward-falling cone segments.

Subsidence of a circular block isolated by outward-dipping ring-fractures, as proposed in the Mull Memoir (Clough, Bailey *et al*, 1924) and later elaborated by Anderson (1936), cannot explain the observed phenomena at Paresis. Stratified deposits on an unfractured block sinking into a cavity of progressively increasing diameter, would not be subjected to any forces intrinsically tending to produce a marginal upwarp. Ring-dyke material, emplaced simultaneously along the opening fracture, might cause updoming but this would, at best, be of a local and irregular nature. By the same token, the component segments of a disrupted cone, subsiding under these conditions, would not be restrained from collapsing and rotating in a completely random manner. Subsidence of a cylindrical block bounded by a vertical ring-fault, though potentially capable of causing slight marginal upwarping due to friction, is an equally unlikely mechanism. For these reasons the postulated vertical or steeply outward-dipping attitude of the faults bounding the cauldrons at Glen Coe (Clough *at al*, 1909), Mull (op. cit., 1924), and Ardnamurchan (Richey and Thomas, 1930) must be doubted. To explain the tilted volcanics at Messum, Korn and Martin (1954, Pl. 17) envisage crustal blocks sinking along vertical ring-fractures which did not, however, penetrate to the surface. The flexure of the overlying volcanics is attributed to their slumping into the cavity thus generated with the development of, presumably, concentric step-faulting near the margin of the cauldron. It is, however, not clear why, once the marginal slump-fractures had formed, the volcanics should not accompany the descending block instead of deforming and fracturing in the nature of a strained elastic material. Kingsley (1931) suggested that the steep centripetal dip of the Moat volcanics exposed in the sunken block at Ossipee, New Hampshire, may be the result of a downward convergence of the ring-fault, "causing crowding of the down-dropped mass". Oftedahl (1953) comes to a similar conclusion in accounting for the inward dip of lavas in the Glitrevann cauldron, Oslo province. In a reappraisal of the British Tertiary volcanic structures, Reynolds (1956) proposed that the configuration of down-faulted, bedded deposits are a more reliable criterion for the three-dimensional form of the ring-faults than surface measurements .of fault planes. She concludes that the form of the fractures bounding the cauldron is essentially conical, opening upwards.

The regular outline of the central cauldron at Paresis suggests that the ring-fracture, which delineated the subsiding block of country-rocks, were continuous to the surface and isolated the central portion of the volcanic cone. The well exposed comendites provide no evidence of concentric tension-fractures, which would be expected if their present configuration was the result of slumping down a parallel-sided cavity with a consequent elongation of their base. The observed features at Paresis are more effectively explained in terms of basement subsidence along inward dipping ring-faults, the overlying volcanics becoming marginally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Damara basement
felspar-rhyolite
basic flows & dykes
quartz-felspar-porph.
agglomerate
comendite
syenite
bostonite
microgranite
approx. scale: 0 1 2 3 4 5 Km.
normal fault
fracture or potential fault
movement of magma
effective pressure

upwarped in the process of descending into a progressively narrowing space. This interpretation raises several questions.

When and how were these conical fractures formed? Implicit in the emplacement of the Paresis rhyolites as ***nueés ardentes*** is the existence of a magma highly charged with gas under great pressure immediately prior to eruption. Anderson (1936) has postulated that under such conditions, inward dipping conical fractures will tend to form in a homogeneous environment. If, as is highly probable, this occurred at Paresis, we have at an early volcanic stage existing fractures which predetermined the form and behaviour of the subsequently collapsing crustal blocks (Fig. 3).

How was the conical block able to subside into a narrowing cavity? The potential space represented by the conduit clearly supplies the answer to this question. A substantial vent diameter is indicated by the extent of the comendite domes which occupy an area roughly 4 Km. by 3 Km. Making allowance for a possible pear-shape in the domes, and considering the dimensions of many diatremes, the suggested diameter of about 2 Km. at the vent, rapidly tapering to an average conduit diameter of 1 Km. seems a reasonable estimate. The power of a particle-charged gas-stream to greatly enlarge the conduit is vividly described by Perret's (1924) account of the Vesuvius 1906 eruption. Over a period of several minutes he observed the diameter of the crater increase from 175 m. to 700 m., and its depth from 75 m. to 500 m. It is not difficult to envisage a series of closely following paroxysmal eruptions similarly enlarging the comendite conduit immediately before subsidence. Escher (1927) has demonstrated experimentally that a rising gas-stream will produce a cylindrical conduit followed by slumping along conical planes.

We may, then, consider a hypothetical cone bounded by a ring-fracture dipping inwards at 75°, having a cylindrical conduit of 1 Km. along its axis, and an original surface diameter of 10 Km. (the latter being an average value for the base of the rhyolites in the central cauldron when straightened out), representing a formalized instance of cauldron subsidence. Simple calculation shows that such a block could drop 2,000 m. before the conduit was reduced to zero diameter, whereas a 2 Km.-diameter conduit would enable it to drop about 4,000 m. The many irregularities and unknown factors involved in cauldron formation obviously limit the significance of these estimates to indicating an order of magnitude, and to showing that the proved depths of subsidence in other investigated ring-complexes can easily be explained by the collapse of a downward tapering block. Moreover, this concept answers the question: Why does the subsiding block stop once it is isolated by ring-fractures? (Billings, 1945). Clearly, because the slumped segments become wedged in the reduced aperture. Escher's computations (1929) of cauldron dimensions deserve attention for, though diagrammatic in their treatment, they emphasize a systematic relationship between the depth of the magma reservoir, the character of the volcanic activity, and the resulting subsidence. The similarity of form and dimensions shown by ring-complexes in widely scattered areas, is far greater than could be expected if their formation was primarily determined by the haphazard sinking of a block whose diameter depends, in the final analysis, on the fortuitous shape and size of the magma reservoir, as implied by Billings (1943), Richey (1932b), and others.

Subsidence along inward dipping ring-fractures also offers an explanation for the emplacement of the endogenous domes and intrusive comendite which, as has been previously shown, were contemporaneous with cauldron formation. Shortly after the final emission of comendite flows, subsidence commenced and resulted in a narrowing of the conduit. Slumping of the walls may have constricted the conduit somewhere along the middle, so that increasing pressure on the remaining comendite would have caused the higher, more viscous portion of the column to be squeezed out as endogenous domes. Lack of brecciation indicates that they remained plastic until after emplacement, which can be attributed to a low heat-loss within the recently erupted cover. Comendite in the lower, and hotter, levels of the column responded to the centripetal pressure by migrating up along well developed fractures and fault-planes, emplacing itself generally as composite dykes together with the syenite. The rarity of contamination in these dykes indicates that the two components ascended along separate channels, converging only near the surface.

How does the syenite ring-dyke fit into this structural pattern? Xenoliths or contamination are not in evidence and a possible mode of emplacement by piecemeal stoping, as proposed by Daly (1933) and Billings (1943), cannot be entertained for the Paresis syenite. Neither is there reason to suggest subsidence *en bloc* of an arcuate segment—the mechanism proposed by Chapman (1935) for the Cape Horn dyke at the Percy complex, New Hampshire, and by Clough *et al* (1924) for the Beinn and Ghraig dyke of Mull. The features shown by the Gemsbok Valley syenite dyke indicate that it was emplaced into a potential cavity, the shape and extent of which appear to have been determined by the collapsing block-segments. Displacement of the subsiding block to one side of the cavity, resulting in an arcuate gap which may be filled with ascending magma, has been suggested to explain such dykes in Mourne (Richey, 1928), the Bellknap and Pliny complexes, New Hampshire (Modell, 1936; Chapman, 1942) and elsewhere. With an important modification regarding the form of the central block, this concept can account for the contrasting characters of the syenite bodies in Gemsbok Valley and on the North-West Margin. For, if it be assumed that the subsidence of the conical block was accompanied by a general tilting towards the northwest, the opening of an arcuate fissure along Gemsbok Valley must have been reflected by a corresponding increase in pressure against the northwestern wall. That this was probably the situation is indicated by the wedging and dislocation which characterise the syenite emplacement on the Northwest Margin.

The eccentric position of centre C-1, with respect to the ring-fault (RF-2) has been previously ascribed to its location on the flank of the P-1 cone (Fig. 2, D) resulting in a directional bias in the distribution of its products. An additional, and perhaps more significant, reason may be that the axis of the cone enclosed by the ring-fault had a primary tilt to the northwest. Supporting evidence is provided by the steep to vertical dips of the intrusive comendite marking the ring-fault at the northern end of Gemsbok Valley, and the moderate dips of down-faulted rhyolites on the opposite side of the Central Massif. Sections show the axial tilt to be of the order of 7° from the vertical. This inherently unstable orientation of the conical block would tend to cause an asymmetric collapse towards the northwest, following the reduction of pressure in the conduit.

Implicit in the foregoing interpretation of the cauldron subsidence at Paresis, is the structural identity of the bostonitic "cone sheets" and the syenitic "ring dyke". Both were emplaced along conical fractures formed under conditions of high magmatic pressure. But, whereas the syenite is restricted to the steeply

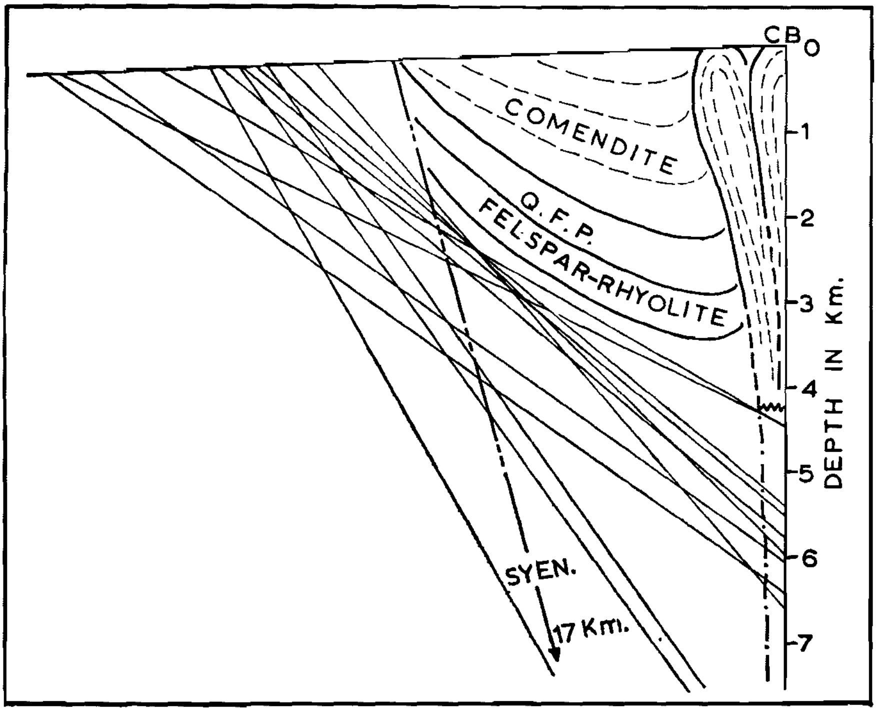

Fig. 4

Fig. 4 — Schematic projection of bostonite dykes on to vertical plane. Each line represents a single dyke. The syenite conical dyke (ave. diameter taken as 9 Km., dip 75°), for comparison, has a focal depth of about 17 Km. CB is the focus of the bostonite dykes—1 Km. northeast of volcanic centre C-1

dipping fault-zones which began forming before eruption, the bostonitic dykes have a moderate dip and occupy post-collapse fractures showing no signs of displacement. In order to emphasize their essentially similar form, the bostonite and syenite dykes are both more appropriately referred to as "cone-dykes"—the difference in their dimensions and associated intrusive features clearly reflecting the nature of their respective channels, rather than a difference in the emplacement mechanism. One can only speculate about the persistence of their dips at depth, but a projection on to a vertical plane shows that, of twelve dykes for which dip measurements are available, nine have focal depths of 4.5 Km. to 6.5 Km. below the present erosion level (Fig. 4). For the syenite dykes, on the other hand, a focal depth of about 17 Km. is indicated (using an average diameter of 9 Km. and a postulated dip of 75°). Although these figures cannot be taken at their face-value, a significant difference in the source-levels of the syenite and bostonite cone-dykes is strongly suggested. This inference is in complete accord with the younger age and more advanced differentiation of the bostonites relative to the syenite. Syenite magma was tapped at a lower level by the conical subsidence-fault, RF-2, whereas the higher and more acidic portion, encountering the conduit-blocking base of the comendite domes, responded to magmatic pressure by emplacement as narrow dykes along conical fractures with shallow dips.

Having adduced evidence indicating the existence of a fractionated magma in the upper levels of the reservoir, it is now possible to consider the emplacement of microgranite along ring-fault RF-3. We must recognise that the (apparently) steeply dipping fault, with a diameter of 17 Km., would tap a considerably deeper and more basic level of the reservoir than RF-2. Petrochemically, the microgranite is the end-member of the acid intrusive sequence and there is no reason to dissociate it spatially from this position. By analogy with the bostonite cone-dykes (having a mean focal depth of 6 Km. ±), it is therefore suggested that the microgranite also represents a conical dyke, but originating at a higher conduit level (4-5 Km.?). It is considered likely that the microgranite, moving along a gently inclined path, encountered ring-fault RF-3 and, abruptly changing its course, rose up along it (Fig. 3, F).

The structural pattern of the circular dykes at Paresis is thus seen to be closely analogous to that postulated by Von Eckermann (1948) for the cone-sheets at Alnö. Underground data from bore-holes and tunnels (Von Eckermann, 1958) confirm the originally proposed configuration of the Alnö dykes and lend support to the present interpretation.

It has been shown earlier that the emplacement of the cone-dykes was closely associated with the collapse of the comendite volcano. The fact that dyke formation at Paresis was accomplished by an upward pushing magma—also postulated by Billings (1943) for the New Hampshire ring-complexes—is in direct conflict with Anderson's (1936) proposal that block subsidence is a response to a decrease in magmatic pressure. It is difficult to reconcile the local evidence with the theory that caldera formation is directly attributable to the withdrawal of magmatic support, whether by means of surface discharge and large-scale dyke emplacement (Williams, 1941) or by pulsations in the magma reservoir (Williams, 1953). Subsidence of the central cauldron clearly occurred in the face of high magmatic pressure. This seems to favour Escher's (1929) concept of cauldron formation, because the behaviour of block segments collapsing into an effectively vacant conduit would be more directly determined by simple gravity considerations than that of a block in a closed system, completely surrounded by magma of only slightly lower (?) specific gravity.

The foregoing line of reasoning applies in all respects to the formation of the first cauldron, in the Paresis-South Area. The absence of volcanic domes—if they were originally present—may be ascribed to their disintegration by the climactic explosions which marked the end of the felspar rhyolite cycle and laid down the thick deposits of agglomerate.

IV. ACKNOWLEDGMENTS

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Fig. 1

Alternation of welded air-fall pyroclastics—ash to lapilli—with massive felspar-rhyolite (Paresis-North Ridge)

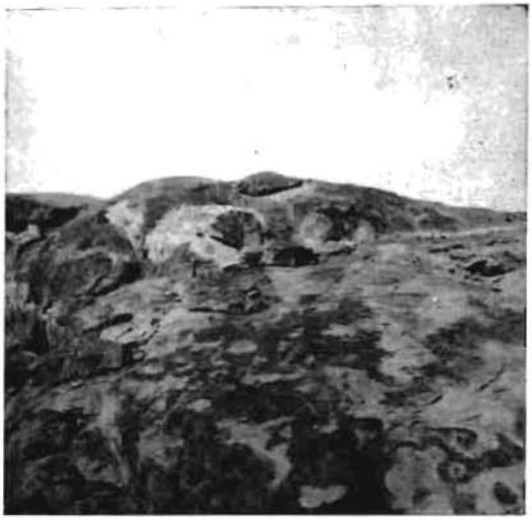


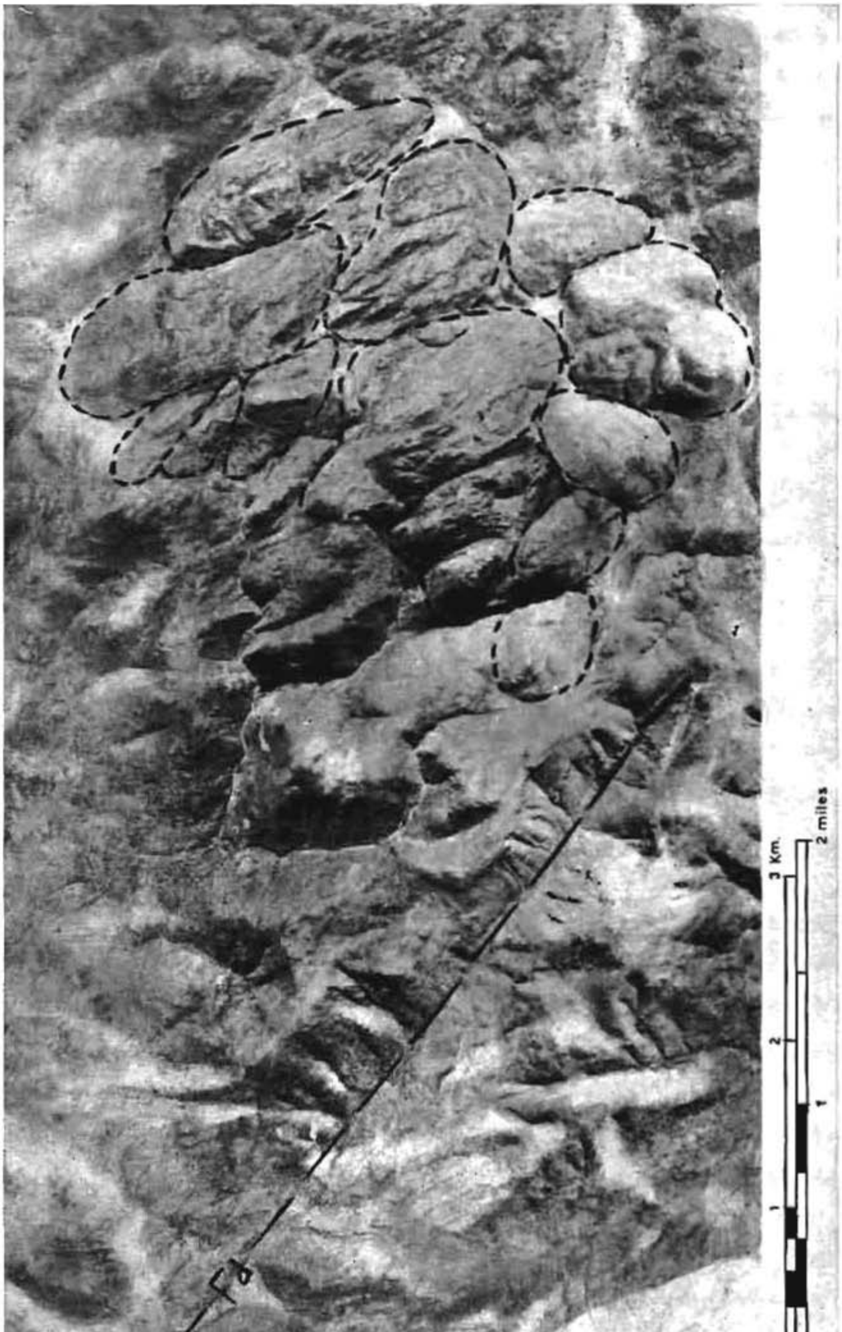
Fig. 2

Endogenous dome, Central Massif. Lighter patches are caused by exfoliation

Fig. 3

Paresis-South Plateau, looking south. Flows of massive, columnar-jointed quartz-felspar porphyry overlie incoherent air-fall agglomerate

RHYOLITE FLOWS AND VOLCANIC DOME



Southern portion of the Central Massif showing volcanic domes. Note concentric ridges and NW-NNW jointing. (Uncontrolled mosaic by Aircraft Operating Company of South Africa, Ltd)

VOLCANIC DOMES OF THE CENTRAL MASSIF

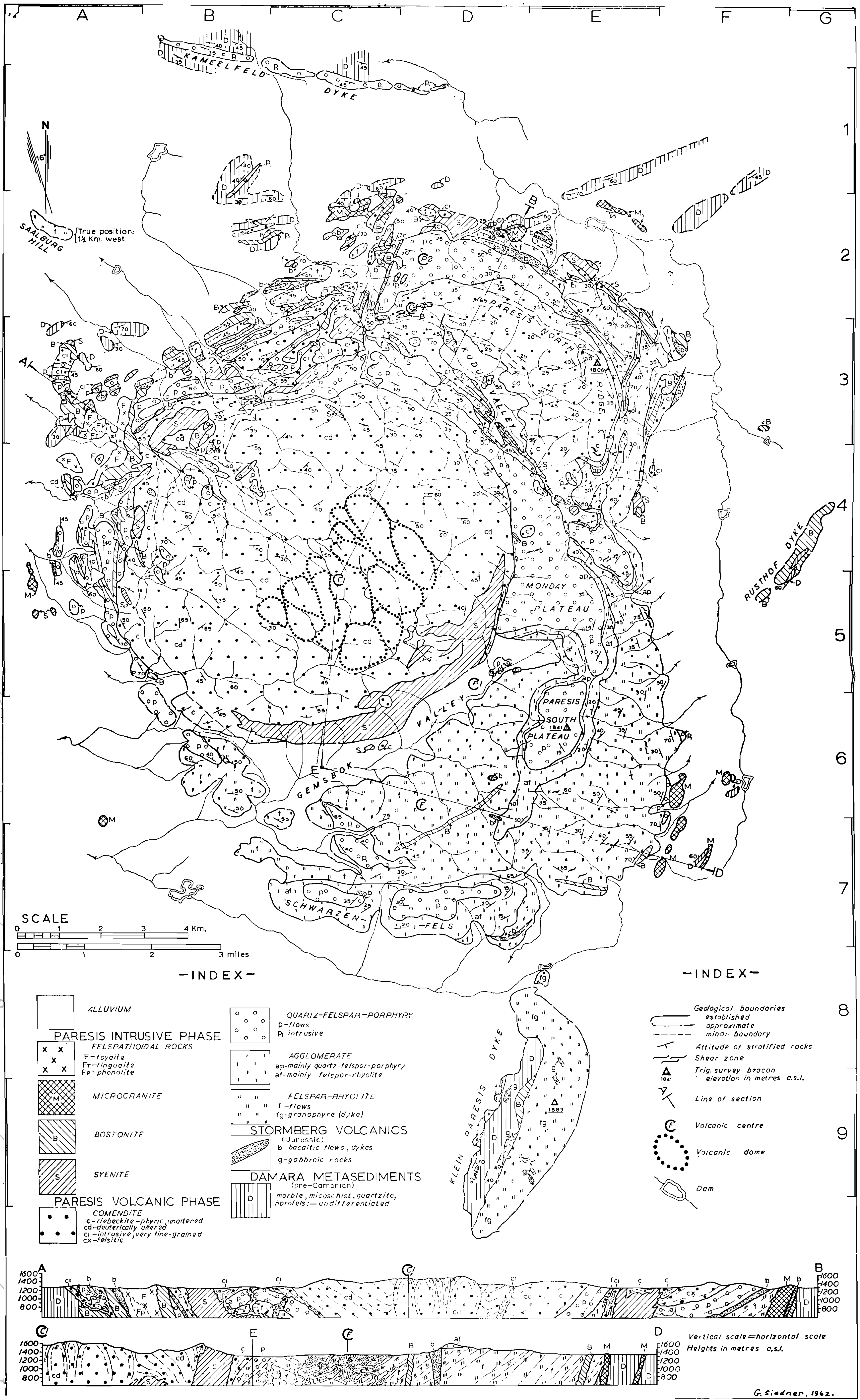
Fig. 1

Aerial view of the Paresis-North Ridge, looking south. Tilted rhyolites alternate with basalts (dense vegetation) at the base of the Ridge. Microgranite plug in left foreground; Kudu Valley at upper right

Fig. 2

Eastern part of the Paresis Complex; aerial view, looking south. Inward-dipping volcanics of the Central Massif in foreground; Monday and Paresis-South Plateaux in centre; the Klein Paresis dyke in the far centre

AERIAL VIEWS OF PARESIS


A
B
C
D
E
F
G
1
2
3
4
5
6
7
8
9
N
16°
SAALBURG HILL
True position: 1½ Km. west
KAMEELFELD DYKE
PARESIS NORTH RIDGE
KUDU VALLEY
MONDAY PLATEAU
PARESIS SOUTH PLATEAU
1841
1806
GEMSBOK VALLEY
SCHWARZEN-FELS
RUSTHOF DYKE
KLEIN PARESIS DYKE
1883
SCALE
0 1 2 3 4 Km.
0 1 2 3 miles
-INDEX-
ALLUVIUM
PARESIS INTRUSIVE PHASE
FELSPATHOIDAL ROCKS
F-foyaite
Ft-tinguaite
Fp-phonolite
MICROGRANITE
BOSTONITE
SYENITE
PARESIS VOLCANIC PHASE
COMENDITE
c-riebeckite-phyric unaltered
cd-deuterically altered
ci-intrusive, very fine-grained
cx-felsitic
QUARTZ-FELSPAR-PORPHYRY
p-flows
pi-intrusive
AGGLOMERATE
ap-mainly quartz-felspar-porphyry
af-mainly felspar-rhyolite
FELSPAR-RHYOLITE
f-flows
fg-granophyre (dyke)
STORMBERG VOLCANICS
(Jurassic)
b-basaltic flows, dykes
g-gabbroic rocks
DAMARA METASEDIMENTS
(pre-Cambrian)
marble, micaschist, quartzite, hornfels:—undifferentiated
-INDEX-
Geological boundaries
established
approximate
minor boundary
Attitude of stratified rocks
Shear zone
Trig. survey beacon
elevation in metres a.s.l.
Line of section
Volcanic centre
Volcanic dome
Dam
1600
1400
1200
1000
800
Vertical scale=horizontal scale
Heights in metres a.s.l.
G. Siedner, 1962.