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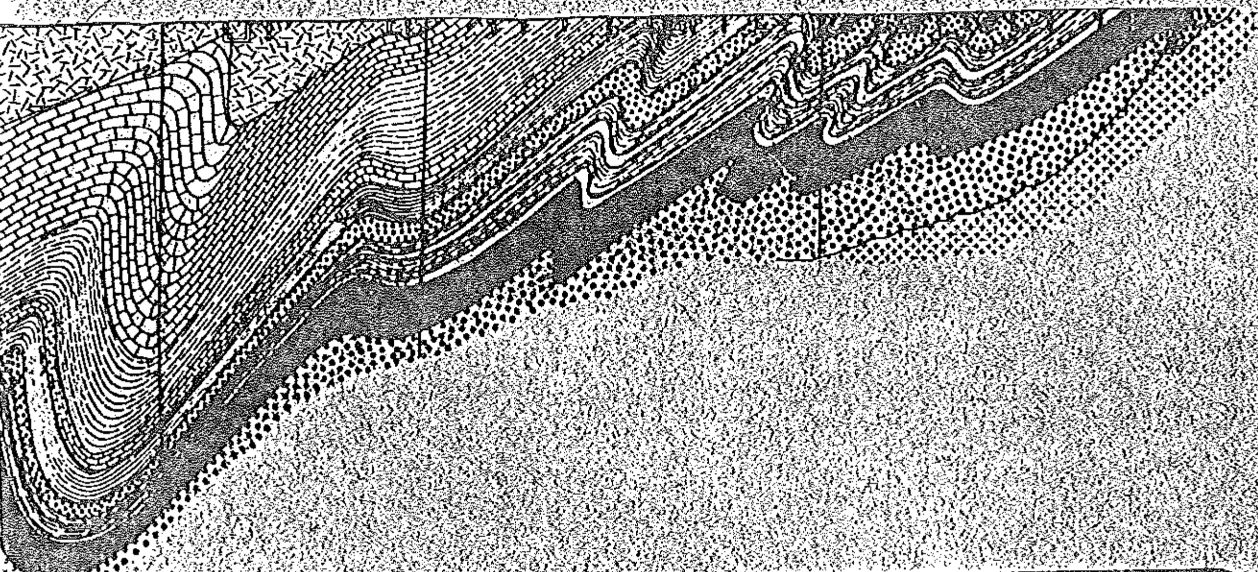

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STRUCTURAL INVESTIGATIONS IN THE BARBERTON MOUNTAIN LAND, EASTERN TRANSVAAL

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STRUCTURAL INVESTIGATIONS IN THE BARBERTON MOUNTAIN LAND, EASTERN TRANSVAAL

ABSTRACT

This paper describes the techniques and results of a structural analysis of a complexly folded region situated in the ancient Archean rocks of the Barberton Mountain Land. The orientations of various small-scale structures are recorded; they are found to be arranged in an orderly fashion and are geometrically related to the major structures. Three successive periods of deformation are recognised. Each successive deformation led to the imprinting of both large-scale and small-scale new structures on the pre-existing structures and also to the deformation of the old structures.

The first deformation gave rise to many folds whose steeply-inclined axial planes were probably initially oriented in a northeast - southwest direction. The second deformation resulted in the widespread development of slaty cleavage and schistosity which cuts obliquely across the first folds. Large, very steeply plunging folds were formed locally during this deformation, a feature which is believed to have resulted from the special orientation of the bedding surfaces before the second deformation. The Nelspruit Granite was probably implaced at this period. The third structures deformed the previously formed slaty cleavage, and large folds of this generation are recognised in the north of the area investigated. It is thought that the important arcuate form of both the first folds and of the superimposed slaty cleavage was developed at this time. The geometry of the arcuate structure is analysed; the folds related to it have orientations which depend on the inclination of the surface before the structure developed. Sets of conjugate shear folds were superimposed on the slaty cleavage, and these might have been synchronous with the development of the arcuate structure. The orientations of the principal axes of stress which formed them are calculated.

Gold mineralisation is believed to have been structurally controlled by second phase structures, and it predates the development of the third set of structures.

Structures resembling algal forms from the Fig Tree Series are described and illustrated.

The relationships of the Swaziland and Moodies sediments to the surrounding granitic rocks are discussed. The granitic and gneissic rocks are believed to represent, in part, a basement complex on which the sediments were deposited. This basement has been metamorphosed and intruded by other granites which postdate the formation of the sediments.

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STRUCTURAL INVESTIGATIONS IN THE BARBERTON MOUNTAIN LAND, EASTERN TRANSVAAL

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STRUCTURAL INVESTIGATIONS IN THE BARBERTON MOUNTAIN LAND, EASTERN TRANSVAAL

INTRODUCTION

A. GENERAL

During recent years considerable advances have been made in the studies of the structures of deformed rocks. In particular, it has been realised that the geometrical arrangements of small-scale structures, minor folds, lineations and cleavage are not random, and that careful recording of these structures shows that, not only are they ordered in a systematic way, but that they are also geometrically related to large-scale structures (Cloos, 1946; Wilson, 1961). The analysis of the regional pattern of small-scale structures has led to advances in knowledge of the structural and tectonic history of many fold belts (e.g. Berthelsen, 1960; Sutton, 1960; Ramsay, 1963A), and one particularly important outcome of these studies has been the realisation that the structures in most folded regions are the result of a succession of superimposed deformations. Although these phases of crustal movement may be of orogenic dimensions, it is commonly found that, even within the framework of one orogeny, the total deformation is the result of smaller pulses of deformation.

The paper which follows presents the results of the application of some of these techniques to one of the ancient fold belts of South Africa, and is the outcome of a visit made by the author to the University of the Witwatersrand during 1961. It was on the suggestion of the Senior Research Fellow of the Economic Geology Research Unit, D.A. Pretorius, that the Barberton region was selected for a study. This area has good surface exposure and, in addition, extra three-dimensional structural data are available from the many mines in the district. It was decided that the most useful results would probably come from a study of the regional structural pattern over a fairly large area, and to this end the region was mapped mostly by traverses, with more systematic outcrop mapping in regions of complexity.

B. ACKNOWLEDGEMENTS

The rapid progress of the field work was made possible by the assistance of various organisations, and I am also indebted to many individual persons. First, I should like to thank Professor T.W. Gevers for the invitation to visit South Africa, and D.A. Pretorius and members of the Economic Geology Research Unit for an enormous amount of helpful discussion on these and other structural problems in South Africa. The

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visit was made possible by a bursary from the Royal Society and Nuffield Foundation Commonwealth Scheme, supplemented by further financial assistance from the Economic Geology Research Unit.

Much help and co-operation were received from the staff and organisation of E.T.C. Mines Ltd., and especially from B. Genis, General Manager, K. Gribnitz, Chief Geologist in the Barberton area, and S. Voges and P. Volrath, geologists on the Consort and Sheba mines. The map compilation shown in Fig. 2 was greatly assisted by the incorporation of unpublished information from E.T.C. Mines Ltd. This includes some of the results of surface mapping around the Consort Mine taken from a map originally made by J.J. van de Berg, J.J. Schoeman and A.F. Lombard, around the Sheba Mine from a map by C.J.J. van Vuuren, and in parts of the Clutha area from a map in preparation by R. Cooke at the time of the investigations. Acknowledgement is also made of discussions with O.R. van Eeden and help from the Geological Survey of South Africa.

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STRATIGRAPHY

A. INTRODUCTION

The principal new facts and the conclusions that are presented in this paper refer, for the most part, to the structural history of the Barberton area. The structural interpretations depend to a certain extent on the arrangements of the various sedimentary, volcanic and igneous rock units, and a summary of the stratigraphic history of the region will be presented first. Except for a few points of detail, this differs little from descriptions presented in a special publication of the Geological Survey (van Eeden and others, 1956), and for further details the reader is referred to this work.

The general arrangement of the principal rock units is shown in Fig. 1; narrow belts of intensely folded sedimentary and igneous rocks are surrounded on all sides, and probably underlain, by a vast volume of granite and granite-gneiss. Although these areas of ancient sediments and volcanic strata are sometimes known as the Archean "schist belts", this terminology is not very precise. In the part of the Barberton area reinvestigated the regional metamorphism of the rocks does not rise higher than phyllite grade, whilst over the greater part of the area the pelitic sediments would be more precisely described as slates or cleaved shales. In view of their great age, in part at least in excess of 2,900 million years, the low-grade metamorphic nature of the sediments is remarkable. Although the rocks have been folded and deformed several times, the arenaceous members of the succession show well-preserved sedimentation structures over much of the area - cross-bedding, graded-bedding, ripple-markings - structures that have been invaluable in establishing the correct order of succession of the various sedimentary units.

The detailed mapping by the Geological Survey, outlined below, led to the establishment of a two-fold major stratigraphic grouping of the rocks in the Barberton area:-

(a) Moodies System

This, the youngest group of sediments, consists, for the most part, of quartzites, conglomerates and silty shales which probably rest unconformably on the underlying rocks.

(b) Swaziland System

(i) Fig Tree Series

This consists of shale, grit and greywacke with banded cherts, banded ironstones and some volcanic rocks. Hall (1918) regarded these strata as a lower part of the Moodies succession but in view of the quite different type of sediments and the evidence for an unconformity between the Moodies and Fig Tree sediments, the Moodies System was redefined and restricted to the upper sediments only.

(ii) Overwacht Series

This is made up of acid and basic lavas lying beneath the Fig Tree Series and it is uncertain whether there is an angular unconformity between the two units.

These various sediments and lavas were intruded by a complex of basic and ultrabasic igneous rocks known as the Jamestown Complex. Serpentinites and basic types were implaced first and were followed by the intrusion of a large mass of hornblende granite (Kaap Valley Granite). After this igneous phase all the rocks were deformed, extensively invaded by granite magma (Nelspruit Granite) and thermally metamorphosed. The later history of erosion followed by the deposition of the Godwan Formation and Transvaal System is apparent only outside the area of this investigation, and certainly post-dates the deformations to be described below.

Later work by the members of the staff of E.T.C. Mines Ltd., has suggested modifications of the lower members of the stratigraphic scheme proposed by the Geological Survey. In the course of re-investigation of certain areas west-south-west of Barberton, O. Kuschke noted that the dolomitic limestone that had been included in the Moodies System by the Survey, together with black shales and ironstones, formed a large part of what had been originally mapped as Jamestown basic intrusions. He named this, the oldest group of sediments, the Oorschot Series after the farm Oorschot 29 where they were first recognised. Gribnitz and others (1961)have cast doubts on the separate existence of the Jamestown Complex andhave suggested that the serpentinites and other basic rocks form an intimate part of the Onverwacht Series, and that the talc-carbonate and tremolite-carbonate schists should be regarded as members of the Oorschot Series.

The stratigraphic scheme adopted in this paper (Fig. 2) is a somewhat simplified one. For the purpose of the present mapping, the main grouping employed by the Survey was adopted, although in practice it was

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found that the deformed derivatives of the Onverwacht volcanic members were indistinguishable from the sheared talc-amphibolite-carbonate phyllites of parts of the Jamestown Complex: they have therefore not been separated in Fig. 2. The present observations confirm Gribnitz's opinion that there seem to be no justifiable reasons for the separation of these rocks.

In the presentation of the stratigraphic details below, it has sometimes been necessary to locate the outcrops with accuracy, and this has been done by means of a kilometre grid system, with the southwest corner of the map as grid origin.

B. STRATIGRAPHIC DETAIL

(a) Lower Swaziland System

Included in this grouping are the members of both the Onverwacht and Oorschot Series and some rocks allegedly belonging to the Jamestown Complex. The group consists of dark greenish-grey volcanic rocks, mostly of andesitic or basic type, with subordinate acid types (only seen in this area as very thin sheets southwest of Hislop's Creek 044016). Much of this group, in the area mapped, is of pyroclastic origin (085014). As noted by the Survey (van Eeden and others, 1956), these volcanic rocks are often in a highly sheared condition with much talc-chlorite phyllite. In Fig. 2 the sheared volcanic rocks have been differentiated from the relatively undeformed material. These sheared rocks may pass into talc-chlorite and talc-chlorite-carbonate phyllites which contain no obviously unaltered igneous material, and which appear indistinguishable from some of the phyllites within the area originally mapped as Jamestown Complex (van Eeden and others, 1956, p.117). Among these volcanic rocks and associated phyllites are bands of dolomitic limestone of the type included by Kuschke in the Oorschot Series. The best exposures of these dolomites are seen in and around Hislop's Creek (037007, 032004, 052014) and they also occur to the north of the Clutha Mine (041118). A thin band of coarsely-crystalline marble (probably thermally metamorphosed) may be seen along the Havelock road close to the contact of the Kaap Valley Granite (015010). Deformed sediments also occur near the volcanic rocks around Hislop's Creek (051013, 056004), and along a belt extending from 056007 to 088019. These consist of reddish-brown and green slates with bands of chert and jasper, and are probably the same as those placed by Kuschke in the Oorschot Series.

(b) Upper Swaziland System

This group comprises the Fig Tree Series and consists, for the most part, of a great thickness of cleaved, fine-grained and coarse-grained greywacke and slate, with well-developed horizons of banded chert, jasper and ironstone. The original thickness of the series seems very variable and it is difficult to measure with any accuracy because of the complexity of the folding and the development of strong internal deformation all over the area. On the southern limb of the Ulundi Syncline (Figs. 8 and 9) the present thickness of the greywacke is at least 7000 feet, and this is certainly a minimum value for the top of the succession is not seen, and the measurements do not take into account the tectonic flattening which these strata have suffered. Probably the original thickness here was at least 10,000 feet. On the northern side of the Eureka Syncline the present thickness of Fig Tree sediments does not exceed 2000 feet, while on the western side of the same fold the Fig Tree Series is missing altogether. It is not certain how much of this reduction in thickness is original, for the rocks are very sheared and there may be strike faulting along this zone.

The greywackes that make up most of the succession are generally of a fine-grained type and show fairly regularly graded-bedded units of from two inches to twelve inches in thickness (Fig. 4). On weathered exposures at the surface this grading is sometimes difficult to detect, but on fresh rock surfaces it is generally discernable, and underground in parts of the Sheba Mine it is very striking. In the coarser greywackes, like those found on the southern limb of the Ulundi Syncline, the graded-bedded units may locally reach ten feet in thickness, and the coarsest clastic particles at the base of the units may reach five inches in diameter. Locally there are beds of slump breccia with angular and sub-angular chert and greywacke material up to six inches in diameter, excellent deformed examples of which may be seen in the bed of the Kaap River (076128) and near the Clutha Mine (037110).

The bedding surfaces are extremely regular. Cross-bedding is very rare in this formation and, when developed, is always of the small-scale "ripple-drift" type. Few of the sedimentation structures associated with turbidites occur in these greywackes. No definite slump-bedding contortions have been seen and no bottom structures appear to be developed in the graded units. It seems unlikely that the absence of sedimentation structures can be attributed to the effects of tectonic deformation for delicate sedimentation structures are still preserved in parts of the overlying Moodies shales, rocks which have suffered the same general deformation as the Fig Tree Series.

Within the greywackes and slates are bands of chert. This is generally a well-banded variety and is of great importance from both a structural and economic viewpoint. Van Eeden realised the necessity of mapping these horizons with care for, although they are numerous, they form the only recognisable markers in the Fig Tree succession.

Economically they are important for they form the locus of much of the gold deposition in the area.

There are several types of chert bands. At the base of the Fig Tree Series is a particularly persistent marker band, the so-called "Zwartkoppie Horizon", consisting of a fairly thick grey chert (up to 150 feet in thickness) underlain by a laminated green chert, known as the "green schist" which locally reaches 500 feet in thickness. The "green schist" always stratigraphically overlies the talc-carbonate phyllites ("grey schists") of the Lower Swaziland System. The origin of the "green schists" is uncertain; from microscopic examination, Koen (quoted in van Eeden and others 1956, p. 56) has suggested that they represent highly altered igneous rocks. Van Eeden and others (1956, p. 55) have suggested that they represent mylonites derived from greywackes, while Gribnitz and others (1961) are convinced that the majority of "green schists" and "grey schists" are deformed derivatives of dolomites, limestones, arenaceous limestones and cherts of the Oorschot Series. From the present observations it is suggested that the "green schists" represent deformed laminated cherts of secondary origin derived from the replacement of greywacke and, perhaps, some calcareous sediments and talc phyllite. At many localities the "green schists" are seen to consist of a network of intersecting and irregularly curving green, white and grey chert veins.

In the north of the area bands of chert are again developed at the contacts of the talc-amphibole-carbonate phyllite and the overlying Fig Tree Series. At the Woodstock Mine and in outcrops just north of the Kaap River, these basal cherts consist of grey and green cherts of normal "Zwartkoppie" type, but farther north, at the Consort Mine, the contact cherts (the so-called "Consort Bar") are dark brown on account of the presence of finely-divided crystals of biotite. These cherts lie within the thermal aureole of the Nelspruit Granite, and, although they look unlike those of normal "Zwartkoppie" type, they appear to lie in the same structural position.

From the close association of grey cherts and jaspers with banded ironstones in the main part of the Fig Tree succession, it would appear that there is some primary sedimentary association of these rock-types. In the lowest chert formations, however, although most of the individual chert sheets are arranged parallel to the bedding planes of the surrounding sediments, locally individual bands do cross-cut the bedding at moderate or high angles (Sheba Mine 21 level, 5 fracture, peg 4709 + 102 feet west), and many of the iron-stained cherts show very obvious replacement forms controlled by fractures which cross-cut the bedding (Fig. 3; also van Eeden and others 1956, p. 58).

A band of rather peculiar, strongly deformed "oolitic" rock was found at one place within the greywackes underground on the Sheba Mine

(Plate 1 A and B). When this rock was first encountered, it was realised that the deep-water environment generally attributed to the deposition of greywackes did not correspond with the shallow-water environment of oolite formation, and it was thought that the "oolitic" material might represent slumped material transported from the margin of the basin of sedimentation. When thin-sections of the rock were made, it was found that the "oolites" show a rather peculiar internal structure; they usually have a central core made up of finely-crystalline silica, and a peripheral zone of rather cloudy greyish material consisting mostly of silica and sericite. This cloudy marginal zone is often further subdivided into small cell-like units, while locally it is breached into an aperture. Several palaeontologists have looked at this material and, without exception, all have commented on the resemblance of these forms to organic structures. Dr. G. Thomas, of Imperial College, London, has further suggested that these originally nearly spherical objects most closely resemble primitive algae (Calcisphaera) of the type described from Palaeozoic strata. In view of the great age of these sediments (more than 2900 million years) the recognition of evidence of living forms seems surprising, and it is hoped to continue these investigations further in order to determine more exactly the morphology of these objects. Similar "oolitic" structures have been recognised on the farm Schoongezicht 81, about 13·5 miles southwest of Barberton (van Eeden and others, 1956, p. 59).

Volcanic rocks have previously been described in the upper part of the Fig Tree Series (van Eeden and others, 1956, p. 60 - 70). These have not been shown in Fig. 2 because much of the material originally classified as being of volcanic origin now appears to be of sedimentary origin. The above authors (p. 67) noted that the basal portions of the lava resemble greywackes and that in places it is bedded with lenses of shale. From the present observations it seems that much of the material is bedded, coarse, felspathic greywacke. In the upper part of the "lava" horizon is a nodular rock which has been interpreted as a pillow lava. At some localities where this is well-exposed (southeast of the Clutha Mine 042110), the nodules are associated with angular fragments of chert and jasper while the matrix of the nodular structures looks like a greywacke and is clearly bedded. This rock may well be a slump breccia, the nodules being disrupted lumps derived from a previously deposited greywacke. The nodules are flattened in the plane of slaty cleavage (which cross-cuts the bedding). This preferred orientation appears to be the result of tectonic deformation, and its significance will be discussed in a later section.

(c) Moodies System

The sediments of the Moodies System were probably deposited under quite different conditions to those of the underlying Fig Tree

Series. The dominant sediments are conglomerates, quartzo-felspathic sandstones, siltstones, and shales with local developments of banded ironstones. Where undeformed, the conglomerates are seen to be poorly-bedded and poorly-sorted boulder beds consisting of well-rounded pebbles, mostly of black and white banded chert and greywacke of identical types to those seen in the Fig Tree Series, together with occasional pebbles of banded ironstone, jasper and a granite of unknown provenance. Pebbles of basic rocks resembling those in the Onverwacht Series are rare and it seems likely that these decomposed rapidly during transport.

The quartzites in the succession are well-bedded and sometimes contain thin lenses and sheets made up of small pebbles. Cross-bedding is exceedingly abundant, and many of the bedding surfaces show well-developed ripple-marking.

The shales and siltstones are also well-bedded with a thinly-laminated texture and often with small-scale current-bedding. Mud-cracks are locally developed (see also van Eeden and others, 1956, p. 214).

All the sedimentary features described above indicate deposition in shallow water and there is a sudden change in both rock-type and condition of sedimentation from the Fig Tree Series to the Moodies System. Although no angular unconformity has been observed between the two groups of rocks, the presence of pebbles of Fig Tree types in the basal Moodies conglomerates suggests considerable erosion of the Fig Tree Series, perhaps even uncovering of the basement locally, before or during the deposition of the Moodies System.

The most complete succession of Moodies rocks is to be seen in the Eureka Syncline (Fig. 8 for the location of this fold), and on the south limb of this structure the maximum measured thickness of sediments is about 10,000 feet. The flattening seen in clastic particles suggests that the original sedimentary thickness here was probably of the order of about 15,000 feet. Within the Eureka Syncline the succession consists of three main quartzite members (the lowest having a thick development of basal conglomerate) and three pelitic members. For further details the reader is referred to van Eeden and others (1956, p. 71 - 88).

(d) Kaap Valley Granite

The southeastern margin of this circular granite mass is included in Fig. 2. This is a hornblende granite, usually very homogeneous, which locally shows feebly-developed planar flow structure. Inside the mass, xenoliths of basic rock are occasionally seen and are sometimes very angular in form suggesting that the basic material was already jointed before the emplacement of the granite. There is no doubt that the mass is intrusive and, at exposures on or near the contact, thin sheets of microgranite penetrate the surrounding phyllites of the Lower Swaziland System. In the area in Fig. 2 the granite contacts are not well-exposed and the marginal intrusive phenomena may be best studied along the side of the road to the Agnes Mine, two miles southwest of Barberton. Weakly-developed thermal alteration is confined to a very narrow zone around the mass and its effects are often difficult to separate from those of the low-grade regional metamorphism. All the structural features of this granite seem to indicate that it is a relatively "high level" igneous pluton, probably of "bubble-like" form.

(e) Nelspruit Granite

Of the main mass of the Nelspruit Granite only the margin near Joe's Luck (086137) was mapped in any detail, although many small apophyses and sheets of granite and granite pegmatite were investigated between Joe's Luck and the Consort Mine. This granite shows a well-developed composition-banding, or foliation, which is locally folded (135137) (Plate 2). The sedimentary rocks of the Moodies System and Fig Tree Series are extensively metamorphosed and recrystallised around the granite (see van Eeden and others, 1956, p. 157 - 160 for details), and, although they are often veined by granite and pegmatite, the contacts with the main granite are sharp (Plate 3). Locally there may be some indication of transference of material across a very narrow contact zone, but there are no extensive transitions between granite and sediment. Northwards across the contact at Joe's Luck, a zone of metamorphosed sediments with sheets of granite is succeeded by a zone of granite with schlieren of metamorphosed sediments before the main granite is reached.

It has previously been suggested that the Nelspruit Granite is the product of granitization (van Eeden and others, 1956, p. 133) which resulted in the assimilation of large volumes of rocks of the Swaziland System and Jamestown Complex. Although it is possible that the bulk of the granite may have originated by metasomatic replacement, the nature of the contacts seems to imply that the contact zone originated by the mechanical injection of successive sheets of magma which were guided to a large extent by the bedding surfaces of the surrounding metasediments.

A fairly thick, persistent sheet of deformed granite is found just to the north of the Clutha Mine (041116). Although this has been correlated with the Kaap Valley Granite on the Geological Survey map, some of its features, particularly the porphyritic felspar crystals, seem to resemble more closely those of the Nelspruit Granite.

The granites and associated pegmatites are often in a very strongly deformed condition, frequently showing a schistosity and a very intense

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linear fabric (087136) More detailed descriptions of these structures and their significance will be given below.

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STRUCTURAL HISTORY

The main aim of this investigation was to attempt to determine the history of development of the fold and fault structures over an area of about 70 square miles. It soon became apparent that there had been several phases of deformation within this region. As each successive deformation was developed the affects were two-fold : first, a new set of structures was imprinted on the rocks; second, the structures already in existence were deformed in some way. From a geometrical analysis of these interfering sets of major and minor structures it was found that at least three separate regional deformations had affected the area.

A. METHODS OF INVESTIGATION

The geological mapping was carried out on aerial photographs at a scale of about 1 : 10,000 and the details later transferred to a base map made from the photographs at a scale of 1 : 20,000. The orientations of the following structural features were recorded:-

(a) Planar Structures

(i) Bedding.

In most of the sedimentary rocks bedding is easily discernable; only in some of the conglomerates within the Moodies System is it difficult to detect. In the strongly cross-bedded parts of the Moodies quartzites only the regional orientation of the bedding was recorded.

(ii) Composition-banding or Foliation

This is developed in certain gneissic parts of the Nelspruit Granite (Plate 2). The granite has a planar structure, some bands being richer in ferromagnesian constituents (mostly biotite) than others. It is not certain whether this is a primary flow feature or a "ghost" structure inherited from granitized inhomogeneous sediments. From the study of a small part of the edge of this granite mass, the former seems to offer the most probable explanation of this structure.

(iii) Slaty Cleavage and Schistosity

Throughout the area most of the more argillaceous sediments and some of the arenaceous sediments have a well-developed slaty cleavage which cross-cuts the bedding, generally at a low angle (Fig. 4, Plate 4). This planar fabric is developed best in argillaceous beds, which, when struck with a hammer, tend to break along a series of planes parallel to the cleavage. Sometimes the orientation of the cleavage planes varies slightly in beds of different lithology, a phenomena generally known as cleavage diffraction.

Slaty cleavage is a well-known structure in deformed rocks. Some of the earliest work on its significance was done by studying the shapes of fossils in Paleozoic slates (Sharpe, 1847). He found that the rocks had been strained and had suffered a change in shape as a result of tectonic processes, and that the maximum contraction within the rock was in a direction perpendicular to the cleavage. He also noted that the cleavage planes contained the direction of maximum elongation. These conclusions have been verified by investigations of other deformed objects in slate. Sorby (1853, 1908) investigated the shapes of originally spherical green coloured spots in purple slate in North Wales and showed that these had been flattened intoellipsoids so that the principal axis of minimum length was arranged perpendicular to the cleavage and that the other two principal axes of the ellipsoid lay within the cleavage plane. Classic studies of deformed oolites and fossils in the Swiss Alps by Heim (1878, 1921) and in the Appalachian mountains of North America by Cloos (1947) have confirmed these results. Cloos has also shown that slaty cleavage only becomes visible to the naked eye when one direction in the rock has suffered a compressive strain of about 30 per cent.

In any analysis of deformed rocks it is usual to recognise what are known as the principal axes of strain. Where an initially spherical object has been deformed into an ellipsoid (the strain ellipsoid), these axes of strain are the principal axes (X, Y and Z) of the ellipsoid such that $X > Y > Z$ (Fig. 5). Slaty cleavage is parallel to the principal plane XY of the ellipsoid and within the cleavage the direction of maximum elongation, X, is found.

Rocks may be plastically deformed in several ways. Only the final product of the strain is observed and, from this total or finite strain, it may be difficult to ascertain the actual movements which led to that specific strained state. For example, a strain ellipsoid may be developed from a sphere by non-rotational strain, that is where the principal strain axes remain constantly oriented throughout the deformational history of the rock (Fig. 5), or the ellipsoid may be developed by shear or rotational strain, where the principal axes of successive strain ellipsoids change orientation with progressives deformation (Fig. 6). The final strained products of deformation may be indistinguishable (Figs. 5 D and 6 D). With the second type of deformation it may be possible to see the shear planes or to deduce their position from other structural evidence. If this can be done then it is possible to establish another set of mutually perpendicular tectonic axes based on the actual shear movements that led to the deformation (Fig. 6). These, denoted a, b and c, can be defined as follows:-

a is the direction of translation or slip in the shear surface.

b is perpendicular to a in the shear plane (ab plane).

c is perpendicular to the shear plane ab.

The orientation of the axes of the folds that result from deforming rocks by irrotational strain need have no special orientation with regard to the strain axes X, Y and Z (Ramberg, 1959; Flinn, 1962), and, although the axes of folds developed through deforming rocks by shear lie in the ab plane, they need have no special orientation with regard to X, Y and Z, or to a and b (Clifford and others, 1957; Weiss and McIntyre, 1959; Ramsay, 1958, 1962A; Weiss, 1959). Probably most rocks are deformed by a combination of irrotational and rotational strain (Ramsay, 1962A).

In the Barberton area, objects of known original shape in the sediments are rather rare. The originally spherical objects in the "oolitic" rock seen in the Fig Tree greywackes at the Sheba Mine have been deformed into ellipsoids. These ellipsoids have their shortest axes arranged normal to the cleavage as described by Sorby (1853), and their longest axes lie in the cleavage and plunge directly down dip. A series of thin-sections were made, each containing two of the principal strain axes X, Y and Z, and one hundred measurements were made of each of the ratios X/Y and X/Z in the two-dimensional strain ellipsoid sections. Using these measurements, it has been possible to compute the shape of the strain ellipsoid and therefore the amount of strain. The ratios of the principal axes are 0·74 : 1·00 : 2·1. As the initial diameters of the spherical particles are not known, it is not possible to compute the volume change, but it would appear from these ratios that the deformation is of a "constriction" type (Flinn, 1962), that is, there has been compressive strain within the cleavage and that the intermediate strain axis (Y) is shorter than the original diameter of the spherical particles.

It might be argued that quantitative investigations of strain in rocks, like those described above, produce only trivial local detail and that attention to these features only obscures the large-scale structural pattern, However, it will be shown later that the regional variations in strain, as deduced from the orientation of slaty cleavage planes and of the maximum elongations of clastic particles, are all orderly. Any tectonic synthesis must endeavour to explain these systematic variations.

In phyllitic rocks the micaceous minerals often have a nearly perfect parallel arrangement (schistosity) in planes which lie parallel to the slaty cleavage in adjacent sediments. This can be seen especially well in the western part of the exposure in the railway cutting north-north-east of the Woodstock Mine (041128) (Fig. 13).

In the Nelspruit Granite and within its thermal aureole, a schistosity structure similar to that seen in nearly phyllitic rocks is developed. Individual mica flakes show a preferred orientation within planes which cross-cut the bedding in sediments or the composition-banding in the granite (Plate 2).

(iv) Shear Cleavage

In many of the phyllitic rocks situated between the Woodstock and Consort mines the main schistosity (iii above) is cut across by closely-spaced planes which are parallel to the axial planes of minute crenulations. In thin-section (Fig. 7) these planes are seen to be parallel to the limbs of the microfolds. The cleavage is produced by the parting of the rock along a zone where the micaceous minerals within the schistosity have been brought into parallelism on the fold limbs. An important distinction between shear and slaty cleavage is that a rock with shear cleavage will break only along certain planes (i.e. the microfold limbs) whereas a slate will cleave along any surface parallel to the cleavage.

This structure is well-developed along the east bank of the South Kaap River (036129), and also in the western part of the railway cutting just north of the Woodstock Mine (041128) (Fig. 13).

(v) Axial Planes of Folds

Wherever possible the orientation of the axial planes of minor folds was recorded. This is often a difficult structure to measure with accuracy, but in view of its importance in any discussion of the fold geometry, care was taken to obtain as many accurate observations of orientation as possible.

(b) <u>Linear Structures</u>

(i) Fold Axes

Wherever minor folds were developed, the trend and angle of plunge of their axes were measured, and the style of the folding recorded (wavelength, amplitude, shape of fold hinge, geometrical relationship of schistosity or cleavage to the fold).

(ii) Mineral Orientation Lineation

In many of the rocks of both igneous and sedimentary origin mapped in the northern portion of the area, a well-developed parallel orientation of the longest axes of individual mineral constituents gives rise to a prominent linear fabric in the rocks. The best localities to study this structure in phyllites occur north of Noordkaap (036129). It is also well-developed in the thermally altered sediments and in the Nelspruit Granite near Joe's Luck (086136). The linear structure lies in the cleavage or schistosity plane, and, where developed in sediments, it also lies at the intersection of the schistosity and bedding planes. If well-developed in the Nelspruit Granite, it is positioned at the intersection of schistosity and composition-banding.

(iii) Cleavage-Bedding Intersection

Over much of the area underlain by sediments of the Moodies System and Fig Tree Series, the intersection of the cleavage and bedding gives rise to a more-or-less prominent linear feature. Sometimes this linear structure lies parallel to the axes of nearby small-scale and large-scale folds, sometimes not. The significance of these various relationships to fold structures will be discussed below.

(iv) Elongation of Clastic Particles

The individual clastic grains within the arenaceous sediments often show an elongation as a result of deformation (Fig. 2). Without exception, this linear elongation is contained within the planar fabric of the slaty cleavage. Where the internal deformation of the rocks was severe, individual pebbles in conglomerate bands are strongly flattened in the cleavage planes, while all the long axes are arranged in a parallel or sub-parallel fashion within the cleavage (Fig. 12, Fig. 14 and Plate 5). The direction of maximum elongation rarely lies in the bedding, but in the few instances when it does, the elongation lies parallel to the cleavage-bedding intersection (iii above); normally there is a significant, but not always large, angle between these structures. A three-dimentional diagram showing the general arrangement of these structures is illustrated in Fig. 14.

It is found that certain of the structures described above appear to be genetically connected with a single deformation, whereas at some localities several sets of related structures are developed. Where several sets are present the development of one set of structures appears to have been synchronous with the deformation of other sets.

Where the style and orientation of minor structures keeps constant over fairly large areas (as for instance over much of the Eureka Syncline), mapping was carried out along traverse lines. Where the structural style was more complex, or where several phases of small-scale structures were developed, outcrop mapping produced more important results. Some time was spent underground in the Sheba and Consort mines, but in order to determine the regional structural pattern most of the field investigation was taken up by mapping surface exposures. Fig. 2 presents the results of the investigation. Most of the structural data are shown on this map, but in those areas where detailed mapping was undertaken Fig. 2 shows a simplified compilation.

At least three main periods of deformation affect the rocks. The nature of these deformations will be described and the evidence for the successive phases of movement will be outlined. The geometrical analysis of the folds and of the orientation of the minor structures has been accomplished by plotting the data on equal-area, lower-hemisphere stereograms. The locations of the major fold axes were determined by the pi-diagram method (Ramsay, 1962D). Not all the individual stereogram plots will be presented here, but most of the information that has been obtained from them has been abstracted and plotted in Figs. 8, 20, 25 and 32.

B. FIRST STRUCTURES

The most important structures that formed during the first recognised phase of deformation are a series of large folds (Fig. 8), one of the largest being the Eureka Syncline. The well-preserved sedimentation structures in the Moodies quartzites leave no doubt whatsoever of the true synclinal nature of this structure. The fold now has a curving axial plane which dips steeply to the south, southeast or east. The axial plane and both limbs dip towards the concave side of this curve, and analysis of the bedding-plane orientation (Figs. 27 and 28) shows that, apart from a small sector near the Clutha and Woodstock mines, the dips on the southern limb are steeper than those on the northern limb (see van Eeden and others, 1956, p. 77). The plunge of the Eureka Syncline is everywhere towards the west, southwest or south, often at high angles (Fig. 8), and even though in the southwestern section the limbs converge towards the south, the plunge may still be southwards. This feature might be the result of greater shortening of the crust here by plastic compression compared with that seen further to the north, but it seems more likely to have been produced by an original decrease in thickness of the individual members of the Moodies succession towards the southwest. The plunge depression in the synclinal axes appears to lie just north of the Havelock road. The southeast limb of the syncline is truncated by an important dislocation known as the Sheba Fault, and this limb is now structurally overlain by rocks of the Swaziland System.

South and east of this fault the mountains are composed almost entirely of tightly-folded rocks of the Fig Tree Series, a region of considerable structural complexity. By mapping various chert horizons and the Lower Swaziland rocks and by observing the "way-up" of the graded-bedded greywackes, a series of first folds has been established (Fig. 8). Immediately to the southeast of the Sheba Fault is a zone where isoclinal folds abound and which has been called the Sheba Anticlinorium. Extensive underground mining has shown that, without doubt, this region is made up of a number of isoclinal anticlines having cores of talc-carbonate phyllite surrounded by green laminated cherts and grey cherts, and overlain and separated by synclines of finely-grained slaty greywacke (Fig. 9) (Koen in van Eeden and others, 1956, p. 55). These anticlines sometimes show a perfect symmetry in the arrangement of the various members of the succession on either side of the core, but often the inverted part of the chert succession is thinned, or discontinuous, or it may be missing altogether. Gribnitz and others (1961) have suggested that this is the result of shearing and boudinage. Observations of the graded-bedding confirm that the greywackes "young" away from the green and grey cherts, but that the inverted fold limb is often severely thinned. Where the chert horizons can be traced over the fold hinges there is often an indication of a strong flexure, or buckle, component in the fold, but there is no abnormal fracturing or brecciation. It would appear that during this folding these rocks all behaved like viscous substances with the cherts slightly more viscous (more competent) than the surrounding greywackes, and with the talc-carbonate rocks behaving as a "decollement" zone of very low viscosity.

Some distance to the southeast of the Sheba Anticlinorium is a particularly well-developed, nearly isoclinal syncline (the Ulundi Syncline). Much of the southern limb of this fold is made up of an abnormally thick development of coarse greywacke, but to the east there is a similar belt of coarse grits and greywackes on the northern limb of the fold (van Eeden and others, 1956, p. 66). If these were deposited in the same sedimentation basin it might mean that the isofacies lines are cross-cut by the later tectonic lines, a feature well-known in the Mesozoic rocks of the Swiss Alps (Arnold Heim, 1916; Trümpy, 1960). The core of the Ulundi Syncline does not appear to show the minor fold complications of the Sheba Anticlinorium. To the southeast of the Ulundi Syncline there are two smaller synclines and two anticlines

(Figs. 8 and 31). Geometrically these structures resemble those to the north; anticlines are nearly isoclinal with cores of talc-carbonate phyllite surrounded by the green and grey cherts of the "Zwartkoppie" type, whereas the synclines appear to have a slightly more open form. Locally, the southern limbs of the anticlines appear to be thinned out by thrust.

In Fig. 8 it will be seen that the regional structural pattern of these folds is somewhat peculiar, the fold traces are all arcuate and in the Fig Tree sediments the traces converge towards the southwest. The arcuation of the folds is believed to be the result of a subsequent deformation and this feature will be described in detail later. The first fold axes are mostly near - horizontal in the southeast of the area, but as individual folds are followed towards the southwest they plunge towards the east or east-north-east at moderate to high angles. This would indicate that where the fold traces converge a deeper structural level is exposed at the surface, and it might be tentatively suggested that in the southeast of the area the axial planes of the Sheba, Ulundi and other first folds converge at depth. This is by no means certain however, and alternative explanations could be proposed. Some of the irregularities in the folding might be the result of variations in facies and thickness of the Fig Tree sediments, for the wavelength of flexure folds is controlled by the thickness of the most competent units (de Sitter, 1956; Ramberg, 1960). These irregularities might also be partly due to later deformation.

On the northern side of the Eureka Syncline the earliest structures are difficult to recognise with any certainty on account of the complexity of development of later folds, many fault complications and rather poor exposure. However, two other large first folds are believed to exist here. These are the Woodstock Anticline (Fig. 8) and the faulted remnant of the northern limb of a syncline which is possibly the continuation of the so-called Lily Syncline seen farther to the east outside the limits of this area (van Eeden and others, 1956, p. 72). The Woodstock Anticline has a core of serpentinite and talc phyllites which have been thrust northwards over Moodies conglomerates.

In the Consort area, structurally above the phyllite-metamorphosed shale contact, another sheet of basic phyllites occurs within the main region of Fig Tree sediments (Fig. 10). The lower contact of this upper phyllite sheet is intensively silicified and resembles the "Consort Bar" (Voges, in Gribnitz and others, 1961), and locally the other contact also shows chert development. It is not certain whether this "hangingwall schist" represents a sill-like intrusion into the shales in a first fold anticline analogous to the anticlinal first fold cores in the Sheba Anticlinorium.

To conclude this section on the first structures, some comments will be made on the variation in plunge shown by the various major folds. There appears to be no direct correspondence between the plunge directions of the Woodstock Anticline and the Eureka Syncline, or between the Eureka Syncline and the various folds to the southeast of the Sheba Fault. The following hypotheses are put forward to account for this lack of correspondence:-

(i) It is possible that not all these folds are of the same generation, perhaps those in the Fig Tree Series were initiated before the deposition of the Moodies sediments. It has been noted by van Eeden and others (1956, p. 87) that some of the chert pebbles of Fig Tree type in the Moodies conglomerates show contorted banding interpreted as tectonically formed folds, and it has been suggested that the Fig Tree Series was subjected to folding and erosion before the deposition of the Moodies System (van Eeden and others, 1956, p. 87; Hunter, 1961).

(ii) Another possible hypothesis, in some ways connected with (i) above, is that the Fig Tree sediments were tilted towards the northwest and eroded before or during the deposition of the overlying Moodies System. The axes of fold structures developed by a common deformation of the two unconformable groups of rocks would inherit the different components of dip of the two formations as a difference in plunge. If there were a plastic compression or flattening component accompanying the formation of the folds, the variations in plunge would be likely to exceed greatly the initial variation in dip of the surface (Ramsay, 1962A).

(iii) During the formation of folds it seems unlikely that the regional compression would be taken up by uniform shortening across the folds. De Sitter (1956, p. 242) has discussed the consequences of non-uniform compression. He has shown that regions of abnormally strong compressive strain must coincide with culminations in the folds, while those of small strain coincide with plunge depressions. Zones of abnormally large shortening must be compensated by zones of abnormally small strain; therefore the plunges of folds developing in adjacent zones of large and small strain might be expected to show considerable variation.

C. SECOND STRUCTURES

The most striking rock fabrics developed over the whole area have resulted from the intense strain suffered by the rocks during the second period of deformation. These fabrics are both planar (slaty cleavage) and linear (grain elongation). The significance of cleavage has been discussed above, but further details will be presented concerning the nature of the particle elongation.

Many of the arenaceous rocks of the Moodies System and Fig Tree Series contain deformed clastic grains, and many pebbles in the conglomerates of the Moodies System are strongly deformed. These grains and pebbles are always flattened in the slaty cleavage and their long axes are usually fairly steeply inclined within the cleavage. It is interesting to compare the pebble orientation of undeformed conglomerates with those where the cleavage is strongly developed. Undeformed conglomerates are developed only outside the area shown in Fig. 2, and at one locality about 5·5 miles east-south-east of Barberton the orientations of the longest axes of 50 undeformed pebbles were recorded (Fig. 11). There is a considerable scatter within a great circle zone which represents the plane of the bedding. Plots of the longest axes of 50 deformed pebbles from a roadside exposure south of the Kaap River (Fig. 2, 079130) are illustrated in Fig. 12. There is a very strong preferred linear orientation which lies within the plane of the slaty cleavage and is the result of intense tectonic elongation overprinted on the initial sedimentary fabric.

Quantitative estimates of the strain in these deformed conglomerates is somewhat difficult because of the original shape factor and the initial preferred orientation of these pebbles. The techniques for separating the primary sedimentary factors and the secondary tectonic fabric are to be discussed in another publication. Here, only the results of this analysis will be noted: the dimensions of the individual deformed pebbles at this locality show a great range in variation (1 : 1·2 : 1·4 to 1 : 1·5 : 4·4 to 1 : 5·7 : 7) but this is almost entirely the result of the original shape factor. Several independant calculations give the tectonic strain components ·68 : 1 : 1·45. This represents a minimum value of strain for the whole rock, because the matrix was strained considerably more than the individual chert pebbles.

An easily accessible section where the relationship of the deformed pebbles, cleavage and bedding may be studied is situated in the railway cutting above a prominent meander bend of the Kaap River, just north of the Woodstock Mine (045128) (Fig. 13). In the eastern half of this section the thin bands of talcose phyllite show a well-developed slaty cleavage which crosses the bedding plane, and in the most easterly part of the section individual pebbles are flattened in the cleavage (Plate 5), with their long axes plunging steeply to the west (Fig. 16). The cleavage and bedding intersect to give a rather weakly developed linear structure which plunges to the west and west-south-west at a slightly lower angle than the direction of maximum pebble elongation. These relationships are summarised in Fig. 14.

One rather interesting effect of this rock strain is that the geometry of sedimentation structures is somewhat modified. The Moodies quartzites often show well-developed cross-bedding. Generally, the cleavage and bedding cross each other at a fairly low angle, and the angle of inclination of the cross-beds against the truncating bedding surface is about 15 degrees. In the core of the Eureka Syncline, however, bedding-planes may be found locally which are oriented ar right angles to the cleavage (e.g. on the south side of the acute bend in the road to the Fairview Mine at 041076). At this locality the angle between the foresets and truncating bedding surfaces may reach 60 degrees. The original angle of repose of the foreset beds in the water-deposited sandstones could never have exceeded 30 degrees. The oversteepening is entirely the result of internal deformation (Ramsay, 1961, Fig. 14c, 16c).

In regions where folding and the production of cleavage have resulted from the same deformation, the cleavage is parallel or sub-parallel to the axial planes of the folds. One well-known outcome of this geometrical relationship is that, on beds which are inverted, the cleavage dips less steeply than the bedding planes, and on normal strata the cleavage dips more steeply than the bedding planes (Leith, 1923; Wilson, 1961). At an early stage in this investigation it was found that the application of this rule did not hold throughout this area. For example, at the Clutha Mine the bedding-planes in the Fig Tree and Moodies sediments are practically vertical whereas the cleavage dips at an angle of from 55 degrees to 70 degrees towards the south or south-south-east (Fig. 4). Application of the cleavage-bedding rule would suggest that the base of the succession lay to the south and that the rocks "young" towards the north. All the evidence of the "way up" of the succession based on the interpretation of abundant cross-bedding in the Moodies quartzites and graded-bedding in the Fig Tree greywackes proved that the beds "young" towards the south. It can therefore be deduced that either the structures "face" downwards (Shackleton, 1958; Wilson, 1961) and that the Eureka Syncline is the nose of a recumbent anticline, or that the cleavage is the result of a secondary strain superimposed on the pre-existing folds. There is no doubt whatever of the true synclinal nature of the Eureka fold and the second deduction must therefore be the correct interpretation, the cleavage is superimposed on pre-existing folds. This was verified on both a small and a large scale.

In the Fig Tree greywackes exposed at the surface about one mile west-north-west of Sheba (082080), prominent and well-exposed first folds are found, and a section of one of those folds some 20 feet long exposed in a road cutting is illustrated in Fig. 18. The fine-grained greywackes are strongly cleaved and this cleavage cuts across the axial plane and the limbs of the fold. In the stereogram analysis of this structure (Fig. 19), plots of the normal to the bedding-planes scatter about a great circle from which the axis of folding (F_1) may be deduced. Poles to the slaty cleavage-planes, recorded at various points in this structure, are plotted as crosses, and the average position of the cleavage-plane is located on a dashed great circle. The fold axis does not lie on the cleavage, and the cleavage clearly cuts across both limbs and axial plane of the fold.

The regional orientations of the cleavage-planes are recorded in Fig. 2, on cross-sections in Figs. 9 and 10, and the trends of the structures are illustrated in Fig. 20. The cleavage surfaces have an arcuate form and clearly transect both limbs of individual first folds in the same sense; the cleavage is certainly not parallel to their axial planes. Thus, the structural relationships seen on a small scale described in the last paragraph are an exact miniature of the regional relationships of cleavage to major first folds.

From this evidence it was concluded that the present position of the slaty cleavage is the result of the superposition of some other strain component on the one that produced the first folds in the area.

When two successive strains are superimposed in a plastic substance the result of the combination of the two individual strain ellipsoids is a third strain ellipsoid. There will only be one direction of total maximum compressive strain, and this appears to account for the presence of only one slaty cleavage in these rocks, even though the rocks have been deformed in two acts. From the observed cross-cutting relationships to the later folds (Figs. 18 and 20) it is possible to describe movement plans to account for the strain. Two extreme arrangements of movements which could produce an oblique cleavage are illustrated in Fig. 21. In Fig. 21A and B, the deformation of the first folds has been accomplished completely by irrotational strain; in Figs. 21C and D, by rotational strain (shear) alone. Whether the actual movement that produced this arrangement was an irrotational or rotational type is uncertain, the first hypothesis appears more acceptable because the strain ellipsoids, where observed in the deformed "oolites" and pebbles, tend to be close to the pure shear type or to the constriction type, and not to the flattened or "pancake" type (Flinn, 1962). This is more in accord with the deformation scheme shown in Fig. 21A and B, than that shown in Fig. 21C and D.

Van Eeden and others (1956, p. 181) have suggested that the arcuate forms of the first folds were the result of a deflection of the structures by the "resistent buttress" of the Kaap Valley Granite. If this is correct, then the shears that would develop as a result of the deflection of the folds around this mass are in the opposite sense to those required to explain the relationship of slaty cleavage and first folds (Fig. 21D). If this granite was at some stage in the structural history a resistant, competent mass then this state must have been reached after the development of the slaty cleavage.

During the second phase of deformation very few major folds were developed. The axes of folds formed during the production of slaty cleavage are generally parallel to the intersection of cleavage and bedding (Leith, 1923). The variation in regional orientations of cleavage-bedding intersections are shown in Fig. 20. In the Eureka Syncline these intersections generally plunge steeply to the southwest or south, while in the Fig Tree Series to the south they plunge at moderate to high angles into the southeast quadrant. This difference in orientation of the superimposed linear structures on either side of the Sheba Fault results from an initial variation in the orientation of the first folds on either side of the fault (see Figs. 8 and 31). In individual first folds there is a considerable variation in orientation of the cleavage-bedding intersections, especially on the hinge of the fold (Figs. 18 and 19). However, because of the very tight nature of most of the first folds this variation is not very pronounced. In general, only rather slight differences in orientation of the cleavage-bedding intersections on either limb of the individual first folds can be discerned (Fig. 20). Although slight, these differences in orientation appear to be consistent.

Only in the southeast of the area where thick basal Fig Tree cherts are found, are there any large-scale second fold structures. Here they cause a series of large S-shaped deflections in the strike of the cherts (Figs. 2 and 31). The axes of these folds, and associated small-scale folds, are parallel to the intersection of cleavage and bedding, and plunge very steeply towards the south-south-east. The principles that govern the orientation of the axes of superimposed folds are now well-established; the orientation of the axes is directly related to the inclinations of the surfaces before the development of the second folds (Clifford and others, 1957; Weiss and McIntyre, 1959; Ramsay, 1958, 1962A, 1962B; Weiss, 1959). Where the first fold limbs are steeply inclined, any fold structures on steep axial planes which are superimposed on them will have a steep plunge.

The relative scarcity of large- and small-scale second folds must be in some way connected with the steep inclination of the strata before the second regional deformation. Over most of the area the intersection of cleavage and bedding generally lies fairly close to the X (maximum elongation) strain axes, and the angle between the cleavage- and bedding-planes is small. Competent beds seem to have suffered little compressive strain within the plane of the bedding, most compressive strain being sub-perpendicular to the bedding surfaces. Within the bedding the dominant strain was extensive, and within the most competent members of the succession (Moodies quartzites and coarsest Fig

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Tree greywackes) there are developed tension fissures which are mostly filled with long fibres of quartz oriented perpendicular to the walls of the fissure. These tension cracks are generally sub-perpendicular to the slaty cleavage, have the same general strike as the cleavage, but all dip outwards away from the concave side of the arc shown by the cleavage. It would appear that, because no great shortening occurred within the plane of the bedding, there was no reason why the most competent bands should buckle (Ramberg, 1959; Flinn, 1962). Also, because the cleavage-bedding intersection lies close to the X direction, any regional variation in strain that would produce a differential movement or shear component between adjacent rock masses (Ramsay, 1962A, Fig. 13) would only give rise to shear folds of very small amplitude (Ramsay, 1960, p. 90).

In the cores of first folds, and locally elsewhere, there are sometimes developed very complex minor structures as a result of the interference of the folds formed during the second deformation with those produced during the first. The best localities in which to study these structures are (i) on the road from Barberton to Sheba (033066) at the reading of cleavage "76" on Fig. 2, (ii) on the hillcrest about 1·8 miles north of Sheba (097107), and (iii) just south of the road from Belfast to Fairview (039053). At all of these localities the superimposition of the two generations of minor folds has led to the development of a complex arrangement of interfering domes and basins. Where second fold anticlines cross first fold anticlines domes are developed as a result of the mutual culmination in both sets of fold axes, while at the intersections of two synclines basins are formed (Ramsay, 1962B, p. 467 - 473).

Particularly interesting types of minor structures were developed in some of the Fig Tree cherts during the second phase of deformation. It would appear that the chert bands within the succession were exceptionally competent in contrast with the surrounding argillaceous strata. During the second deformation they generally buckled but, because of their extremely brittle nature, they developed a complex fracture pattern. In Fig. 22A the folded form is obvious, but on the outer part of the fold arc the competent cherts have been fractured and separated by tension, whereas on the inner fold arc the intensive compressive stresses resulting from the space problem led to the formation of complex thrusts. Where the individual chert bands are relatively thin in comparison with the interbedded argillaceous material, or where the folding was more intense, the breaking up of the chert bands may be more complete. Sometimes the form of the fold can still be recognised (Figs. 22B and C), but often complete rupture has led to the development of wide zones of tectonically produced breccia (Fig. 22D). Well-exposed examples of these structures are to be seen in the south of the area, especially to the south and southeast of Hislop's Creek (0522007 and 081021), and large-scale development of these features can be seen in the thick cherts around the major second folds in the locality (088024).

There was a difference in structural behaviour of the chert bands during the first and second deformations. With the development of the first folds it was noted that the chert bands were partly buckled, and partly deformed plastically, with little or no brecciation; this contrasts greatly with their behaviour during the second folding. This variation in structure shows that the competence difference in the same rocks was not the same during these two phases of deformation.

So far, attention has been focussed on the rocks in the central and southern parts of the area. Proceeding northwards from the Eureka Syncline, the metamorphic state of the rocks begins to change, the slates become more phyllitic, while against the Nelspruit Granite high-grade thermal metamorphic affects are evident and argillaceous rocks contain biotite and garnet. The exact relationships of these changes is difficult to determine on account of the presence of great strike faults and because of a reduction in the degree of surface exposure. However, in view of the structural conformity of the slaty cleavage with the schistosity in the phyllites and even within the granite, it would appear that there is a strong argument suggesting that the slaty cleavage and the schistosity are related in time. The granite and the rocks from the thermal aureole show well-developed tectonically-produced structures which were developed while the rock was crystallising, and which suggest that the granite emplacement and the second deformation were broadly synchronous. All the rocks show a fairly well-formed preferred orientation of platy and prismatic crystals, yet in thin section some are seen to be intensely mylonitized. The microgranite and pegmatite sheets that abound near the contact zone are likewise deformed; where the sheets are arranged sub-parallel to the schistosity they have been stretched and show boudinage structure (Fig. 23), whereas, where they cross-cut the cleavage at a moderate or high angle, they have been shortened and folded. In thin-section the deformed nature of these sheets is very obvious; plagioclase twin-lamellae are bent, broken and faulted, quartz crystals are granulated and locally drawn out into long threads, and muscovite plates in the once coarsely-crystalline pegmatites are sheared into small slivers. These mylonites are themselves sometimes strongly folded (Fig. 24). All the macroscopic and microscopic textures suggest that the granite was implaced during the second deformation (hence the strong preferred orientation of many of the crystalline components), but that the same general period of deformation outlasted the high temperature effects of this body, and as it cooled, cataclasis set in.

The cleavage penetrates into the Kaap Valley Granite, but it would appear that this structure was developed in an already consolidated mass.

Although it has been shown that the development of the cross-cutting slaty cleavage cannot be attributed to the strain produced by deflection of the first structures around the Kaap Valley Granite, there are indications that at some period in the structural history the granite did behave as a rather rigid block. In the contact zone of this intrusion, the normal east-south-east-dipping cleavage becomes vertical and within the granite it dips towards the west-north-west (Fig. 2). Stereogram analyses of the orientations of the limbs of the Eureka Syncline close to the granite show that the west limb of this fold (i.e. the one close to the granite) dips towards the east-south-east slightly more steeply than the east limb (see dips to east-south-east in Figs. 27 and 28). At first sight the synclinal fold appears to close upwards, but it seems that the western limb has been oversteepend. It would appear likely that the granite mass has been sheared over the sediments towards the east-south-east, probably at a late stage in the deformational history of the area (a post-cleavage deformation).

To summarise, during the second deformation the following phenomena occurred:

(i) Slaty cleavage and schistosity were developed over the whole area, cutting across both major and minor first folds.

(ii) In general, large folds did not develop over most of the area on account of the initially steep inclination of the strata on the limbs of the first folds.

(iii) The intrusion of the Nelspruit Granite and the development of its thermal aureole appears to have occurred during this phase of movement, but the granite mass cooled and was deformed by late-stage movement.

(iv) The pronounced curvature of the first fold axial planes, of the second cleavage, of tension joints, and of the various axial and linear structures related to the first and second folding appears to post-date the second deformation.

D. THIRD STRUCTURES

The evidence for a phase of folding which was later than the formation of the slaty cleavage is best seen in the northern part of the area. The relationships of cleavage, bedding and pebble elongation north of the Woodstock Mine have already been described (Figs. 13, 14 and 16). If the cleavage planes in the more phyllitic horizons are examined in detail, it is found that they are cut across by a shear cleavage (Fig. 14A). The phyllite and slate bands interbedded in the conglomerates have tiny crenulations on them resulting in a "crinkled appearance" of these surfaces (Figs. 13 and 14). These structures generally plunge at a low angle to the east or northeast, and the axial planes are inclined at low angles towards the north-north-west (Figs. 14 and 16). The development of these structures is closely controlled by the lithology as they are only seen at this locality in the slaty or phyllitic bands. The "crinkle" folds are developed in certain types of pebbles (generally of fine-grained greywackes) within the Moodies conglomerates, and not in others (cherts, coarse greywackes and granites), nor are they present in the surrounding quartzites.

Proceeding westwards along the section shown on Fig. 13 a band of phyllitic slate is seen with a fairly strong linear structure (parallel to the axes of F_2 folds). As the rocks become more phyllitic, so the third structures become more intensely developed, and parts of this section show exceedingly well-developed folded schistosity (Plate 6; Fig. 15). Parallel to the axial planes of these new folds there is a strong shear cleavage, of the type illustrated in Fig. 7, which generally dips at low angles to the north and northwest (Fig. 17).

In most of these rocks there are deformed second fold structures. These deformed lineations can be traced over the hinges of third phase folds (Plate 7; Fig. 15). The special arrangement of the deformed lineations is important; at any one locality they are positioned on a plane (some of these planes are plotted in Fig. 17). This arrangement is characteristic of rectilinear structures which have been deformed by similar folding (Ramsay, 1960, pp. 76 - 90) and it seems very likely, both from the nature of these folds (Plate 6) and the associated cleavage, that the lineations were deformed by shear, with the shear planes parallel to the fold axial planes. If this is so, then the geometry of these deformed structures may be used to determine the axes a, b and c (Fig. 6) of the shear movements. The axial plane of the fold is the shear plane ab, and the plane of deformed lineations also contains the slip direction a. Therefore, by intersecting these two planes, the line common to both, i.e. a, will be obtained (Fig. 15A) (for a more detailed discussion of this method see Ramsay, 1960, p. 89). It is now possible to obtain b (perpendicular to a in the axial plane) and c (perpendicular to the axial plane). A series of calculations of the positions of these axes has been made from the exposures in the railway cutting, and for the other localities around Noordkaap (Fig. 26). An abstract of the main orientation of the a direction is plotted in Fig. 25.

To the north of Noordkaap there are several large folds of this generation. The Noordkaap Antiform is seen north-north-west of Noordkaap, where the schistosity of the talc phyllites has a pronounced "V" shape on the map, closing towards the east. Further north, in the Consort Mine, there is a set of minor and major folds which may be classified into the No. 3 Shaft Synform and the Top Section Synform, separated by a rather complicated antiformal zone (Figs. 10 and 25). These structures are correlated with the third folds because they fold the schistosity in the phyllites, deform the F_2 mineral orientation lineations, and have a shear cleavage parallel to their axial planes. The orientation of the axial planes of these structures changes northwards from Noordkaap; they become steeper, pass through the vertical in parts of the Consort Mine, and dip southwestwards in the Top Section Synform. More work is necessary to establish the complete picture of these structures. Investigations of their geometry are hampered on the surface by lack of continuous exposure.

Two sets of folds are developed underground in the Consort Mine. Early folds, probably F_2 structures with intense mineral orientation parallel to their axes, are deformed by later, F_3, structures. A few calculations of the tectonic axes a, b and c have been made in this mine, but there is scope for further systematic work in unravelling the complexities of the two generations of folds and the geometry of the deformed lineations.

One of the major problems in the area is the significance of the great arcuate structure seen in both the first folds and in the superimposed cleavage. The axial trace of this arc is shown in Fig. 25. The structure is a synform, although the axes of folding plunge very steeply. This synformal structure folds a number of variably inclined surfaces, and the fold axis varies from place to place. The orientation of these fold axes has been determined by plotting the poles to the bedding surfaces on a stereogram (the pi-diagram method, Ramsay, 1962D), a method which is good for detecting variations in fold plunge. The data from the following curved surfaces have been plotted on separate stereograms:-

(i) Fig. 27 : northern and western limbs of Eureka Syncline; fold axis plunges 70 degrees towards 167 degrees.

(ii) Fig. 28 : southern and eastern limbs of Eureka Syncline; fold axis plunges 75 degrees towards 147 degrees.

(iii) Fig. 29 : slaty cleavage in Moodies and Fig Tree sediments; fold axis plunges 70 degrees towards 140 degrees.

(iv) Fig. 30 : poles to the variably oriented surface of the Sheba Fault; these poles lie on a small circle (a cone of apical angle 68 degrees), the axis of which coincides with the axis of folding of the southeastern limb of the Eureka Syncline (Fig. 28).

The plunges of the folds on either limb of the Eureka Syncline converge downwards, as would be expected from the downward closure of the fold, and, although not presented here in detail, similar results come from the computations of the fold axis of the arcuate structure on the limbs of folds in the Fig Tree Series southeast of the Sheba Fault. The poles to the folded Sheba Fault do not lie on a great circle. It seems likely that the locus is a complex surface, such as has been described when two inclined planes are folded together by flexure (Ramsay, 1961, p. 85; Fig. 3; 1962B). The folded Sheba Fault surface has a conical form, which is a special limiting form of the complex surface described above where the intersection of the two planes, flexure-folded together, lies parallel to the fold axis of one of them. The general relationships of all these curved surfaces are illustrated in the dissected block diagram, Fig. 31.

In the massive quartzites of the Moodies System, many of the bedding-planes show well-developed, nearly horizontal slickensides, especially on the southeast limb of the Eureka Syncline, the result of bedding-plane slip during folding. It would appear that the massive quartzite bands in the Eureka Syncline were buckled about steeply plunging axes and that, as this happened, they deformed both the slaty cleavage and the Sheba Fault. The problem is to date this folding. The axial trace of the fold arc has the same general orientation as the third folds north of Noordkaap, but neither the dips of the axial planes nor the plunge of these folds coincide. It is possible that these structures were formed at the same time, but that the rigidity of the Moodies quartzites exerted a strong control over the structures that developed in the Eureka Syncline. They are classified together in Fig. 25 because both structures deform the slaty cleavage and both have axial traces which trend northwestwards along the length of the main Jamestown Belt.

E. CONJUGATE FOLDS

Some sets of rather special minor structures are found in the slate bands of the Eureka Syncline (Fig. 32). These structures consist of monoclinal folds which, when completely developed, are arranged in conjugate pairs. They fold both bedding- and slaty cleavage-planes, and their axial planes are often accompanied by zones of brecciation. Examples of such structures from this region and elsewhere have been described previously (Ramsay, 1962C), and their mechanical significance discussed. They appear to be the result of failure of the rock on two inclined shear surfaces, and a method has been described by which the principal axes of stress may be calculated. The result of applying this method to the conjugate folds of the Barberton region is shown in Fig. 32.

In the northern and central parts of the Eureka Syncline both the maximum and minimum stress axes are sub-horizontal, whereas in the southwestern part the maximum stress axis plunges quite steeply towards the northeast. Conjugate shear folds generally develop as a result of the deformation of brittle rocks. The tear faults north of Sheba are parallel to one of the sets of monoclinal shears in this region, and show the same relative sense of movement. They are thus probably related in time to the shear folds. The orientation of the stress axes shown in Fig. 32 could be explained as the result of the strong compression that would develop on the inner side of the buckled Moodies quartzite. This is not definitely proved, however, and they may have developed at a later stage in the structural history.

The orientation of the axes of the conjugate folds has no special significance with respect to the orientation of the principal axes of strain, the axes developing at the intersections of the shear planes with the bedding surface. In the northern and central parts of Fig. 32 the axes of the conjugate folds are generally steeply inclined (Fig. 33) and are aligned parallel to the axes of refolding of the bedding-planes on the limbs of the Eureka Syncline. In the southwest of the area the orientation of the fold axes is more complex, some being very steep, while others plunge at low angles (Fig. 34).

F. FAULTS

The Barberton area has been extensively faulted. The faults are of several types and were probably developed at different times during the polyphase deformational history of the area. Some of the faults originated early in the fold history, but probably moved again during the later deformations.

Van Eeden and others (1956, p. 171) noted that the major fault structures were longitudinal or strike faults, but that, locally, faults with smaller displacements cross the bedding at moderate or high angles (oblique faults). The great strike faults seem to be developed in the overturned limbs of first fold anticlines. The best examples are the Sheba Fault (in the northern limb of the Sheba Anticlinorium) and a well-developed fault in the northern limb of the Woodstock Anticline. The broad geological relationships of these structures indicate that they are high-angle thrusts (Figs. 9 and 10), although there is sometimes slickenside evidence of other (probably later) components of movement. The Sheba Fault dips less steeply than the bedding-planes in the Moodies and Fig Tree sediments, and also less steeply than the slaty cleavage and axial planes of first folds. Locally, as for example southwest of Hislop's Creek, the fault zone is of great complexity, and to the southeast of the main fault plane there seems to be a zone of imbrication, a large slab of Moodies quartzites and shales is surrounded by, and intricately folded with, volcanic and talcose rocks of the Lower Swaziland System and with Fig Tree sediments (Fig. 2, area centred on 041015). The main strike faults were probably initiated during the first period of deformation, although most appear to have been reactivated during later periods of folding. For example, late movements on the Sheba Fault have brecciated the cleaved slates of the Fig Tree Series. On the 1:50,000 published Geological Survey map of the Barberton area (1956), the Barbrook Fault is shown to be folded by structures classified in this paper as F_2 folds.

At the Sheba Mine there are conjugate sets of fractures (often mineralized) which generally trend northeast (left-hand wrench) or east-south-east (right-hand wrench). If these are interpreted as first-order conjugate shears, the axis of maximum stress that produced them (bisector of the shear planes) lies very close to the perpendicular to the slaty cleavage. Although some movement on these shears has led to deformation of the slaty cleavage, they may be related to the same general phase of deformation as that which produced the cleavage.

Around, and to the north of, Sheba other sets of faults displace the mineralized fractures. It has already been noted that these faults lie parallel to the axial planes of the conjugate shear folds, and all could be related to a single deformation post-dating the slaty cleavage and, perhaps, synchronous with the formation of the arcuate structure of the first folds.

G. RELATIONSHIP OF MINERALISATION TO STRUCTURE

In this regional study only the broad aspects of this problem were investigated. There seems to be little doubt that, although this is a rather complex problem, it would repay careful study from both academic and economic viewpoints. It is certain that the mineralisation is structurally controlled, for the regularity of the orientation of individual ore-shoots precludes any chance arrangement.

In the Consort Mine the mineralisation is almost entirely confined to the chert "bar" that forms the phyllite-Fig Tree Series contact. This chert band is certainly folded by at least two separate phases of folding, and perhaps a folding which predates them is also present. In the northern fold, known as the Top Section Synform, the mineralised bar is locally cut by large pegmatite sheets which post-date the ore (Voges in Gribnitz and others, 1961). South of this synform the pegmatites are locally folded and show intense internal deformation and local mylonitization (Fig. 24). The deformed pegmatites generally exhibit a linear structure parallel to the axes of the F_2 folds, and these linear structures are themselves folded by another set of structures (F_3). The same two sets of folds are obvious in the mineralized "bar" : an intensely rodded structure, parallel to the axes of one set of folds, is refolded by another set of folds. In the Prince Consort Synform the principal stoped-out ore-shoots run down the limbs of the later folds (F_3) at an angle to their axes, but appear to be oriented parallel to the second linear structure. Thus, the geometry of the curving ore-shoots on the flanks of the third folds is similar to that of the deformed F_2 lineations on these folds, and suggests that the mineralisation was guided in some way by the second structures. To summarise:-

(i) The mineralisation pre-dates the development of some slightly sheared sheets of granite pegmatite which were deformed during the second regional deformation.

(ii) The ore-shoots are linear and run parallel to minor structures believed to be parallel to the axes of second folds.

(iii) The ore-shoots are deformed by the third fold structure.

(iv) Arsenopyrite needles associated with the gold mineralisation sometimes show a prefered orientation parallel to the F_2 lineation.

It is therefore suggested that the arrival of the ore fluids was broadly synchronous with the development of the second period of regional deformation, perhaps at a rather late stage in the movement, and thus broadly synchronous with the intrusion of the Nelspruit Granite. This is in agreement with the suggestion put forward by van Eeden and others (1956) that the ore fluids may have been related in space and in time to the intrusion of the Nelspruit Granite.

In the Sheba Mine the mineralisation is mostly controlled by the location of fractures, the main Sheba Fault and in particular the intersections of conjugate fractures with the basal cherts of the Fig Tree Series (Gribnitz and others, 1961). In the section on faulting it was suggested that these particular fractures were formed late during the second period of deformation. The main mineralisation here would therefore seem to have occurred at some time between the later stages of the second deformation and the initiation of the third deformation.

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CONCLUSIONS

The area northeast of Barberton has been subjected to at least three deformations, the affects of which may be summarised as follows:-

Deformation 1

This led to the development of a series of major and minor folds (F_1), the traces of which probably trended northeast-southwest, or NNE. - SSW.

Deformation 2

(i) Further compression with the principal regional compressive strain oriented north-north-west led to the formation of a slaty cleavage and schistosity which was superimposed across the F_1 folds.

(ii) Only locally were new folds (F_2) developed.

(iii) Low-grade regional metamorphism was associated with this deformation.

(iv) The Nelspruit Granite was intruded, thermally metamorphosing the adjacent strata.

(v) The main period of gold mineralisation was probably late during this deformation.

Deformation 3

(i) This led to the folding of the slaty cleavage and schistosity by large- and small-scale folds (F_3) in the north of the region.

(ii) The formation of the great arcuate structure of the first folds and of the slaty cleavage was possibly contemporaneous with the folding of the cleavage seen in the north of the area.

(iii) The development of conjugate folds and faults may have been synchronous with the formation of the arcuate structure (ii), but they may post-date it and be related to a fourth period of deformation.

In view of the close orientations of the positions of the general maximum compressive strain axes of the first and second deformations, it seems likely that these movements may have been phases of a single orogeny. The general orientation of the maximum compressive strain of

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the third deformation is approximately at right angles to those of the earlier deformations. If this feature is not the result of some abnormal local stress condition, it suggests that some considerable interval of time may separate the second and third deformations.

Both the Nelspruit Granite and the Kaap Valley Granite intrude the Swaziland sediments. Can it, therefore, be concluded that all the granitic material which surrounds these sedimentary belts post-dates the deposition of the sediments? Van Eeden and others (1956) have suggested that vast quantities of the Lower Swaziland rocks have been assimilated by the granite, and Read (1951) thinks that the basic rocks in the lower part of the succession formed an effective resistance to the introduction of granitizing fluids and protected the sediments from alteration. It is difficult to understand how regional metamorphic pressures could convert such a vast volume of rock into granite while leaving others only a mile or so from the contact unaltered. The nature of both the contact and the local metamorphism suggests an igneous origin for the granite.

Another related problem is that of the location of the pre-Swaziland basement. The sediments and volcanic rocks must have been deposited on some foundation, yet where is this seen today? It seems that the most likely solution to all these problems may lie in the recognition of the complexity of the granites and gneisses which surround the Barberton belt. As in Swaziland (Hunter, 1961) there certainly are granitic rocks with very variable characters, and it is suggested that much of this granitic material formed a fundamental basement complex on which the volcanic and sedimentary rocks were deposited. Subsequently both basement and cover were deformed together, intruded by other granites and then suffered a common period or periods of regional metamorphism.

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KEY TO FIGURES

Fig. 1 :

General geological relationship of the Archean "schist belt" of Swaziland and the Eastern Transvaal with the surrounding rocks.

Fig. 2 :

Reconnaissance structural map of the area lying northeast of Barberton.

Fig. 3 :

Chert replacing well-bedded, fine-grained greywackes of the Fig Tree Series from the "Zwartkoppie Horizon" underground in the Sheba Mine.

Fig. 4 :

Relationship of cleavage and bedding outside the mine office at the Clutha Mine. A simple interpretation of cleavage and bedding here would suggest that the rocks "young" towards the northwest, but the graded-bedding indicates that, in fact, they "young" southeastwards. This is because the cleavage is a superimposed structure not directly related to the major folds.

Figs. 5 and 6 :

Nomenclature of tectonic axes in strained rocks. An original sphere is deformed by irrotational strain (Fig. 5) or rotational strain - simple shear, (Fig. 6) - into a strain ellipsoid with principal axes $X > Y > Z$. In these figures Y is perpendicular to the plane of the diagram. Where the deformation is by irrotational strain, the orientations of the strain axes throughout deformation remain constant, but in rotational strain they change orientation during the progress of deformation. Where the deformation is accomplished by simple shear, another set of tectonic axes, a, b and c may be described; a is the direction of slip, b is perpendicular to a in the shear plane (perpendicular to the plane of the diagram here), and c is perpendicular to the shear plane.

Fig. 7 :

Folded schistosity in a thin-section of folded phyllite, illustrating the development of shear cleavage parallel to the limbs of microfolds. Taken from north of Noordkaap (036130).

Fig. 8 :

Synthesis of the structure that developed during the first phase of deformation. Lithology simplified : Lower Swaziland System and Jamestown Complex - vertical lines; Upper Swaziland System (Fig Tree Series) - unornamented; Moodies System - stippled.

Fig. 9 :

Geological section from the Kaap River to the Barbrook Fault across the Eureka Syncline, the Sheba Anticlinorium and the Ulundi Syncline.

Fig. 10 :

Geological section from northeast of the Consort Mine to north of the Fairview Mine across the Top Section Synform, the Noordkaap Antiform, the Woodstock Anticline and the Eureka Syncline.

Figs. 11 and 12 :

Stereographic plots of the orientations of longest axes of pebbles in the basal Moodies conglomerates, taken from an undeformed example (Fig. 11), and from a locality where the pebbles are strongly deformed (079130) (Fig. 12).

Fig. 13 :

Structural detail from a section exposed in a railway cutting about 1500 feet north-north-east of the Woodstock Mine, just south of a pronounced meander bend in the Kaap River (see Fig. 2 for location). The orientation of this map (with north pointing towards the bottom of the diagram) is to facilitate its use in the field.

Fig. 14 :

Schematic block diagram to show the relationships of the various planes and linear structural elements in the Moodies conglomerates in the eastern part of the railway cutting section shown in Fig. 13. Fig. 14A illustrates a detail of the nature of the microfold "crinkles" and the shear cleavage developed on the slaty cleavage surfaces in phyllites and in some individual pebbles.

Fig. 15 :

Schematic block diagram to show the relationships of the folded schistosity, third folds, and deformed linear structures at the western end of the railway-cutting section in Fig. 13. Fig. 15A shows the detail of the geometrical relationships of the deformed lineations and axial planes of the F_3 folds, and the method of computing the a direction of the shear folds.

Figs. 16 and 17 :

Stereogram analysis of pebble elongation in western and eastern portions of Fig. 13.

Figs. 18 and 19 :

Relationships of slaty cleavage and first folds, one mile west-north-west of the Sheba Mine (for location see Fig. 2).

Fig. 20 :

Synthesis of the structures that developed during the second phase of deformation. Ornamentation as for Fig. 8.

Fig. 21 :

Analysis of the geometrical affects of the superposition of two successive strains. The blocks are representative of the two-dimensional surface geology, with bedding shown as a thick black line, and cleavage traces (representing the trace of the XY plane of the strain ellipsoid on this surface) as finer lines. After one deformation the bedding has been folded (see A and C) and axial plane slaty cleavage developed. B and D represent two modifications of this geometry as a result of superimposed pure shear and simple shear respectively. Both types of combination of the two deformations (A to B and C to D) produce results where the final slaty cleavage cross-cuts the previously formed folds.

Fig. 22 :

Examples of structures seen in folded chert bands in the Fig Tree Series.

Fig. 23 :

Boudinaged pegmatitic granite sheet from the margin of the Nelspruit Granite near Joe's Luck (084137).

Fig. 24 :

Thin-section of a folded mylonitized pegmatite sheet from the Consort Mine.

Fig. 25 :

Synthesis of the structure that developed during the third phase of deformation. For ornamentation see Fig. 8.

Fig. 26 :

Stereogram plot of the calculated positions of a, b and c from shear folds (F_3) around Noordkaap.

Figs. 27 and 28 :

Analysis of bedding plane orientation on the two limbs of the Eureka Syncline.

Fig. 29 :

Analysis of slaty cleavage in Moodies and Fig Tree sediments.

Fig. 30 :

Analysis of poles to the variably oriented surface of the Sheba Fault.

Fig. 31 :

Block diagram of part of the Eureka Syncline and the structures in the Fig Tree Series south of the Sheba Fault. The block has been pulled into two parts along the Sheba Fault to show the relationship of the cross-cutting cleavage to the fold and fault structures.

Fig. 32 :

Synthesis of the conjugate shear folds.

Figs. 33 and 34 :

Stereogram plots of the axes of conjugate folds developed in the Eureka Syncline.

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KEY TO PLATES

Plate 1, A and B :

Microphotographs of deformed structures believed to be of organic origin from the Fig Tree greywacke in the Sheba Mine.

Plate 2 :

Folded composition-banding within porphyritic Nelspruit Granite near Joe's Luck (085137). In the dark micaceous horizon a moderately well-developed cleavage is to be seen (approximately parallel to the pen). This structure is believed to have been formed during the second deformation.

Plate 3 :

Complexly folded microgranite and pegmatite sheets cutting metasediments in the central zone of the Nelspruit Granite near Joe's Luck (086135).

Plate 4 :

Slaty cleavage cutting through Fig Tree greywackes and slates between the Fairview and Sheba mines (085072).

Plate 5 :

Strongly deformed conglomerate pebbles in Moodies basal conglomerates at the eastern end of the section mapped in Fig. 13.

Plate 6 :

Schistosity intensely folded by F_3 structures in the western part of the area illustrated in Fig. 13. The shear cleavage parallel to the axial planes of these structures is well-developed. This macroscopic structure should be compared with the structures seen in thin-section of the rocks (Fig. 7).

Plate 7 :

Schistosity surface in phyllites folded by many small-scale, nearly horizontal structures, "crinkle folds" of F_3 generation, seen in the western part of the section illustrated in Fig. 13. Linear structures of F_2 generation run parallel to the pen and are deformed by these third folds.

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PLATE 1A.

PLATE 1B.

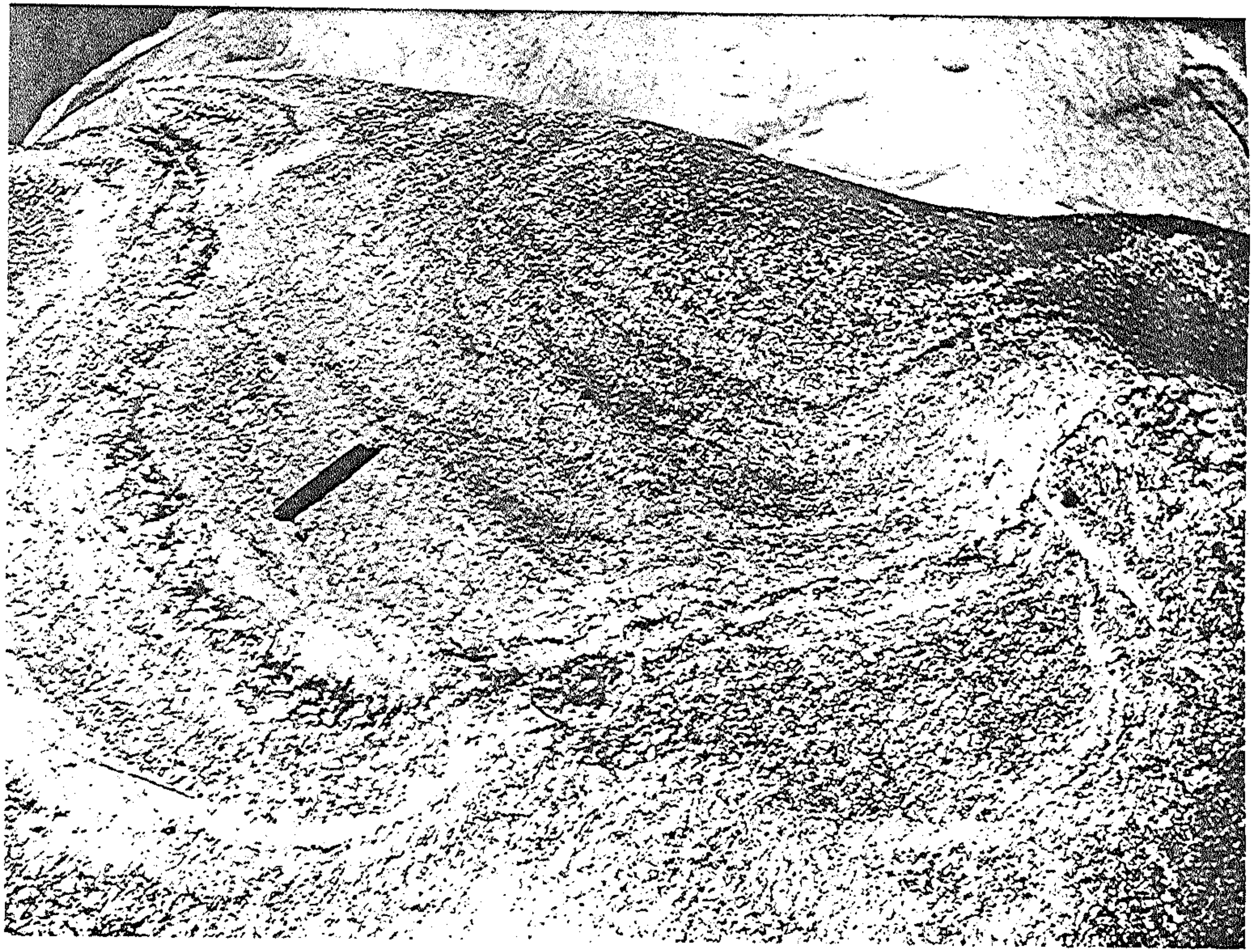


PLATE 2.

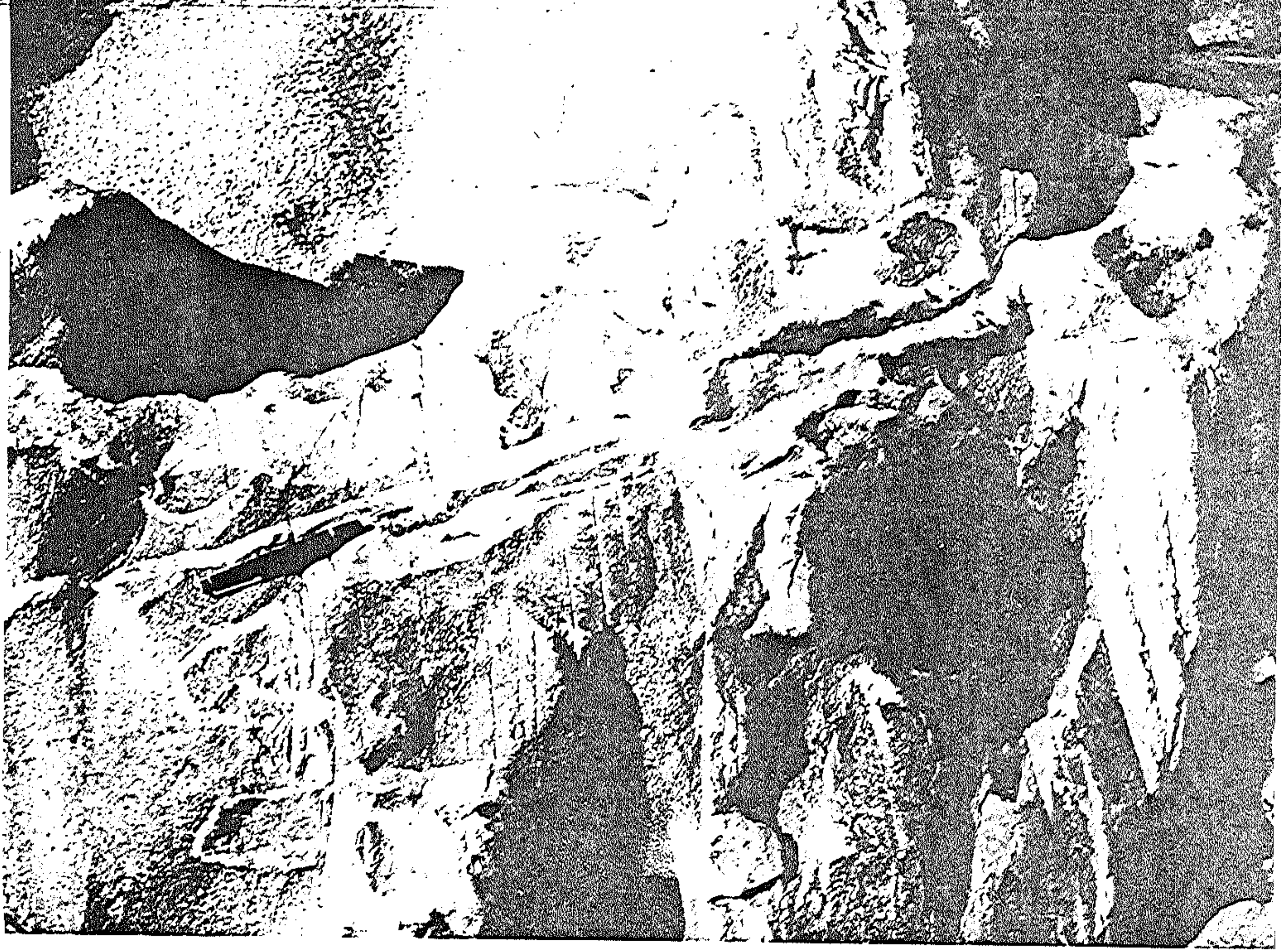


PLATE 3.

PLATE 4.

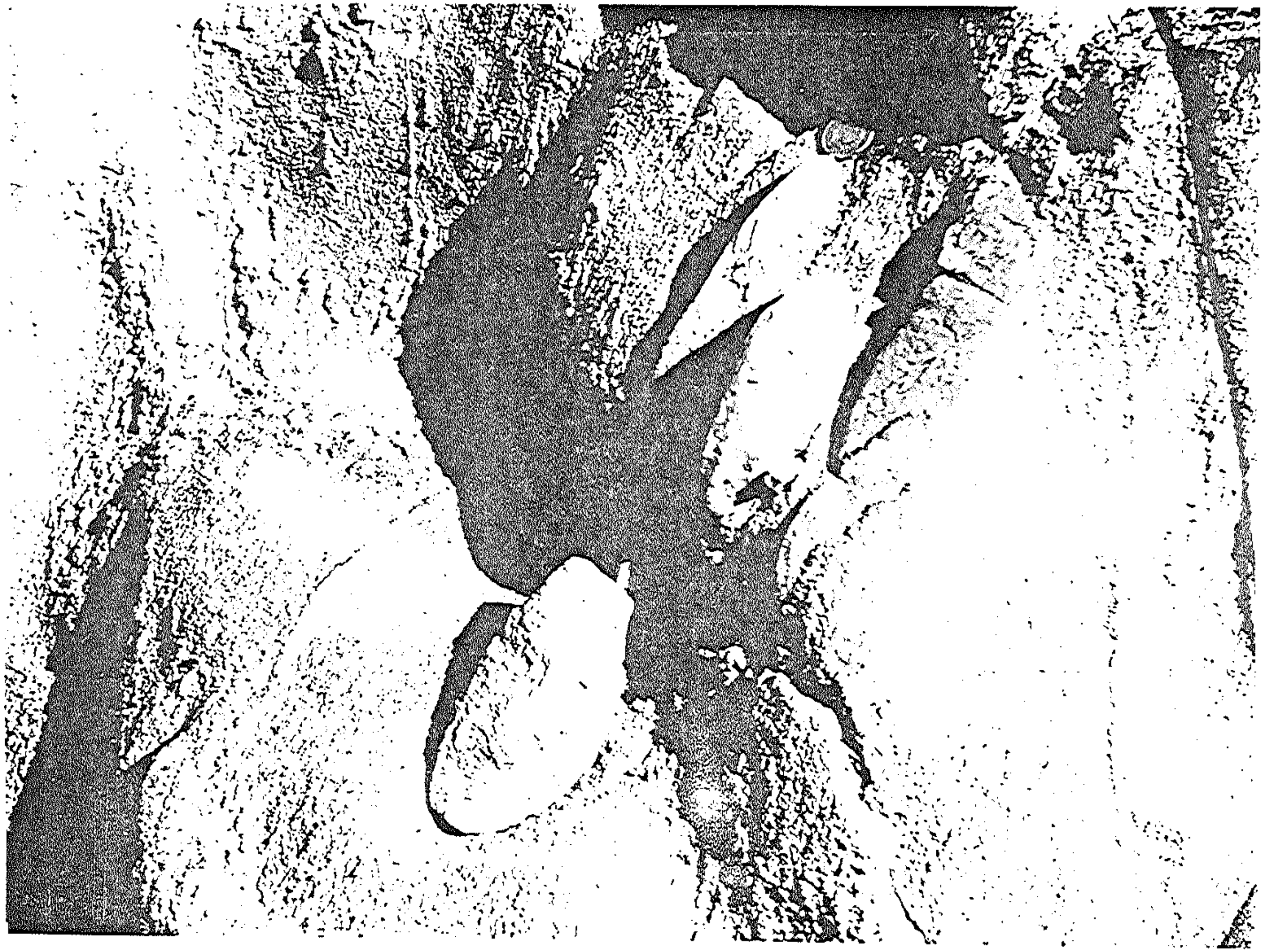


PLATE 5.

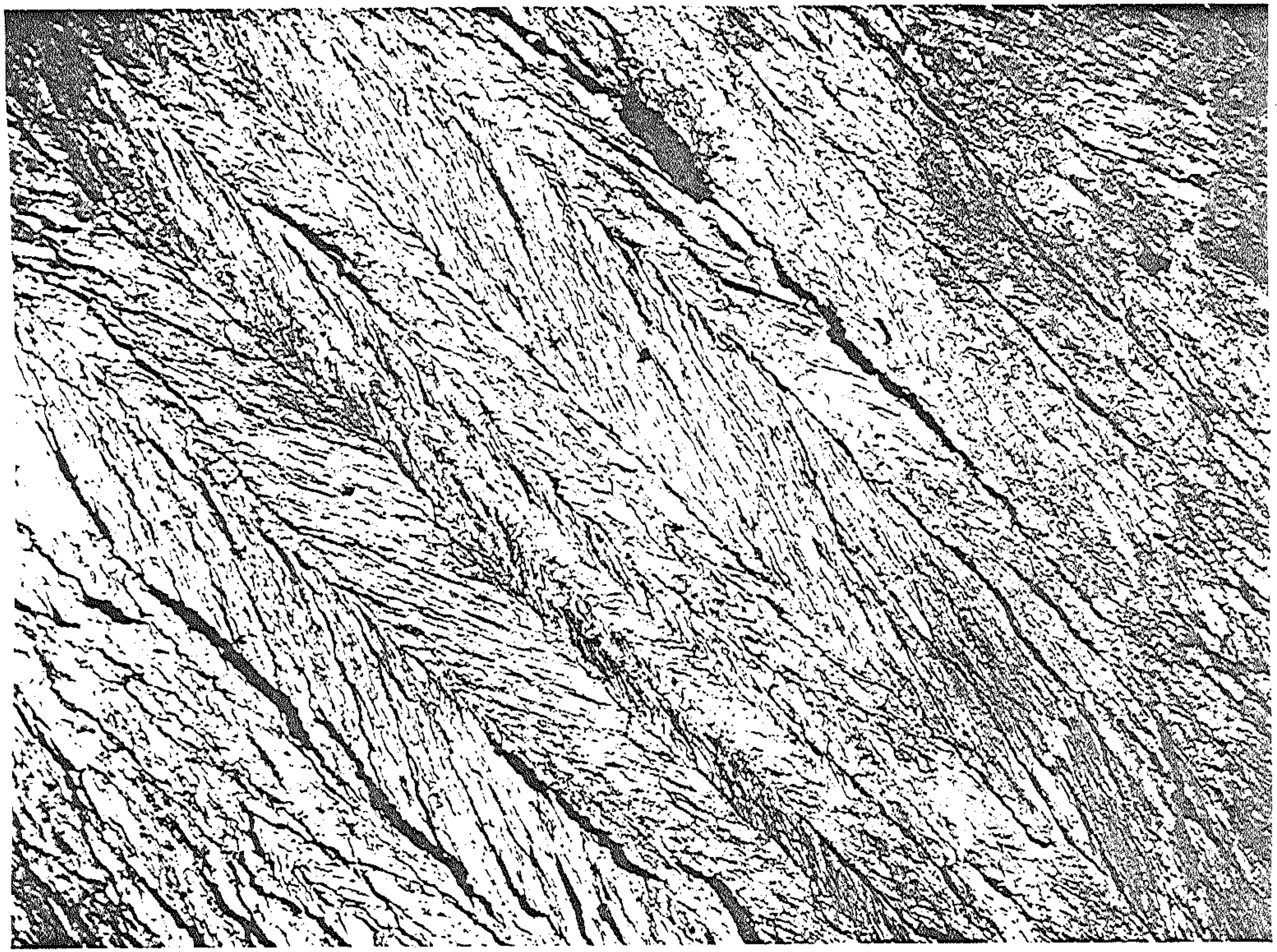


PLATE 6.

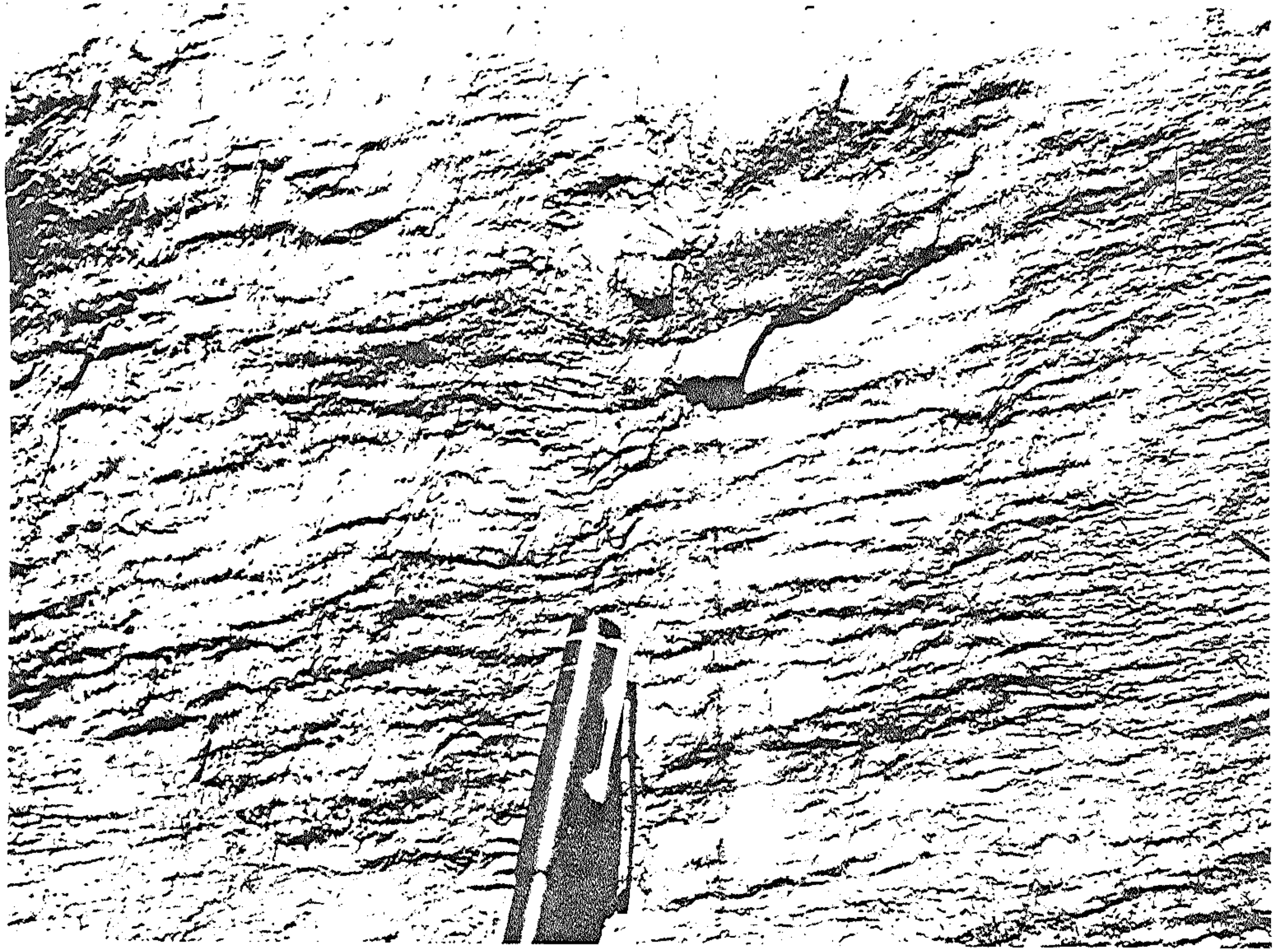


PLATE 7.

Fig. 26 Movement axes of F_3 folds.

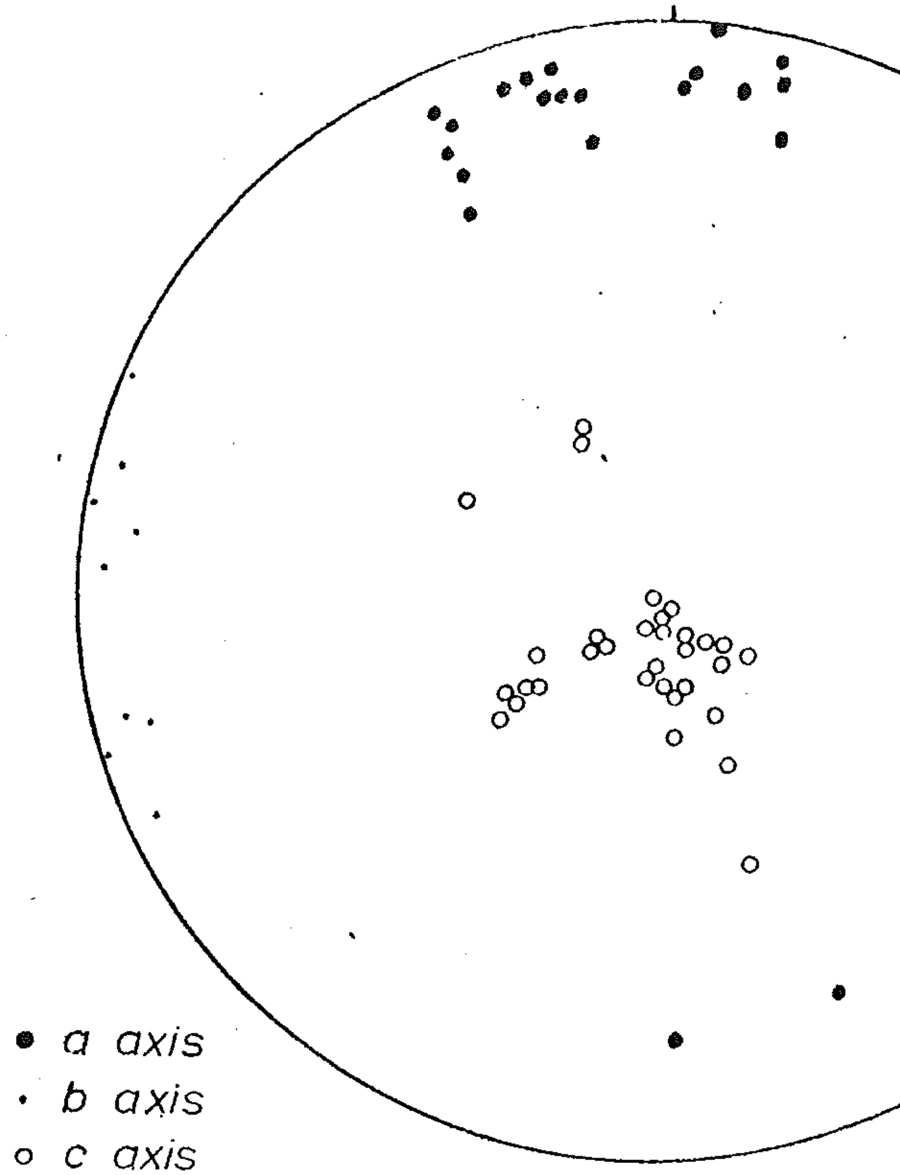


Fig. 27. Refolding of N.W. limb of Eureka syncline.

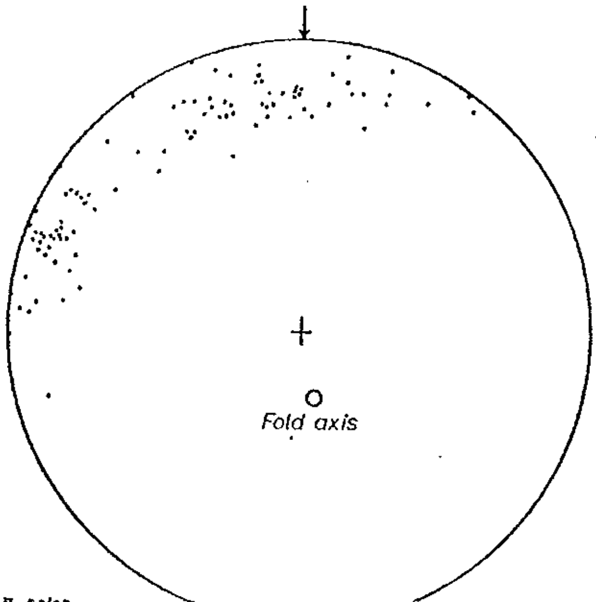
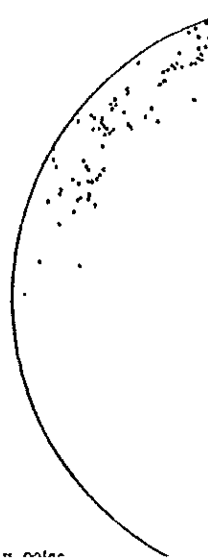
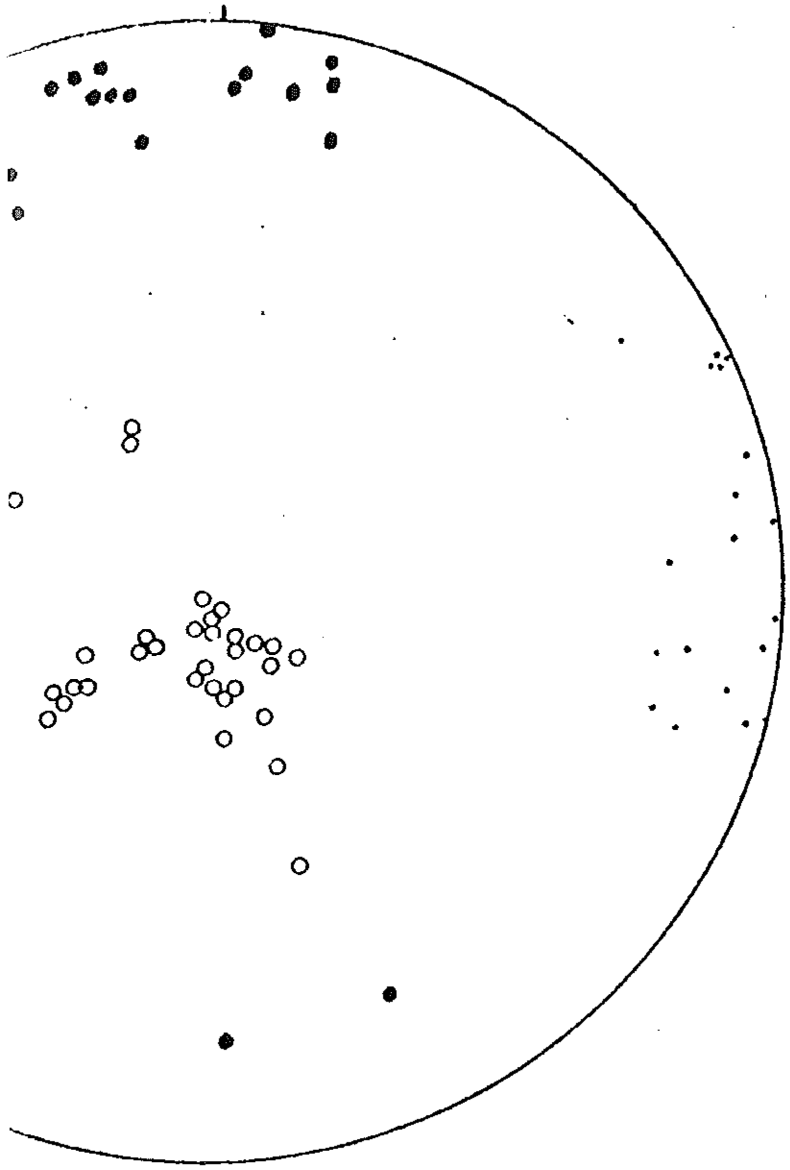


Fig. 28. Refolding of S



xes of F_3 folds.



cline.

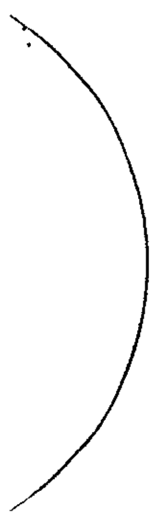
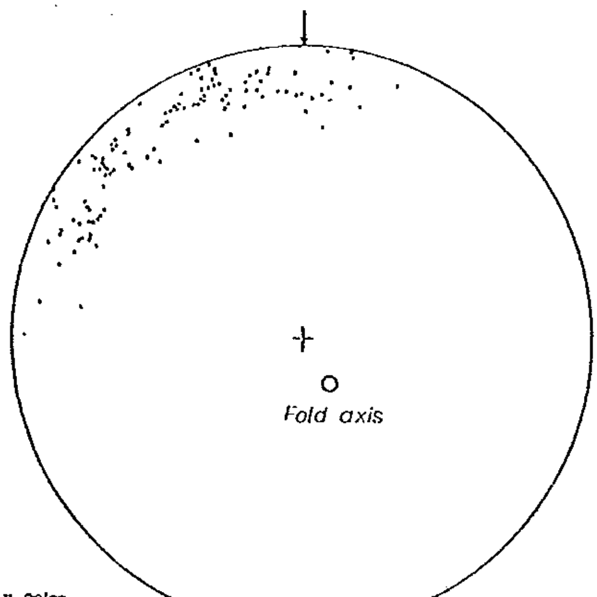


Fig. 28. Refolding of SE. limb of Eureka syncline.



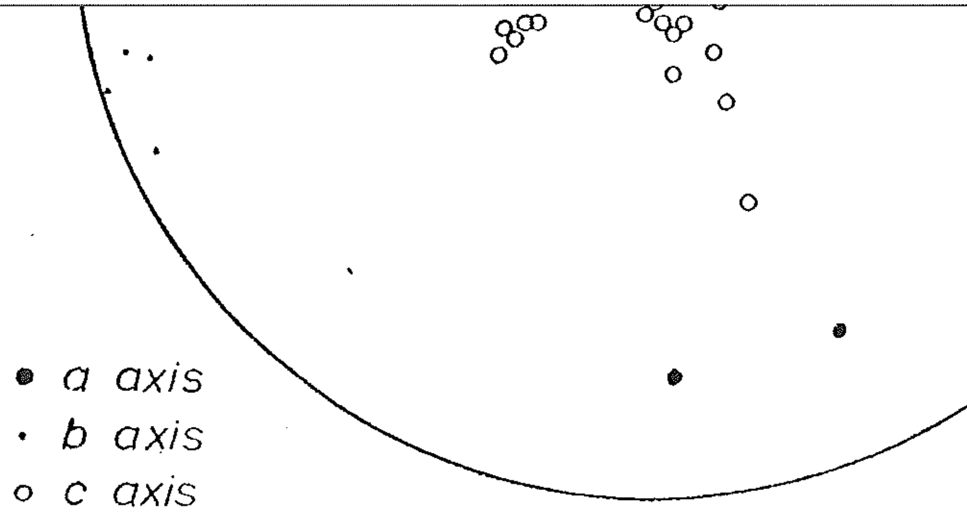


Fig. 27. Refolding of N.W. limb of Eureka syncline.

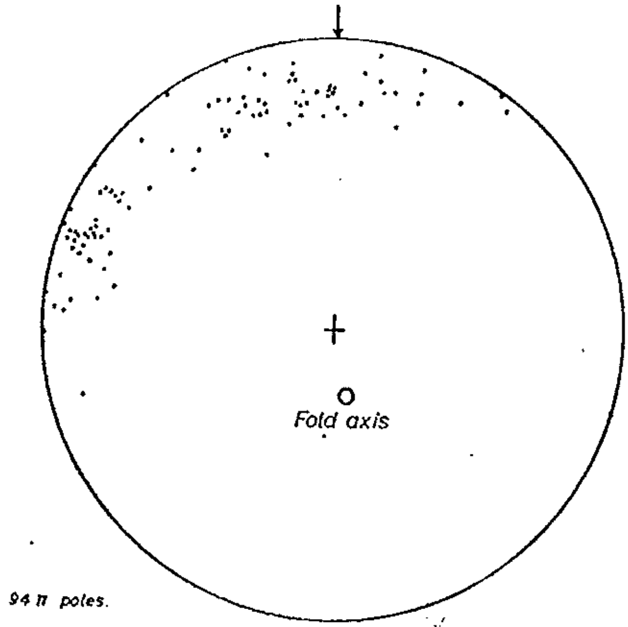


Fig. 28. Refolding of S.E. lim

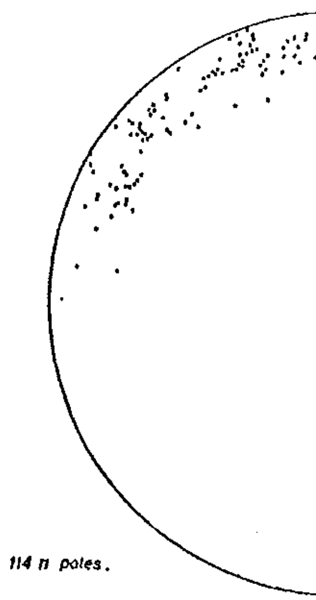


Fig. 29. Folding of slaty cleavage.

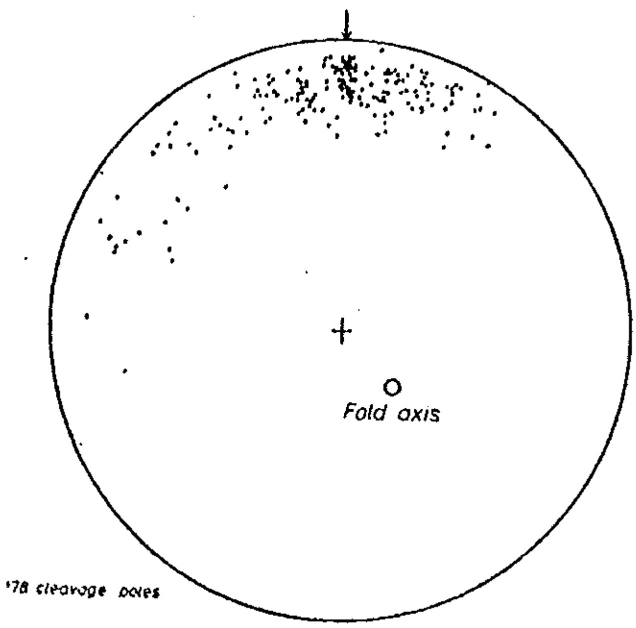
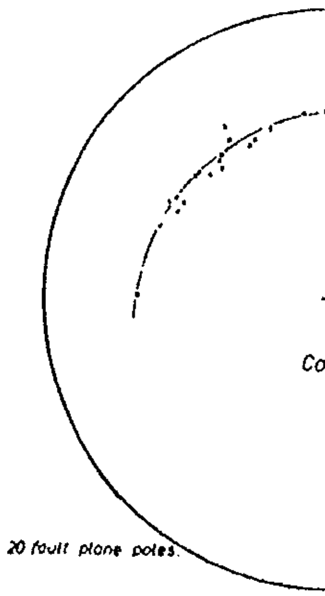


Fig. 30. Folding of the Shebo



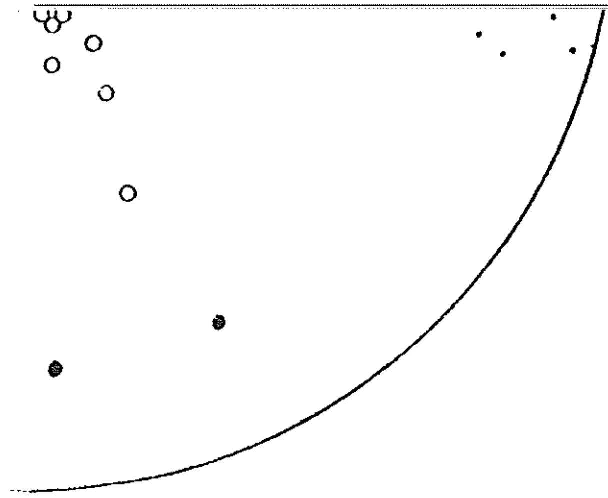


Fig. 28. Refolding of SE. limb of Eureka syncline.

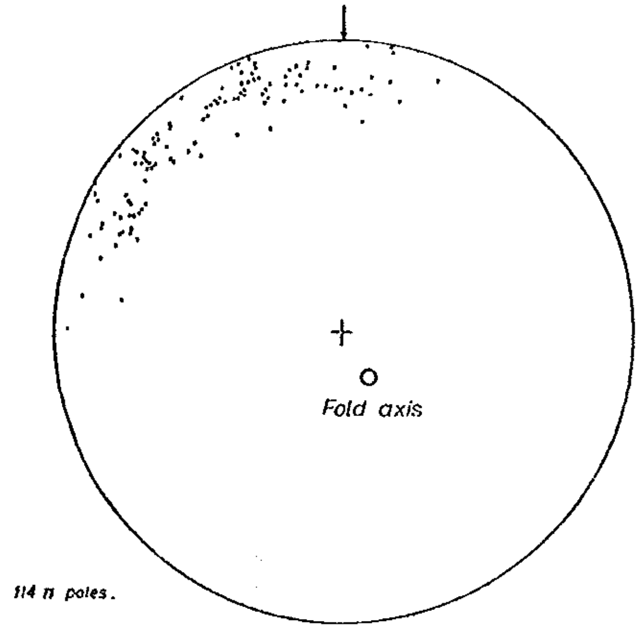


Fig.30. Folding of the Sheba fault.

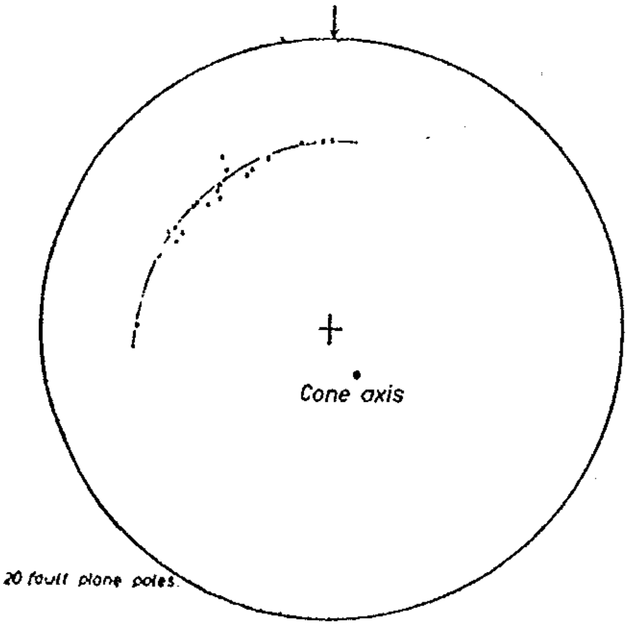


Fig. 31. Schematic block diagram of part

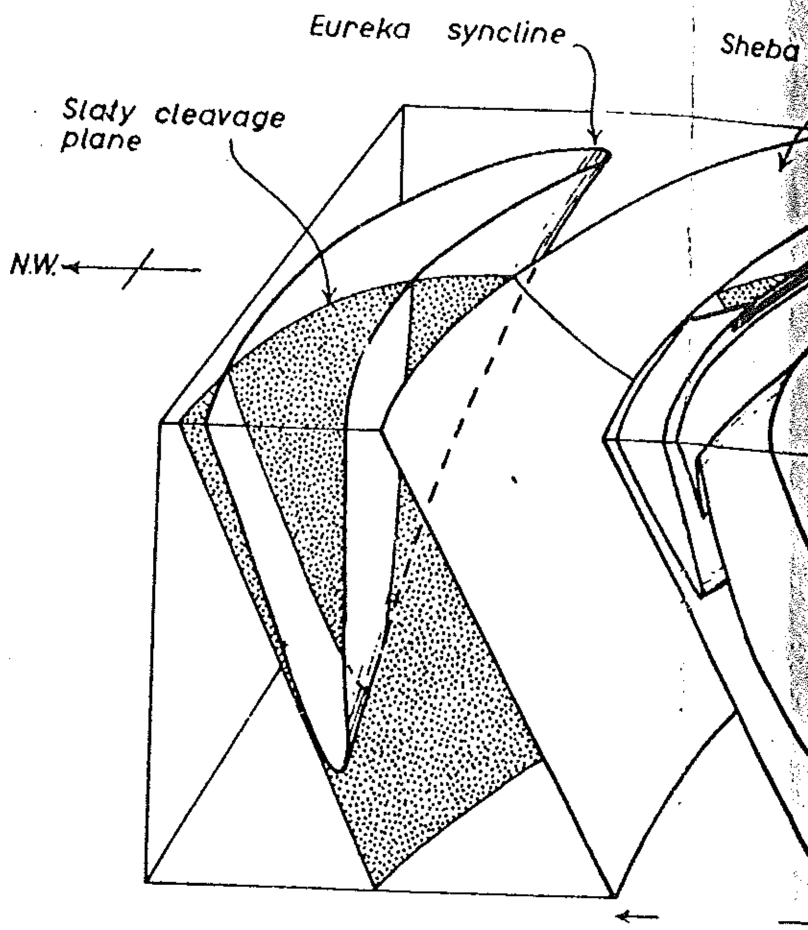


Fig. 33. Axes of conjugate shear folds, 1.

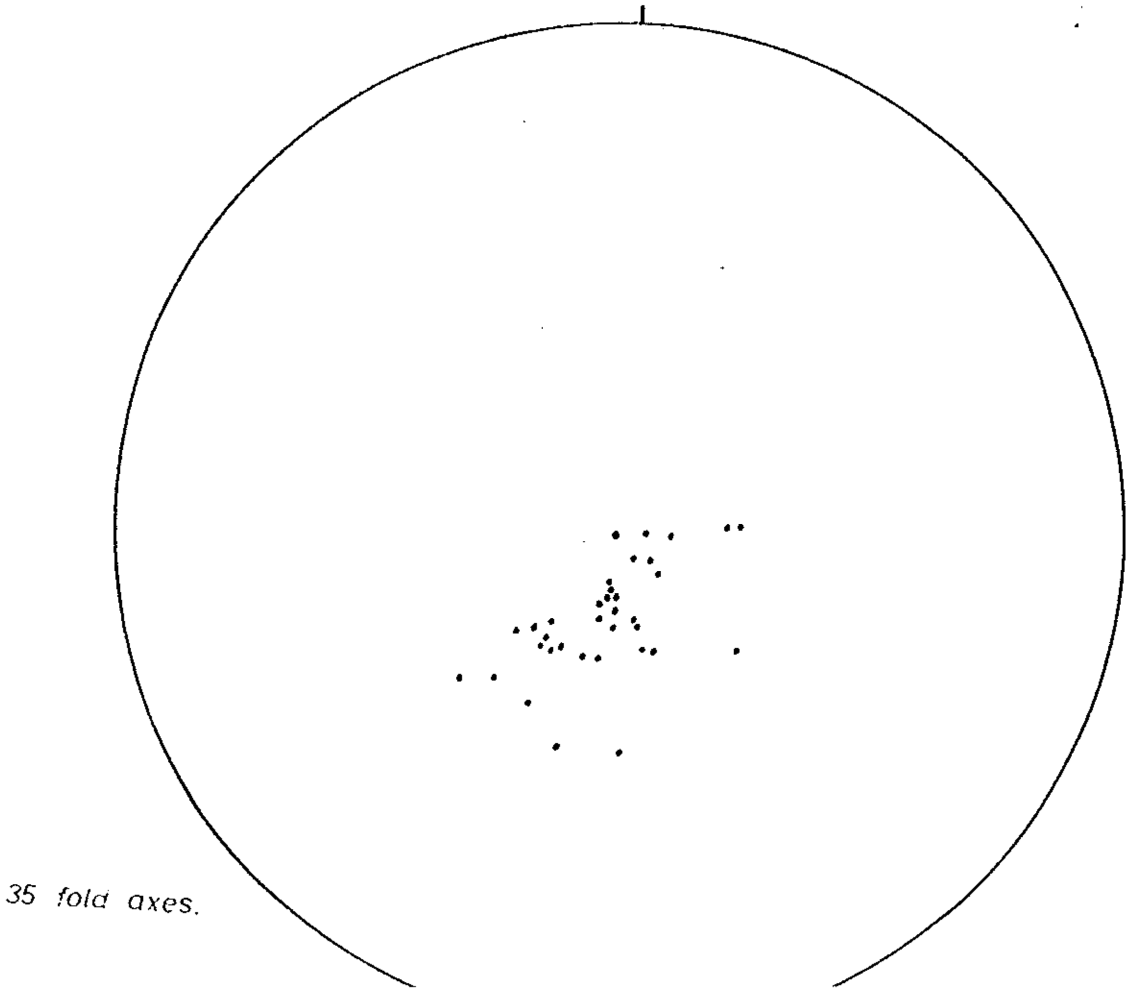


diagram of parts of Eureka & Ulundi synclines.

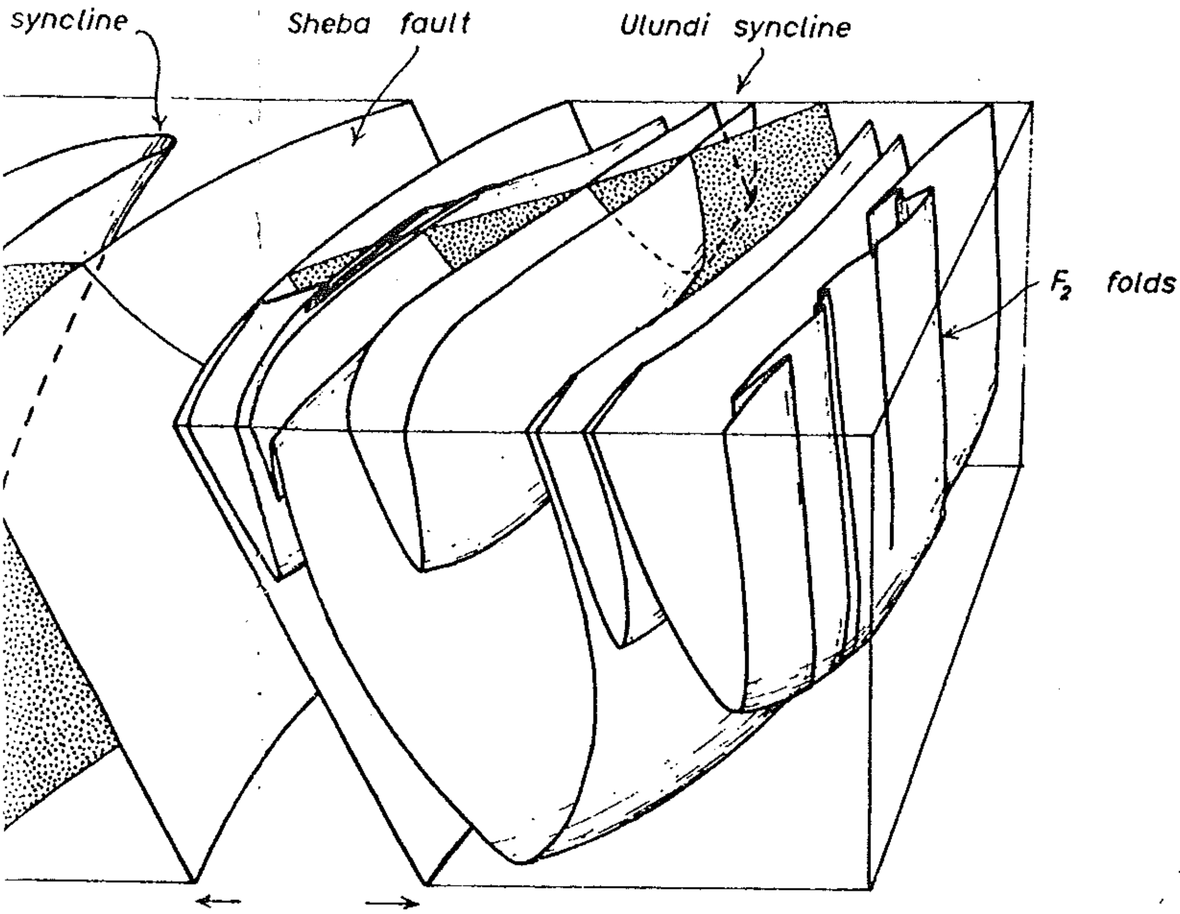
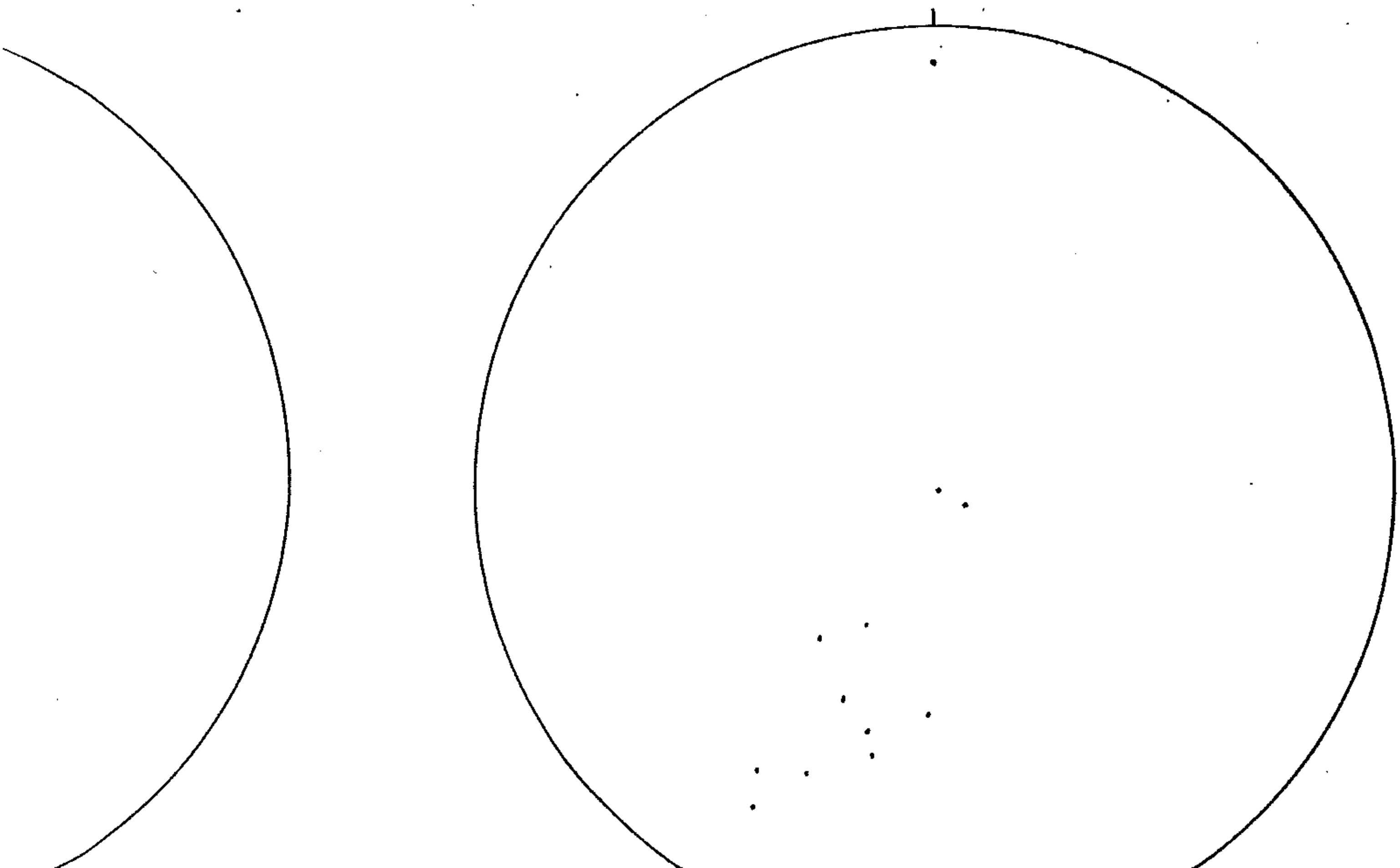


Fig. 34. Axes of conjugate shear folds, 2.



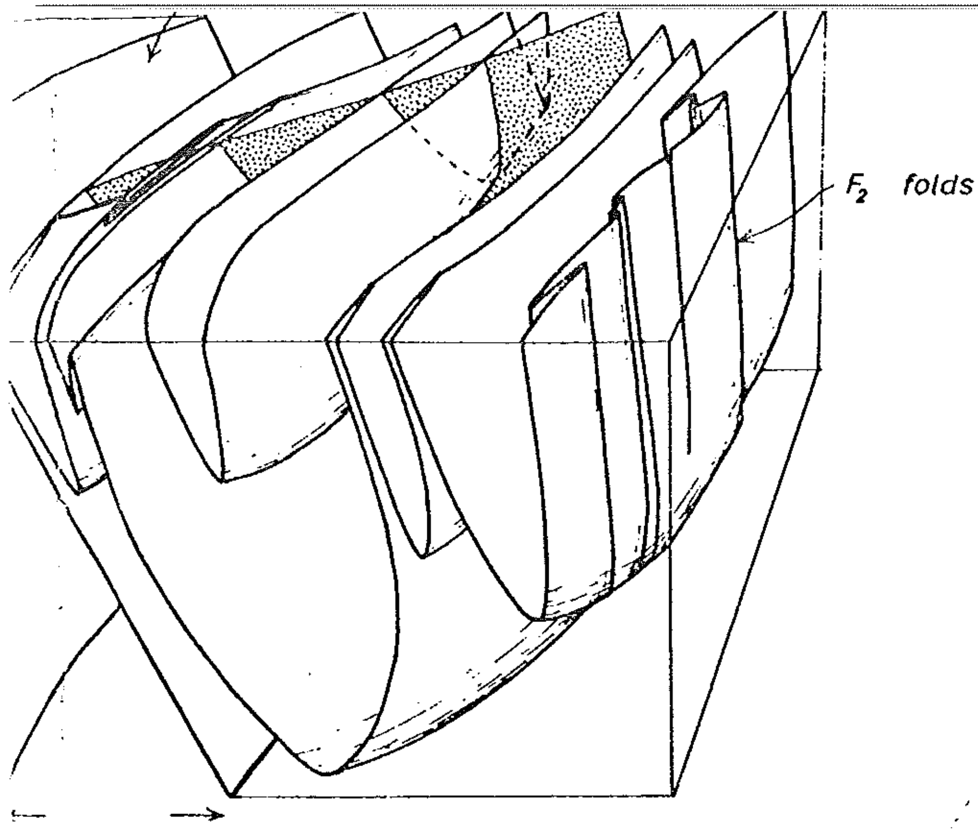
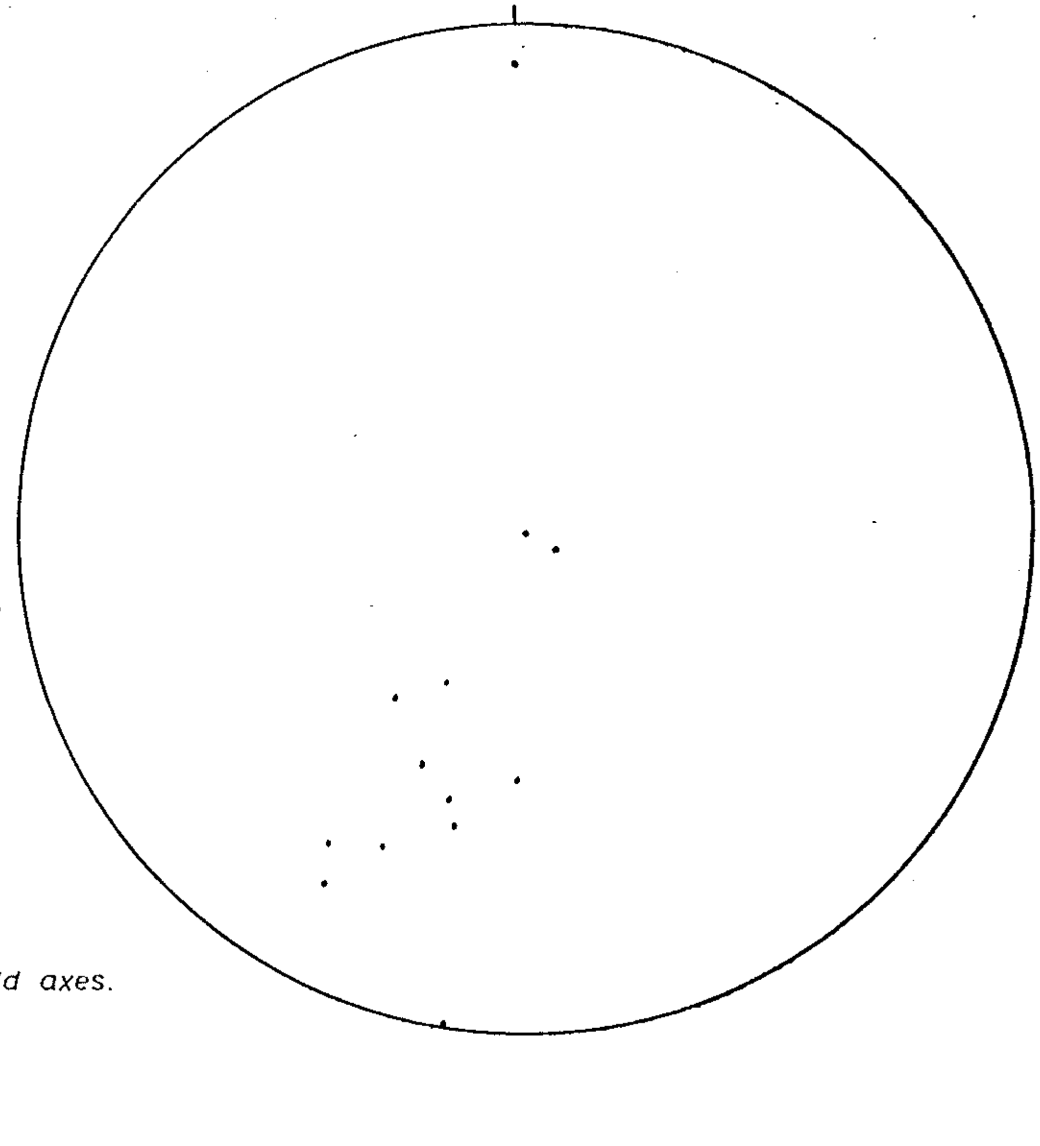


Fig. 34. Axes of conjugate shear folds, 2.



13 fold axes.

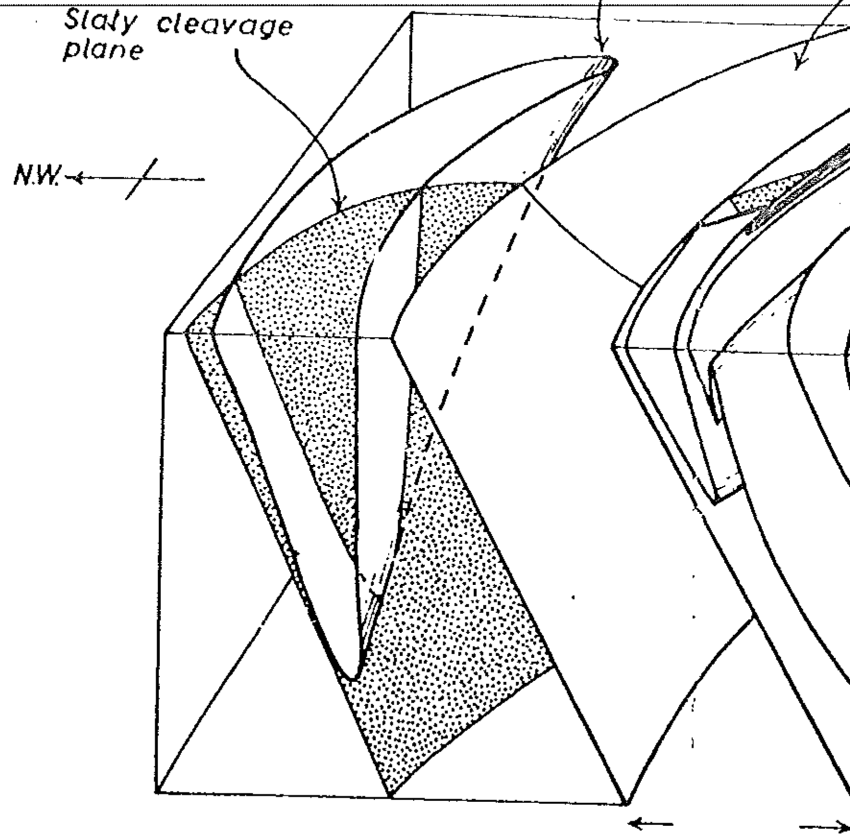
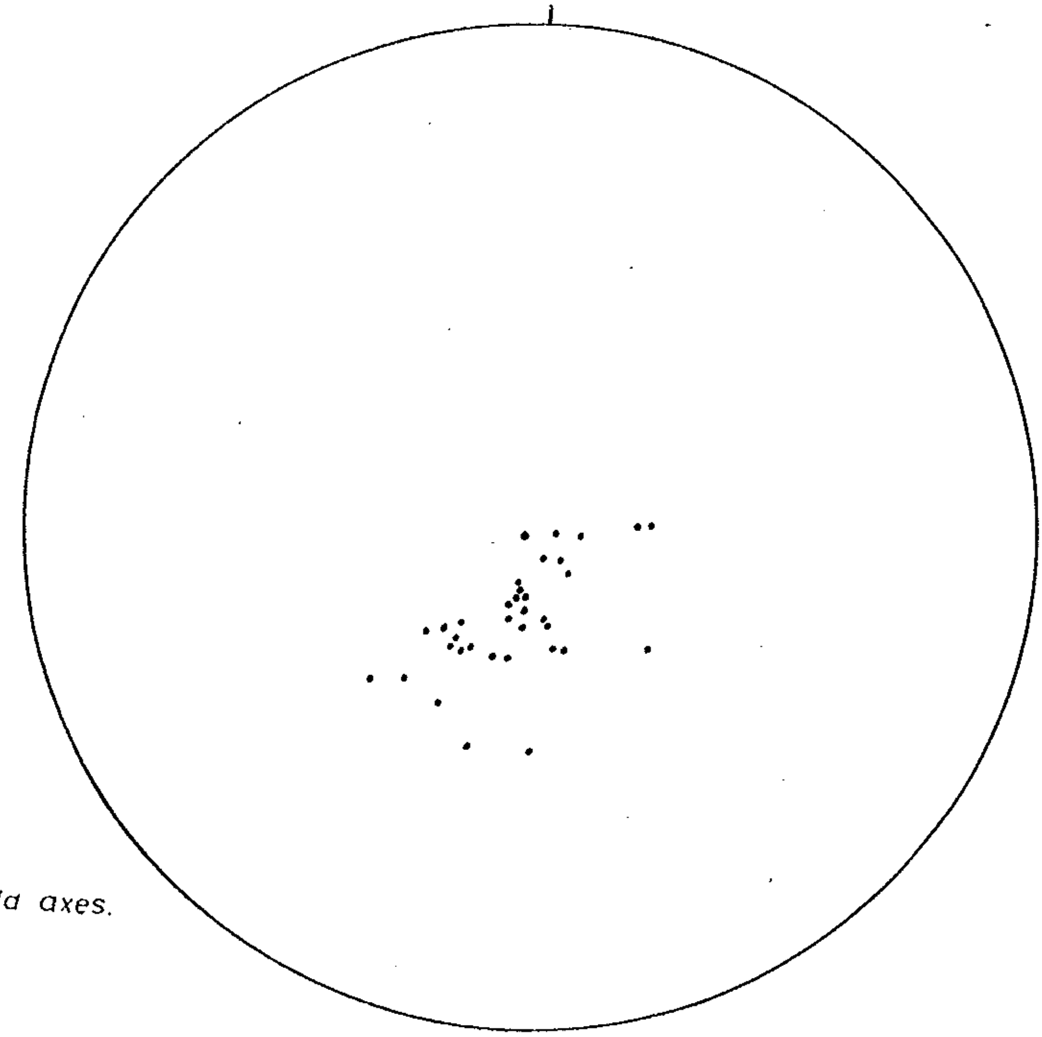


Fig.33. Axes of conjugate shear folds, 1.



35 fold axes.

Fig. 25. THIRD STR

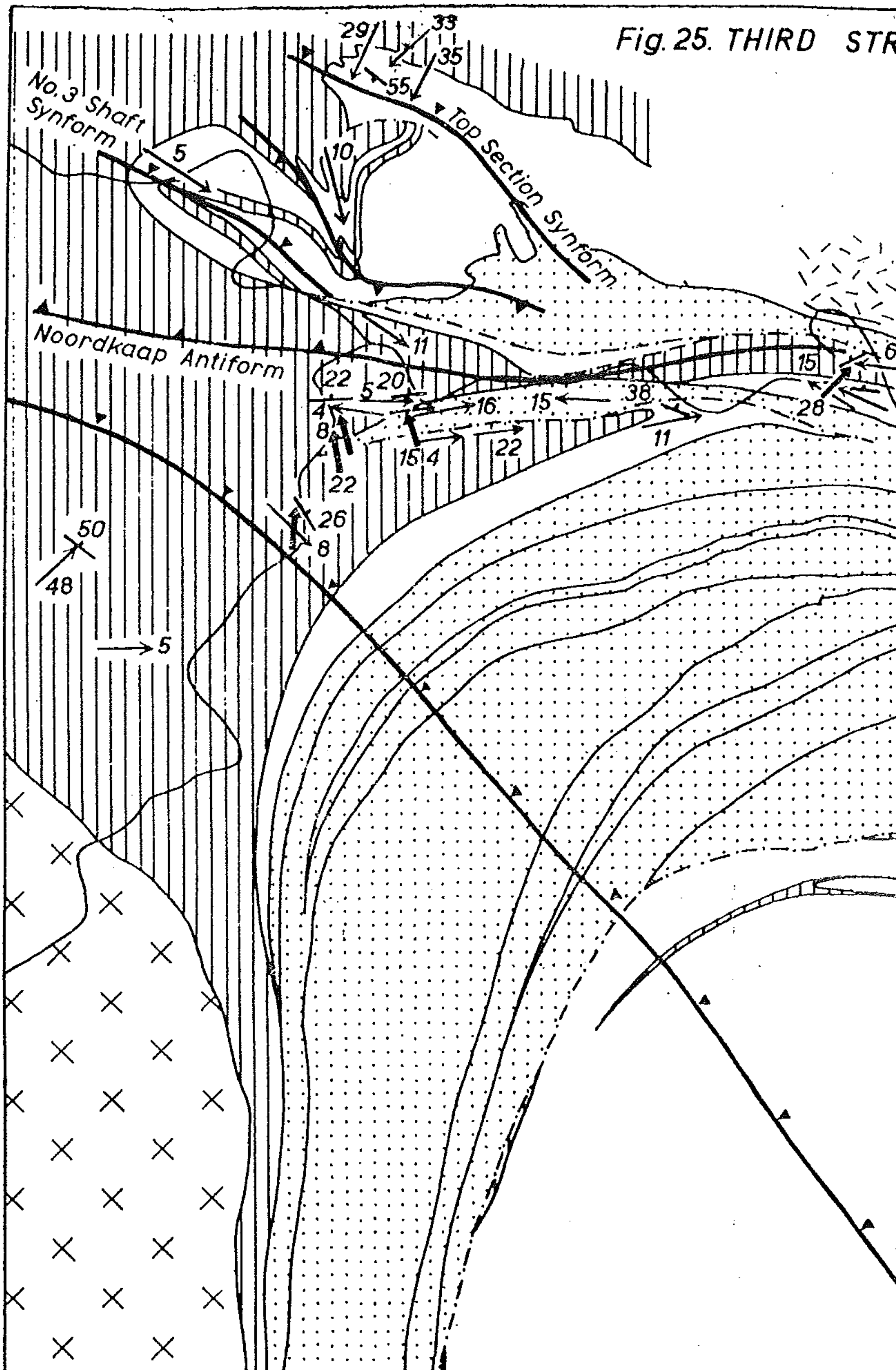
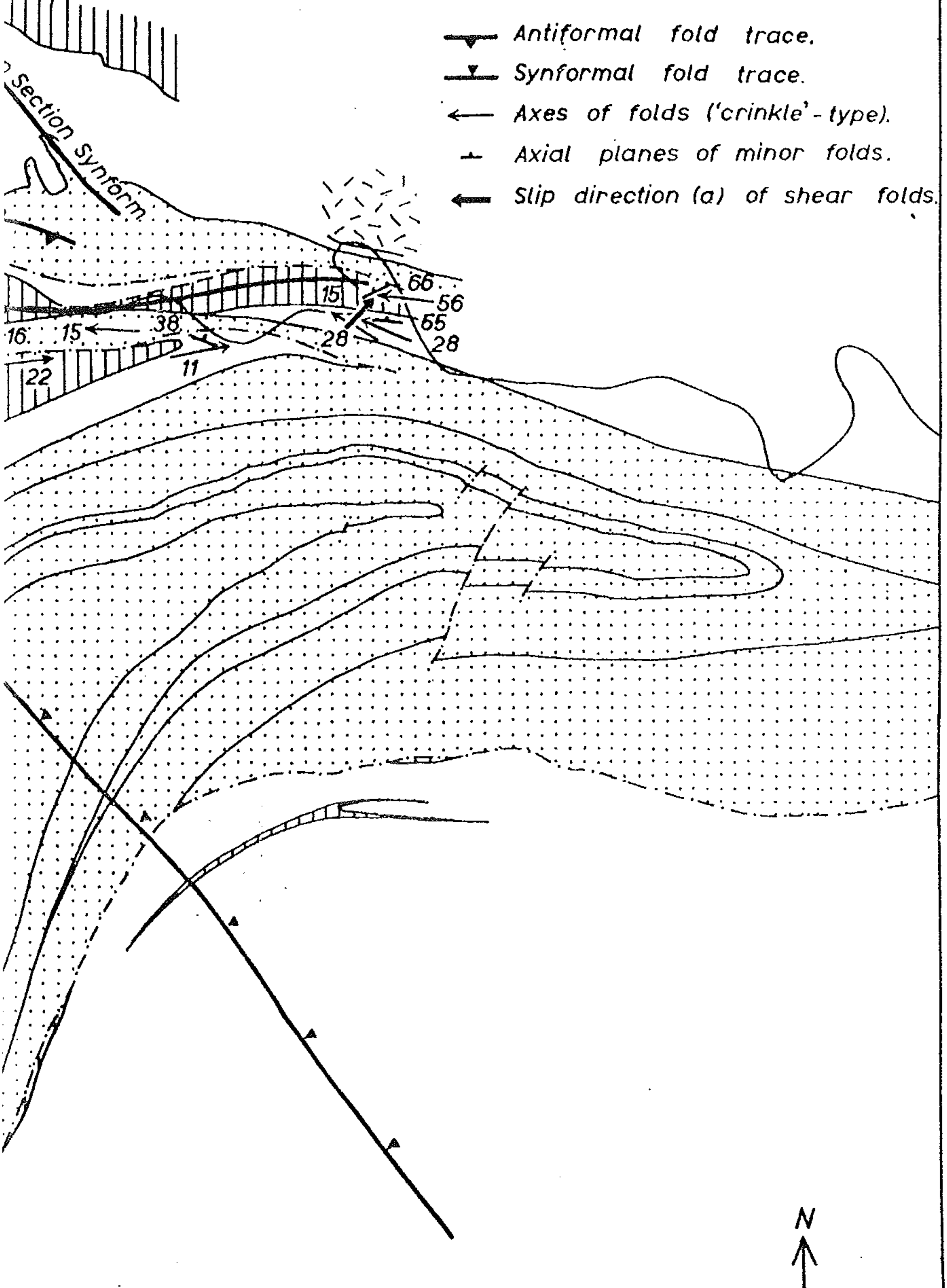
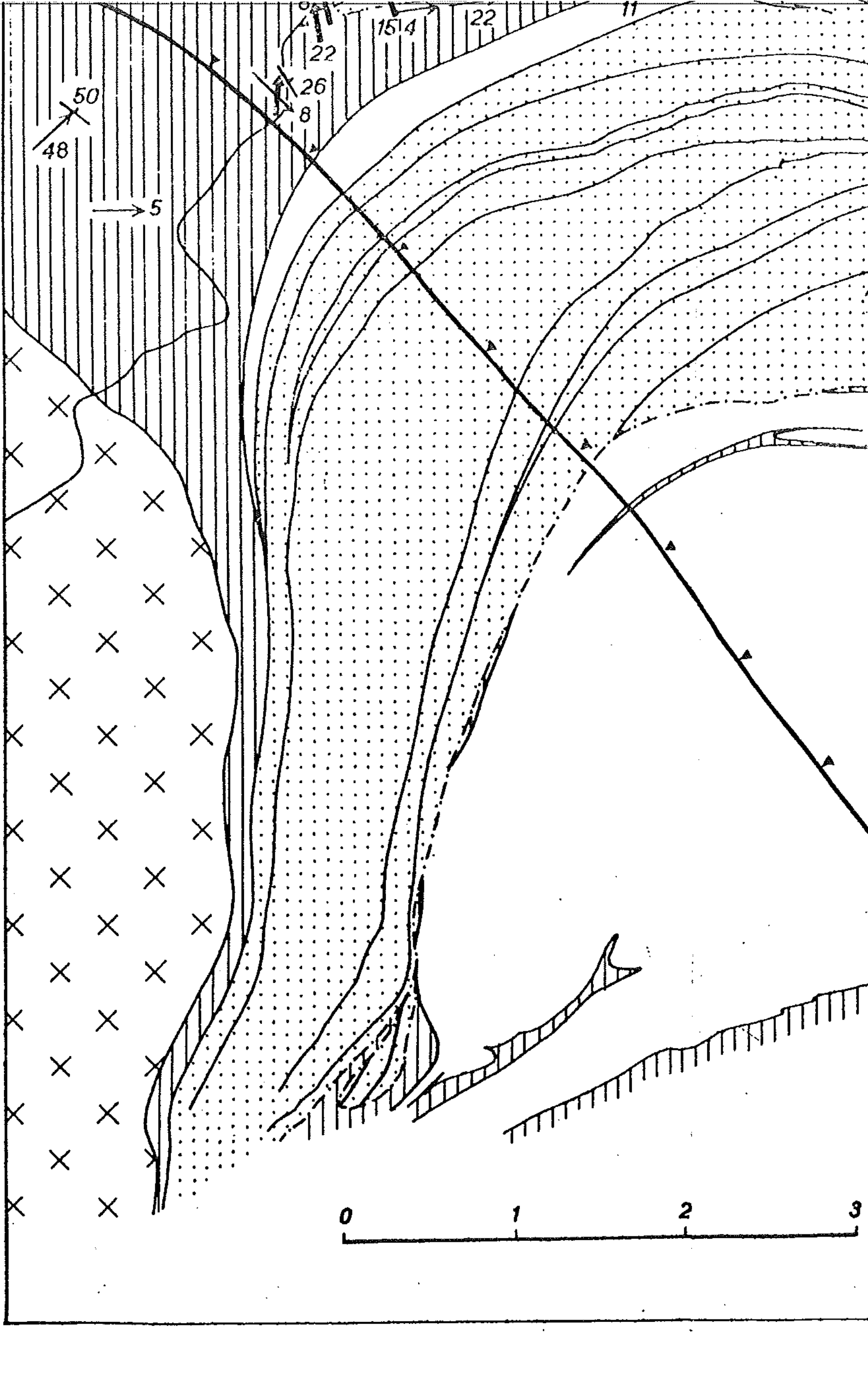


Fig. 25. THIRD STRUCTURES





22
15
4
22
11
50
48
26
8
22
5
0
1
2
3

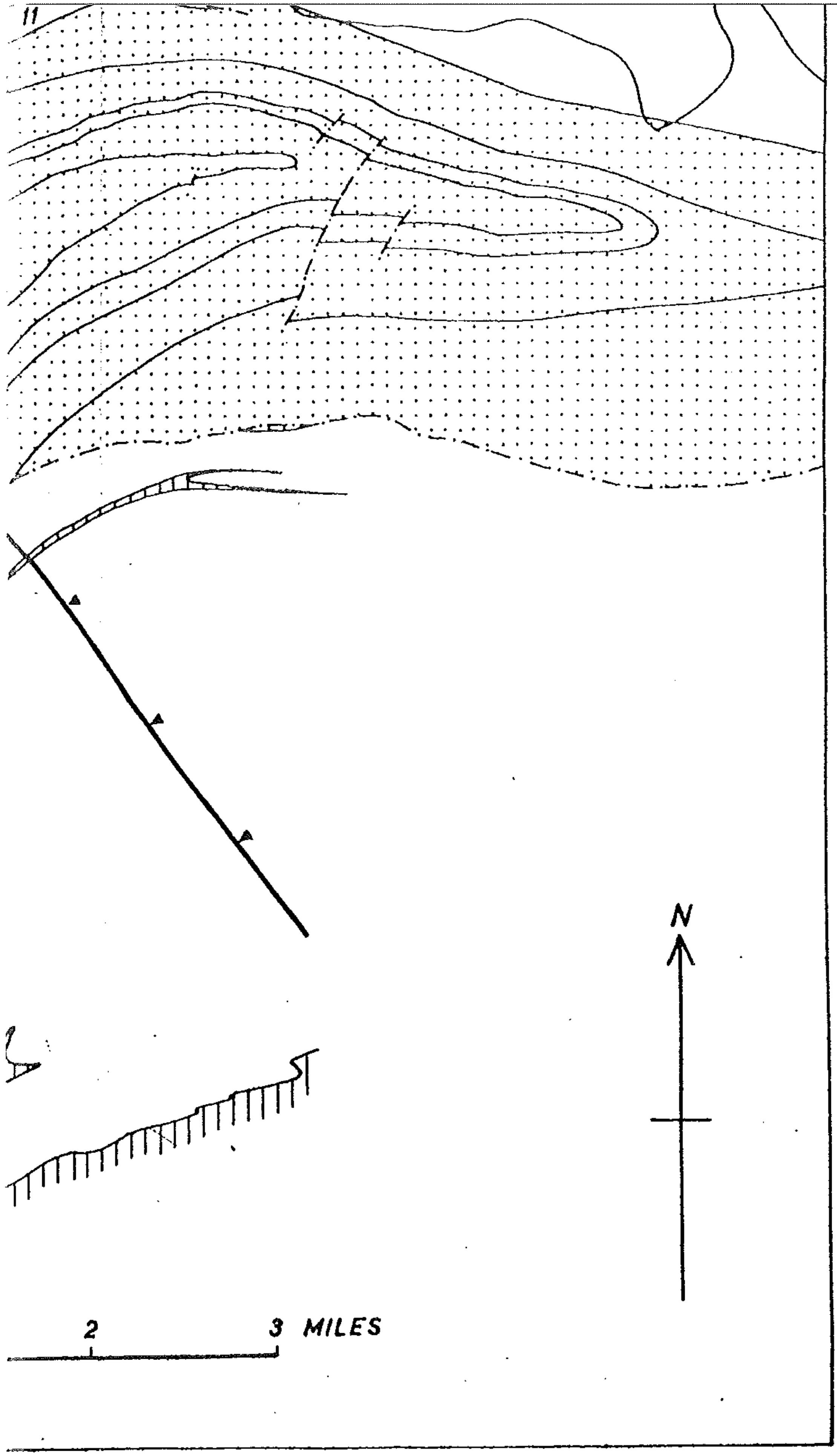

11
N
2
3 MILES

Fig. 22

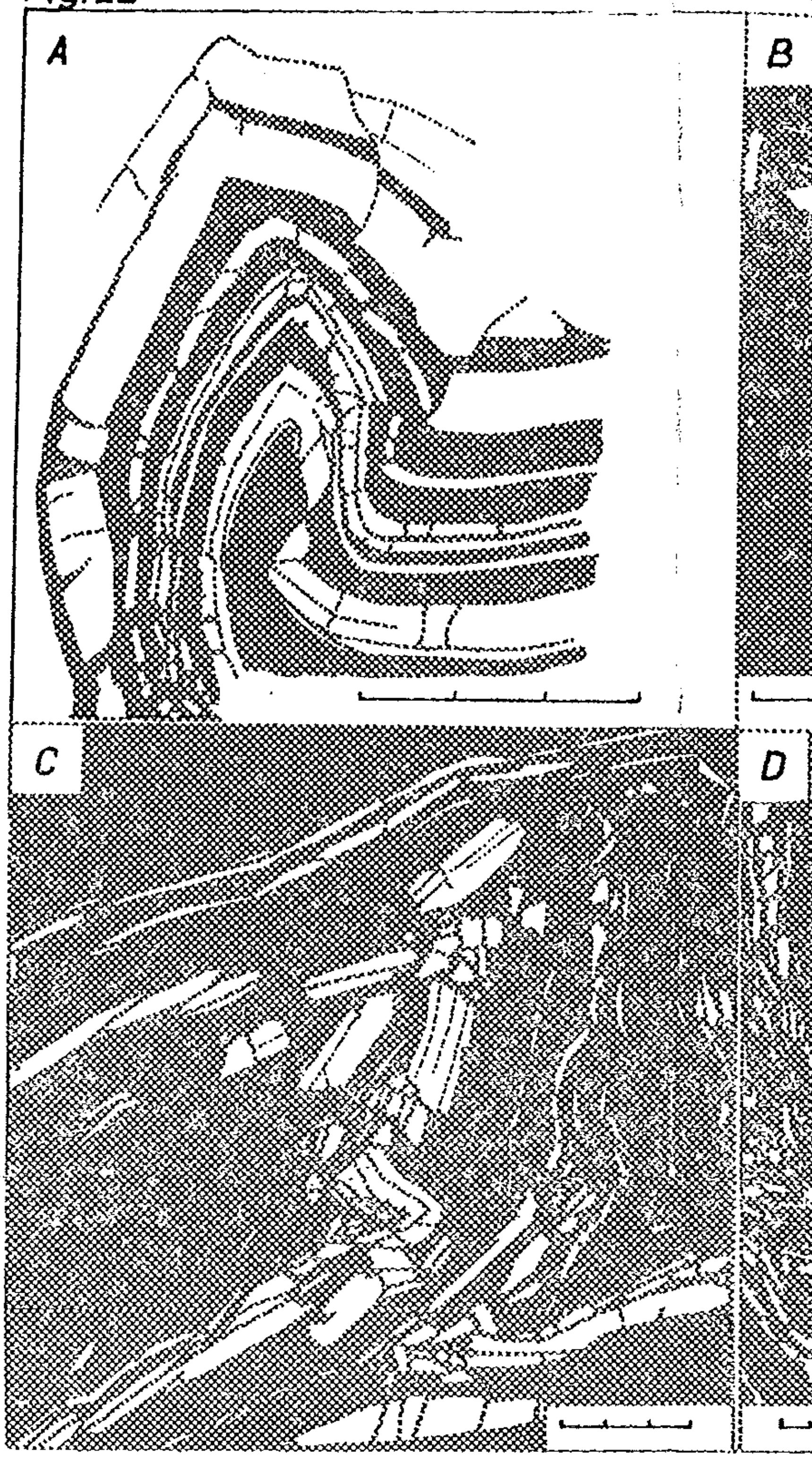


Fig. 23.

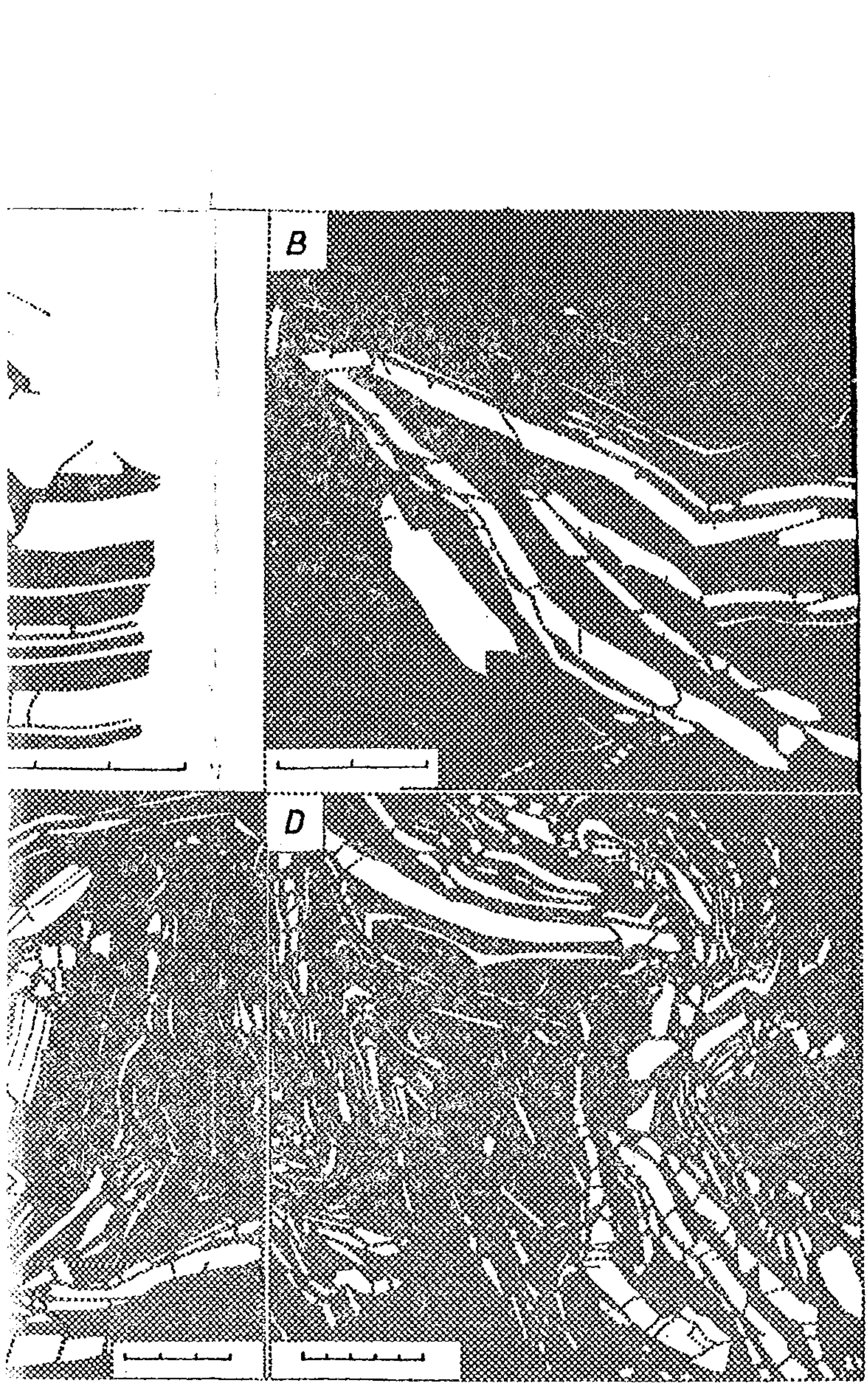
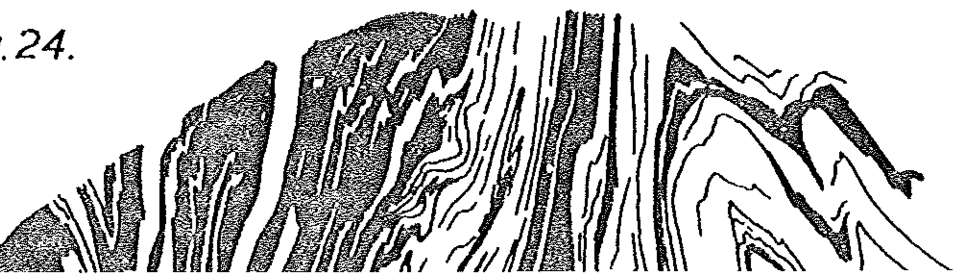


Fig.24.



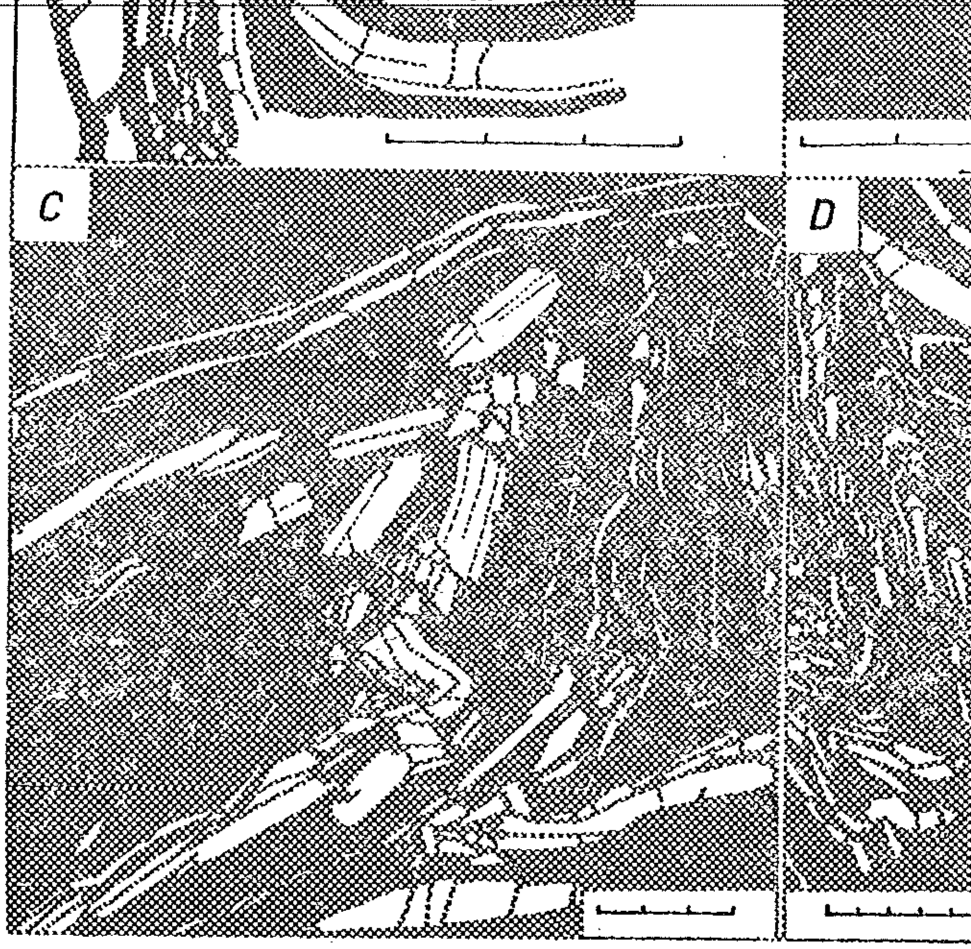


Fig. 23.

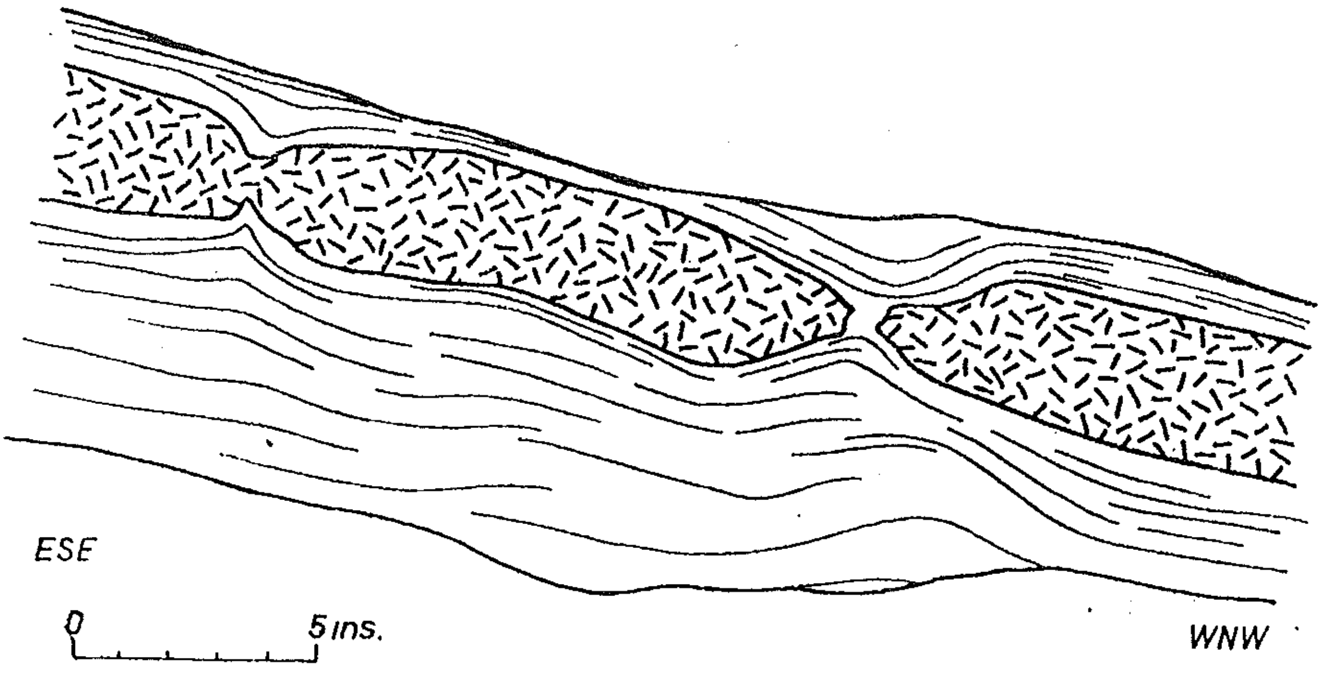
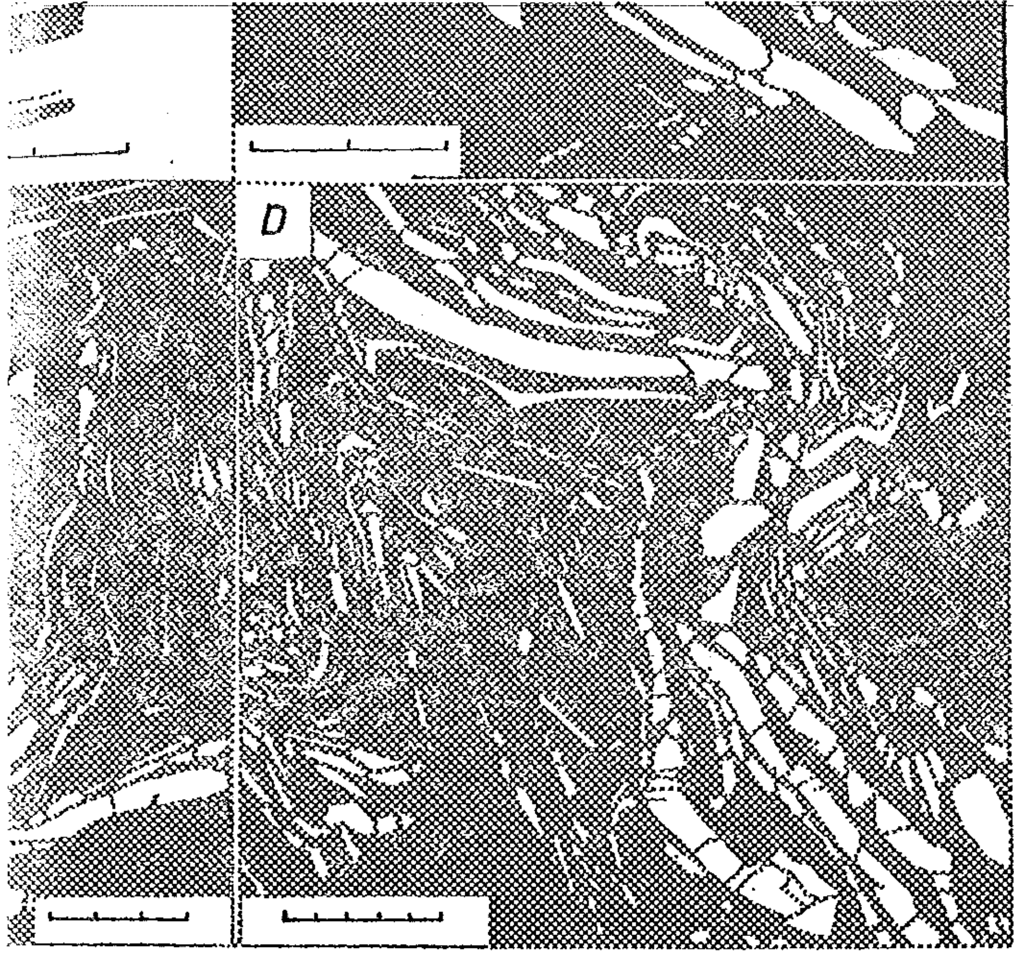


Fig. 2



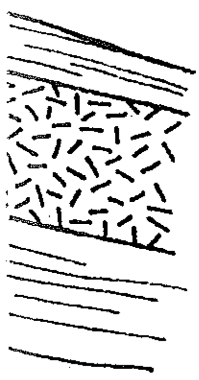


Fig. 24.

Fig. 21. Analysis of superimposed strains

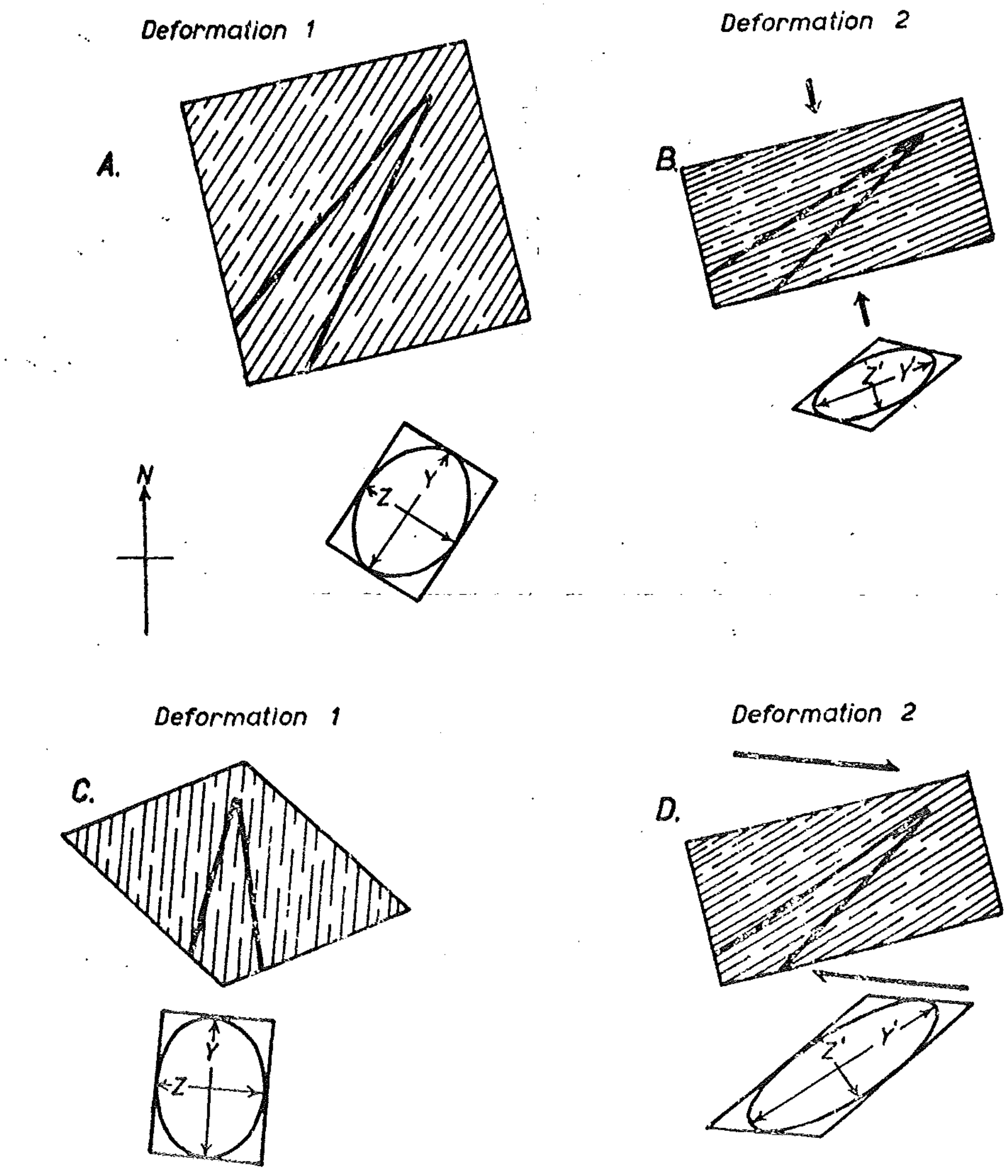


Fig.20 SECOND

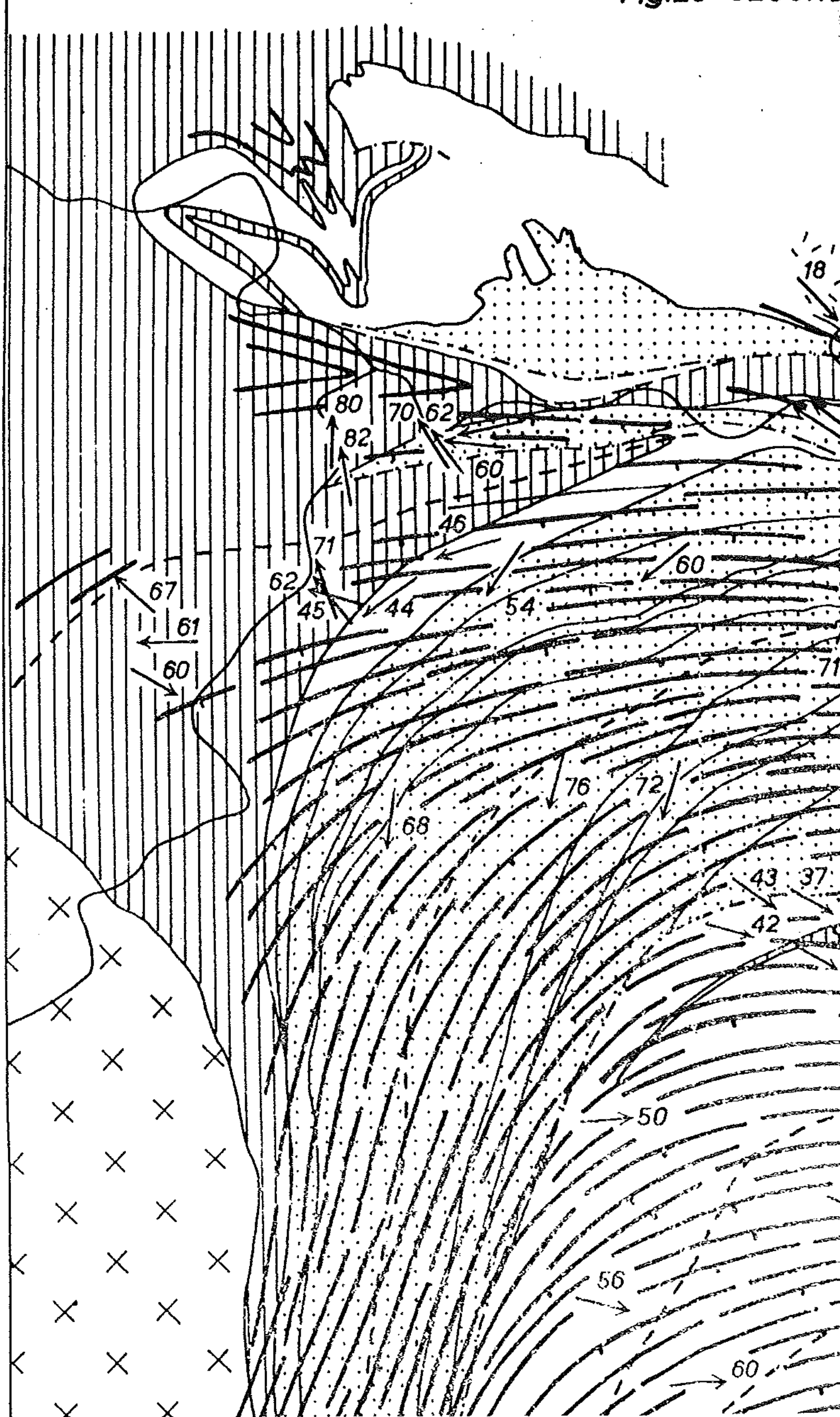
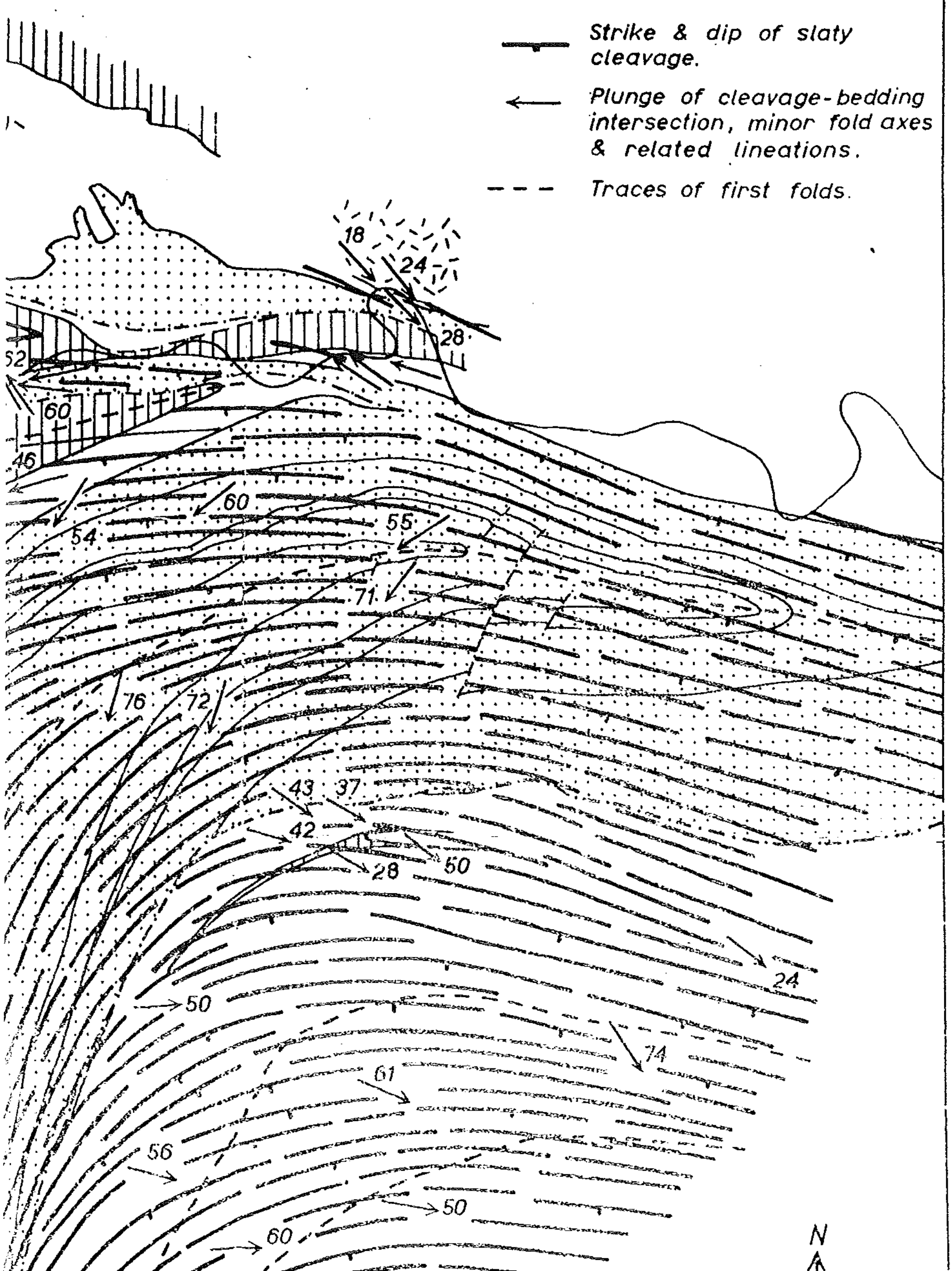
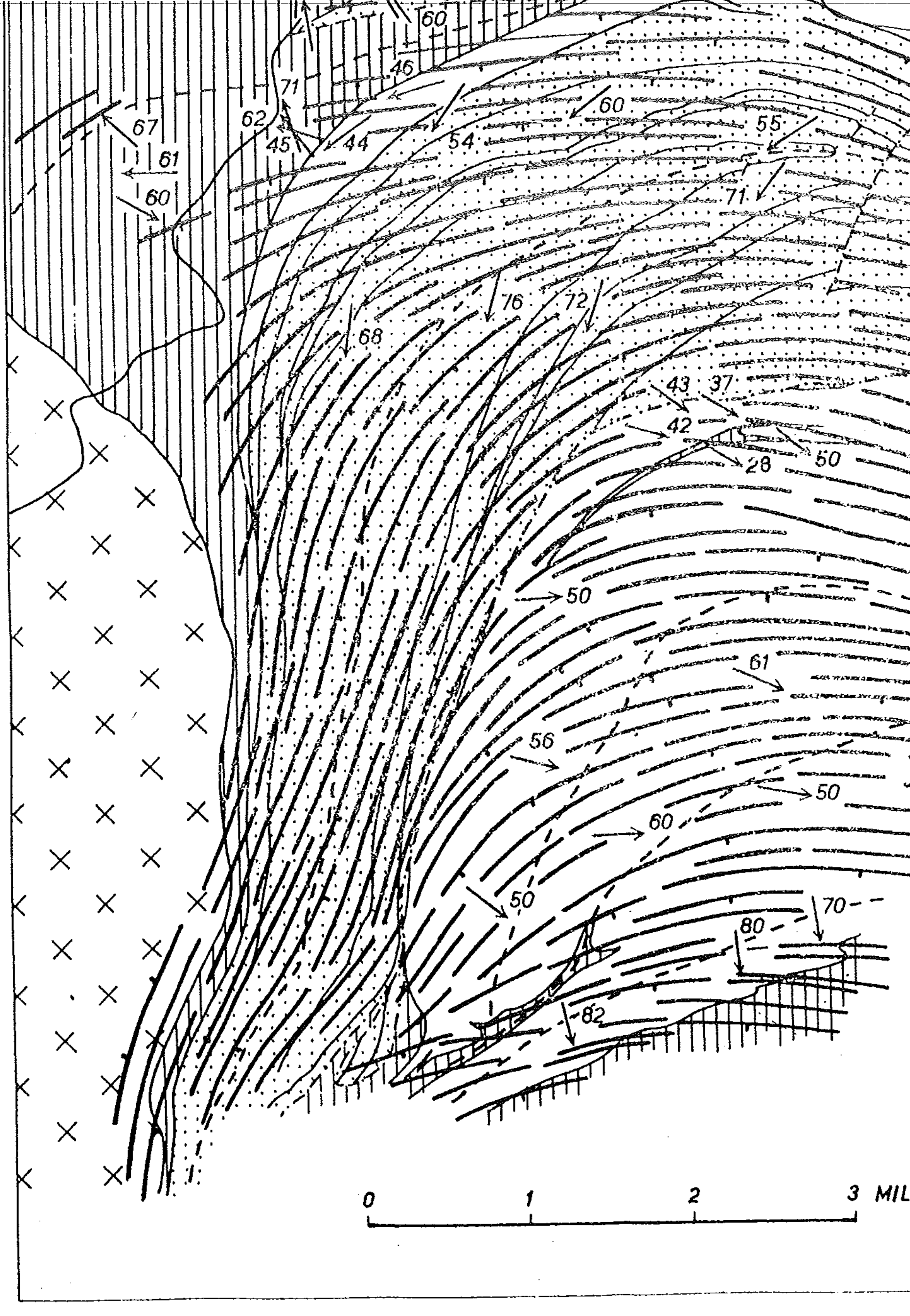
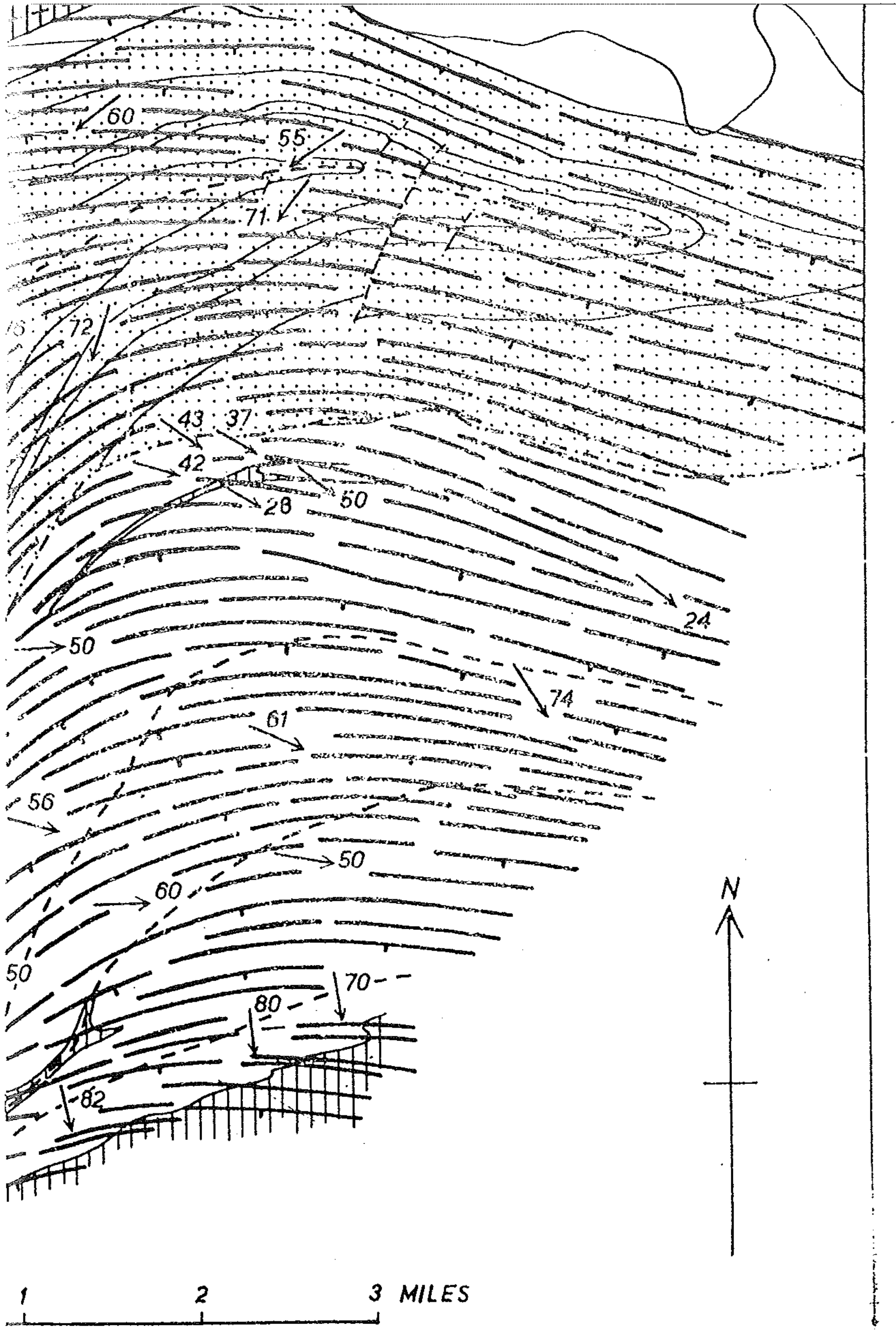


Fig.20 SECOND STRUCTURES





60
46
71
62
45
44
67
61
60
60
54
55
71
76
72
68
43
37
42
50
28
50
61
56
50
60
50
70
80
82
0
1
2
3 MIL



60
55
71
72
43
37
42
50
28
24
50
74
61
56
50
60
50
70
80
82
N
1
2
3 MILES

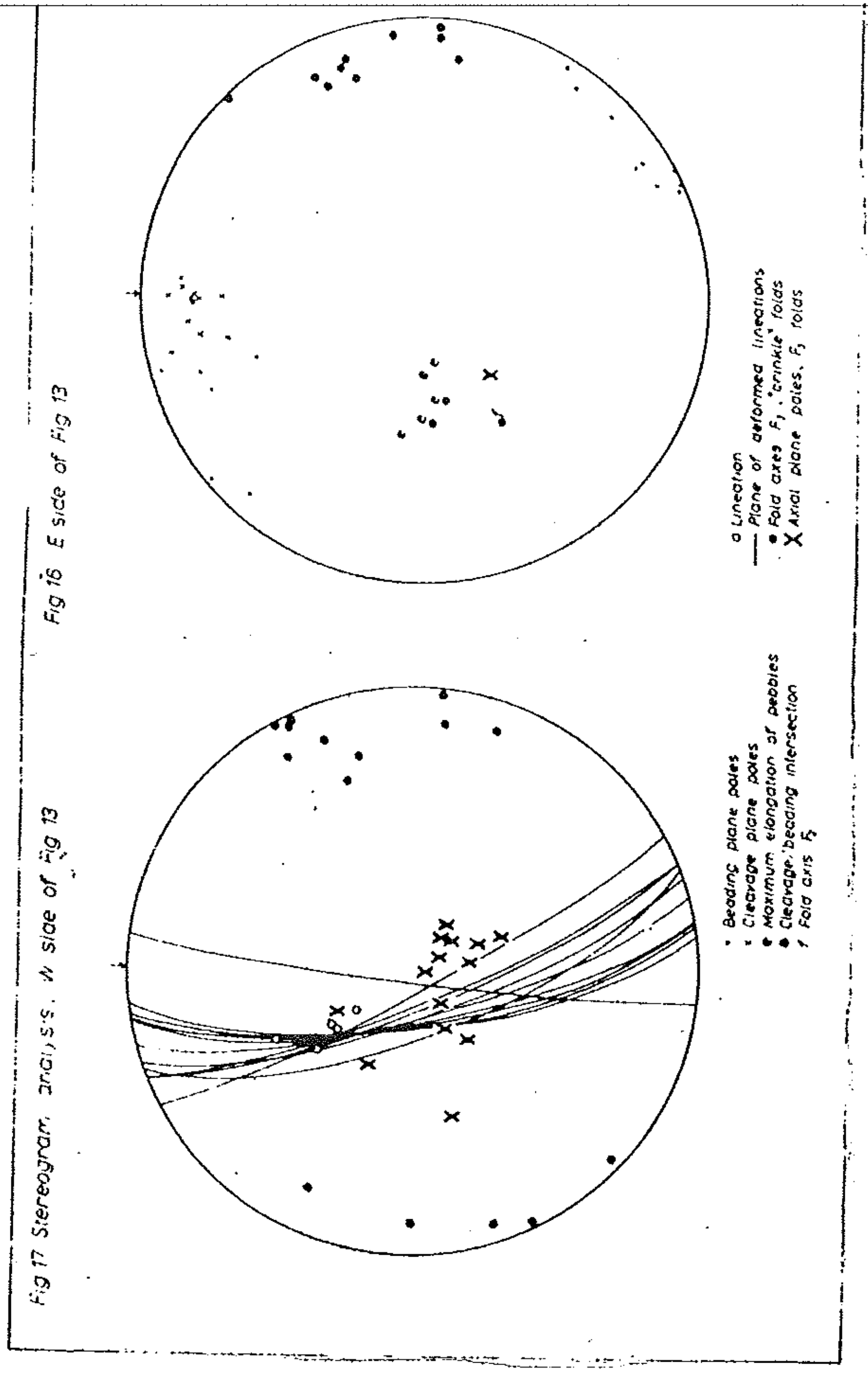


Fig 16 E side of Fig 13

Fig 17 Stereogram analysis. W side of Fig 13

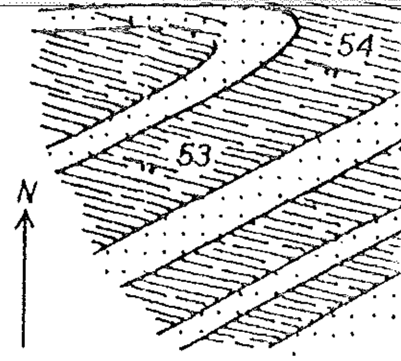
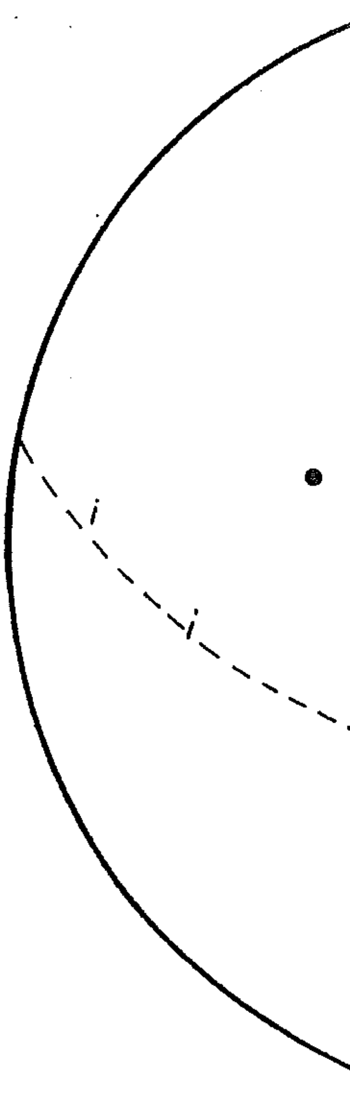


Fig.19.

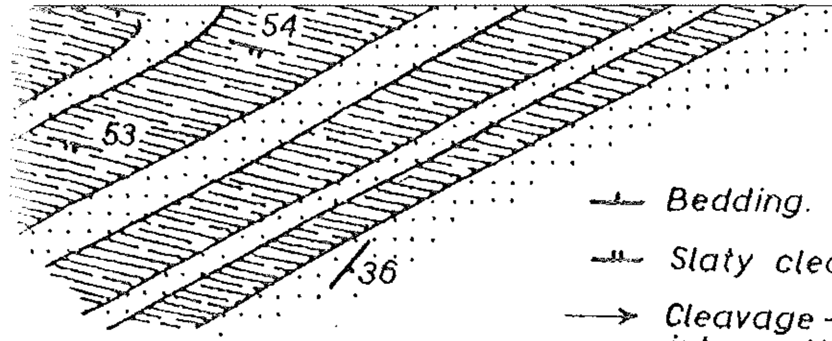


Poles to bed

F_1 Axis of fol

x Poles to clea

i Cleavage–be

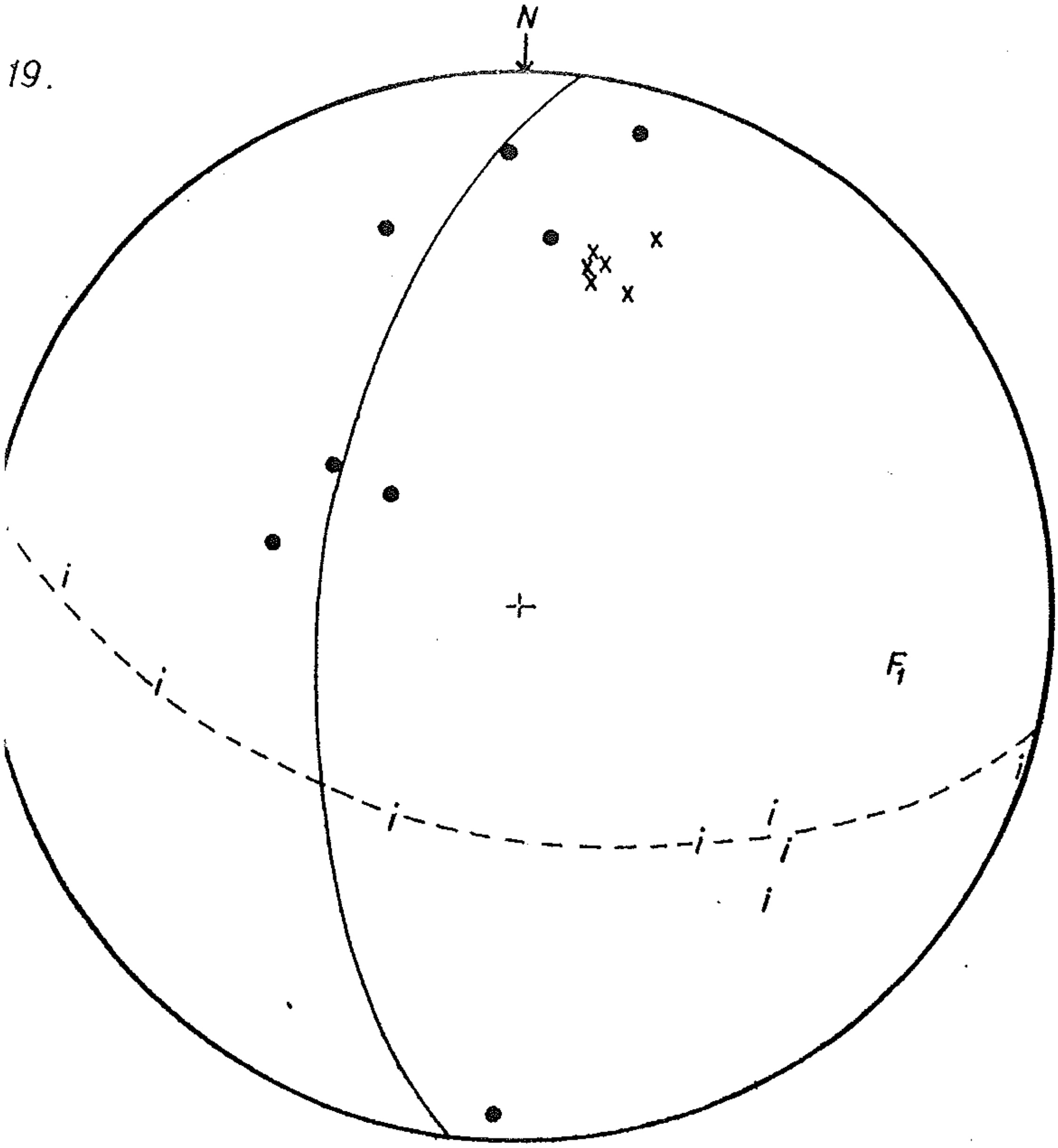


Bedding.

Slaty cleavage.

Cleavage - bedding intersection.

19.



Poles to bedding with great circle.

F_1 Axis of folding.

x Poles to cleavage.

i Cleavage–bedding intersection.

Fig. 18.

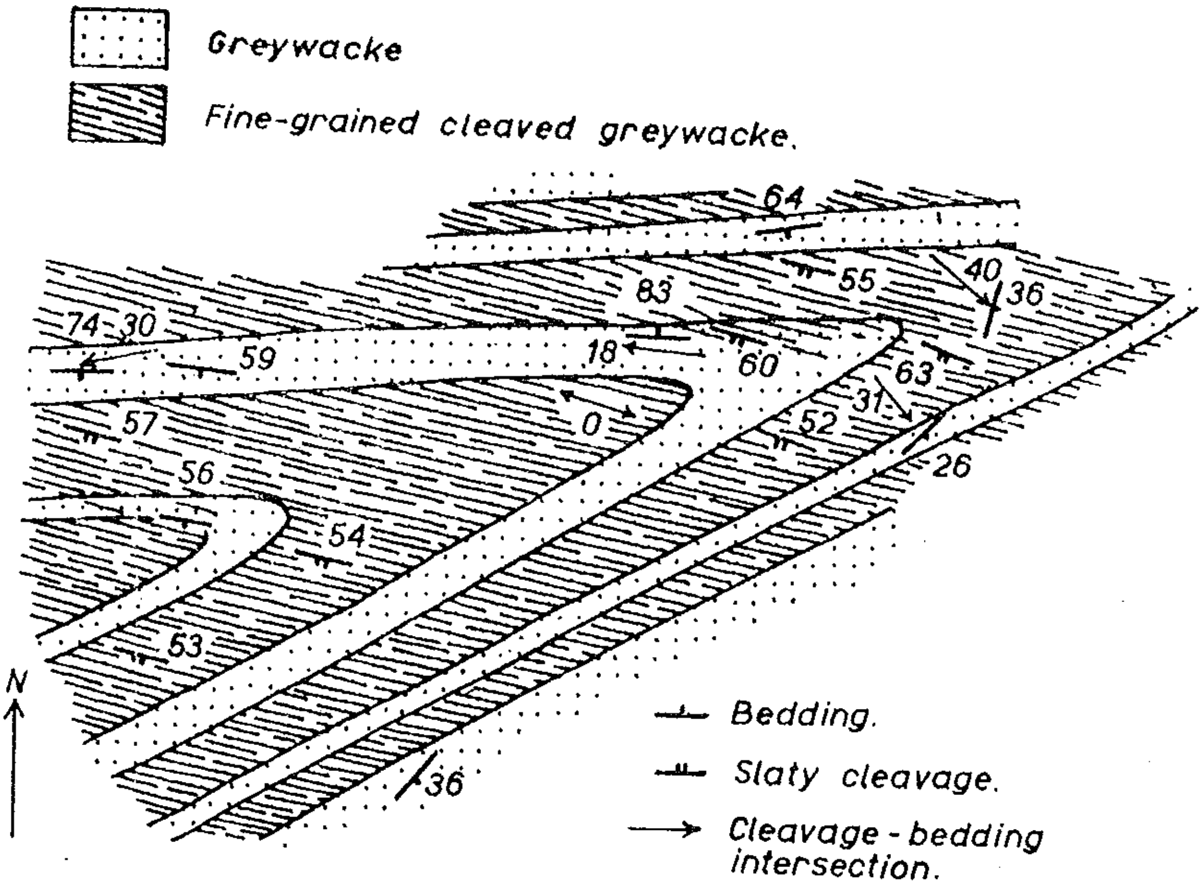
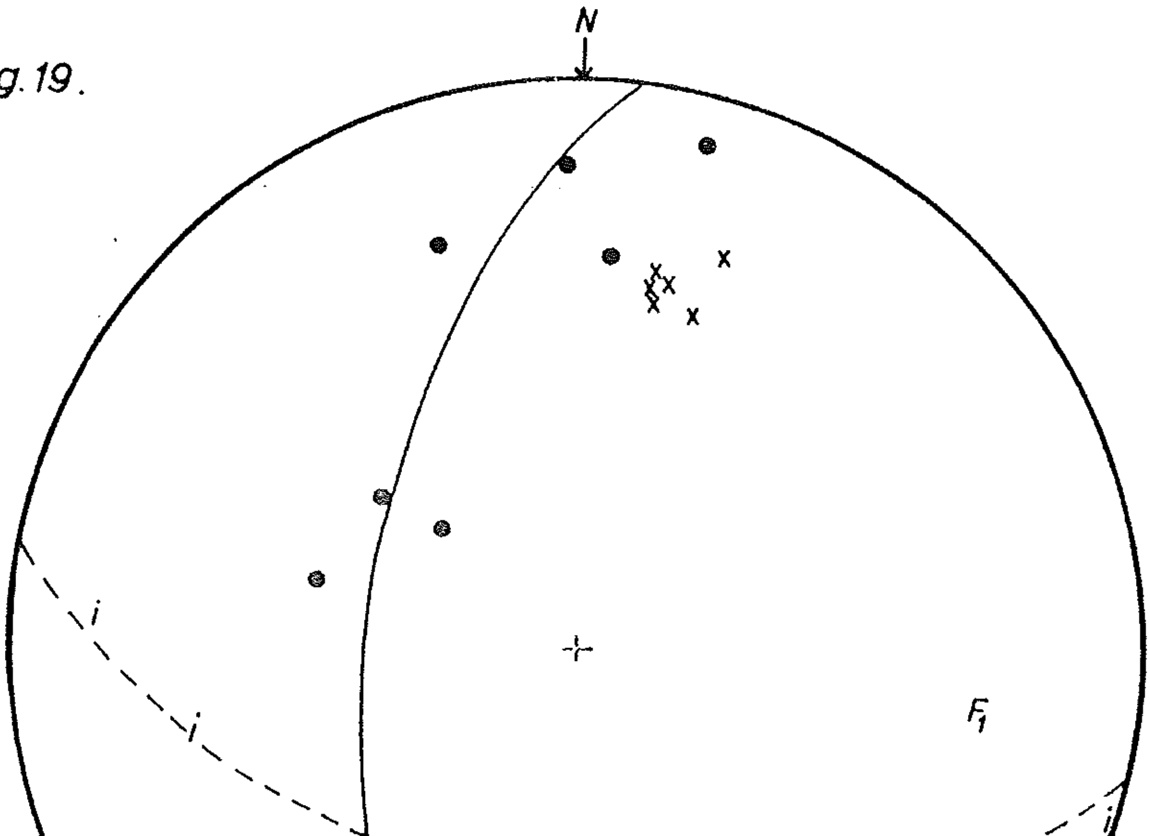

Greywacke
Fine-grained cleaved greywacke.
64
55
40
36
83
74
30
59
18
60
63
57
0
31
52
26
56
54
53
N
Bedding.
Slaty cleavage.
Cleavage-bedding intersection.
36

Fig. 19.


N
i
i
F_1

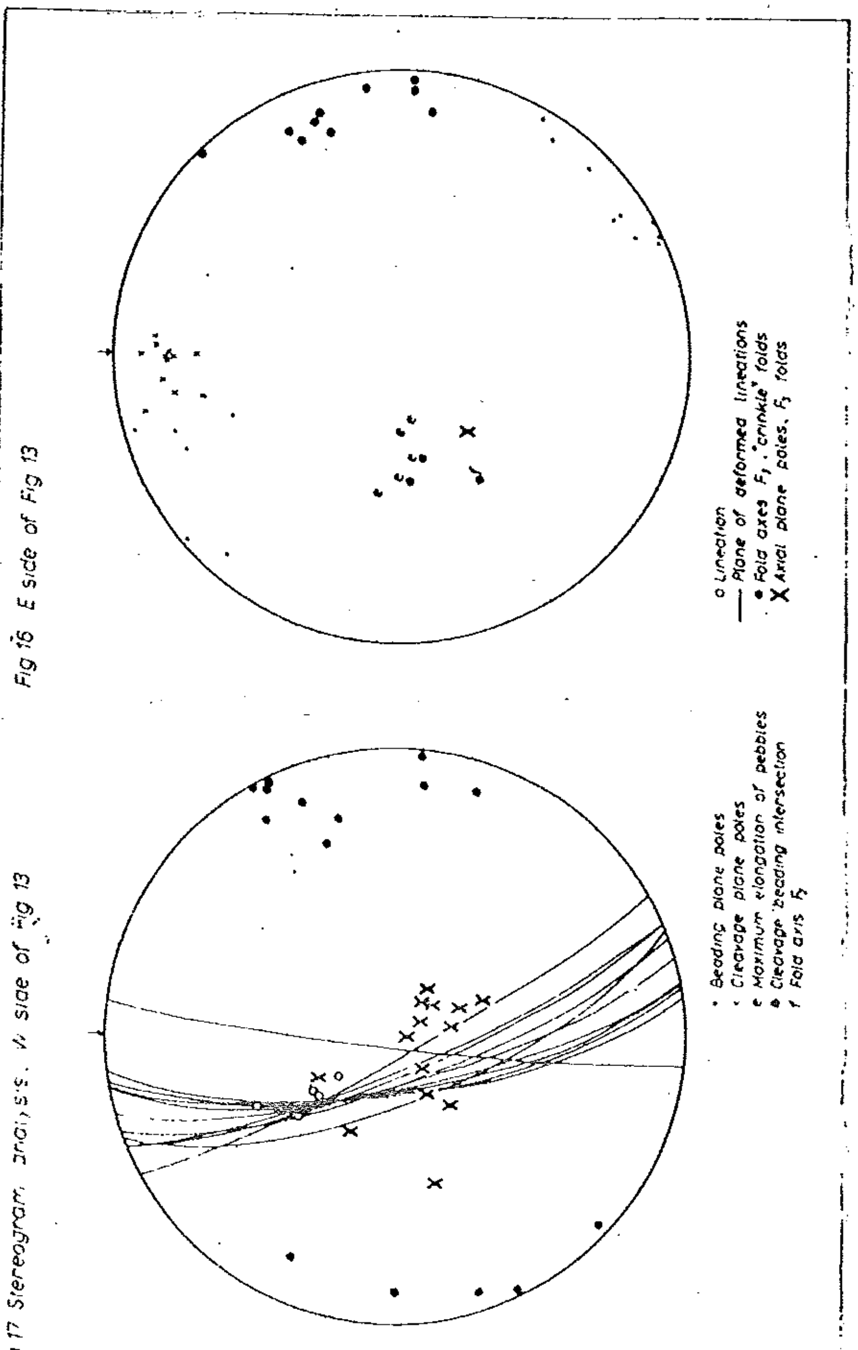


Fig 16 E side of Fig 13

Fig 17 Stereogram analysis. W side of Fig 13

Fig. 18.

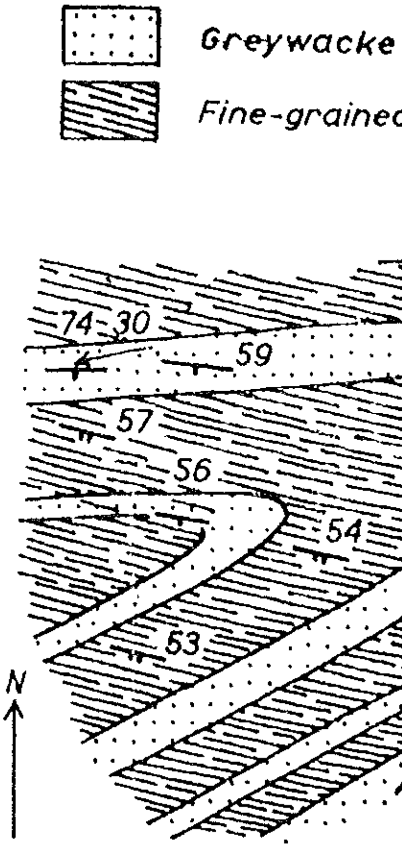
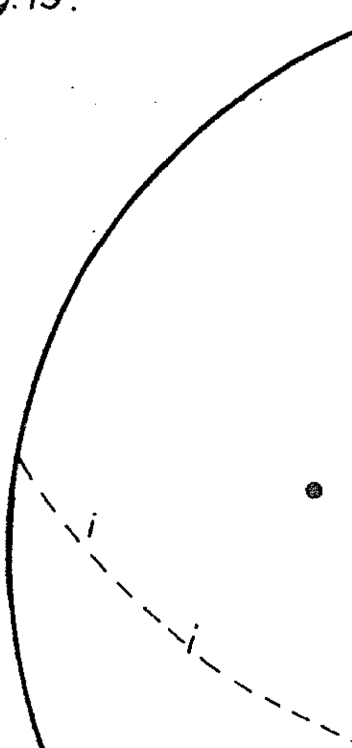
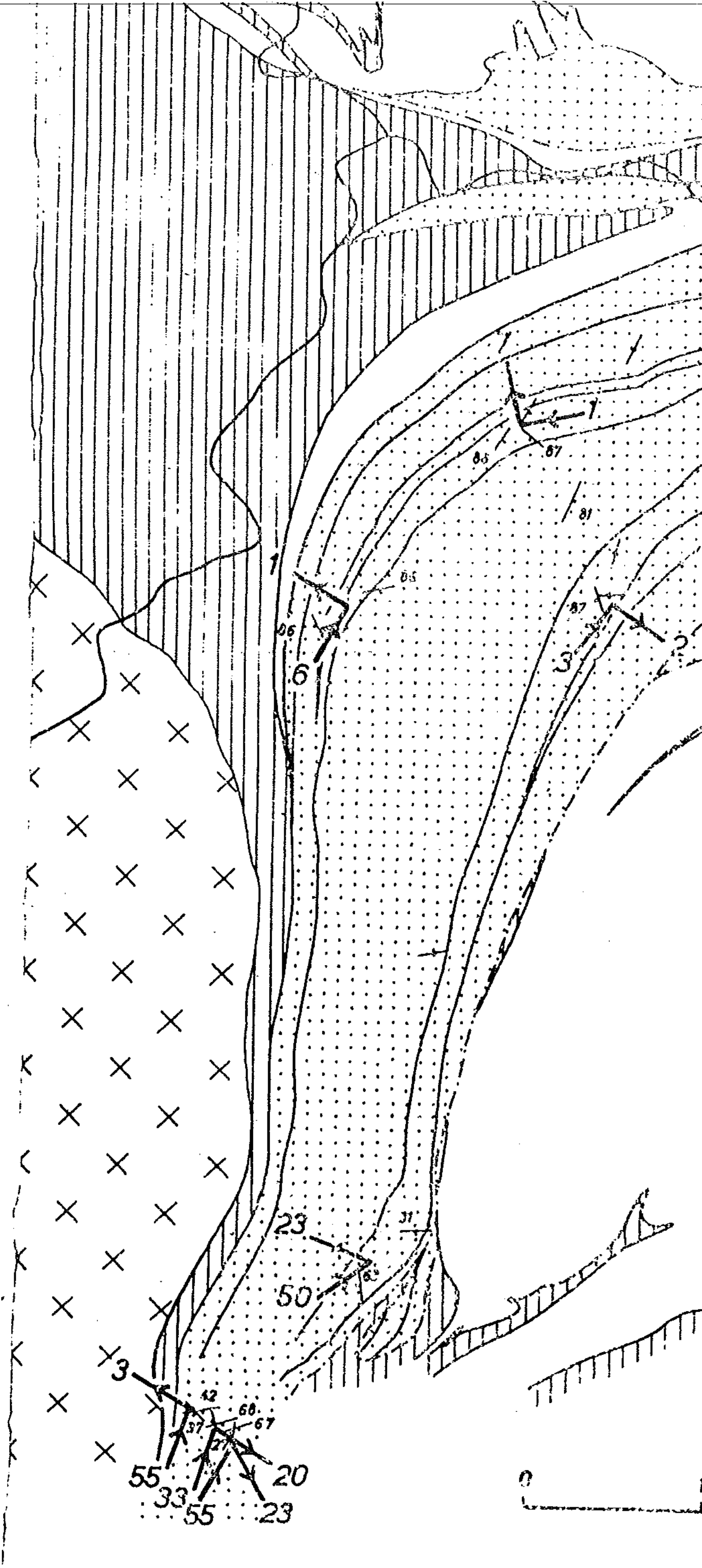


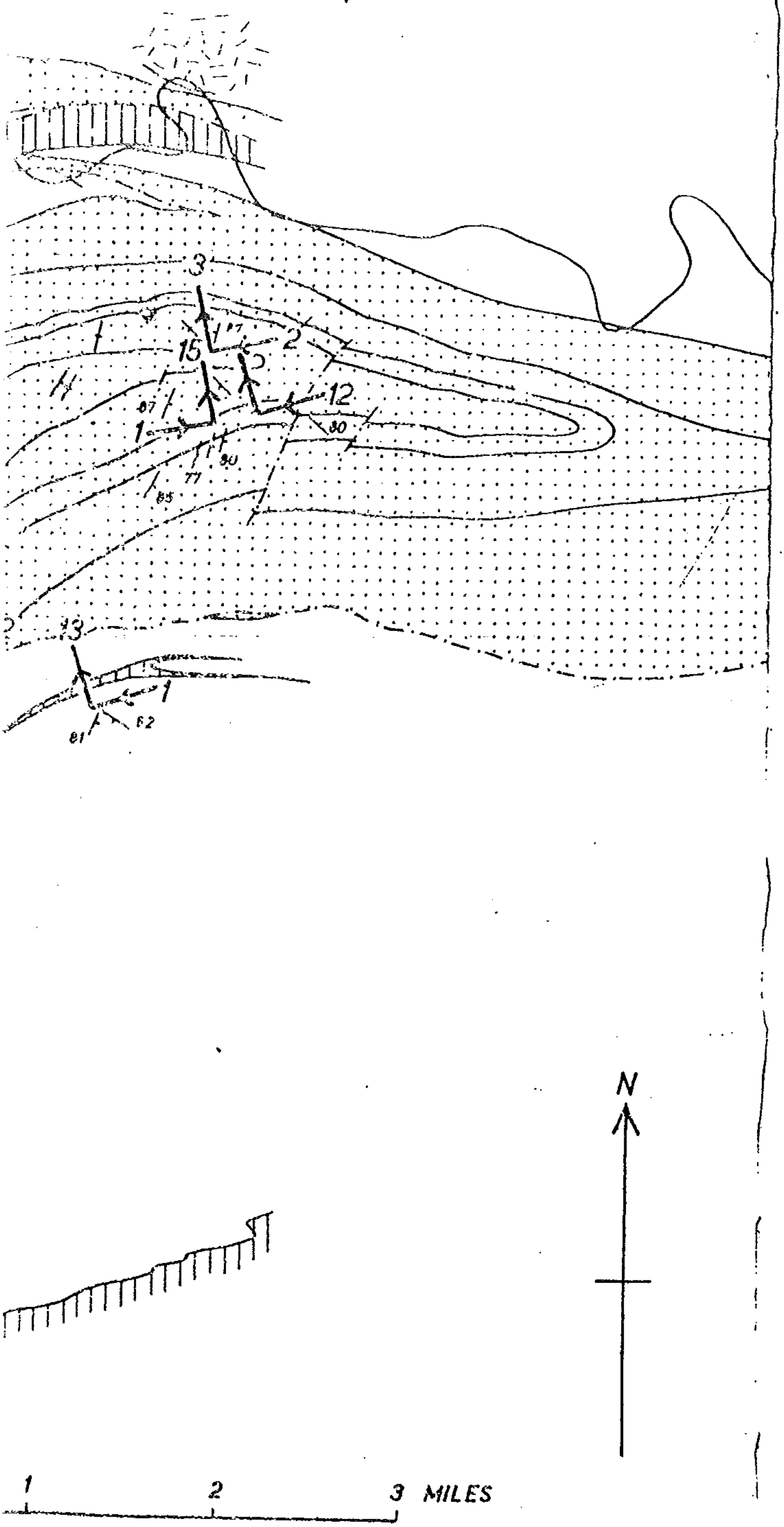
Fig. 19.





1
6
3
23
50
31
3
42
68
67
37
27
55
33
55
20
23
0

& minimum strain.



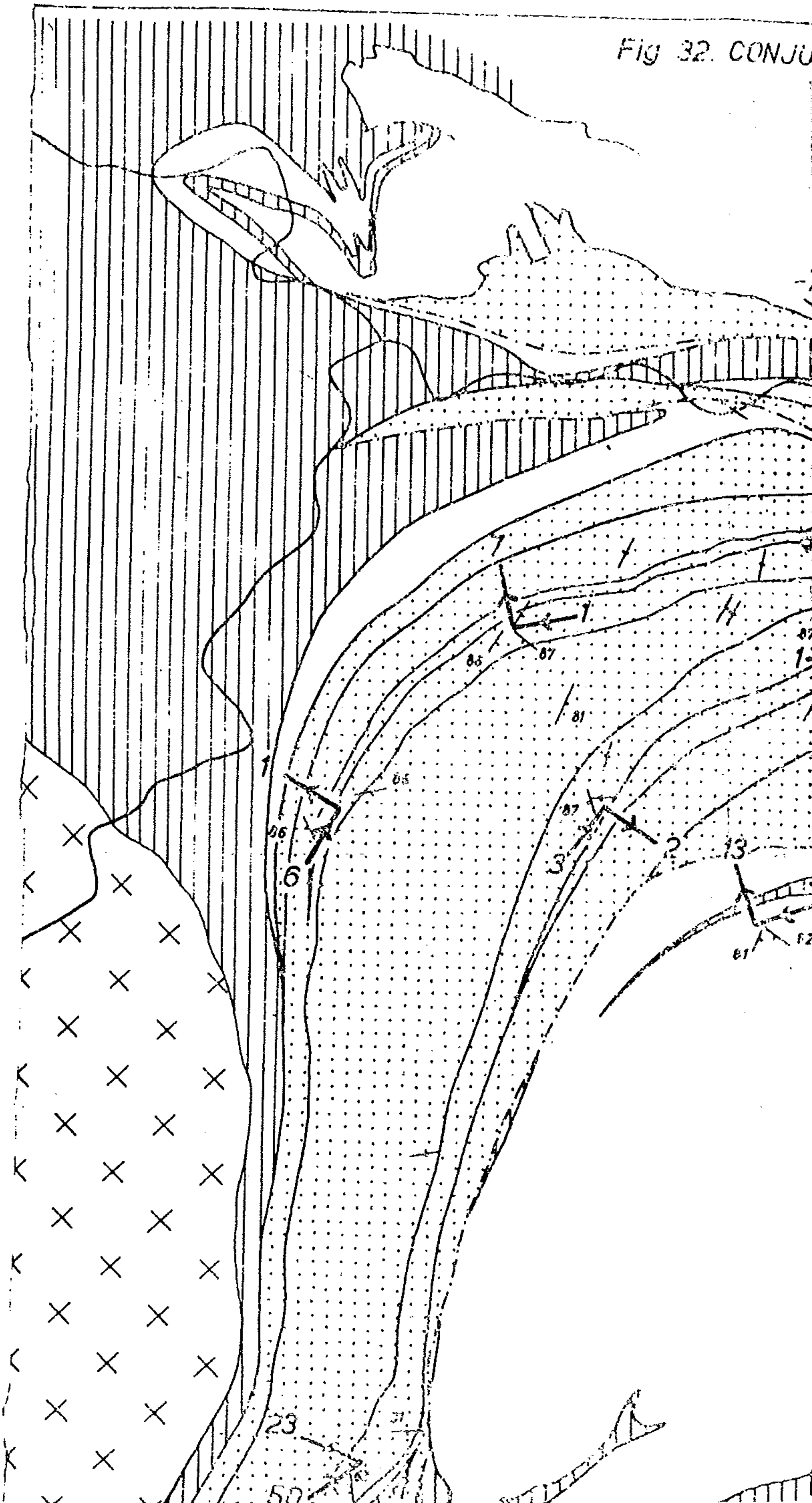


Fig 32. CONJU
1
2
3
6
13
23
81
85
86
87

Fig. 32. CONJUGATE SHEAR FOLDS

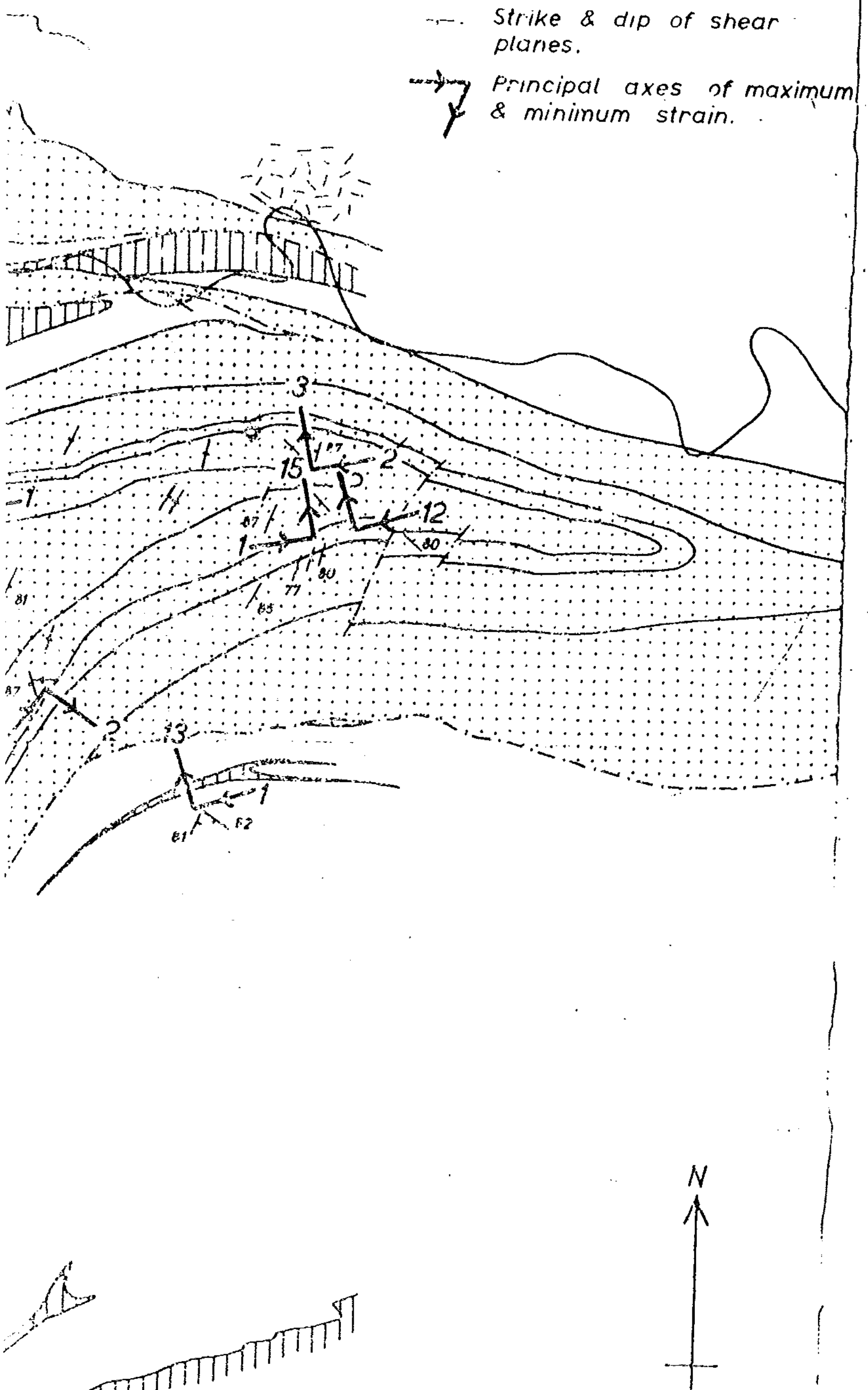


Fig. 14.

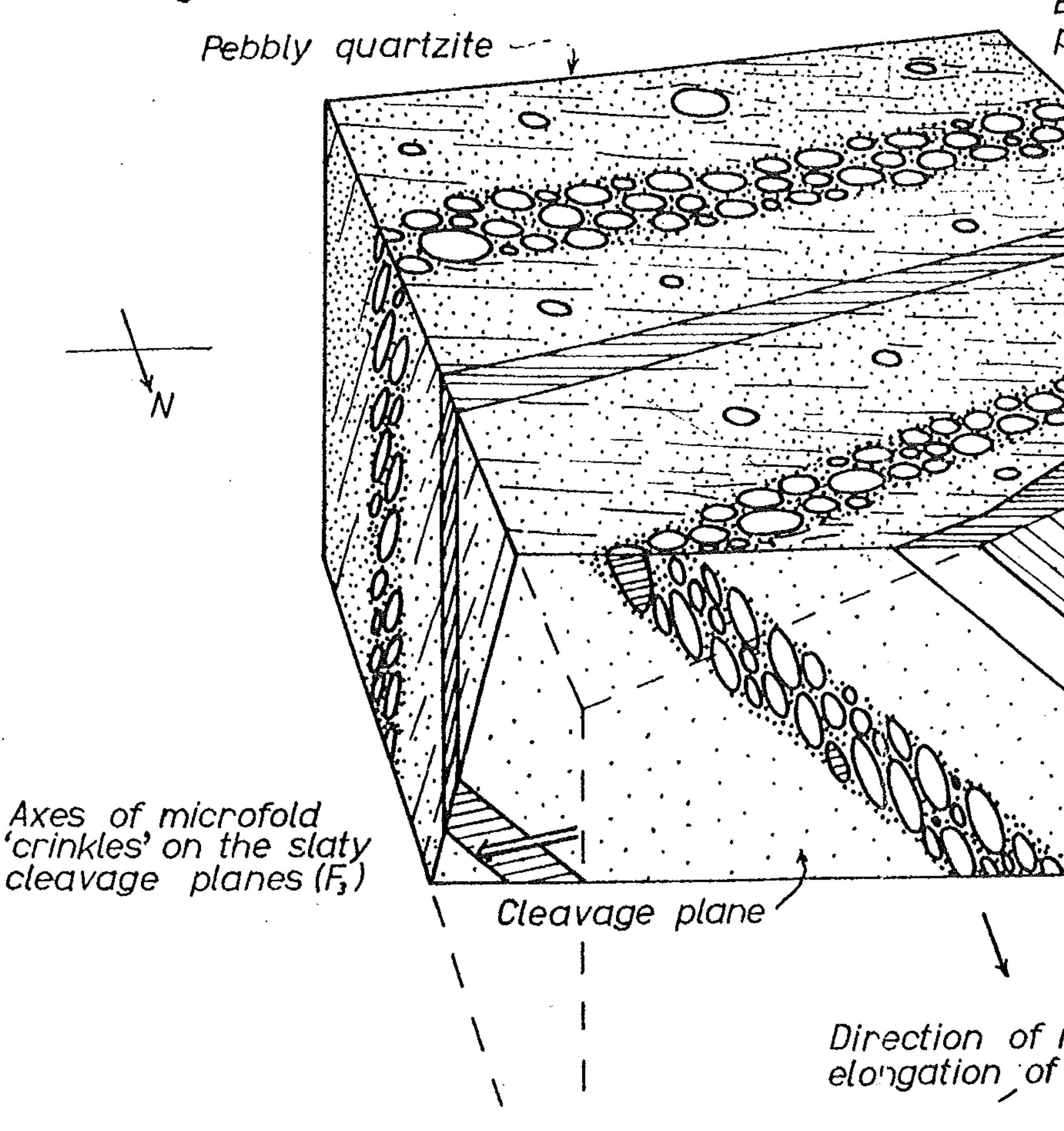
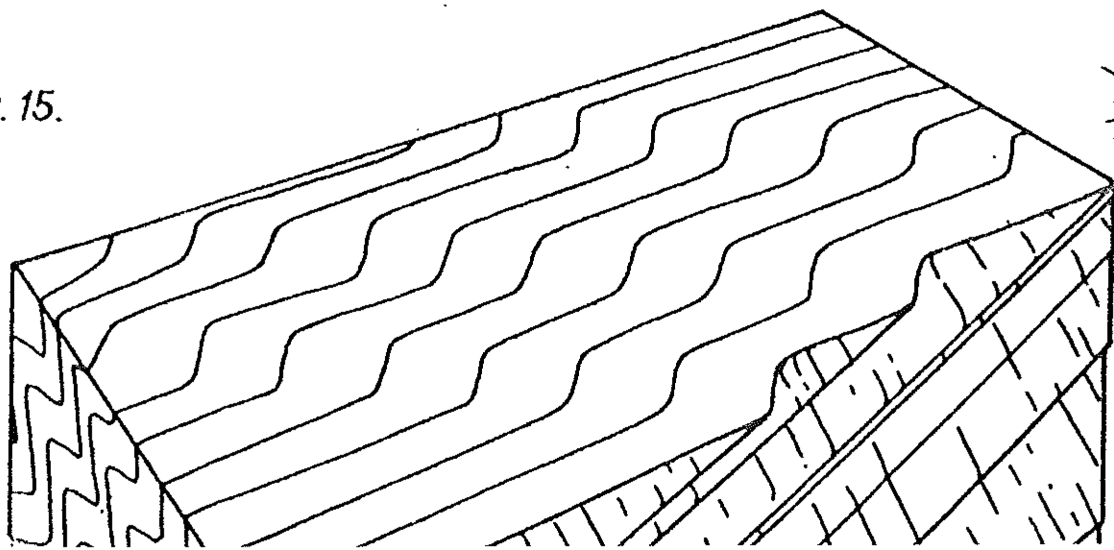
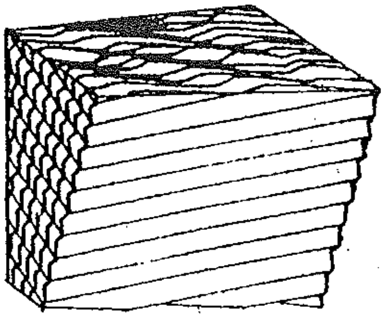
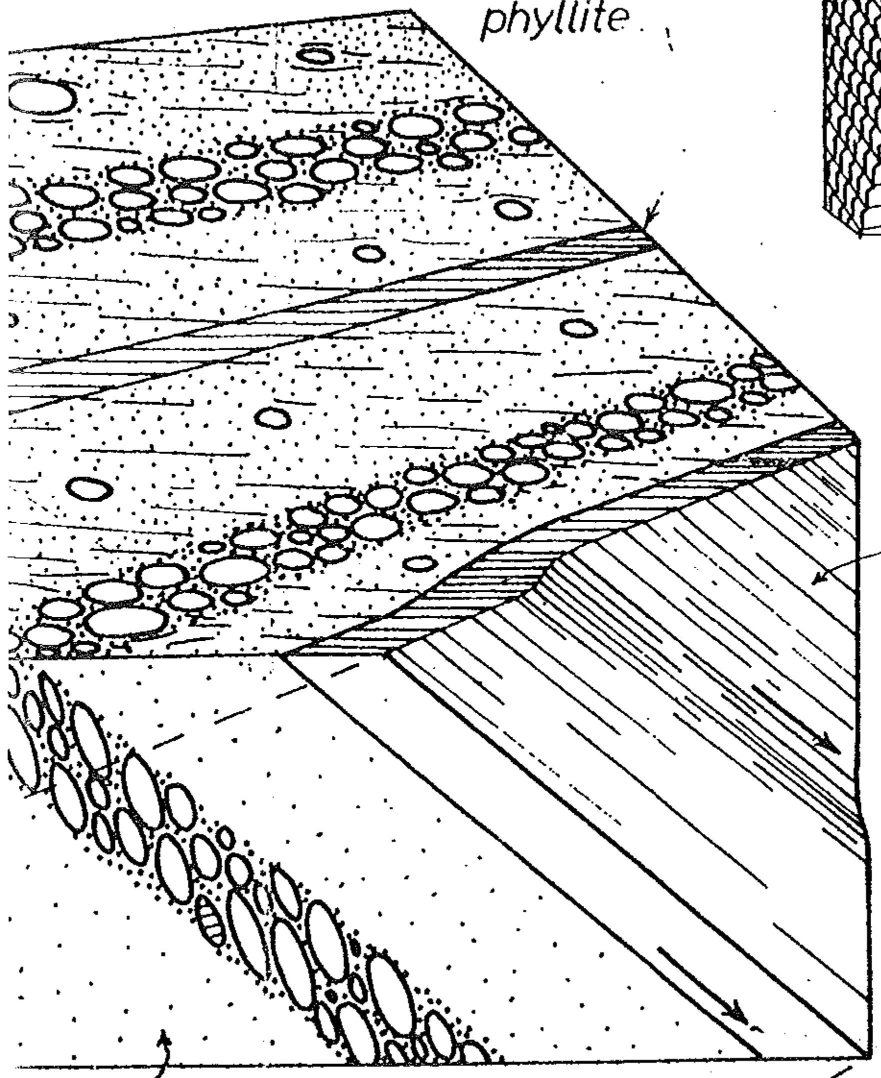

Pebbly quartzite
N
Axes of microfold 'crinkles' on the slaty cleavage planes (F_3)
Cleavage plane
Direction of
elongation of

Fig. 15.

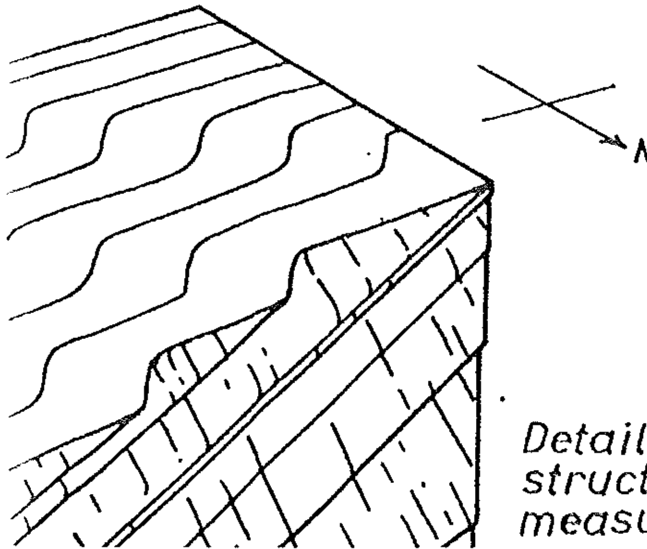




14 A



Band of talc-phyllite
Bedding plane
Axis of minor fold (F_2)
plane
Cleavage/bedding intersection
Direction of maximum elongation of pebbles



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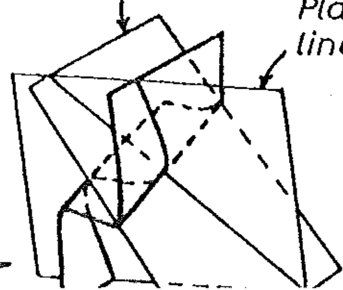
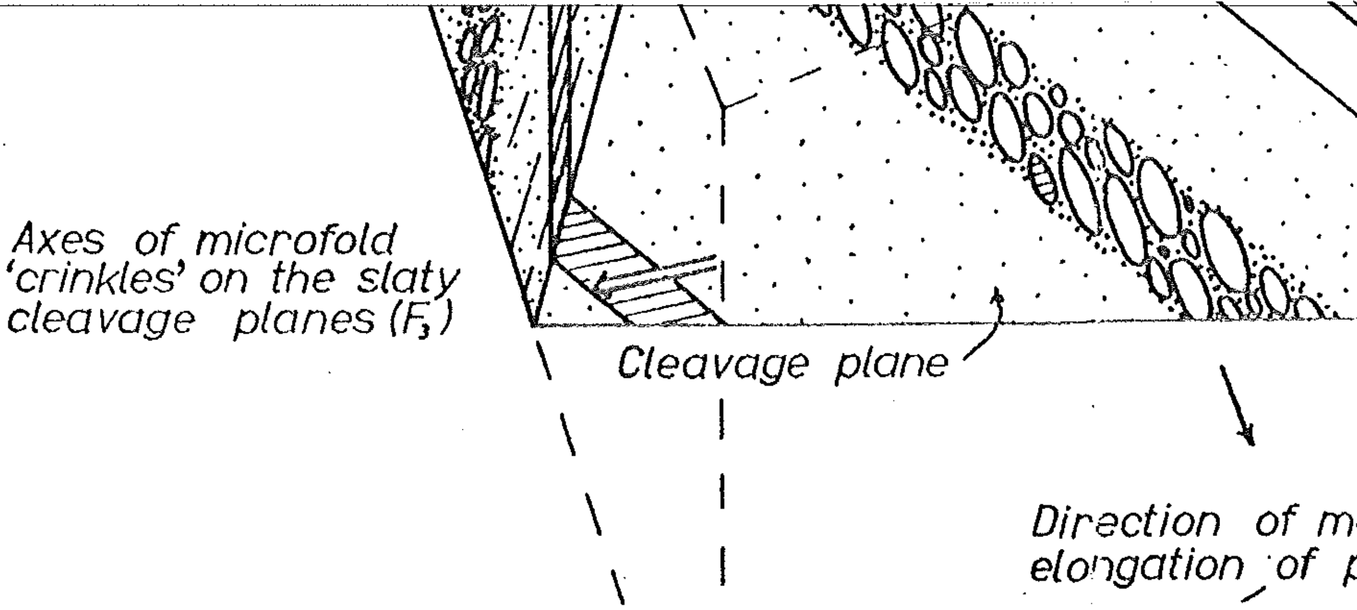
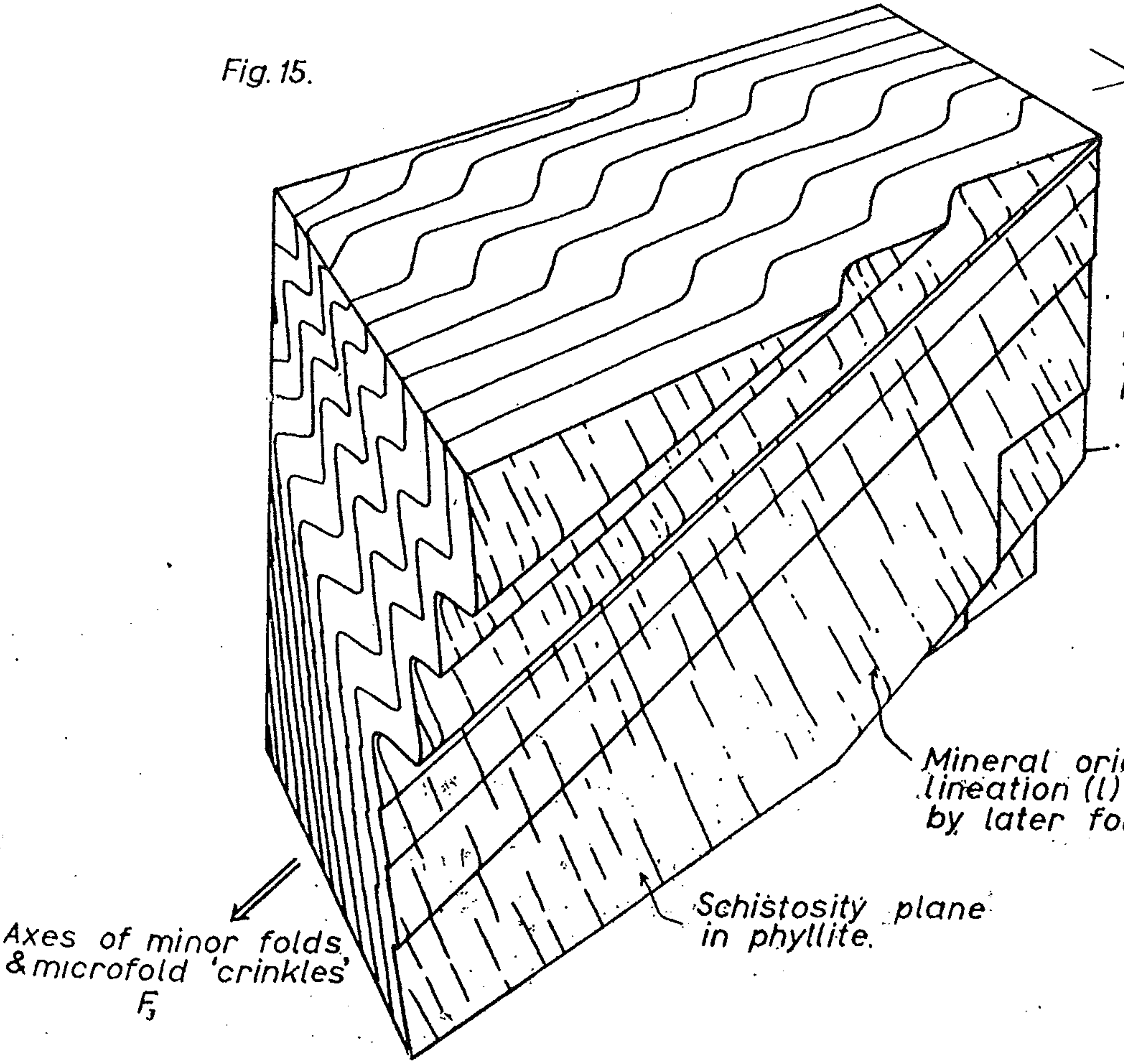


Axial plane of fold
Plane of deformed lineation
Detail to show structures measured.

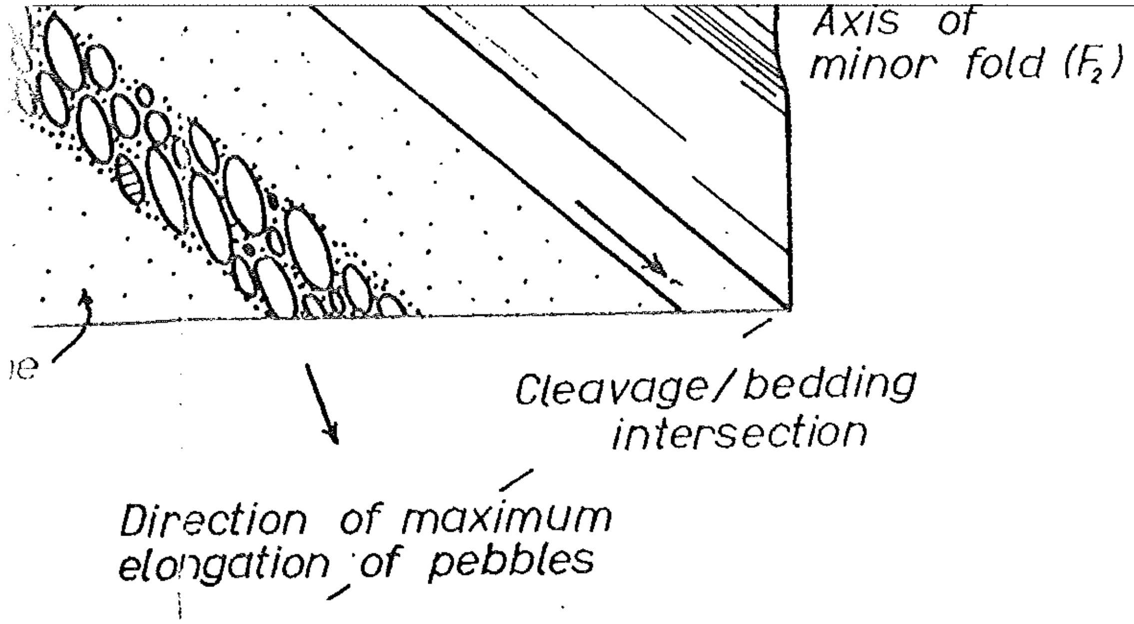


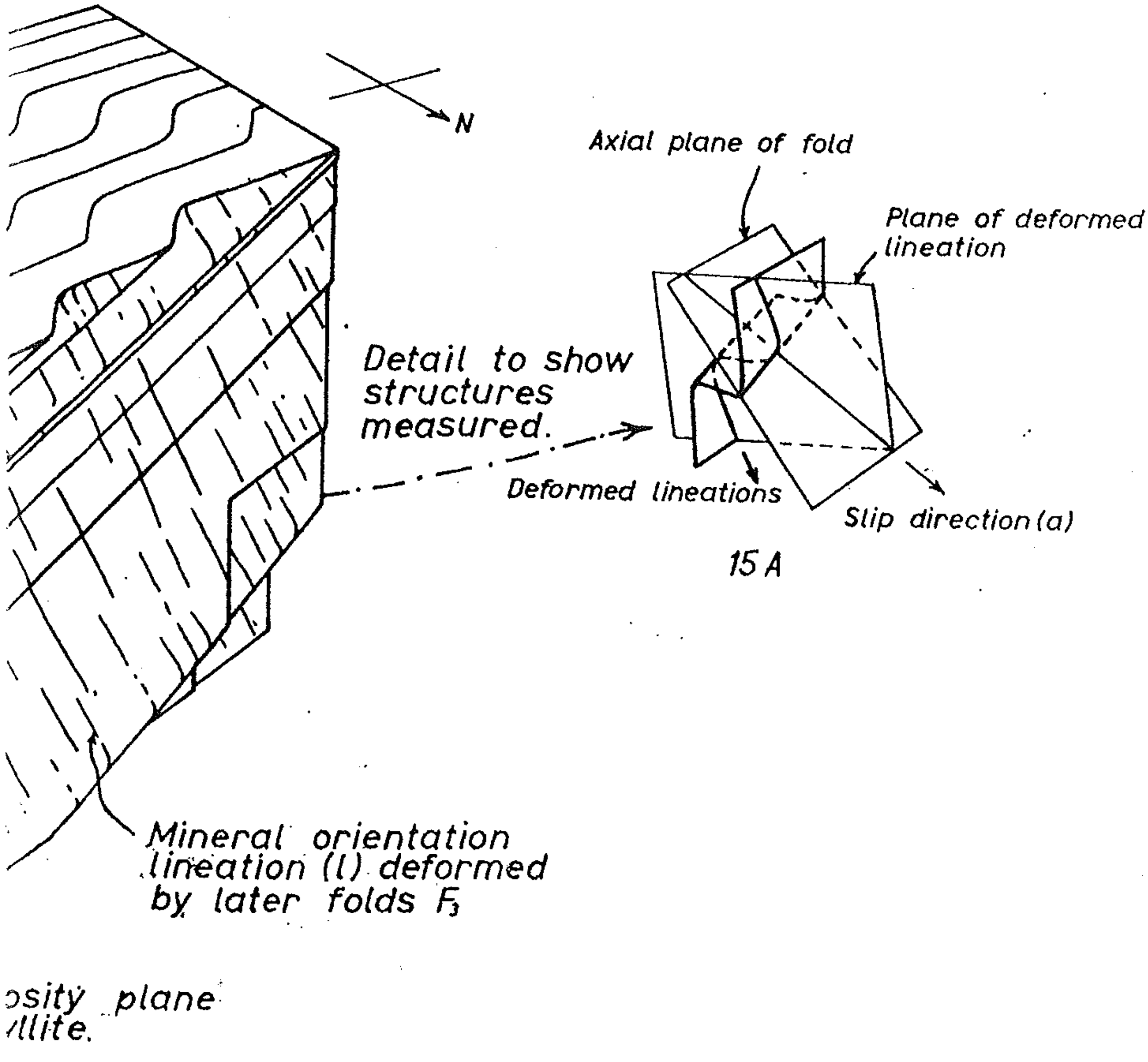
Axes of microfold 'crinkles' on the slaty cleavage planes (F_3)
Cleavage plane
Direction of
elongation of

Fig. 15.



Mineral
lineation (l)
by later
Schistosity plane in phyllite.
Axes of minor folds & microfold 'crinkles' F_3

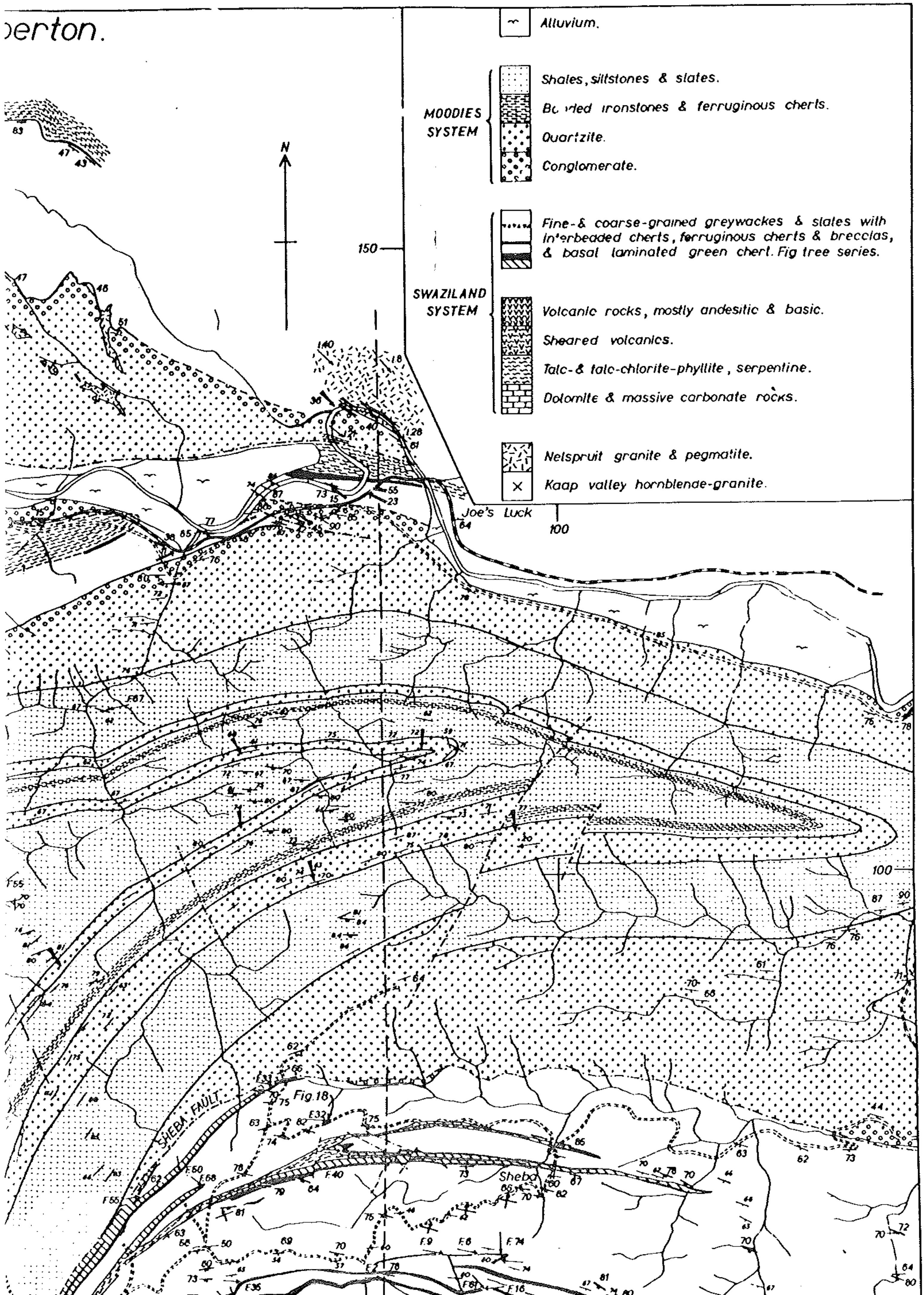

Axis of
minor fold (F_2)
Cleavage/bedding
intersection
Direction of maximum
elongation of pebbles


N
Axial plane of fold
Plane of deformed
lineation
Detail to show
structures
measured.
Deformed lineations
Slip direction (a)
Mineral orientation
lineation (l) deformed
by later folds F_3
osity plane
ıllite.

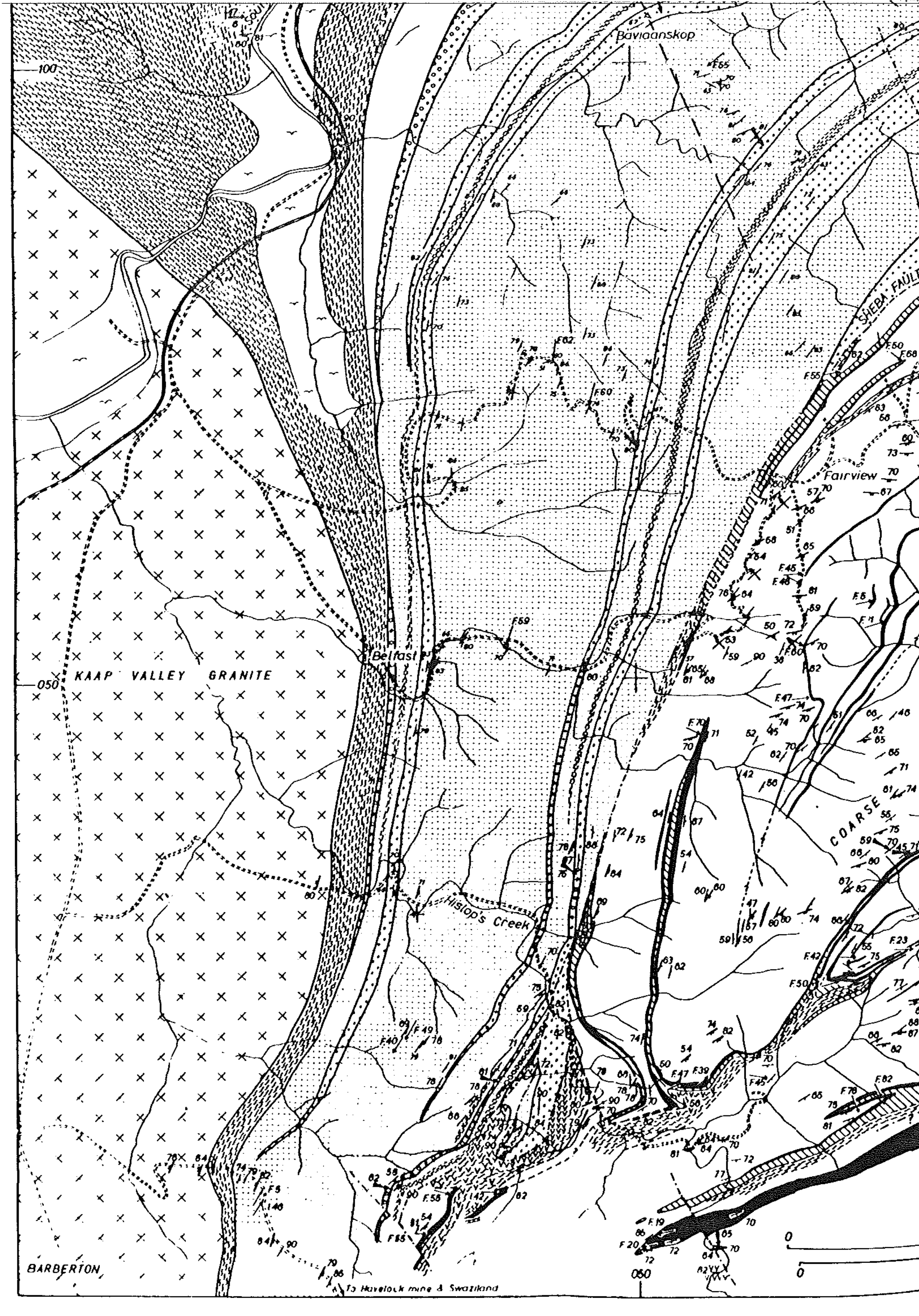
15 A

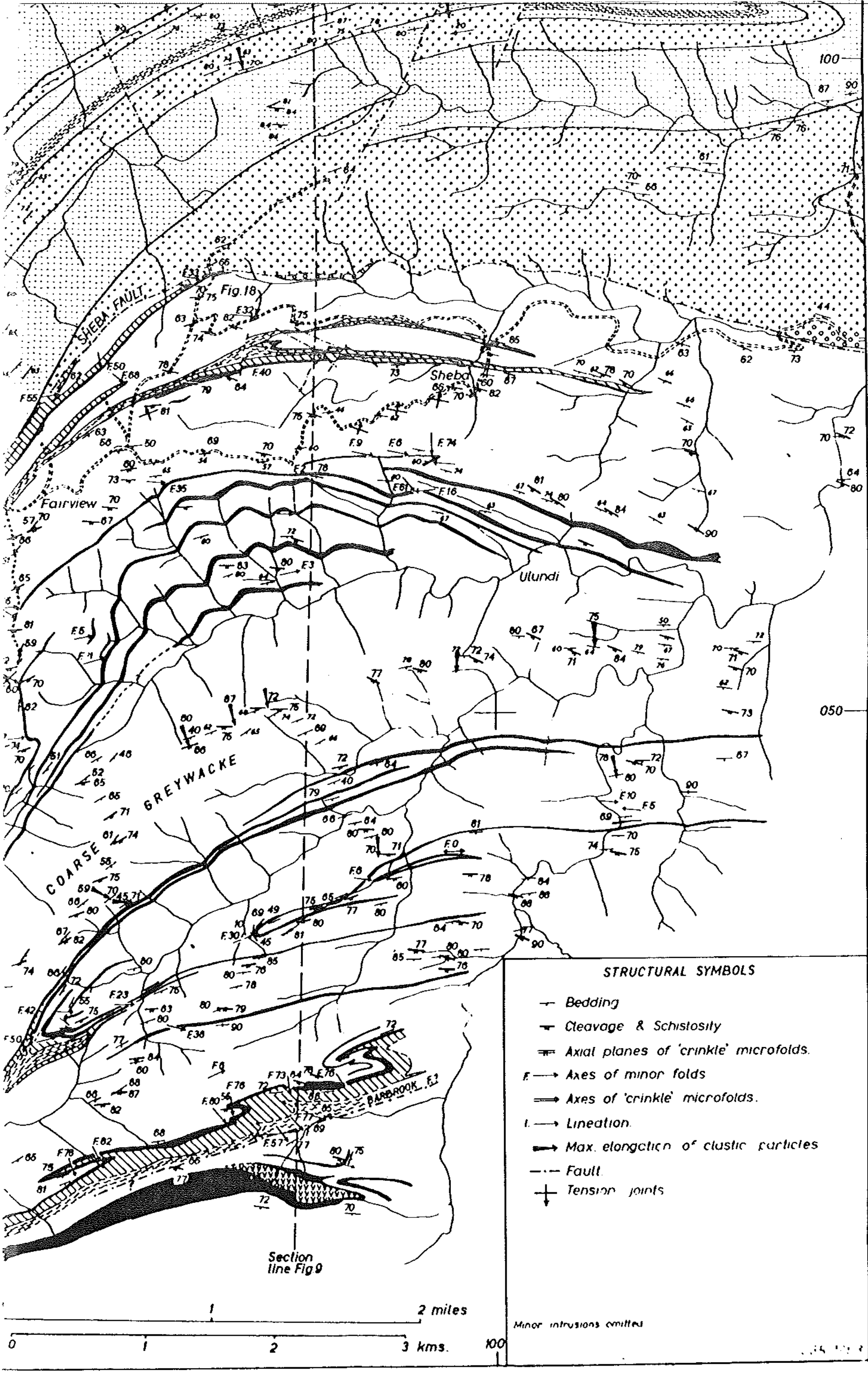
Structural reconnaissance map - N.E. Barberton.

Fig. 2.



berton.
N
Alluvium.
MOODIES SYSTEM
Shales, siltstones & slates.
Banded ironstones & ferruginous cherts.
Quartzite.
Conglomerate.
SWAZILAND SYSTEM
Fine- & coarse-grained greywackes & slates with interbeaded cherts, ferruginous cherts & breccias, & basal laminated green chert. Fig tree series.
Volcanic rocks, mostly andesitic & basic.
Sheared volcanics.
Talc- & talc-chlorite-phyllite, serpentine.
Dolomite & massive carbonate rocks.
Nelspruit granite & pegmatite.
Kaap valley hornblende-granite.
150
100
Joe's Luck
SHEBA FAULT
Fig.18
Sheba


Baviaanskop
100
050
KAAP VALLEY GRANITE
Belfast
Hislop's Creek
Fairview
SHEBA FAULT
COARSE
BARBERTON
To Havelock mine & Swaziland
050
0
0


SHEBA FAULT
Fig.18
Sheba
Fairview
Ulundi
COARSE GREYWACKE
BARBROOK F.1
Section line Fig.9
100
050
STRUCTURAL SYMBOLS
Bedding
Cleavage & Schistosity
Axial planes of 'crinkle' microfolds.
Axes of minor folds
Axes of 'crinkle' microfolds.
Lineation.
Max. elongation of clastic particles
Fault.
Tension joints
1
2 miles
0
1
2
3 kms.
100
Minor intrusions omitted

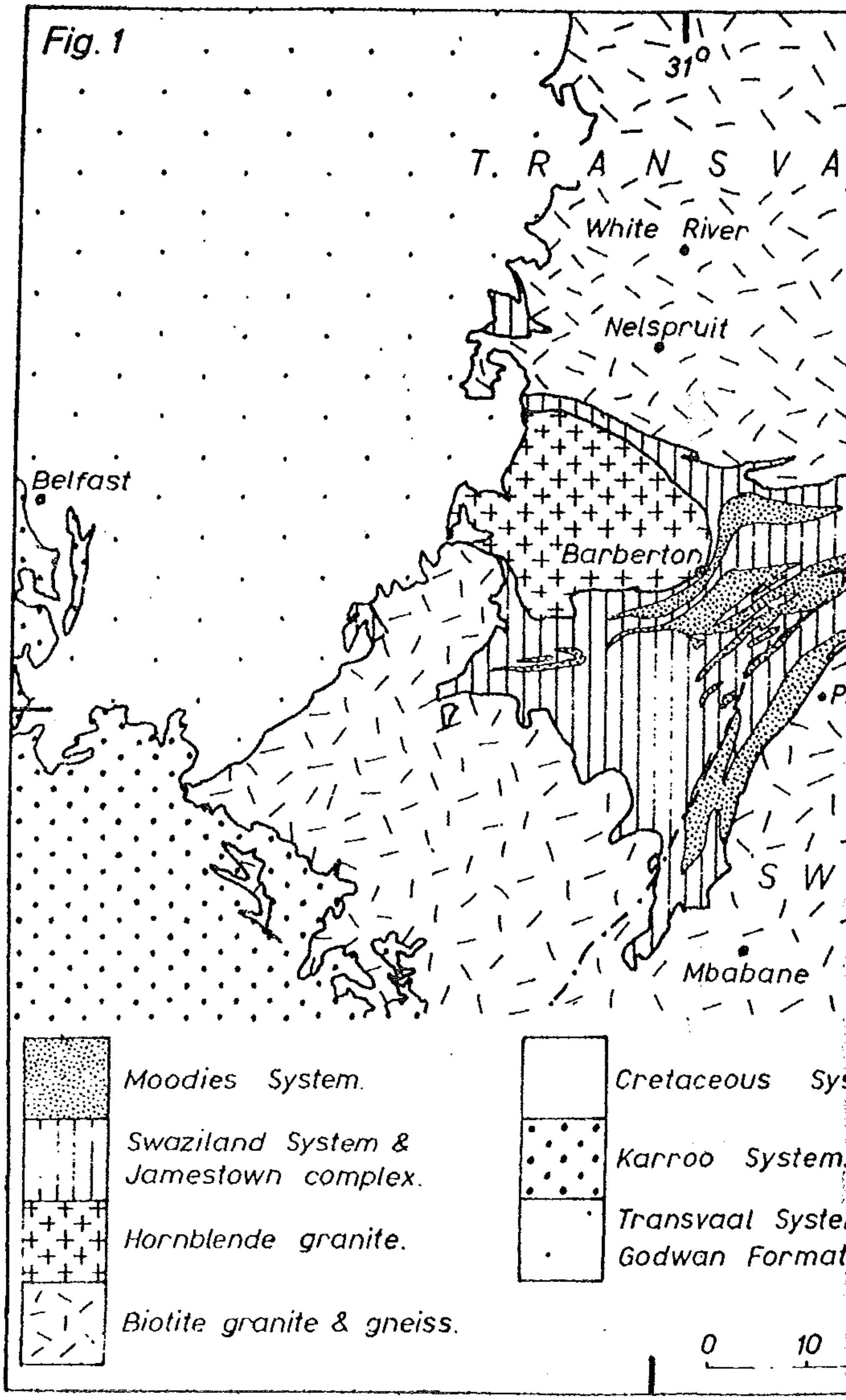
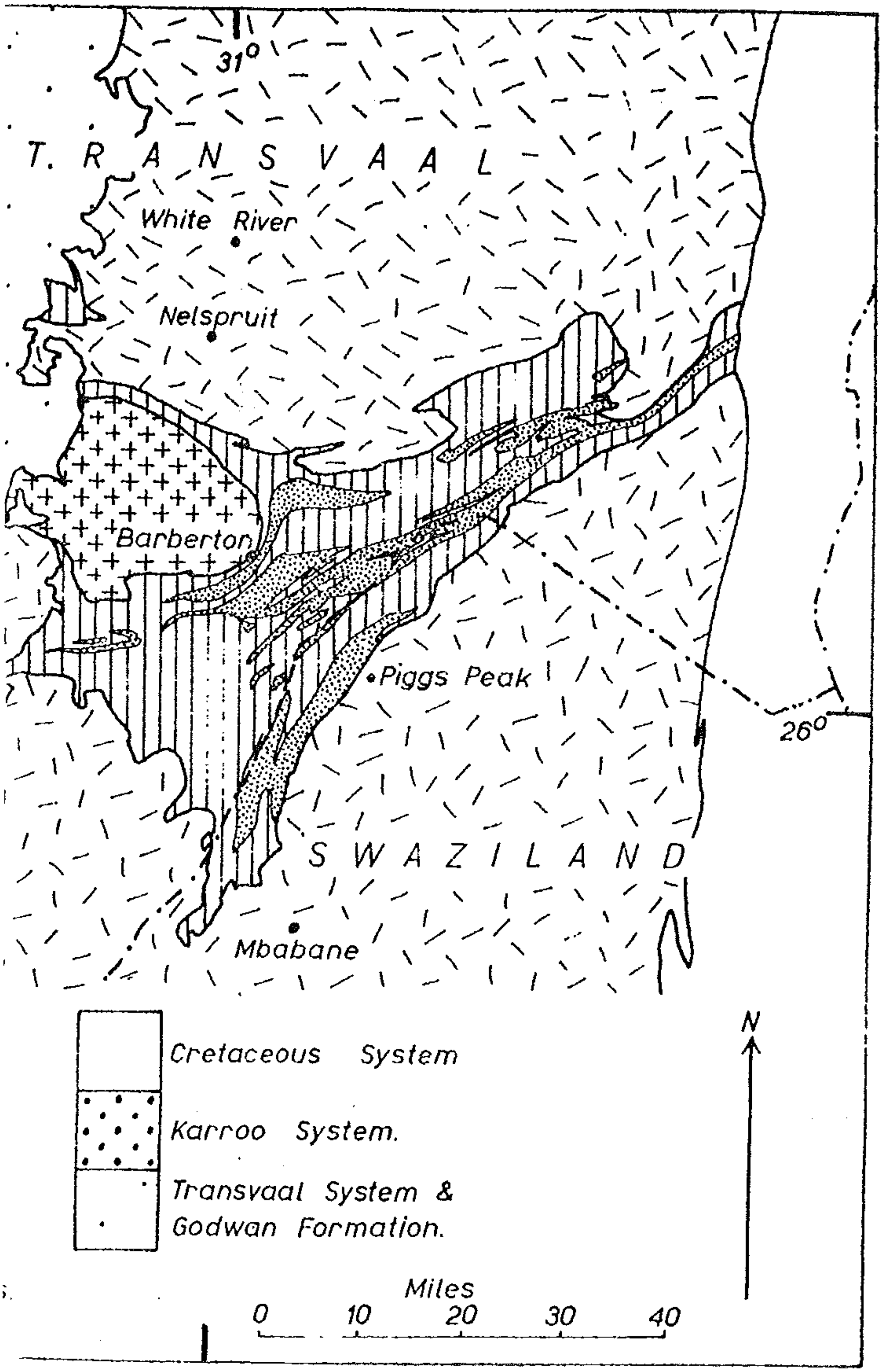


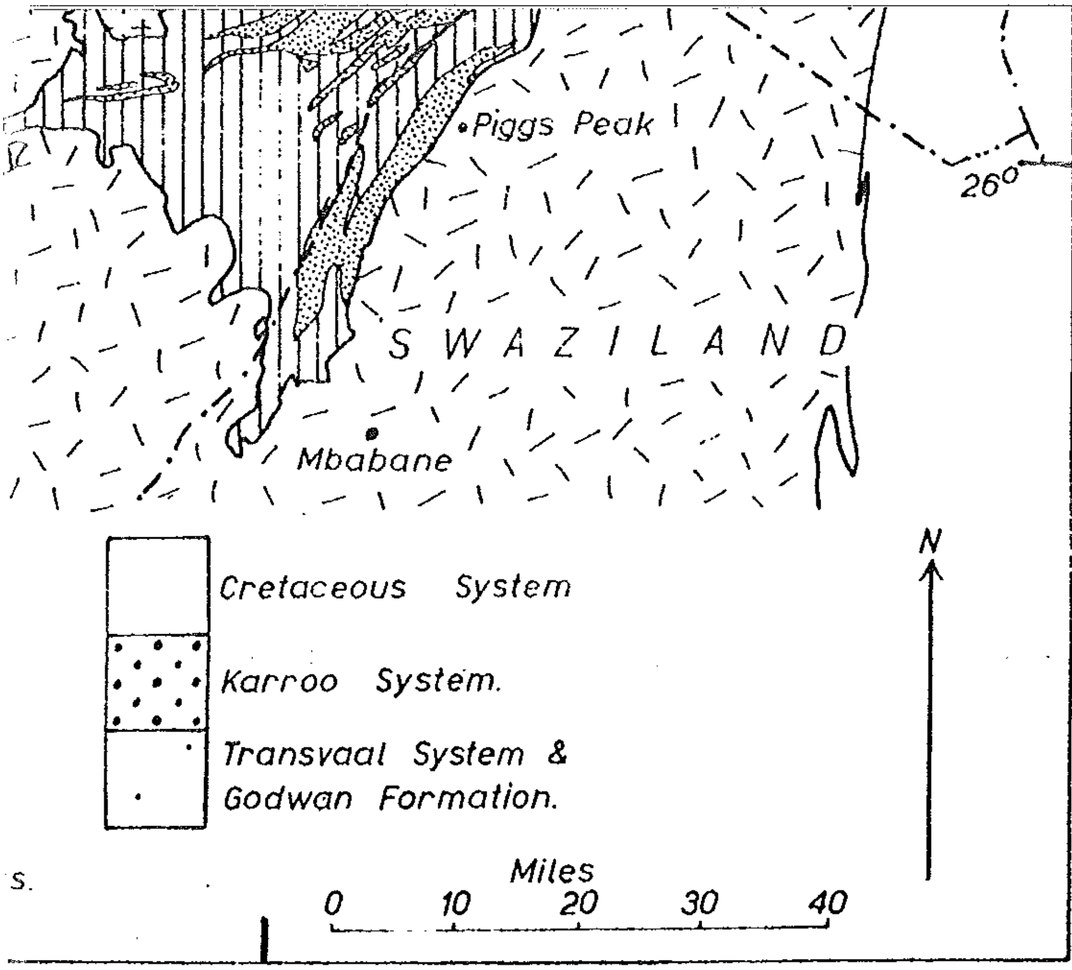
Fig.4.

Fig 3. Chert of replacement origin



31°
T R A N S V A A L
White River
Nelspruit
Barberton
Piggs Peak
26°
S W A Z I L A N D
Mbabane
Cretaceous System
Karroo System.
Transvaal System &
Godwan Formation.
N
Miles
0 10 20 30 40

Fig.4.


Piggs Peak
26°
SWAZILAND
Mbabane
Cretaceous System
Karroo System.
Transvaal System &
Godwan Formation.
N
Miles
0 10 20 30 40

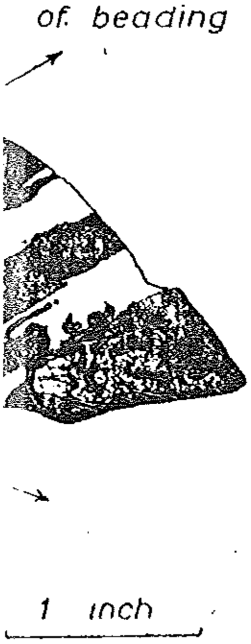
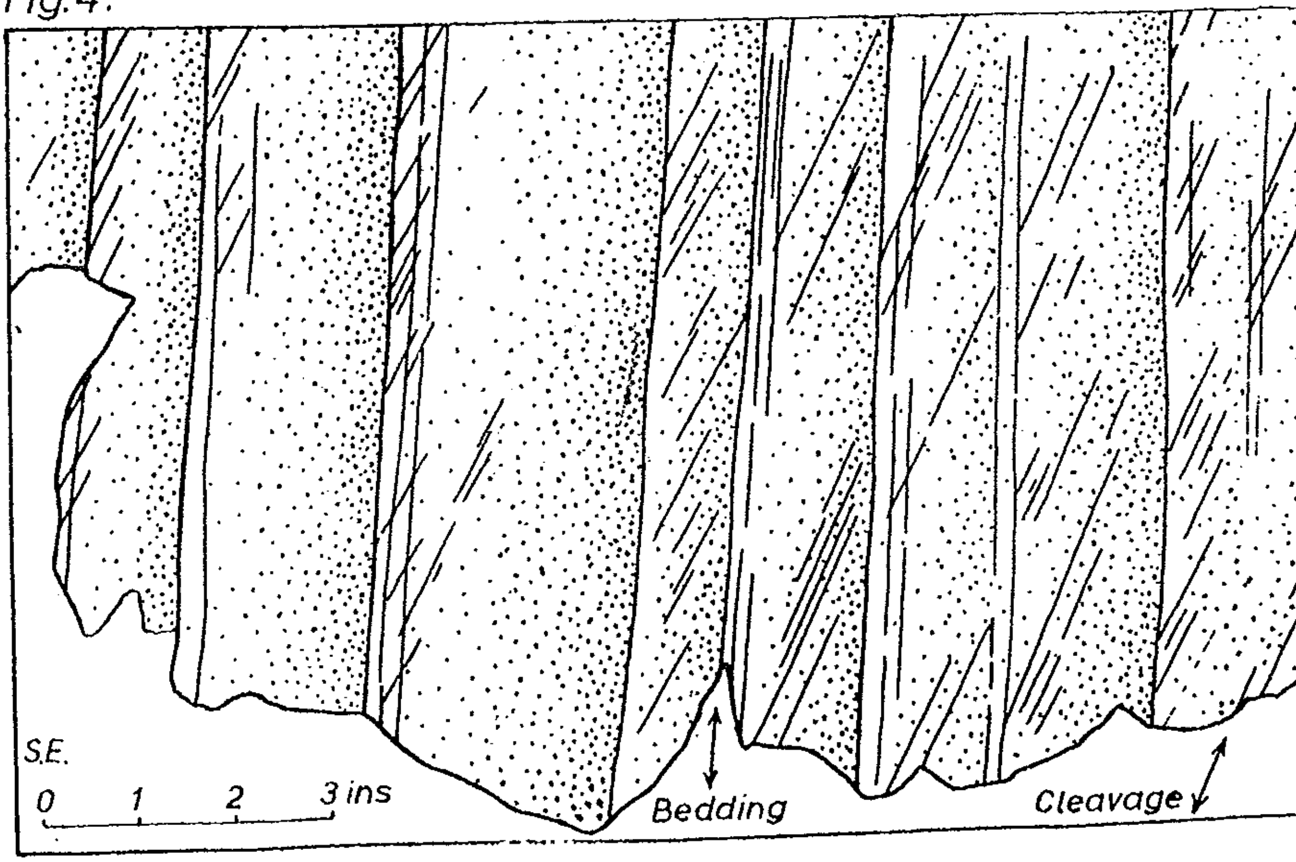

of. beading
1 inch

Fig.4.


S.E.
0 1 2 3 ins
Bedding
Cleavage

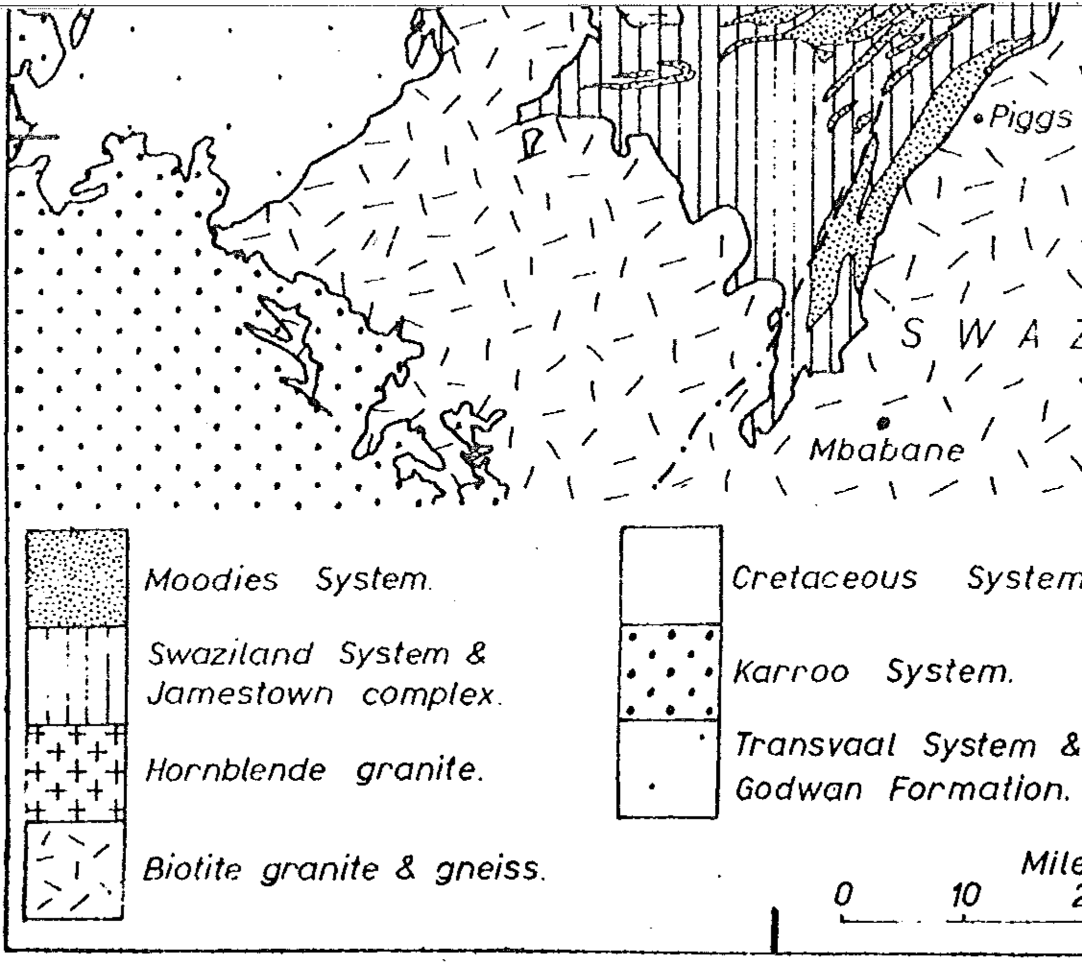


Fig 3. Chert of replacement origin

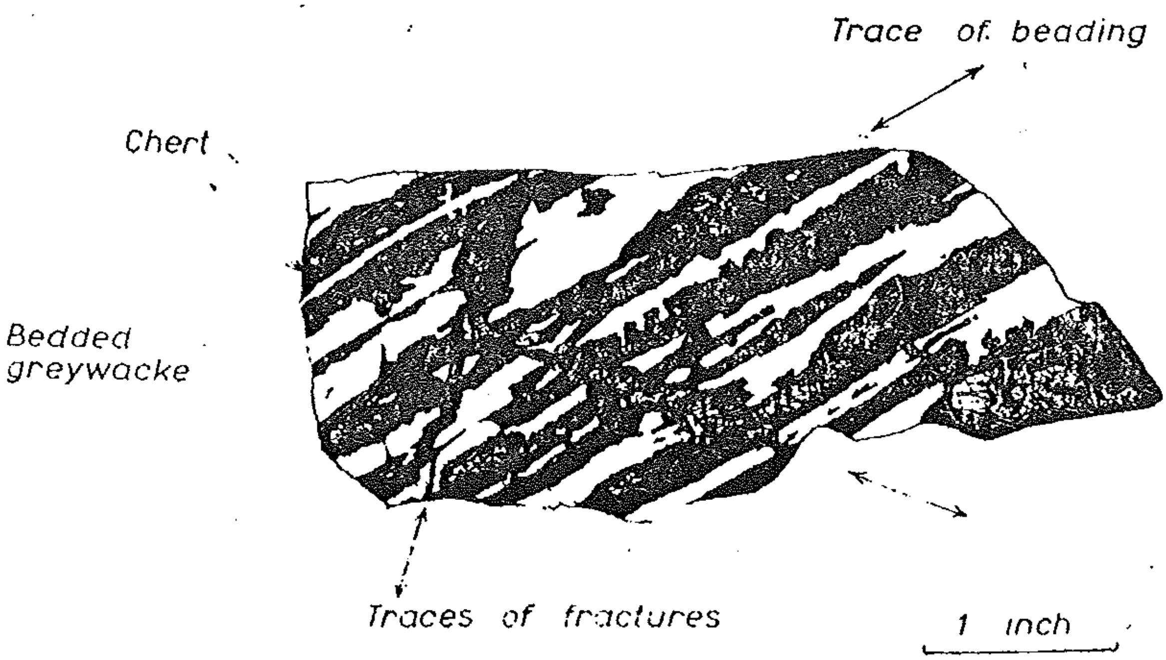


Fig.4.

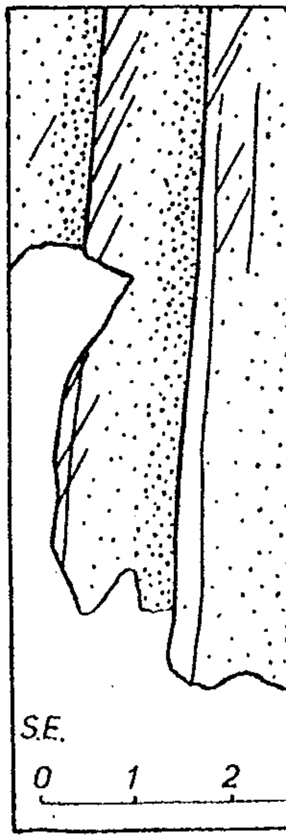
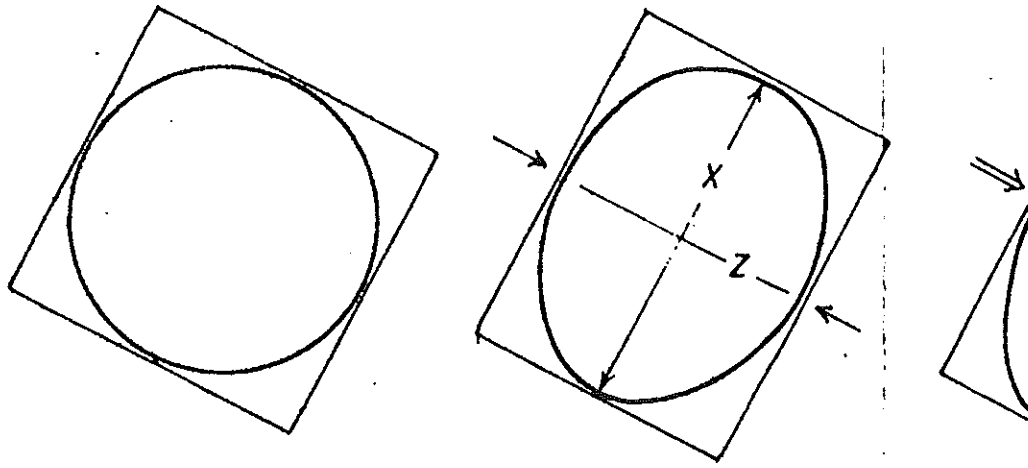
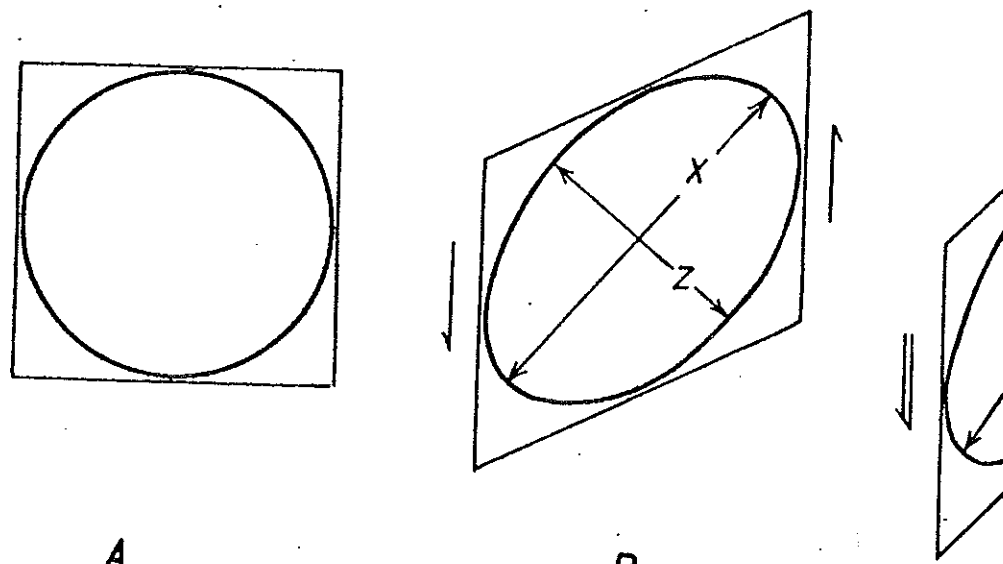


Fig. 5.



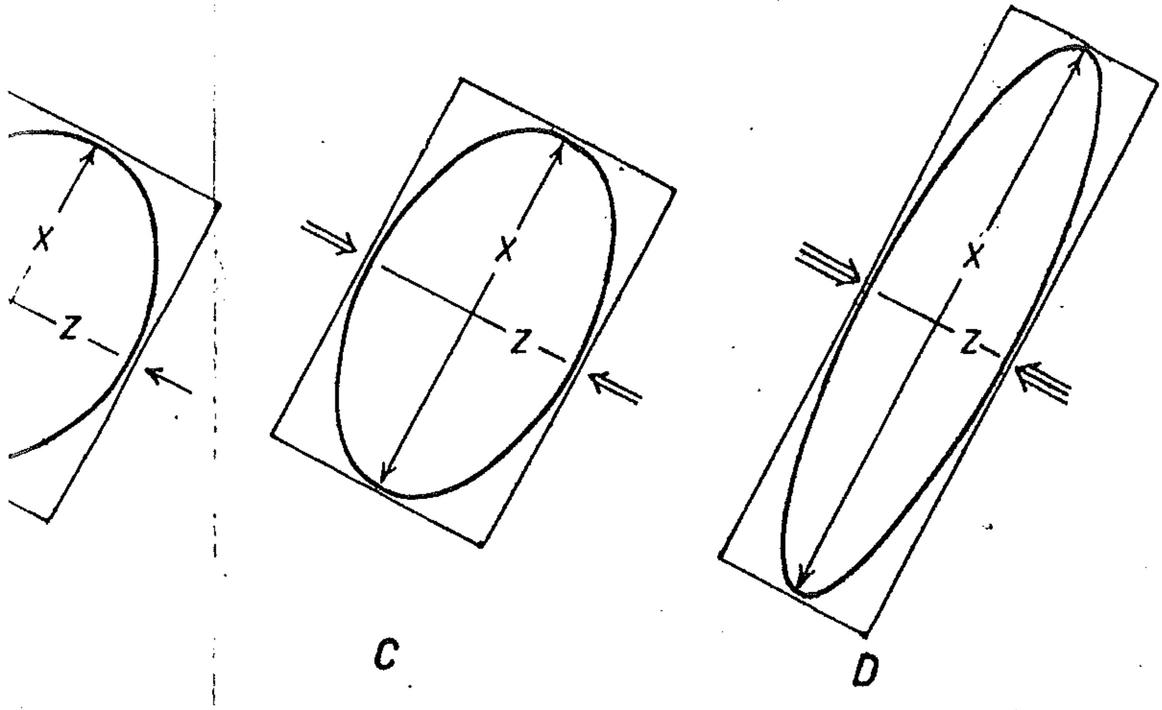
A B

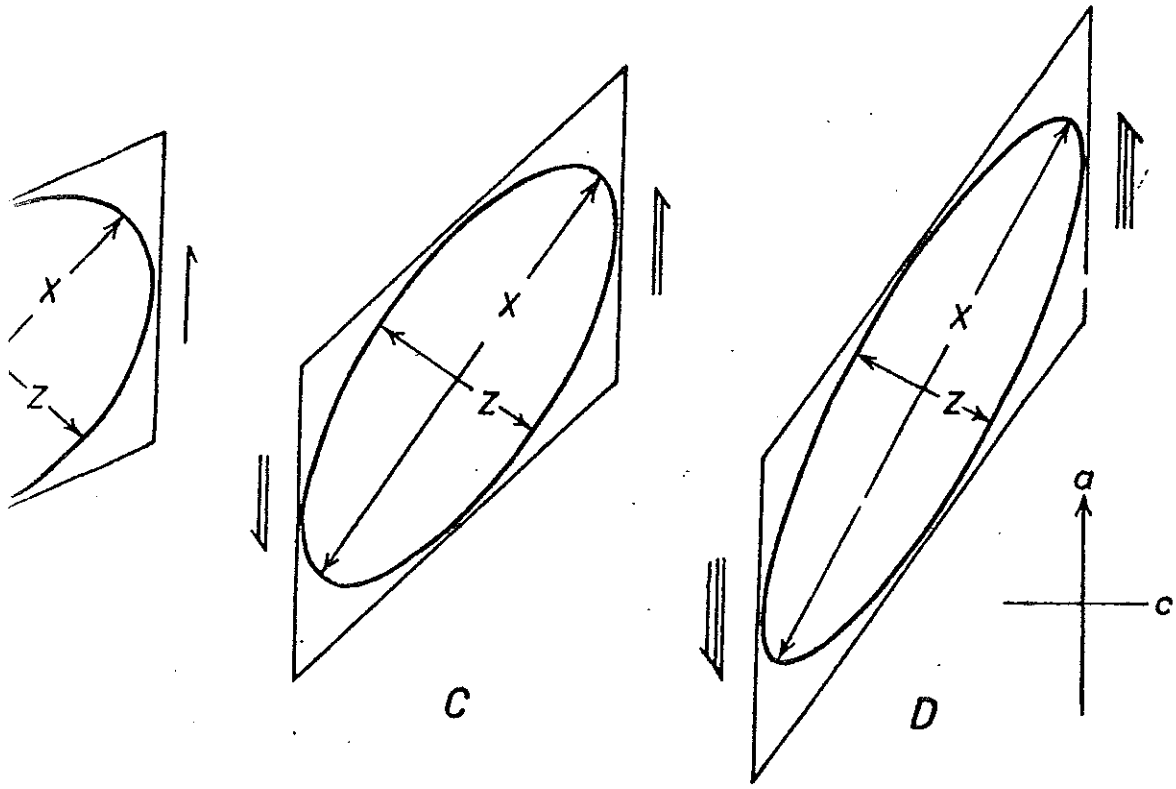
Fig. 6.



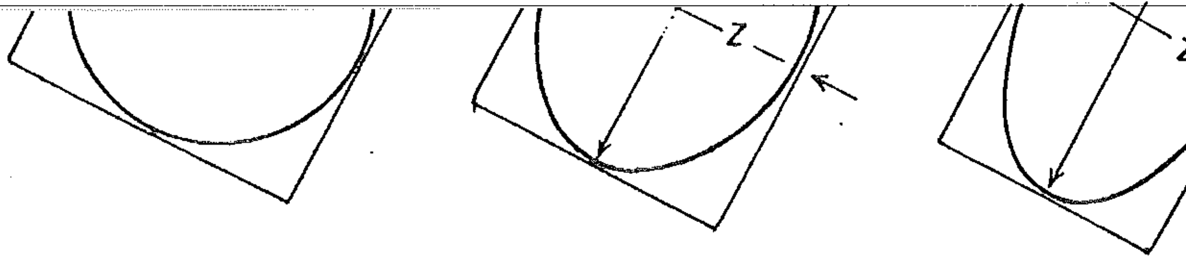
A B

Fig. 7.


X
Z
X
Z
C
X
Z
D

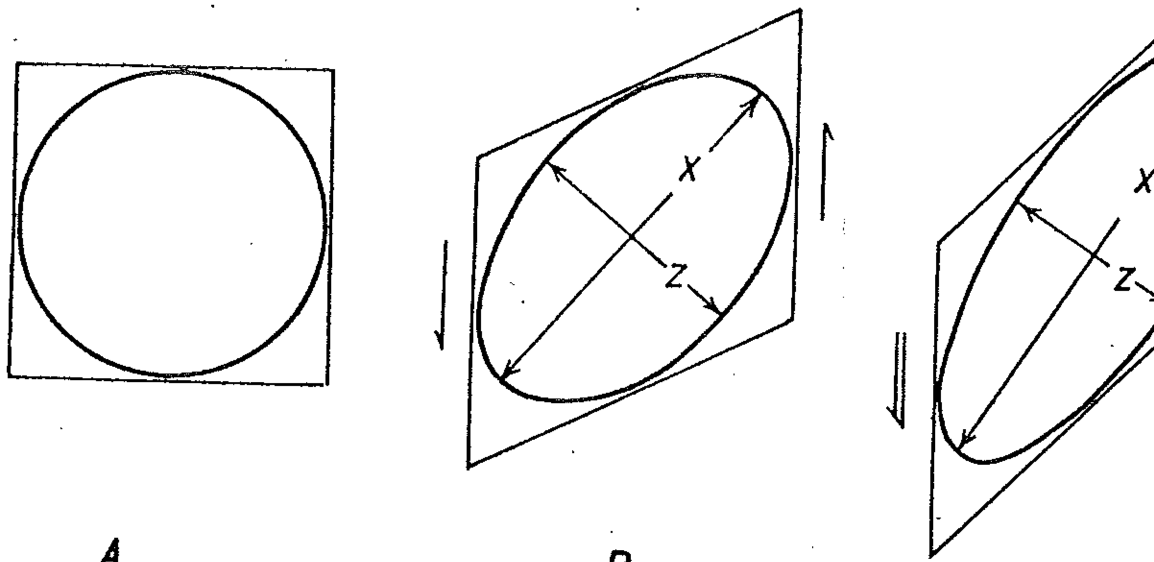

X
Z
X
Z
C
X
Z
D
a
c


Shear
Cleavage



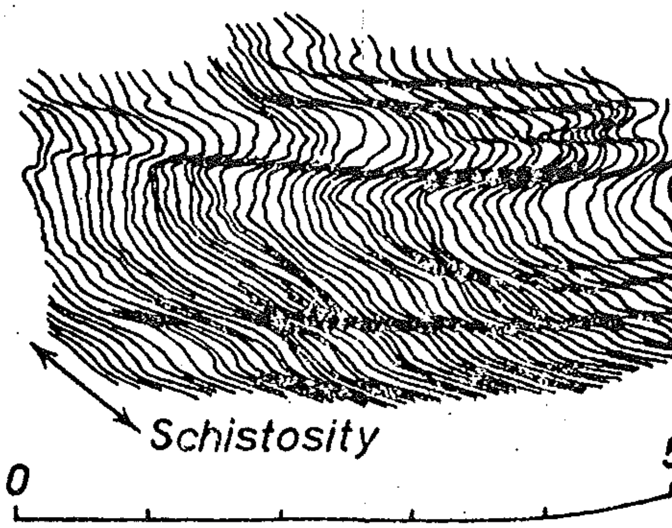
A B C

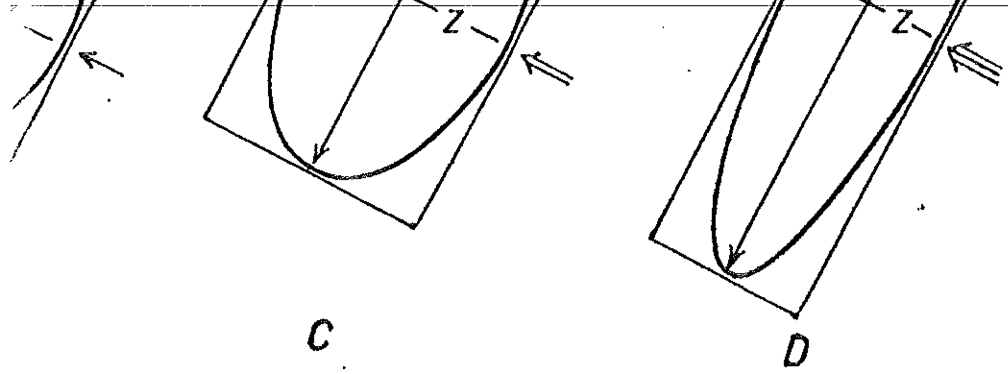
Fig. 6.

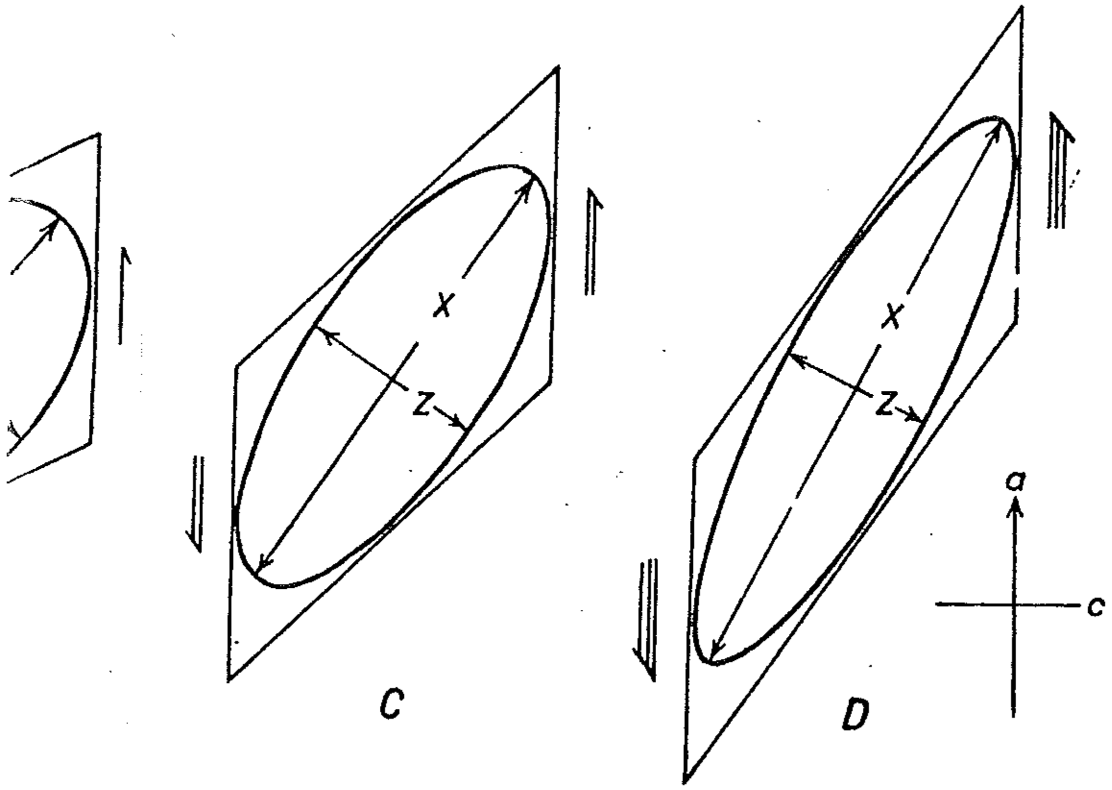


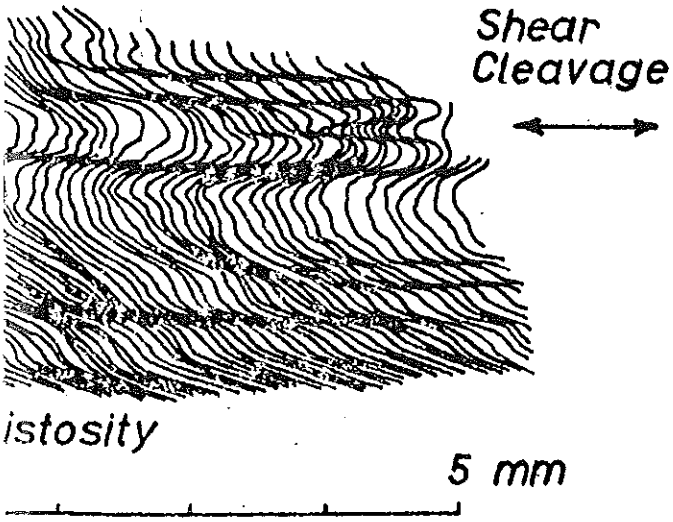
A B C

Fig. 7.




Z
Z
C
D


X
Z
X
Z
a
c
C
D


Shear
Cleavage
istosity
5 mm

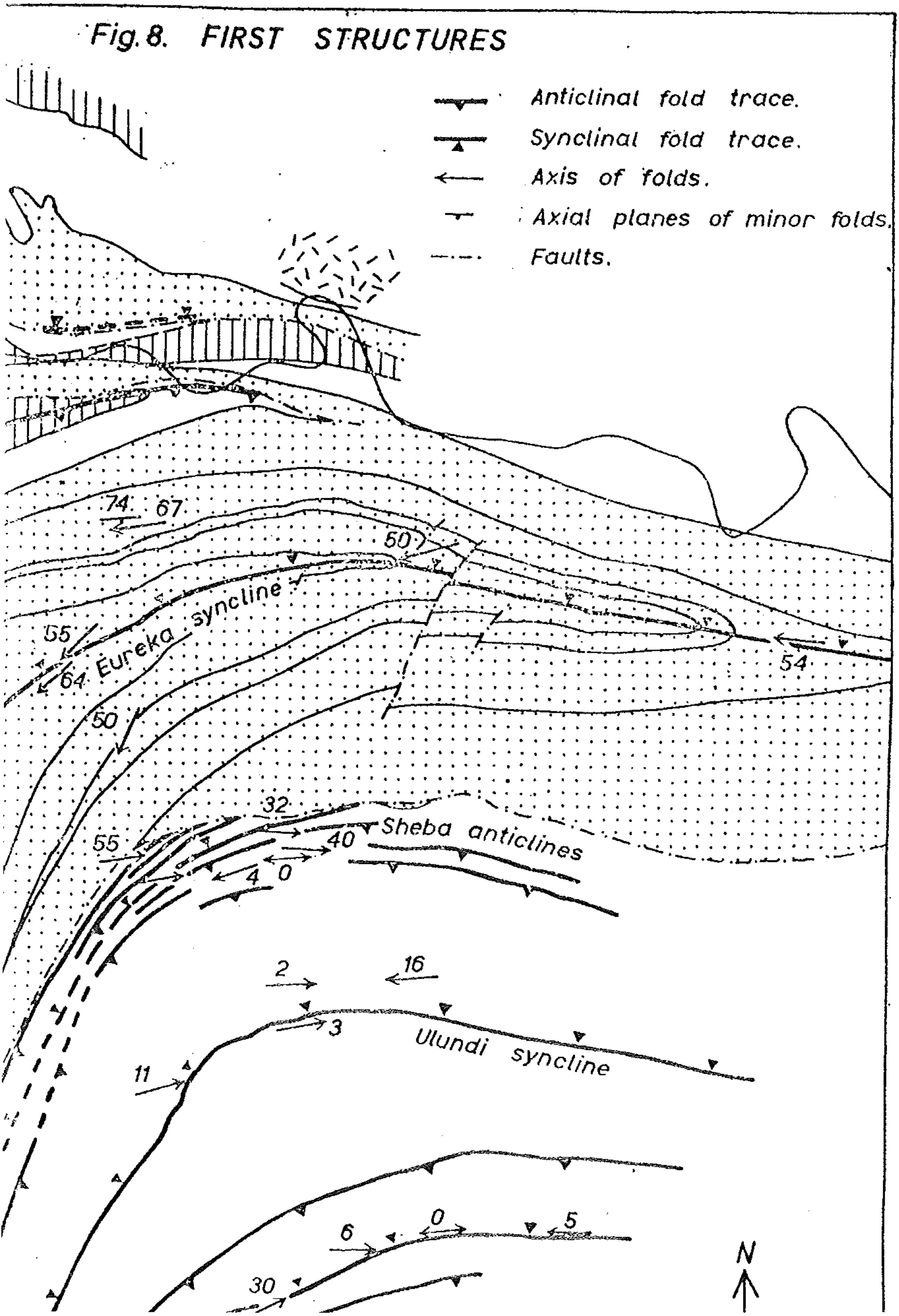


Fig. 8. FIRST STRUCTURES
Anticlinal fold trace.
Synclinal fold trace.
Axis of folds.
Axial planes of minor folds.
Faults.
74
67
50
55
64
Eureka syncline
54
50
32
55
40
4
0
Sheba anticlines
2
16
3
Ulundi syncline
11
0
5
6
30
N

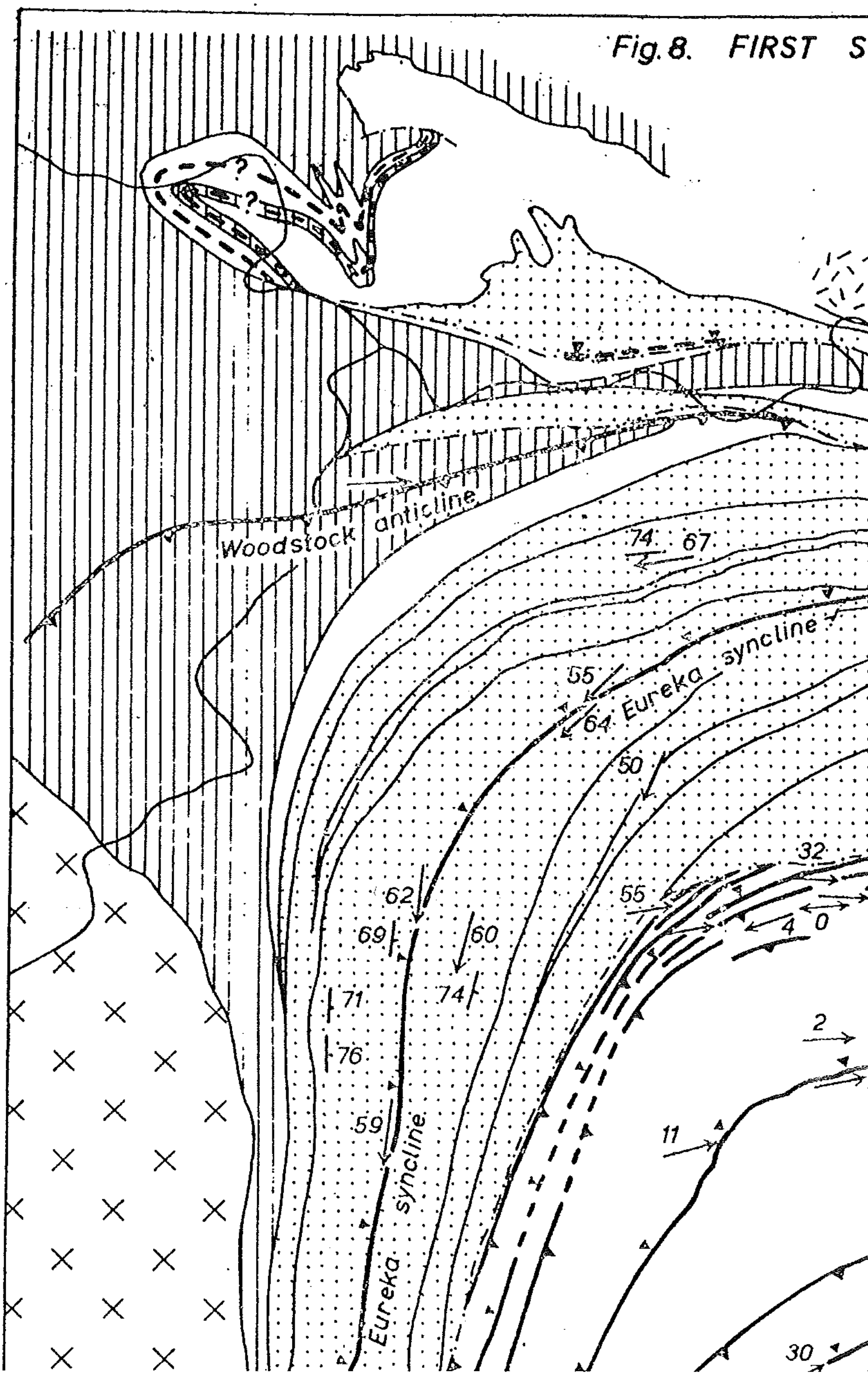
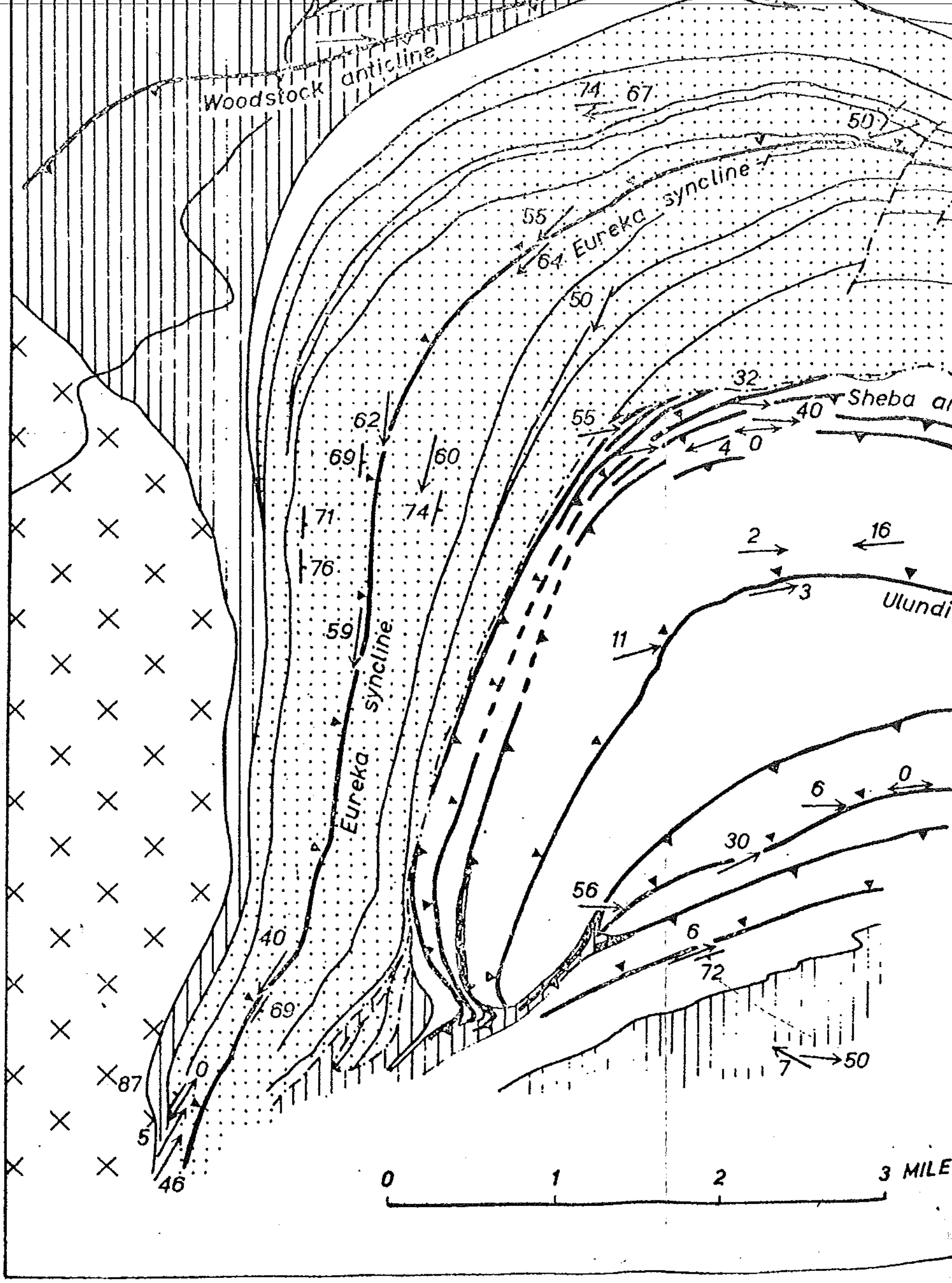
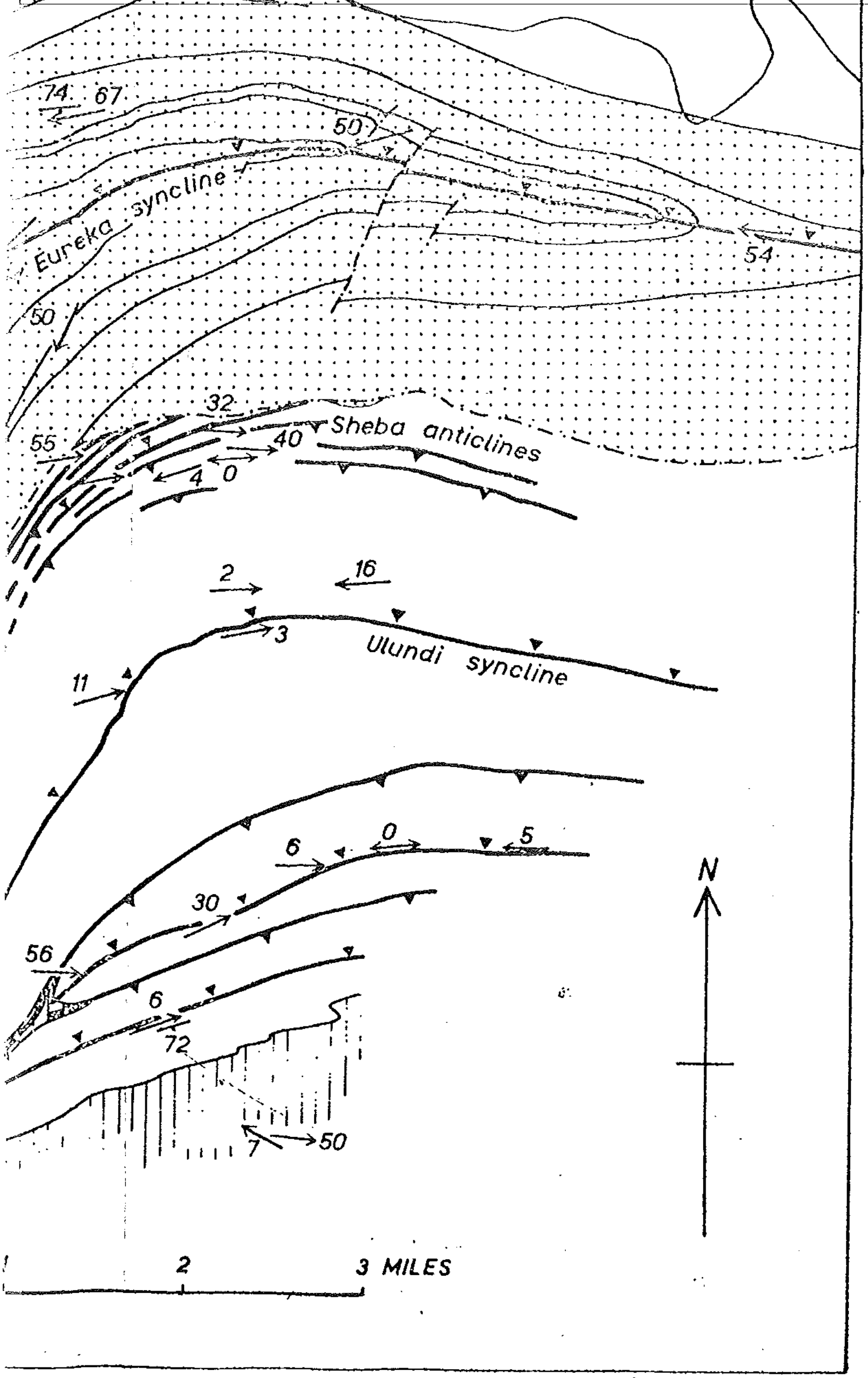


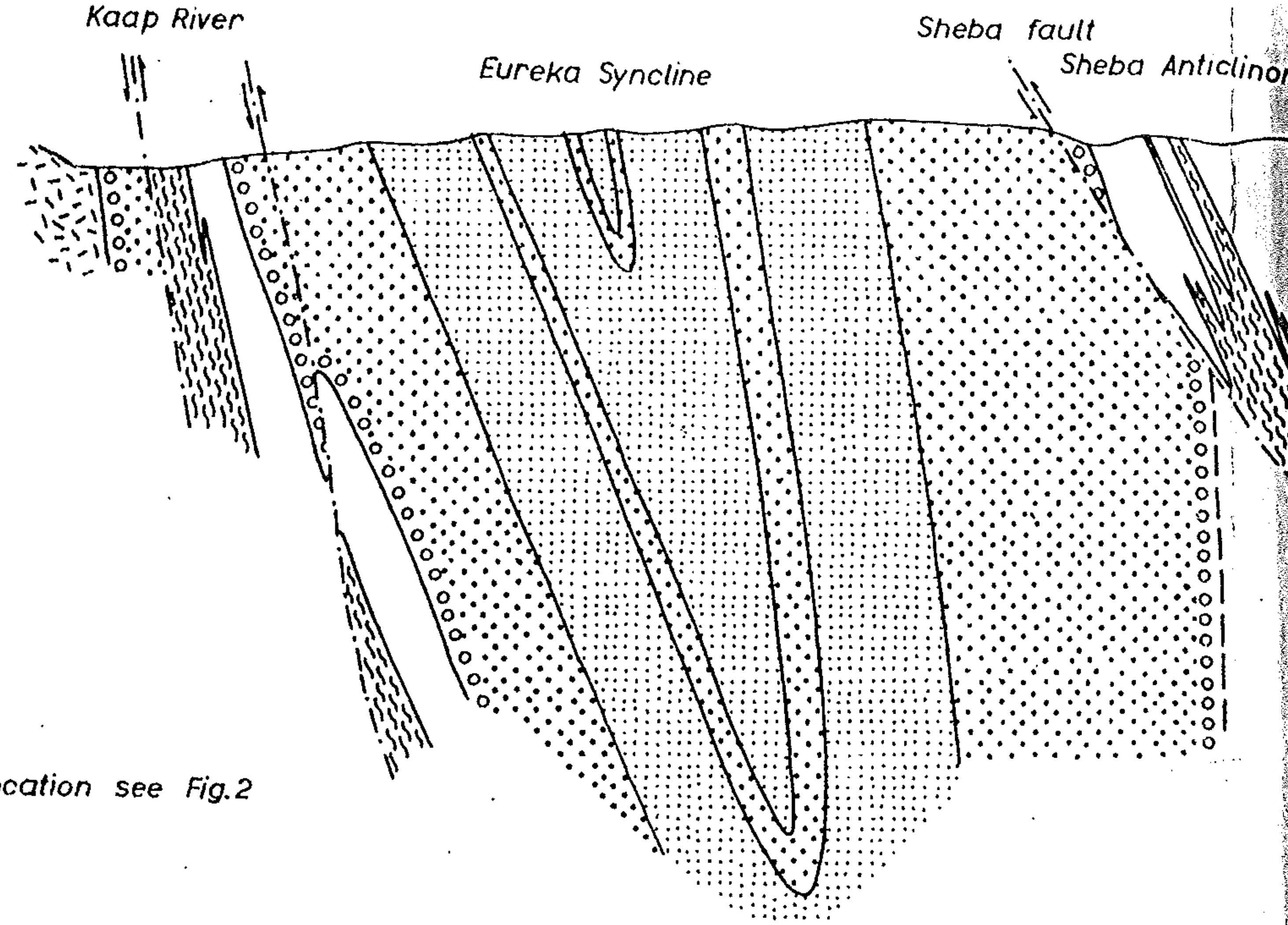
Fig. 8. FIRST S
Woodstock anticline
Eureka syncline
Eureka syncline
74
67
55
64
50
32
55
4
0
62
69
60
71
74
76
2
59
11
30



Woodstock anticline
Eureka syncline
Eureka syncline
Sheba
Ulundi
0
1
2
3 MILE

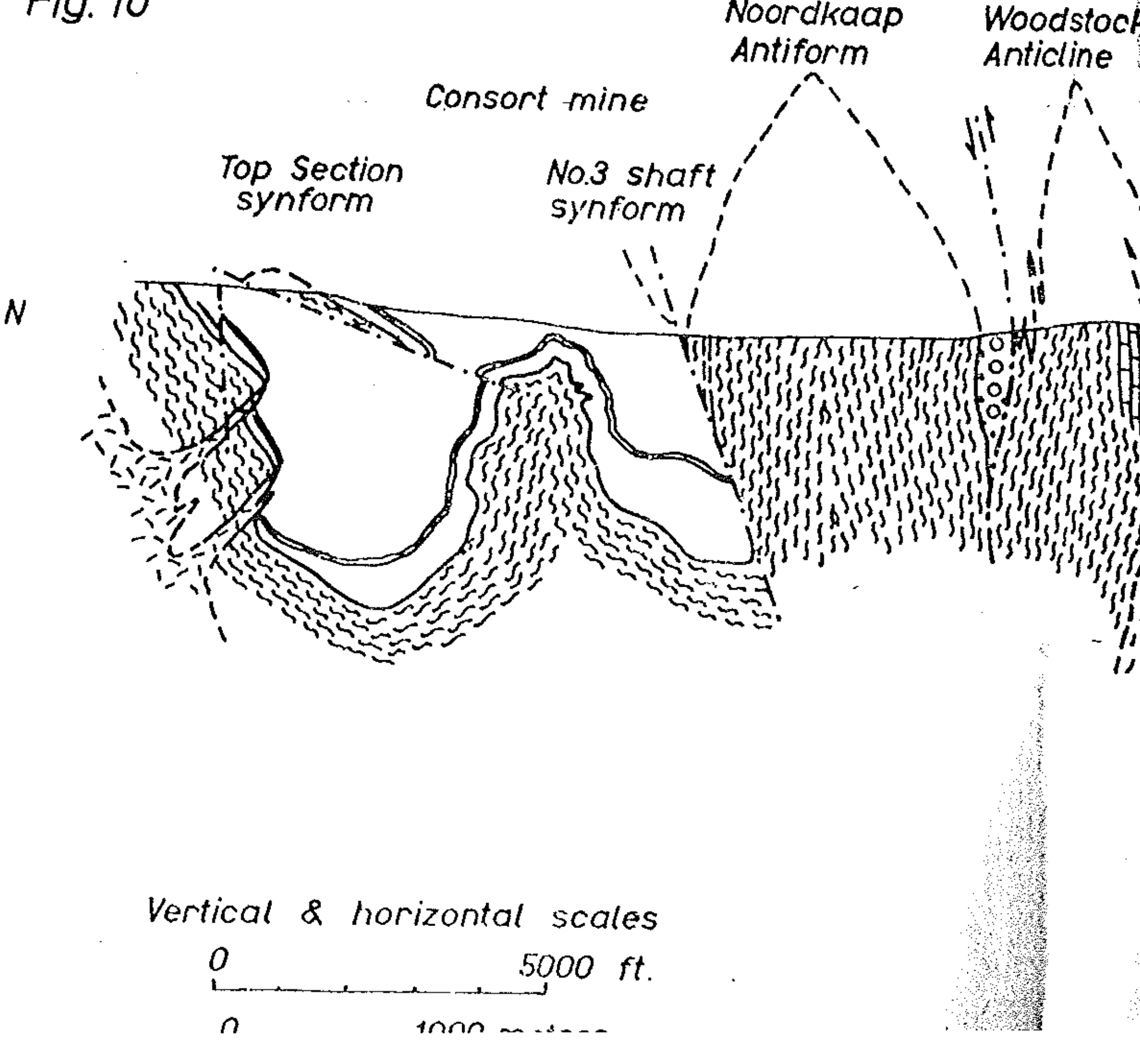


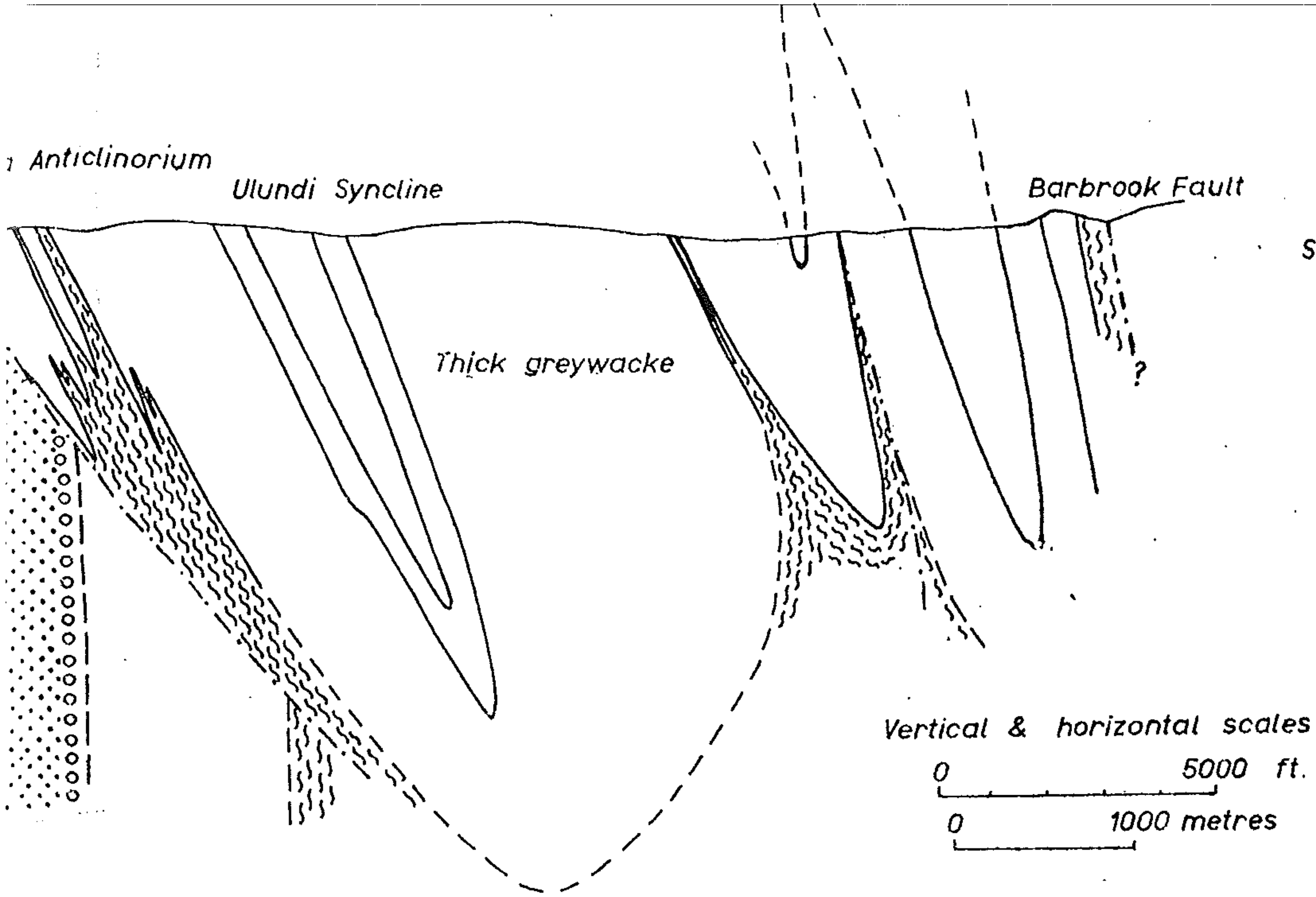
74
67
50
Eureka syncline
54
50
32
55
40
Sheba anticlines
4
0
2
16
3
Ulundi syncline
11
0
5
6
30
N
56
6
72
7
50
2
3 MILES

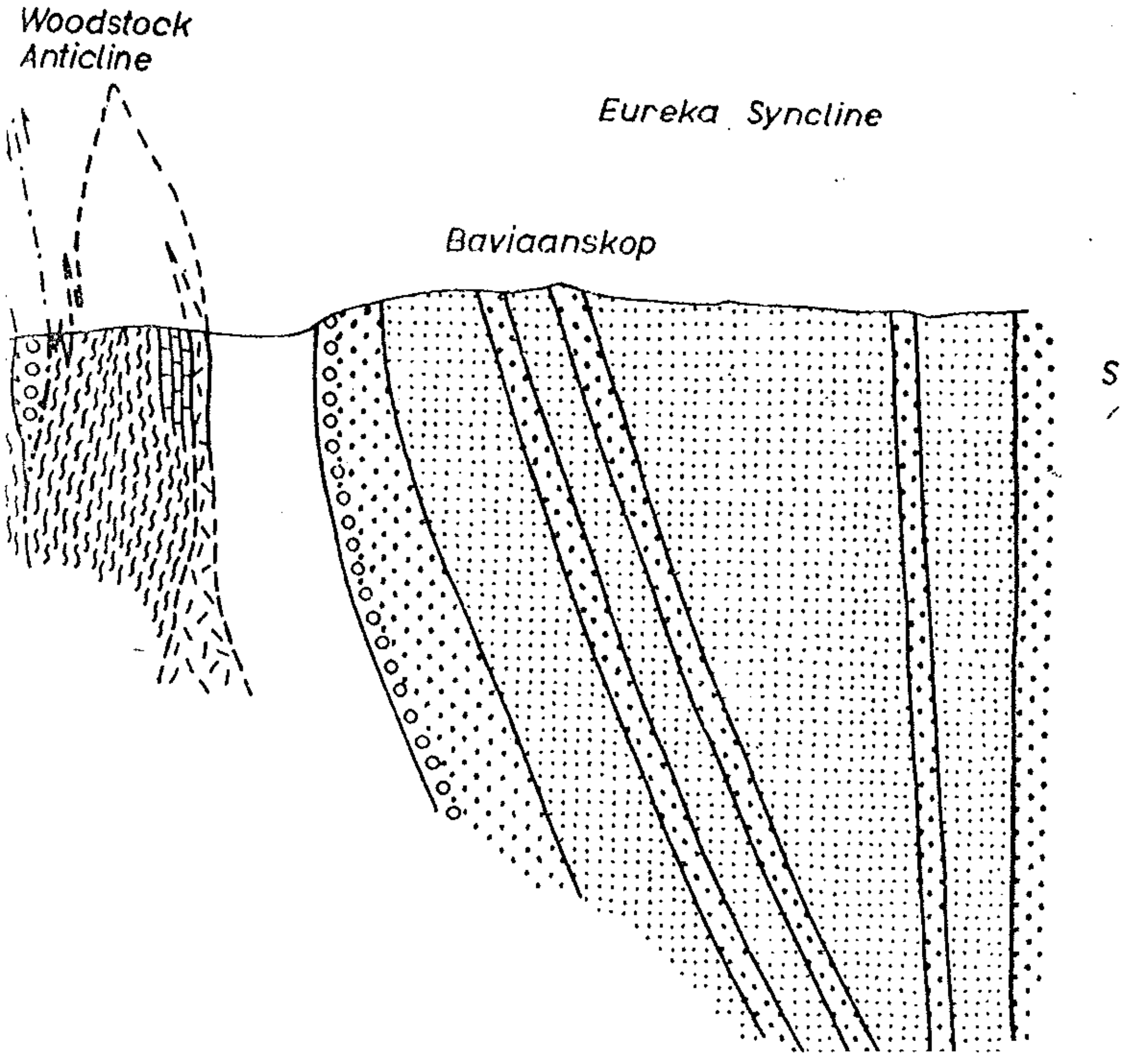


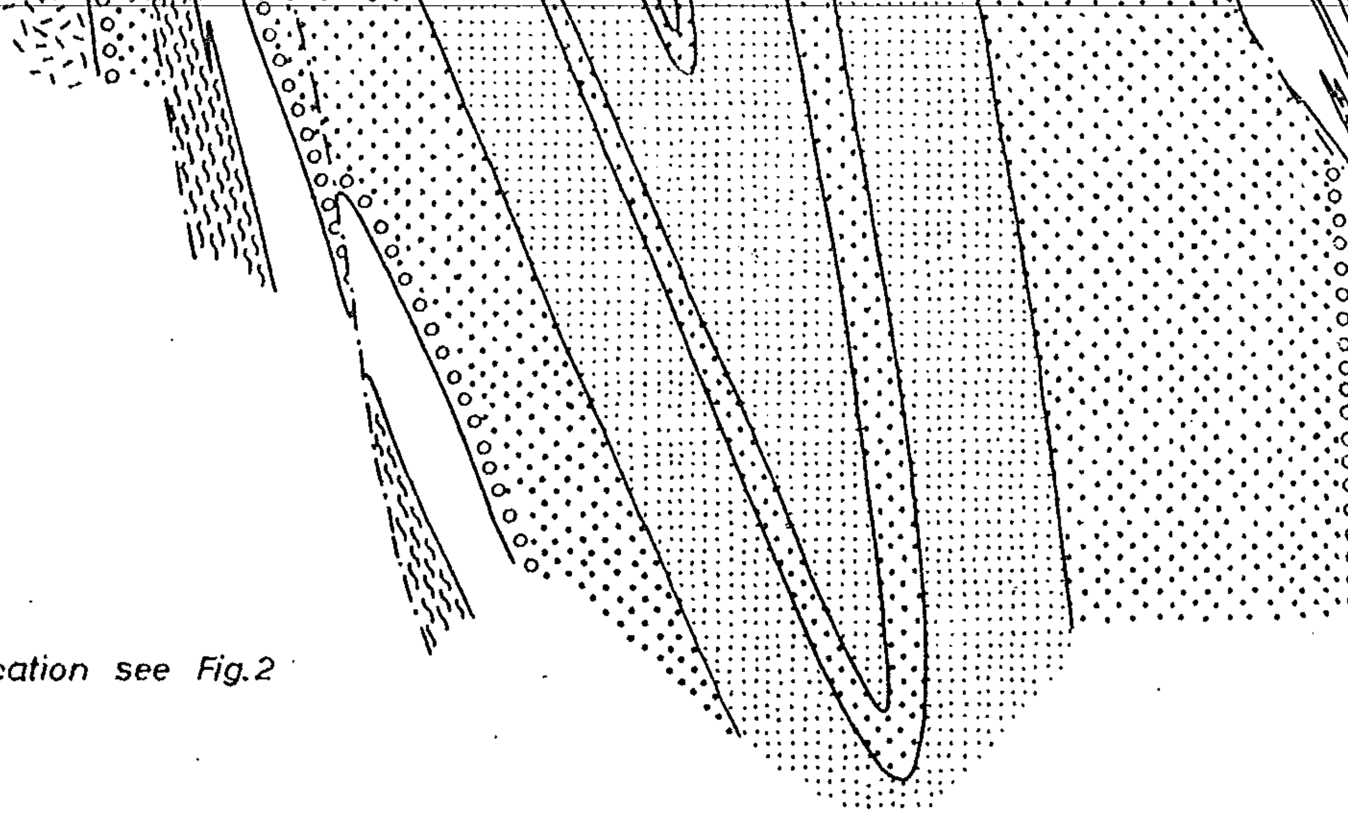
For location see Fig.2

Fig. 10



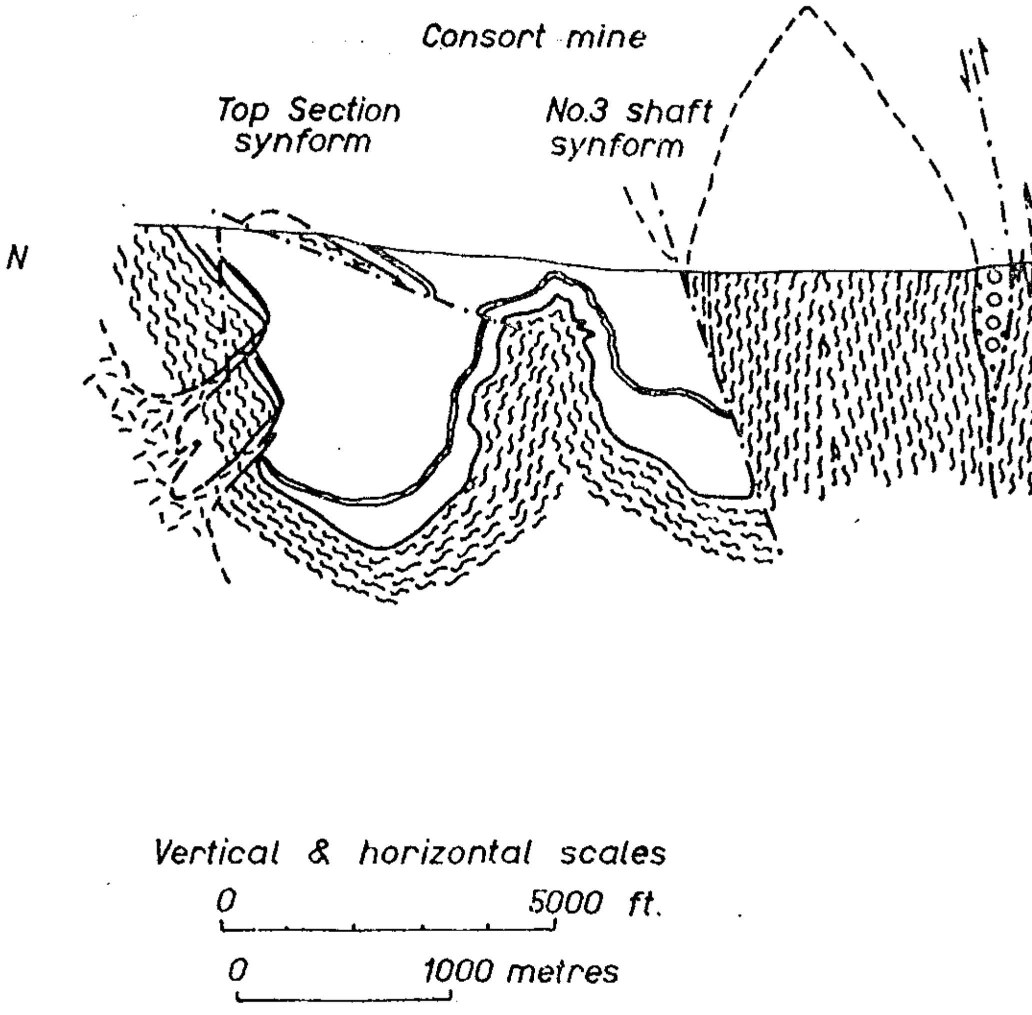

Anticlinorium
Ulundi Syncline
Barbrook Fault
S
Thick greywacke
?
Vertical & horizontal scales
0
5000 ft.
0
1000 metres


Woodstock
Anticline
Eureka Syncline
Baviaanskop
S

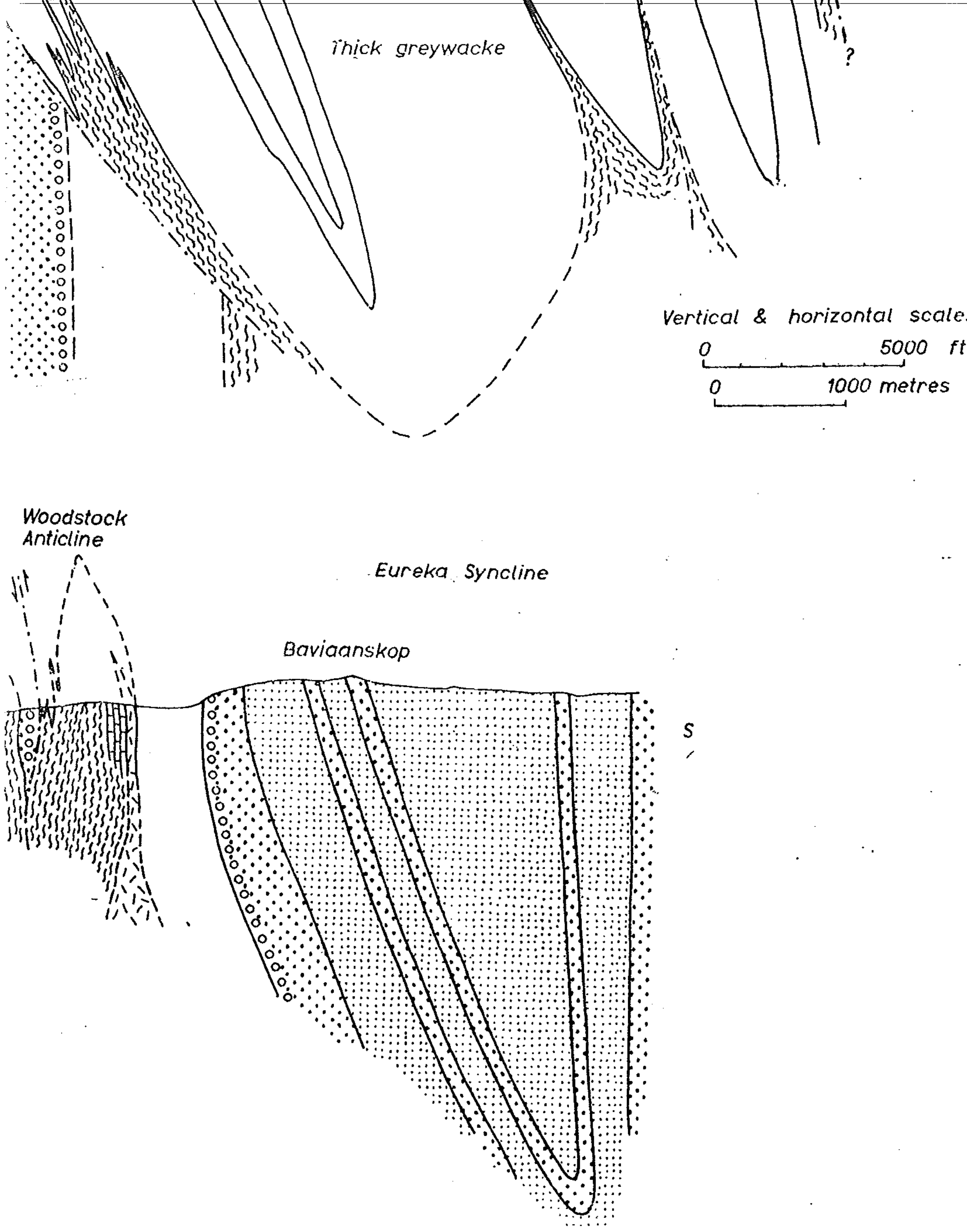


For location see Fig.2

Fig. 10

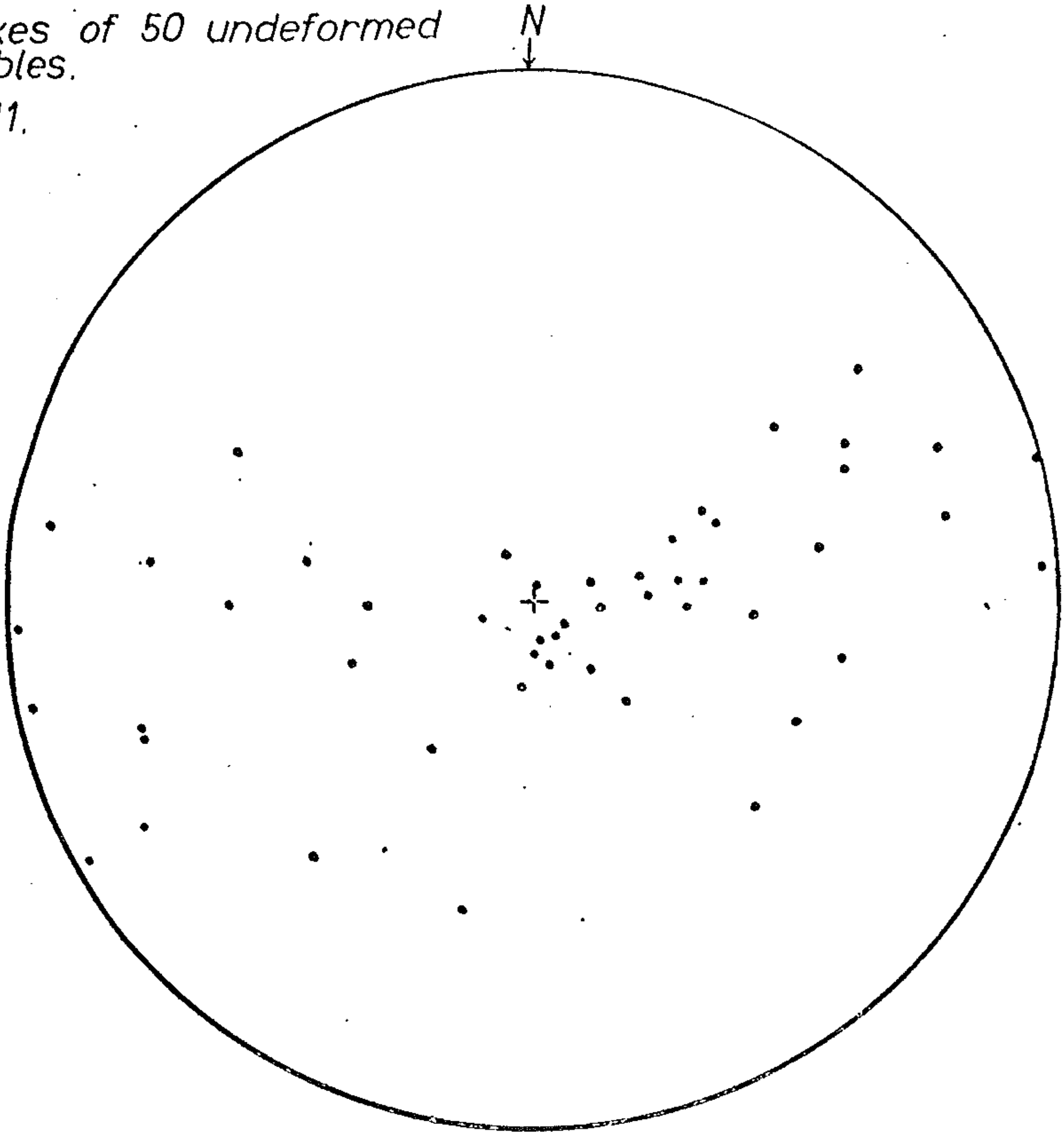


For location see Fig.2


Thick greywacke
?
Vertical & horizontal scale
0
5000 ft
0
1000 metres
Woodstock Anticline
Eureka Syncline
Baviaanskop
S

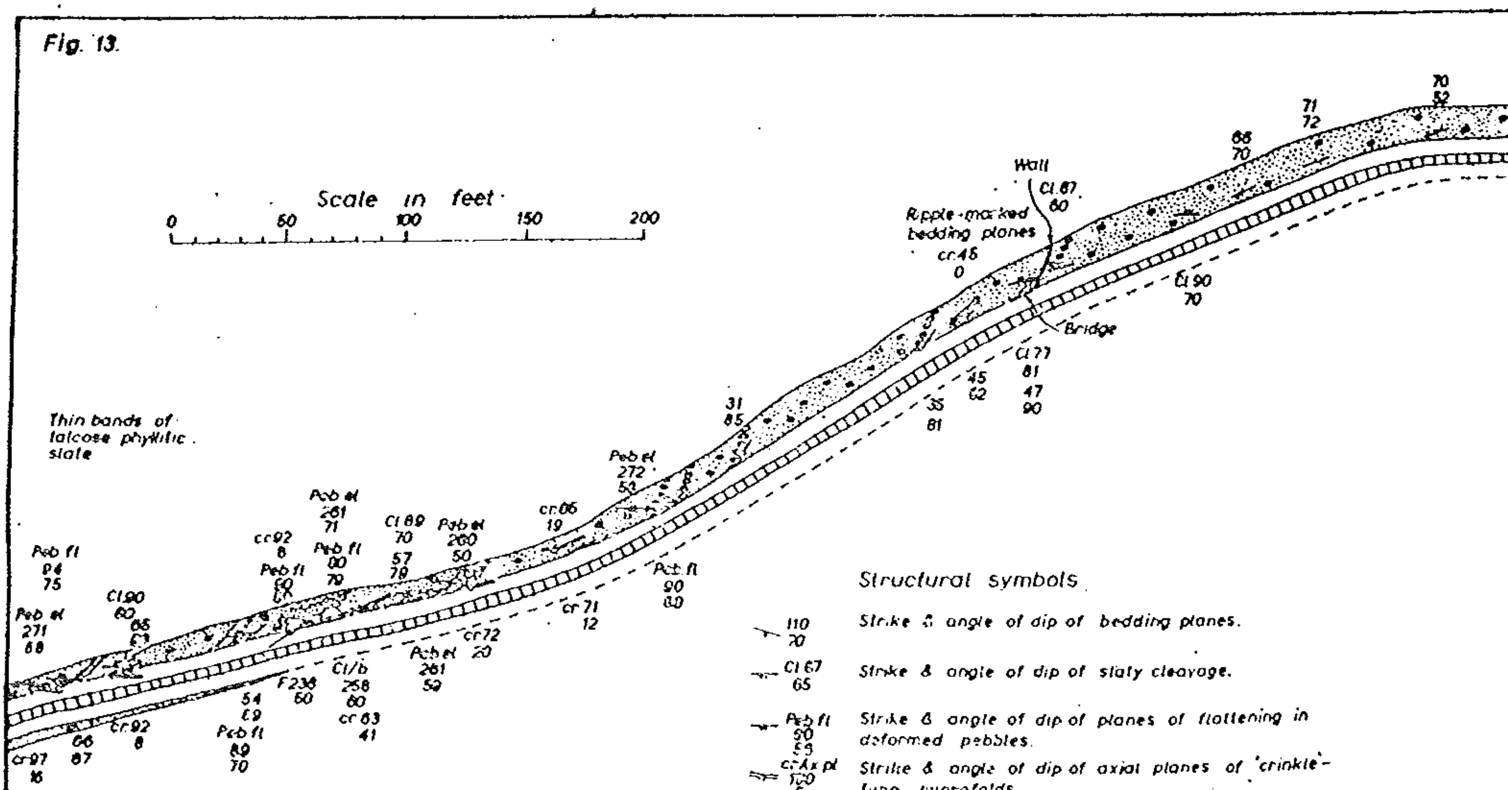
X Axes of 50 undeformed pebbles.

Fig.11.

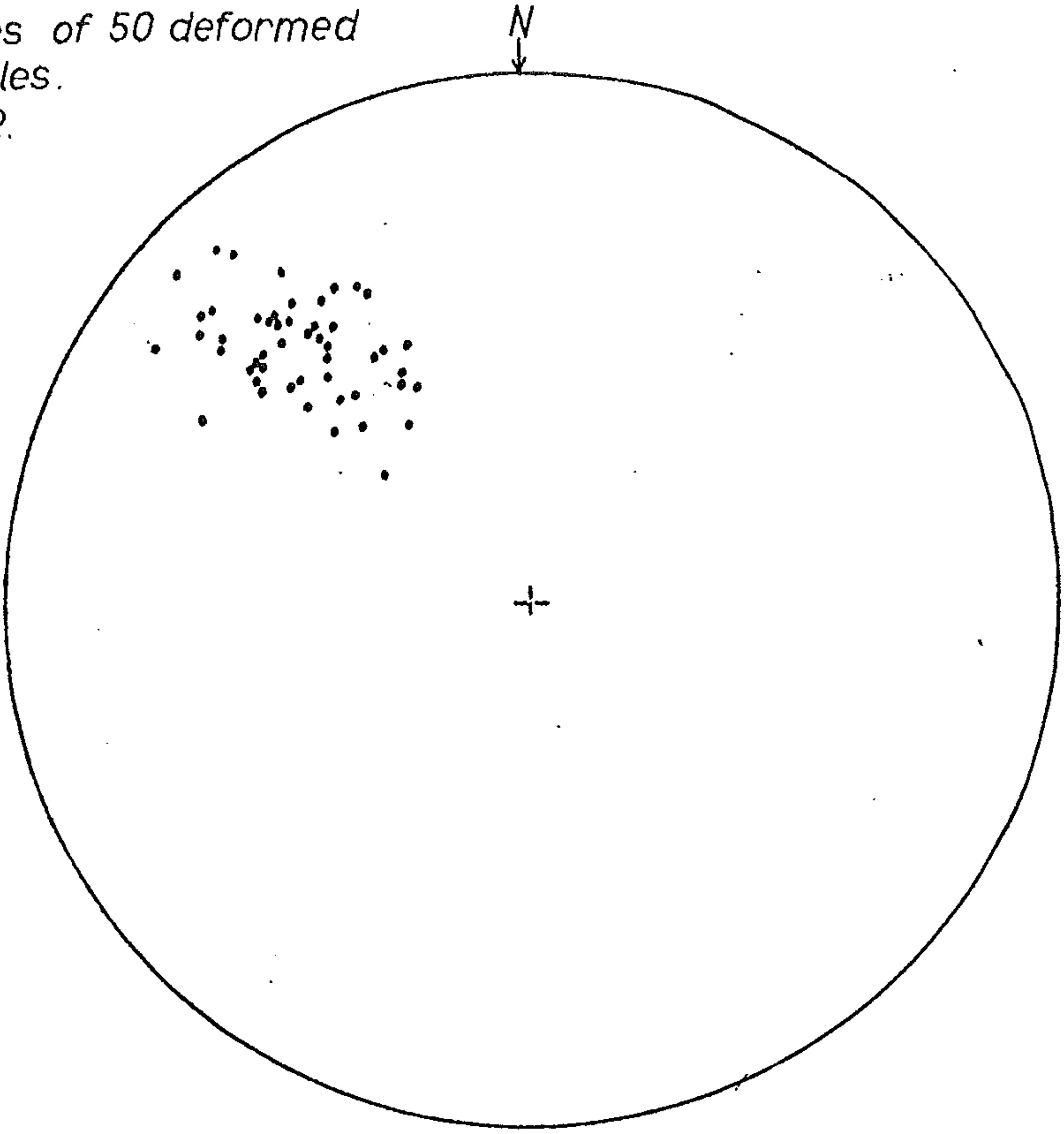


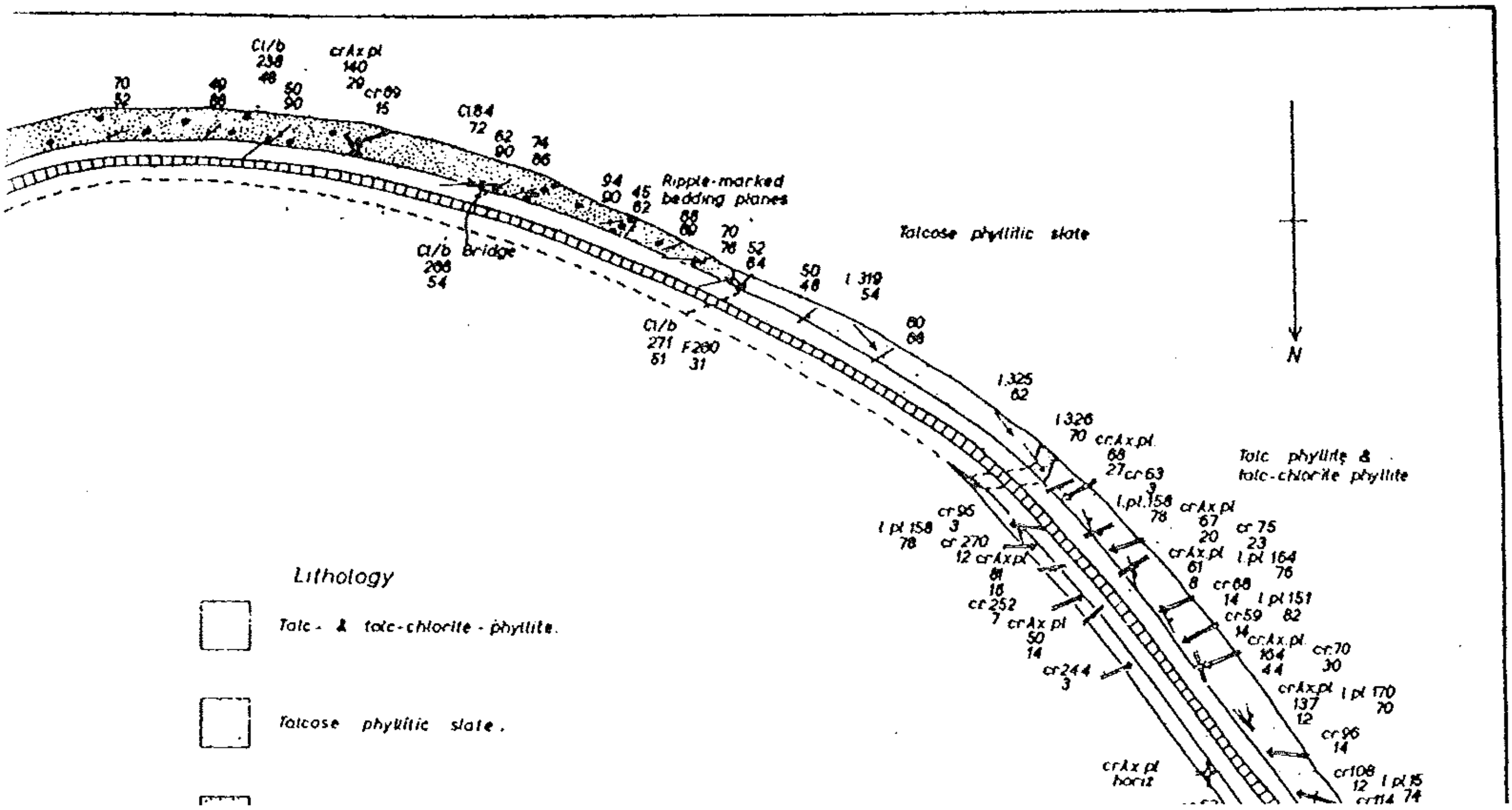
X Ax
pebb
Fig.1

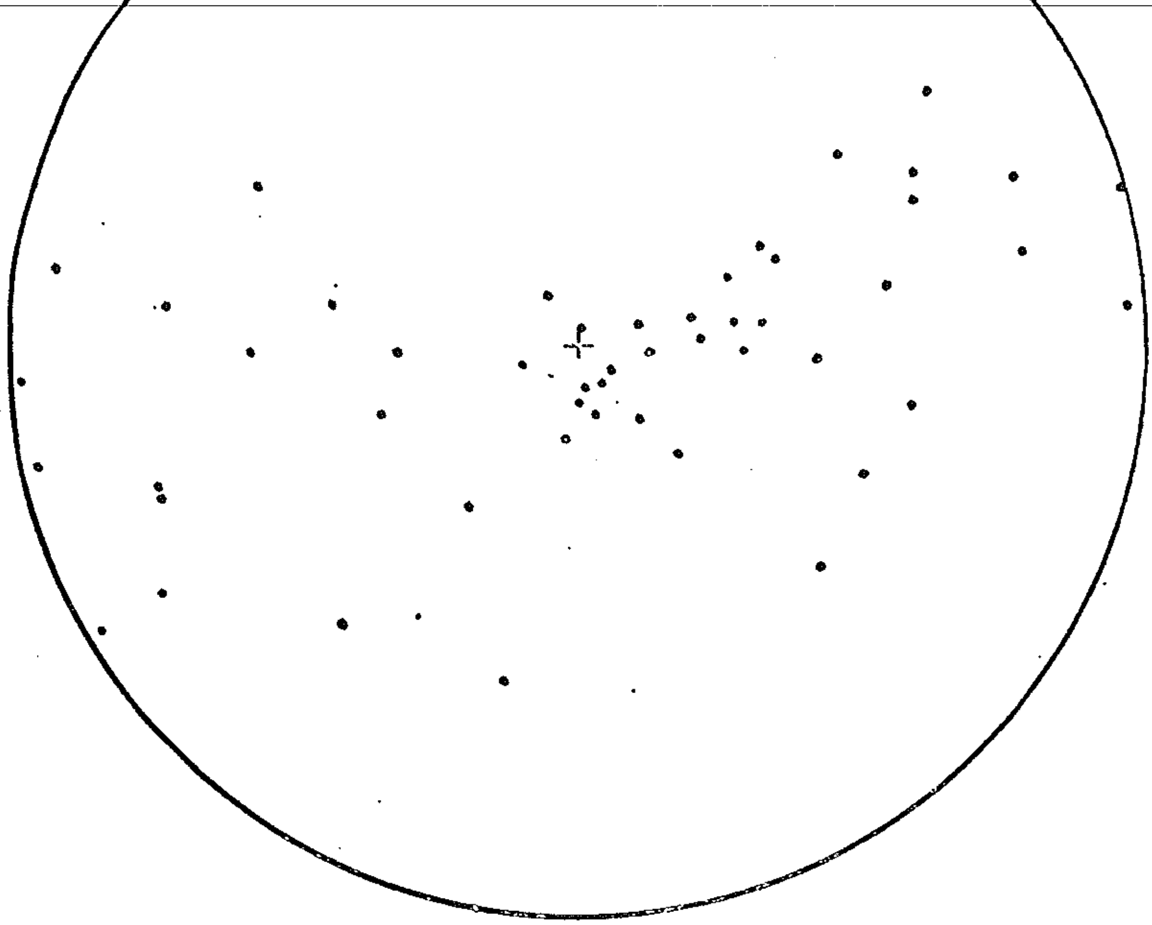
Fig. 13.



X Axes of 50 deformed pebbles.
Fig. 12.







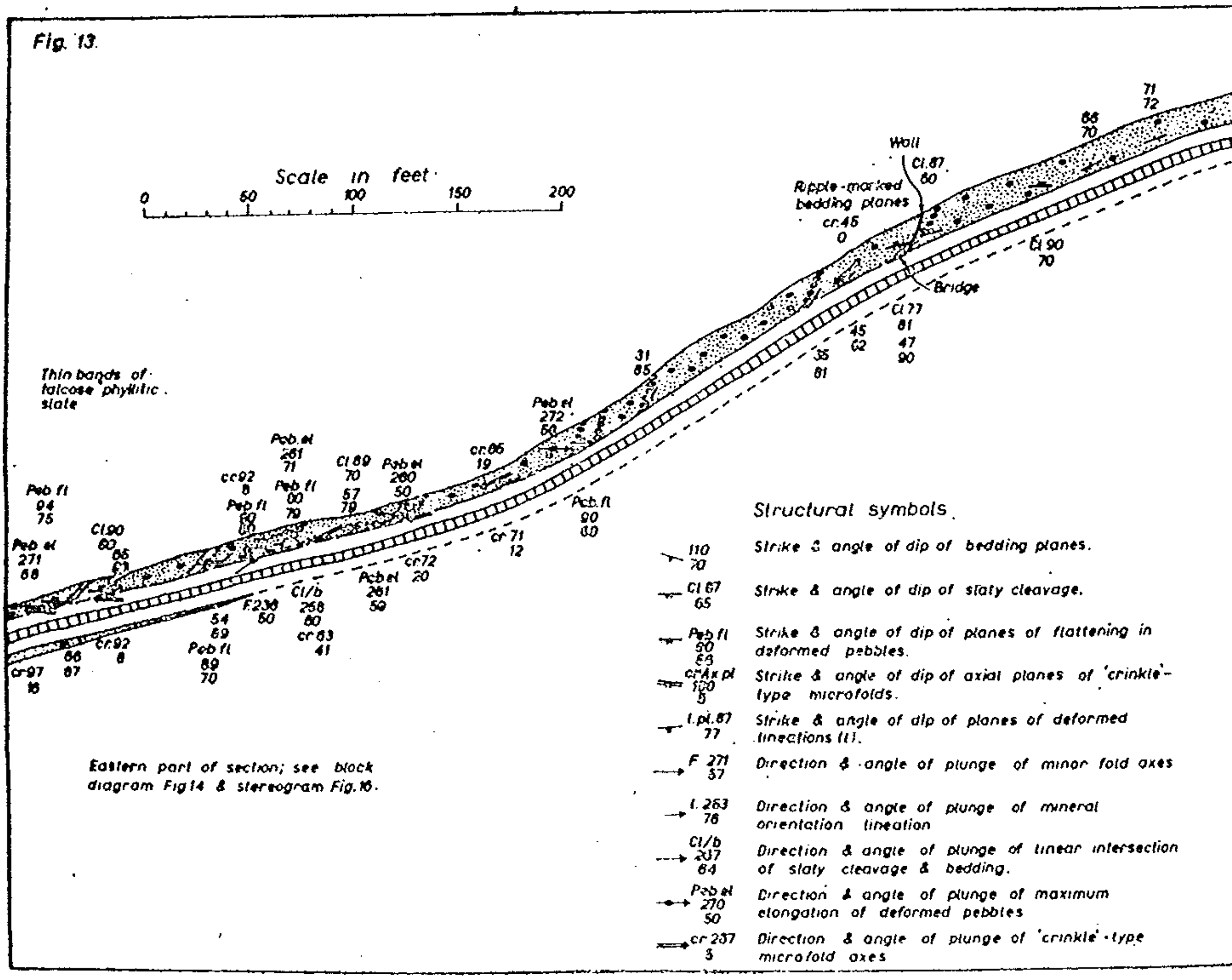
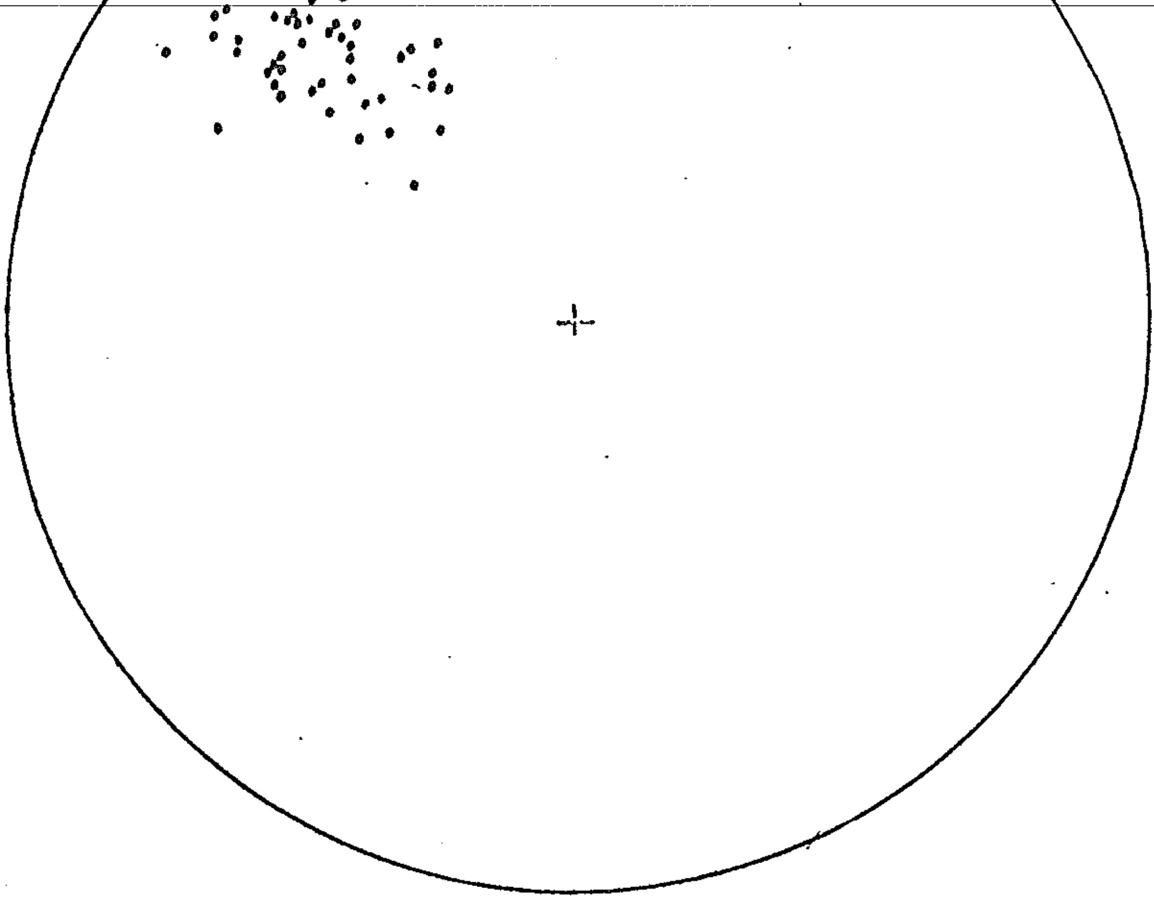
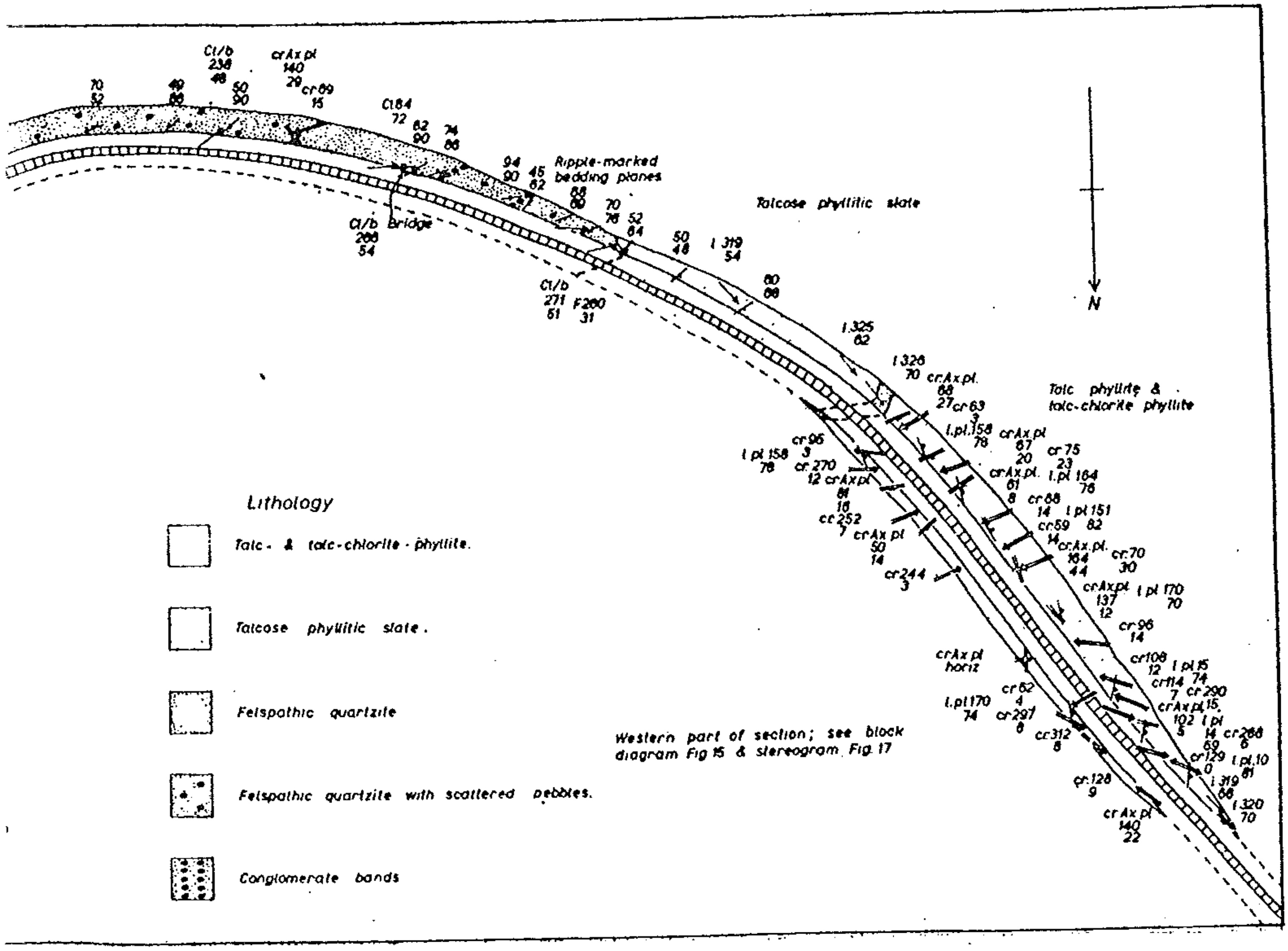


Fig. 13.
Scale in feet
0
50
100
150
200
Thin bands of talcose phyllitic slate
Wall
Ripple-marked bedding planes
Bridge
Structural symbols
Strike & angle of dip of bedding planes.
Strike & angle of dip of slaty cleavage.
Strike & angle of dip of planes of flattening in deformed pebbles.
Strike & angle of dip of axial planes of 'crinkle'-type microfolds.
Strike & angle of dip of planes of deformed lineations (l).
Direction & angle of plunge of minor fold axes
Direction & angle of plunge of mineral orientation lineation
Direction & angle of plunge of linear intersection of slaty cleavage & bedding.
Direction & angle of plunge of maximum elongation of deformed pebbles
Direction & angle of plunge of 'crinkle'-type microfold axes
Eastern part of section; see block diagram Fig 14 & stereogram Fig. 16.




Ripple-marked bedding planes
Talcose phyllitic slate
Talc phyllite & talc-chlorite phyllite
Cl/b Bridge
N
Lithology
Talc- & talc-chlorite-phyllite.
Talcose phyllitic slate.
Felspathic quartzite
Felspathic quartzite with scattered pebbles.
Conglomerate bands
Western part of section; see block diagram Fig 15 & stereogram Fig 17