

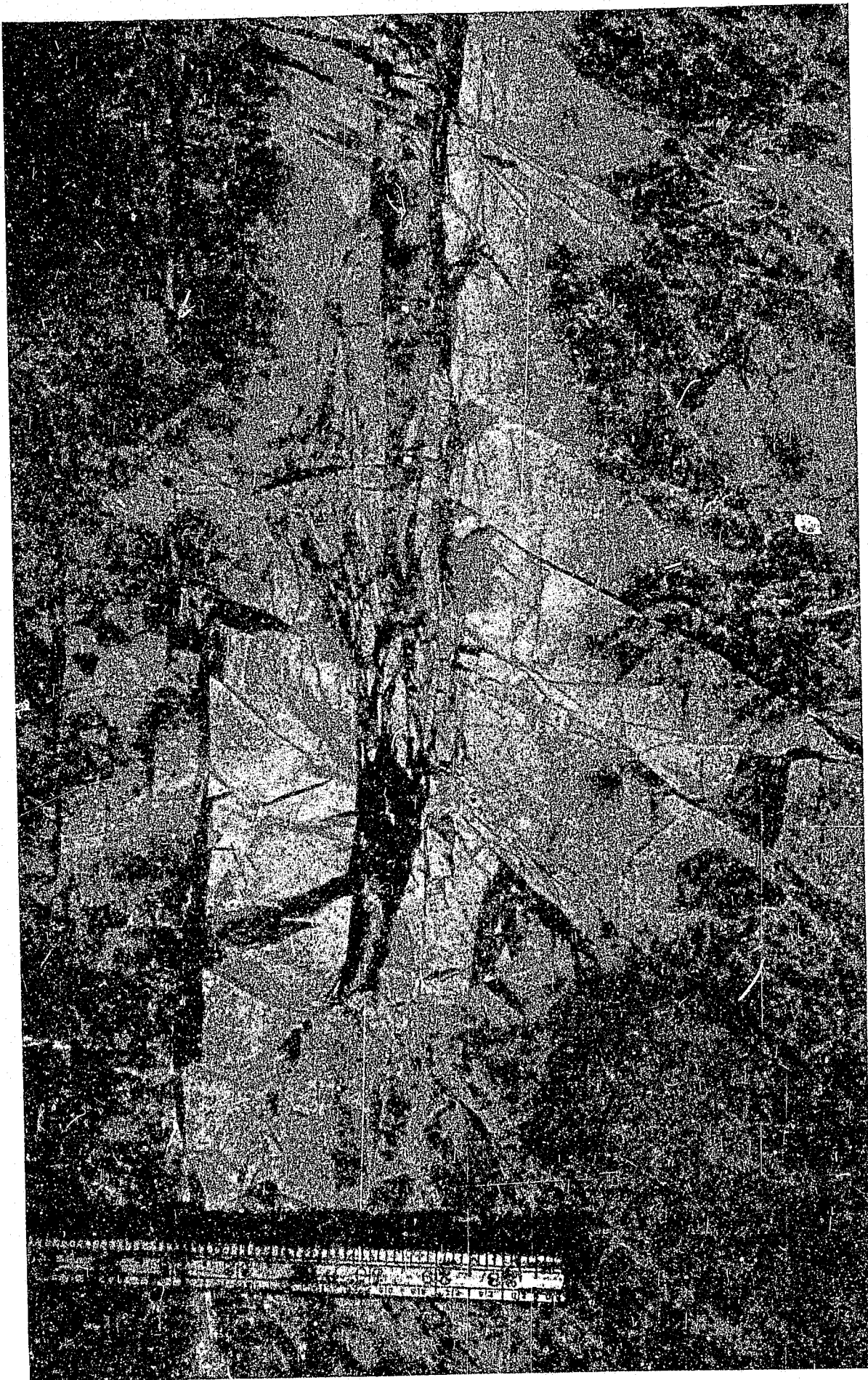


Plate 6.1

Alteration of host aplite adjacent to mineralised extension joints and breccias, towards the bottom of the joint the left hand sidewall has collapsed into the joint filling in a manner very much like the hydraulic brecciation described by Phillips (1972).

Figure 6.1

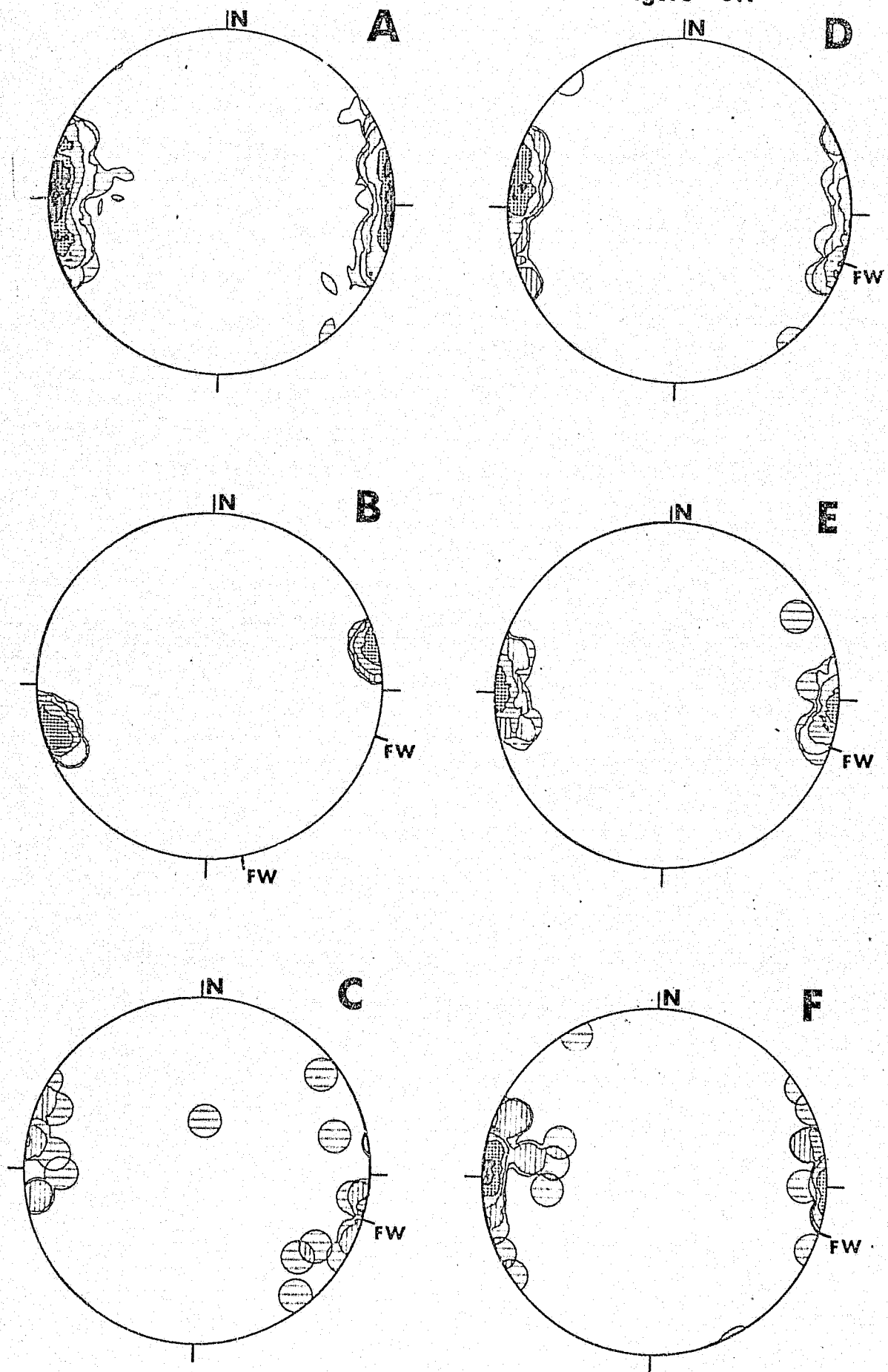


Figure 6.1

Contoured stereographic plot of the measured extension joints
in the aplite from various parts of the Lower Hercules section.

'FW' is the orientation of the footwall drive reference line.

A All measured joint readings

No. of readings	200
Maximum concentration	21%
Contours	1, 2, 5, 10, 15, 20% per 1% area

B Joints measured in No. 1 sill, H73E FWD

No. of readings	37
Maximum concentration	65%
Contours	3, 6, 12, 25% per 1% area

C Joints measured in No. 1 sill, H76E FWD

No. of readings	20
Maximum concentration	20%
Contours	4, 8, 20% per 1% area

D Joints measured in No. 3 sill, H73 FWD

No. of readings	50
Maximum concentration	46%
Contours	2, 4, 10, 20, 30% per 1% area

E Joints measured in No. 3 sill, H74 FWD

No. of readings	33
Maximum concentration	42%
Contours	3, 6, 12, 25% per 1% area

F Joints measured in No. 3 sill, H76W FWD

No. of readings	50
Maximum concentration	54%
Contours	2, 4, 10, 20, 30% per 1% area

Lower hemisphere, equal area projections
plane of projection is horizontal

steps are generally less than two millimetres high and the minor extension joints are impersistent once they no longer form part of the dominant joint.

The striations on the slickensided surfaces vary in orientation, indicating several phases of movement of the joint walls with respect to each other.

Late stage geological movement has been noted for one vertical extension joint, which has displaced a post jointing quartz vein by 4 centimetres. Normally displacements across joints result from dilation due to mining activity.

Joint orientations were measured with respect to the footwall drives in several regions of high joint density in the Hercules section. The results are shown in Figure 6.1, these stereographic projections show that the joints are almost all vertical and strike approximately north - south (005°), perpendicular to the drives and approximately perpendicular to the plane of the sills.

6.1.2 Chemistry

Plate 6.1 illustrates a dilational joint with a marked zone of alteration in the adjacent aplite. A chemical traverse was carried out across one of these zones, the results of which are listed in Appendix B. Briefly, they show a depletion in iron towards the joint, which correlates with the absence of feric minerals and a slight increase in sodium, which presumably reflects an increase in albite.

6.1.3 Origin of jointing

The close correlation between the joints and the sills and particularly the fact that the joints are confined to the central zone of the sills; as well as the strong preferred orientation, suggests that the joints might have formed by thermal contraction in an external stress field. If this is correct then the coefficient of thermal expansion for the central

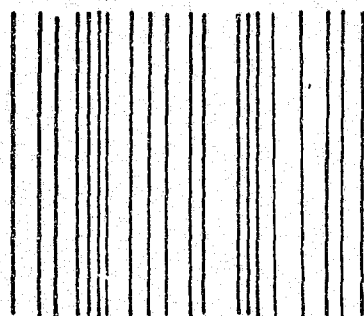
zone must be larger than that for the contact zone since the joints are restricted to the central zone. To check this, approximate thermal coefficients of volume expansion were calculated, using the determined mineralogy and the coefficients quoted in Skinner (1966). The mineralogy was based on the norms and the minerals known to be present; furthermore, the coefficient for mica was approximated by that of hornblende, and assuming random fabric elements using this data, a calculation was made for a temperature change from $800^{\circ} - 200^{\circ} \text{C}^*$ which showed that the contact zone had a larger coefficient than the central zone (2,34 - 2,16%). This makes a thermal contraction hypothesis unlikely for the origin of the joints, a conclusion supported by Jaeger's (1961) observation that jointing due to thermal contraction initiates at the contacts and propagates towards a molten core, which is not known in the sills.

Another hypothesis for the origin of the jointing is one based on the different rheological properties of the contact and central zones. The essence of this hypothesis is that the central zone should be easier to fracture than the contact zone. This situation is not supported by the present behaviour of the aplite, but Griggs et al. (1960) have shown that granite at high temperatures and 500 MPa confining pressure, has a drastically reduced strength which, together with the large temperature difference across the sills, might result in the contact zone being stronger during jointing and therefore the central zone should be easier to fracture. The work of Jaeger (1959, 1961, 1968) on the cooling of magmas indicates that these aplite sills would cool to 600°C in the centre within sixty days, if the magma intruded at 1000°C into country rock at 200°C . This imposes a restraint on the time available for the

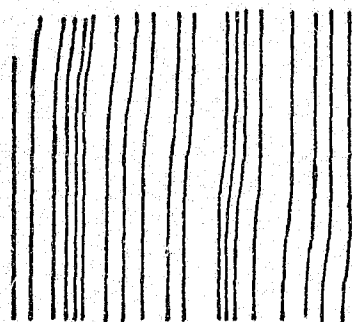
* It is felt that the geothermal gradient at the time of intrusion was higher than the present gradient on the Witwatersrand ($0,0119^{\circ} \text{C/m}$) (Lee and Clark, 1966). The value for Messina in the Transvaal ($0,0269^{\circ} \text{C/m}$, Lee and Clark, 1966) is regarded as a better estimate. The temperature at 9 kilometres would be 240°C .

POSSIBLE STAGES IN THE EVOLUTION OF THE SHEAR BRECCIA

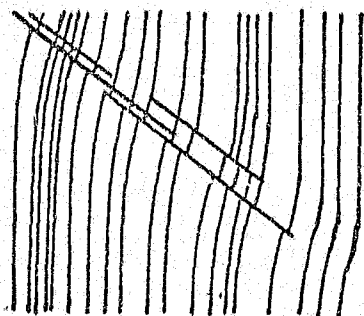
Figure 6.2



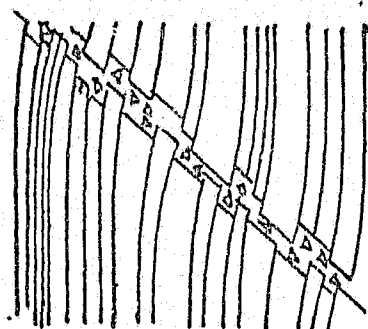
- 1 Aplite with numerous pre-existing extension joints



- 2 The principle load acting in the plane of joints causes some shearing and kinking



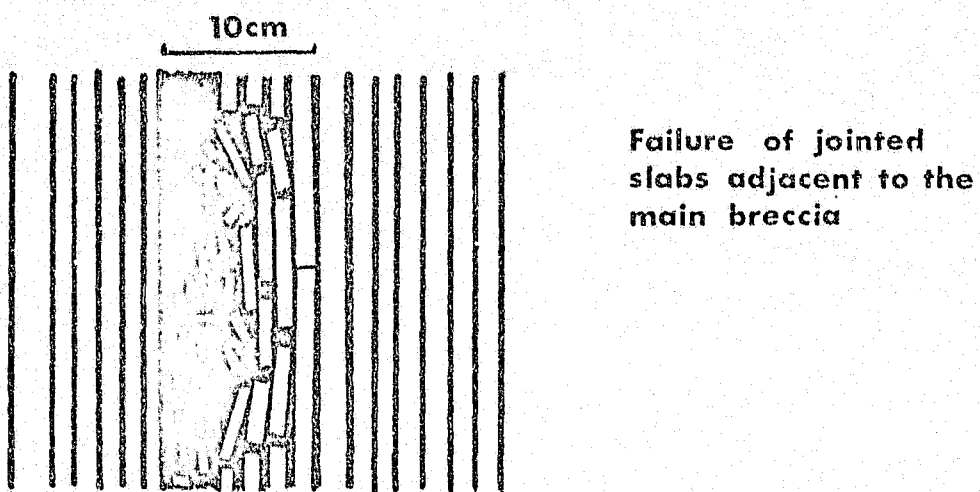
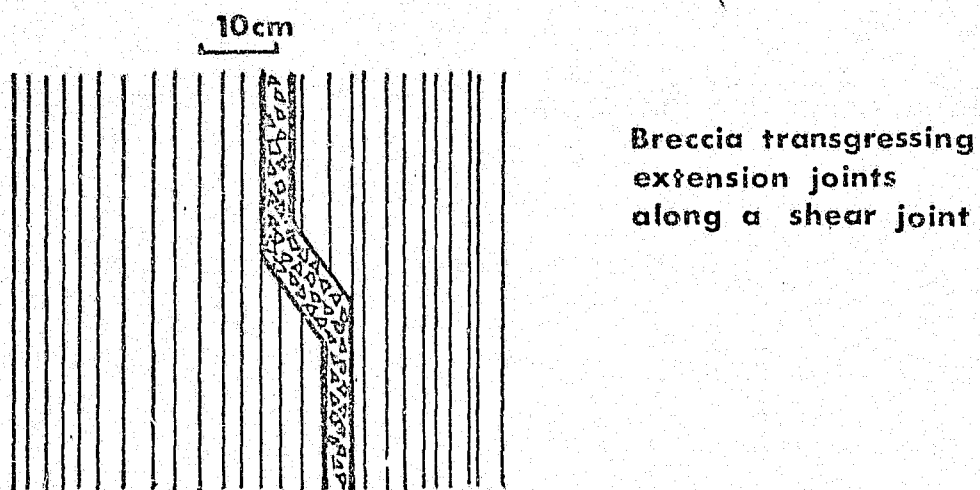
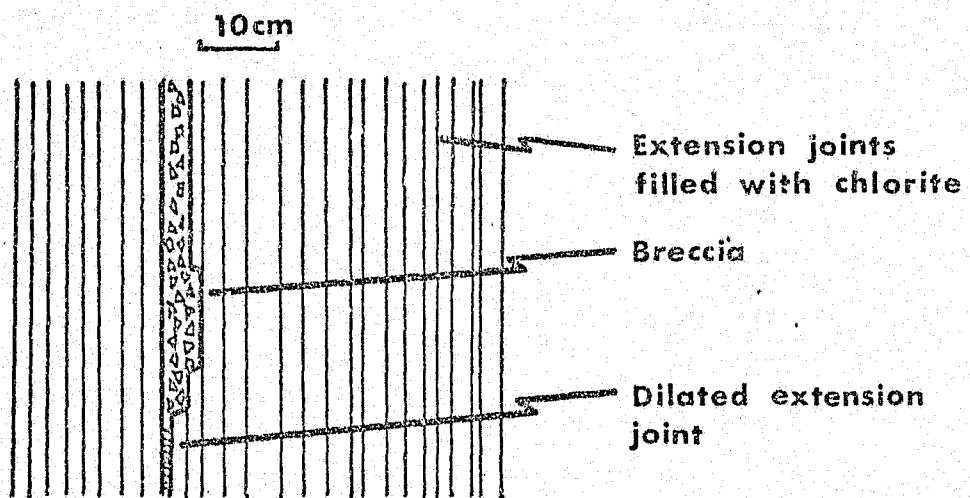
- 3 Development of a number of shear joints



- 4 Rotation and fragmentation of material between the extension and shear joints to form the breccia

TYPICAL FEATURES OF THE EXTENSION BRECCIA

Figure 6.3



jointing to occur at some stage when the contact zone is appreciably lower in temperature than the central zone, and so is stronger.

This suggests that the jointing could have taken place in the very early stages of cooling temperatures above 600°C , and probably involved sweat fluids derived from the cooling magma. These fluids would be pressurised and so would reduce the effective magnitude of the confining stress, resulting in the development of the joint by hydraulic fracturing, as described by Secor (1965), as well as precipitating the minerals in the joints. At the temperatures contemplated in this hypothesis, the adjacent quartzites would have been appreciably stronger than the aplite, their temperature having been only slightly increased by the intrusion, and so joints of the same orientation would be unlikely to form.

6.2 Brecciation

Two types of brecciation are seen in the sills; the more spectacular is always confined within the extension joints, and the other within shear joints.

6.2.1 Shear breccias

These breccias are formed when shear joints intersect extension joints with concomitant disruption and rotation of joint bounded blocks. A possible mechanism is outlined in Figure 6.2. Material within the breccia core is cemented together by chlorite and shows very little comminution. This suggests that little movement along the shear joints occurred after brecciation, and that brecciation must have involved considerable dilation.

6.2.2 Extension breccias

These breccias are found only in areas of high density extension jointing, where they occur as pods within the joint zone. Figure 6.3

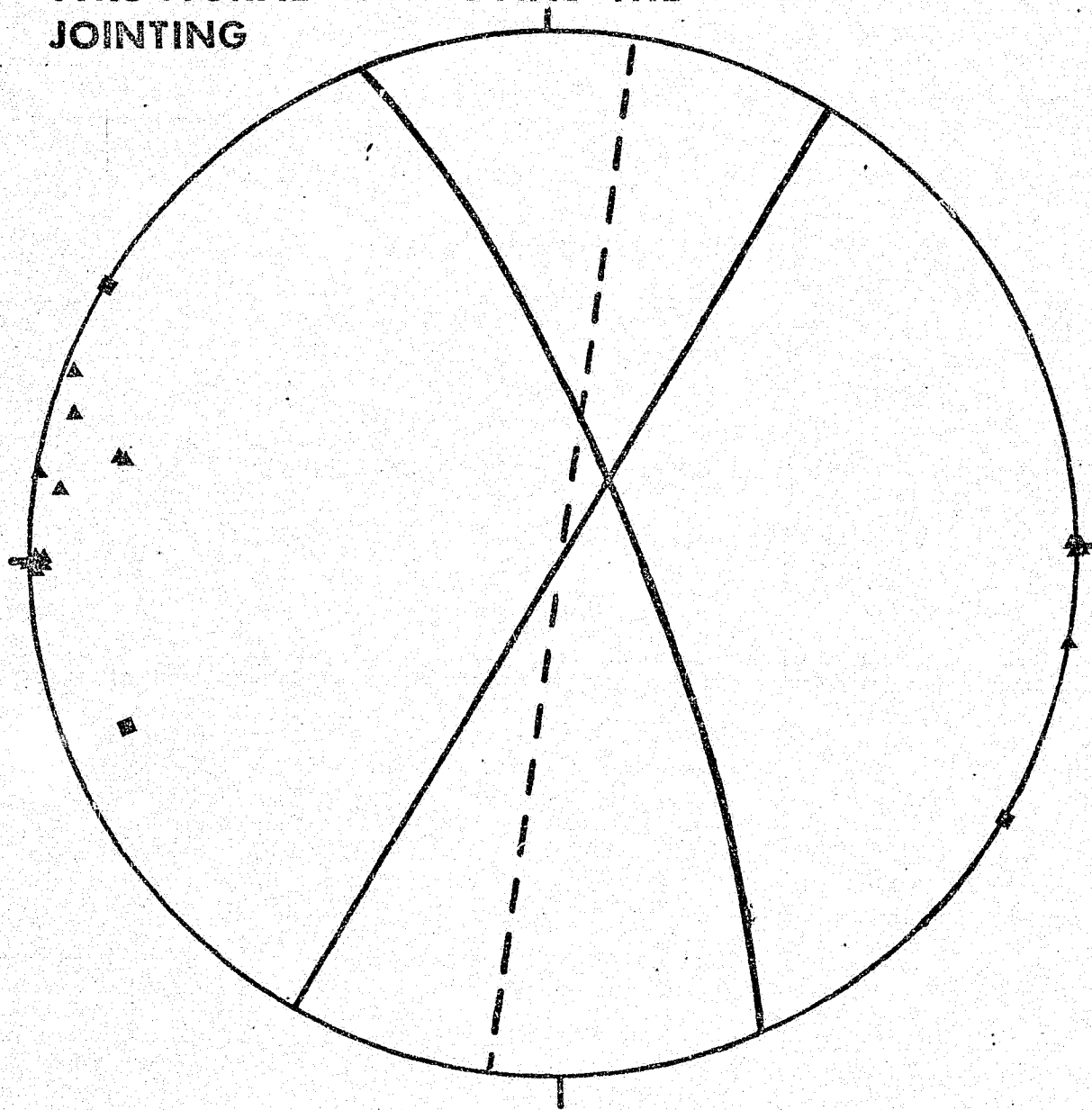


Plate 6.2

The structural fabric that is only known in Sill No. 3 and associated with brecciation. The acute angle between the sets of conjugate planes are bisected by the jointing. The width of the photograph is equivalent to 30 centimetres.

STEREOGRAPHIC PROJECTION
SHOWING THE RELATIONSHIP BETWEEN THE
STRUCTURAL FABRIC AND THE EXTENSION
JOINTING

Figure 6.4



■ Poles to fabric ——— Plane of fabric

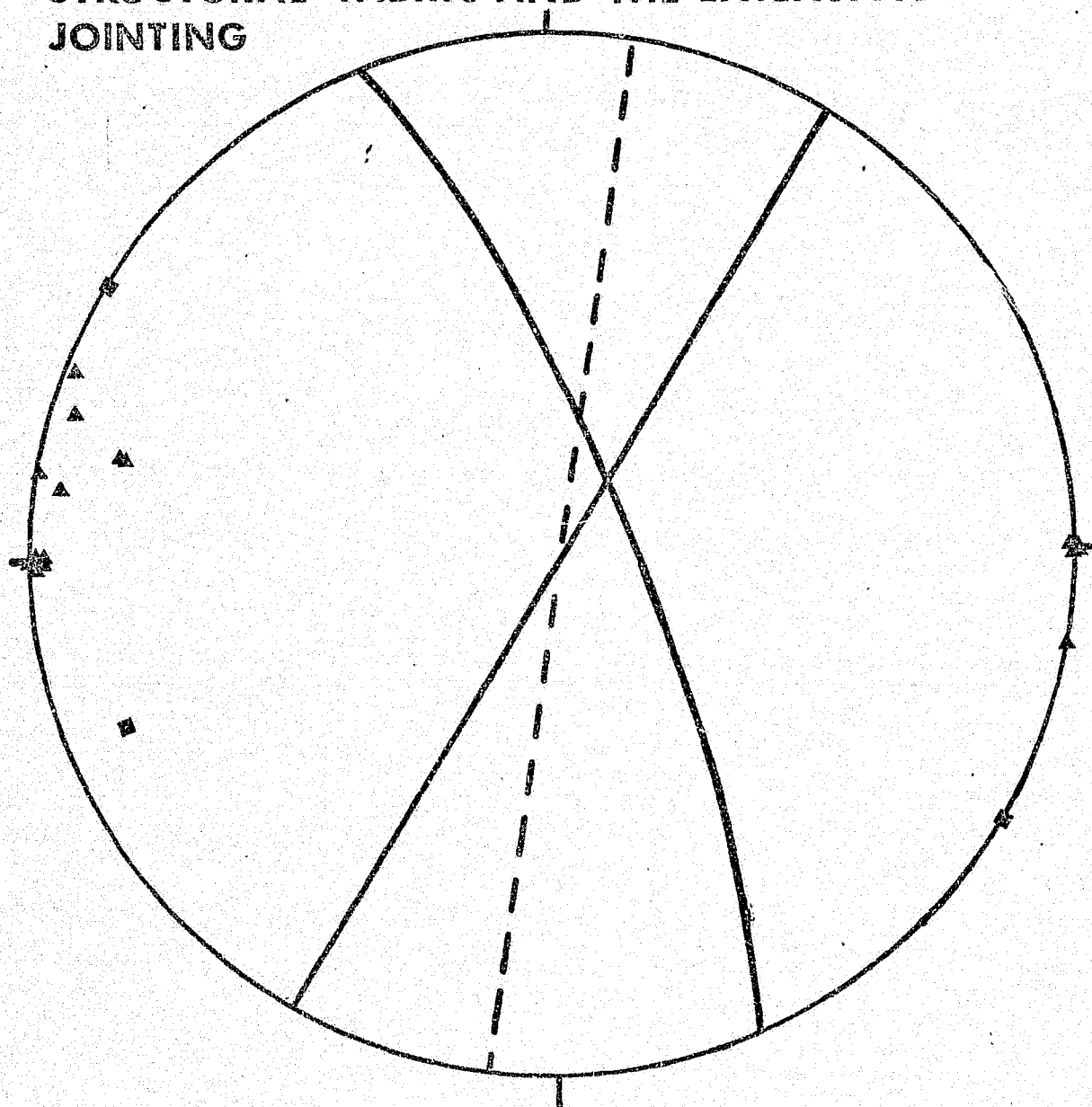
▲ Poles to joints - - - Plane of joints

The great circles were measured on an orientated specimen

Lower hemisphere, equal area projection,
 plane of projection is horizontal

STEREOGRAPHIC PROJECTION
SHOWING THE RELATIONSHIP BETWEEN THE
STRUCTURAL FABRIC AND THE EXTENSION
JOINTING

Figure 6.4



■ Poles to fabric



Plane of fabric

▲ Poles to joints



Plane of joints

The great circles were measured on an orientated specimen

Lower hemisphere, equal area projection,
 plane of projection is horizontal

illustrates typical features of these breccia zones. The breccia comprises angular fragments of aplite which show no sign of shearing or comminution. The spaces between these fragments are partially filled by euhedral crystals of quartz and chlorite and, in lesser amounts, the other minerals listed in Table 3.1. Because the voids are not completely filled, the rock is very friable. Considerable dilation accompanied brecciation and in at least one example (Figure 6.3) the adjacent jointed aplite has been disrupted with possible shortening in the plane of the jointing.

Plate 6.1 shows the side wall of a joint that has partially collapsed into the joint, which suggests that it might be formed by hydraulic brecciation as described by Phillips (1972), where the fluids associated with the mineralisation permeate the adjacent aplite and explode if the pressure in the joint drops.

6.3 Structural fabric*

6.3.1 Description

A strong structural fabric frequently occurs in the brecciated zones of No. 3 sill, where it appears as two sets of closely spaced intersecting planes which give the rock a cross-hatched appearance (Plate 6.2); one set is occasionally better developed at the expense of the other. Sometimes the fabric is bounded by the jointing but in general it is irregularly developed. The average angle between these planar sets is 67° (range 61° - 72°) and this acute angle is bisected by the extension joints, Figure 6.4. The intersection line of the planes plunges at $030/77^{\circ}$ azimuth and plunge.

Under the microscope the planes are seen to be formed by alternating zones richer in muscovite and biotite. The mica flakes in these zones do not show a preferred orientation.

*It has been pointed out to me that the term "structural fabric" is being used incorrectly in this section. The correct term for the structure described here is "planar features".

The texture is best developed in the central parts of the sill and is absent in the contact zone where an undisturbed primary flow structure is found. A similar but less well defined structure has been seen in a sample from City Deep Gold Mine. In this specimen the fabric is formed by mineralised zones, with chlorite along the planar sets. The average acute angle between the planes is $68,5^{\circ}$, range $60^{\circ} - 78^{\circ}$.

6.3.2 Origin

The close relationship between the joints and the planar sets suggests that they formed in the same stress field. If the extension joints form parallel to the $\sigma_1 - \sigma_2$ plane, then the planar sets probably correspond to closely spaced shear joints, very similar to the Lüders' bands described by Friedman and Logan (1973). The sporadic occurrence of the structure probably reflects inhomogeneities in the regional stress conditions, with the fabric only forming when the stress difference and confining pressure were suitable. In Friedman and Logan's (1973) experimental work on sandstone and Solenhofen limestone they found that the maximum principal stress bisected an angle between the planes that varied from $\pm 70^{\circ}$ at zero confining pressure to $\pm 120^{\circ}$ at 300 MPa. Their data have been supplemented by work on the aplite described in the next chapter. Using the curve presented in Figure 8.2 the effective confining pressure operating during the formation of the Lüders' band would be 30 - 40 MPa. As there is no distortion of the fabric near the jointing it is presumed that the fabric was formed before the jointing, very early in the life of the sills when the central zone was more ductile.

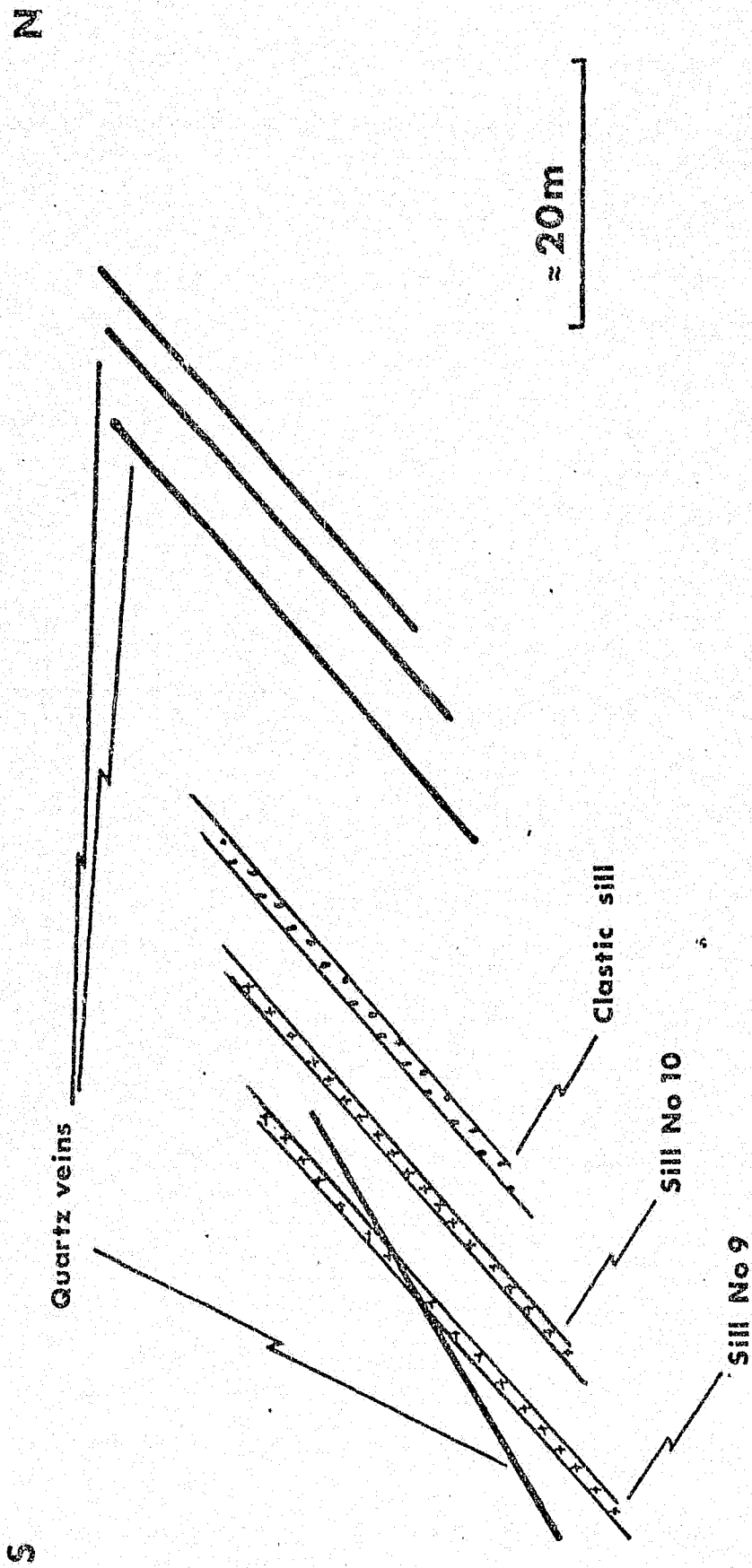
6.4 Quartz-chlorite, quartz-calcite veins

6.4.1 Description

There are two well exposed examples of quartz-chlorite veins, which from their morphology appear to be related tectonically to the sills.

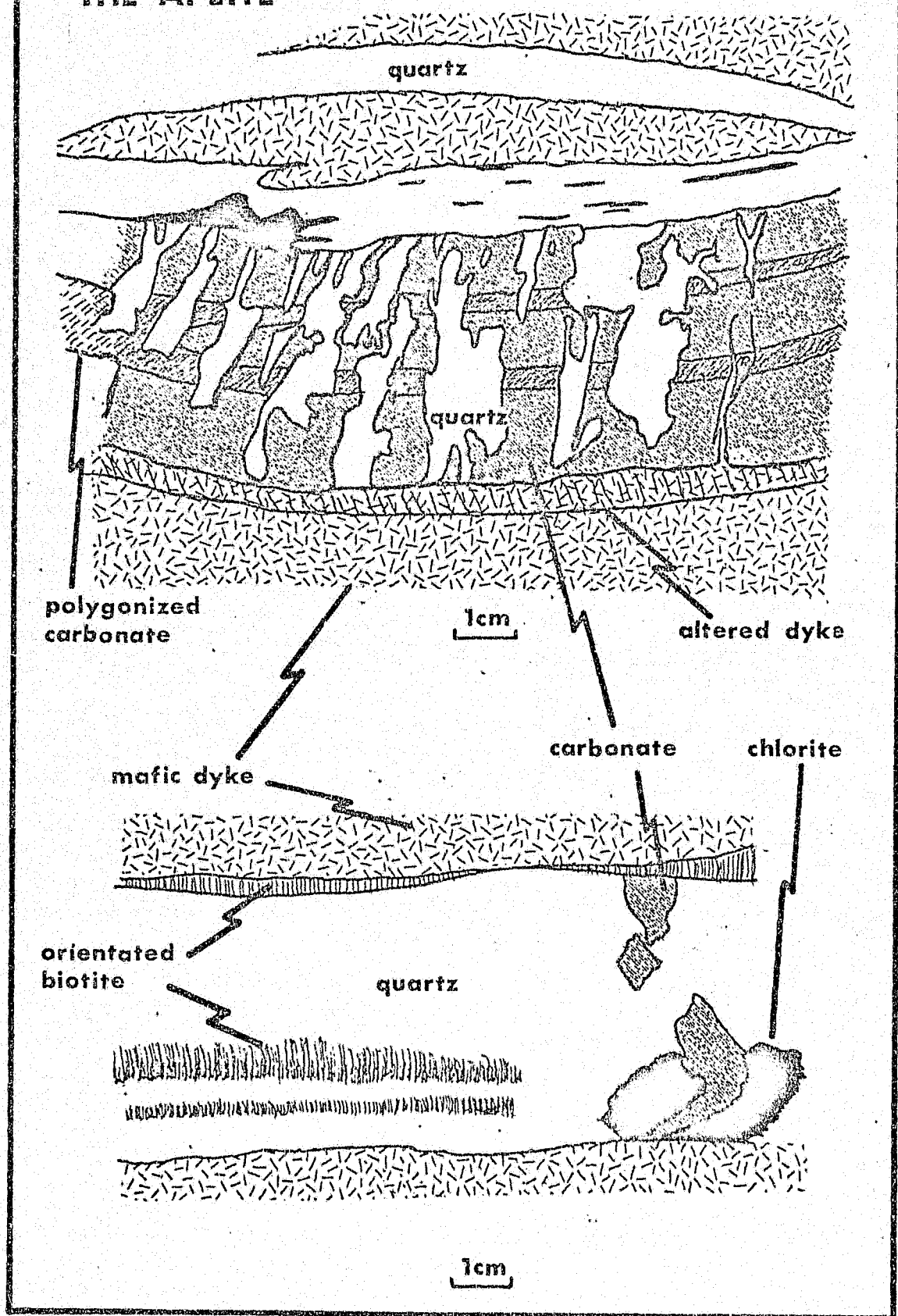
DIAGRAMMATIC SECTION SHOWING THE
RELATIONSHIP OF THE QUARTZ VEINS
TO SILLS 9 AND 10

Figure 6.5



**THE ORIENTATION AND
RELATIONSHIP OF THE MINERALS
IN VEINS ASSOCIATED WITH
THE APLITE**

Figure 6.6



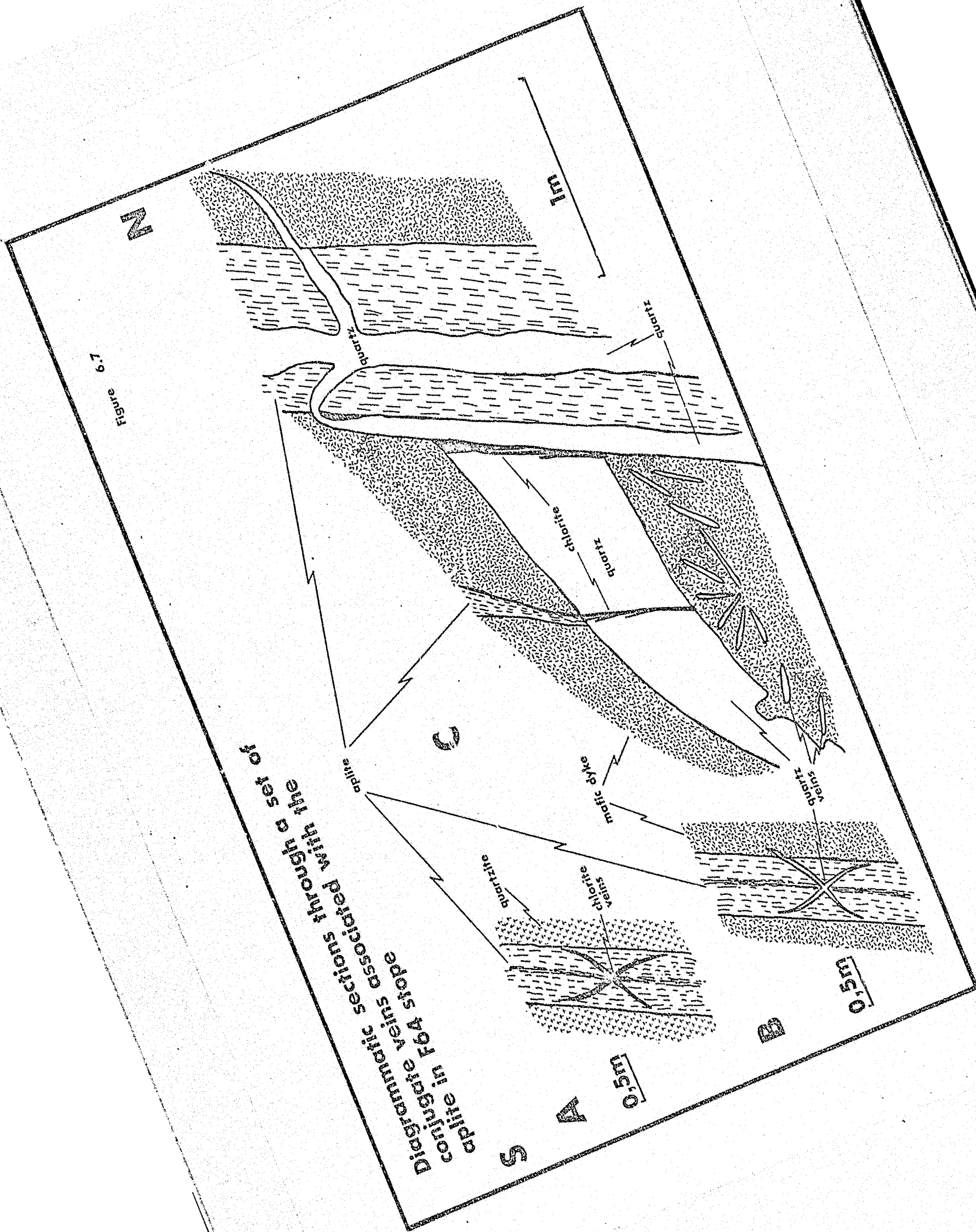
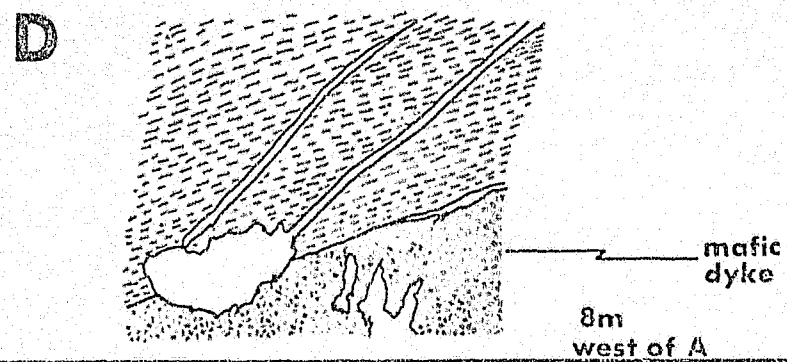
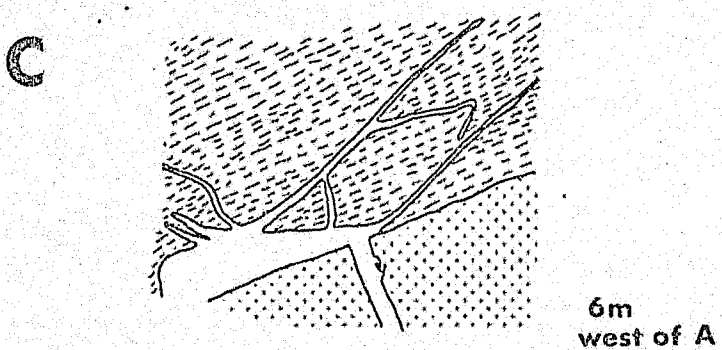
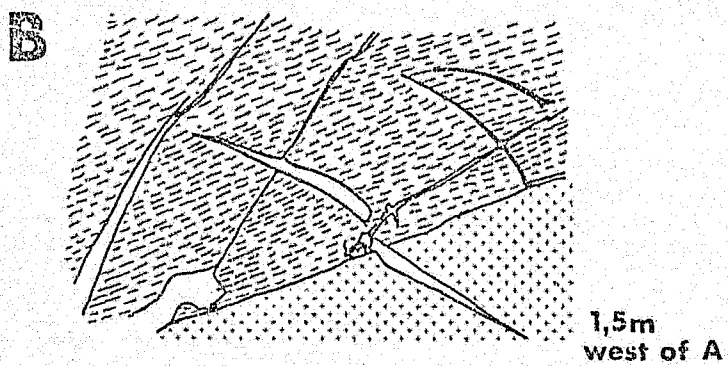
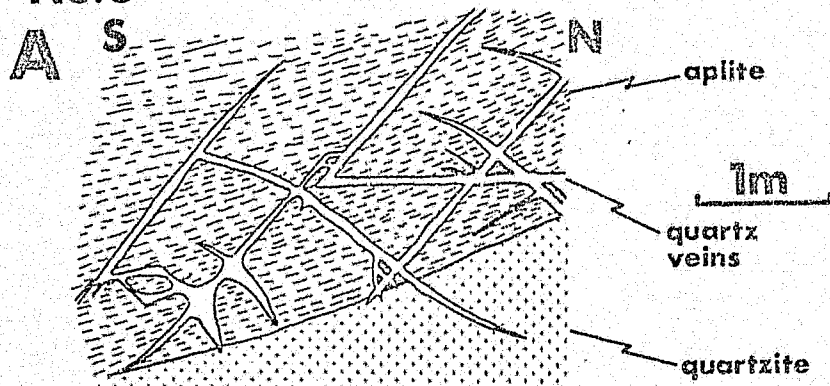


Figure 9

Mapped sections of a network of
quartz veins associated with sill
No.3

Figure 6.8



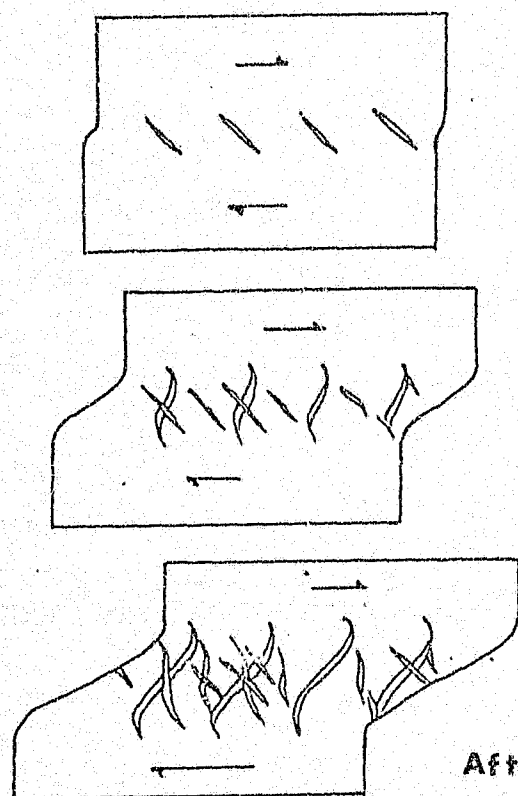
It is thought that they occur late in the history of the sills as they cut the extension joints. The two examples were from F63 - 64 south reef stopes and H76W footwall drive; in the former, the associated sills are Nos. 9 and 10, while the latter is associated with Sill No. 4.

Figure 6.5 is a diagrammatic sketch of the large scale relationships of the quartz veins in the F63 - 64 stope. Here, the quartz and quartz-calcite veins are thin (< 10 centimetres) sheet-like bodies. The quartz is massive and shows no fibrous relations to the walls. By contrast, the quartz-calcite veins are composite veins (c.f. Durney and Ramsay, 1973) in which the minerals show specific growth relationships to the walls, as illustrated in Figure 6.6. Smaller veins of quartz and chlorite occur in association with Sill No. 9. These were observed in the stope at various stages of face advance, as illustrated in Figure 6.7. The first exposure (Figure 6.7A) was of a pair of conjugate-like chlorite veins. These were only four or five millimetres thick, entirely confined to the sills, and the angle between them was bisected by the prominent median line of the sill. Later blasting revealed the network shown in Figure 6.7B, in which the conjugate veins are between 1 - 1.5 centimetres thick, essentially composed of quartz with a thin layer of chlorite on the walls of the veins. The tips of the veins curve towards median line of the sill, and one transgresses the contact. This vein is displaced along the contact but not disrupted. Finally after a longer time interval, a much more complex structure was seen (Figure 6.7C). This last exposure is similar to that seen on a much larger scale in H76W footwall drive which is illustrated in Figure 6.8 at various stages of tunnel advance.

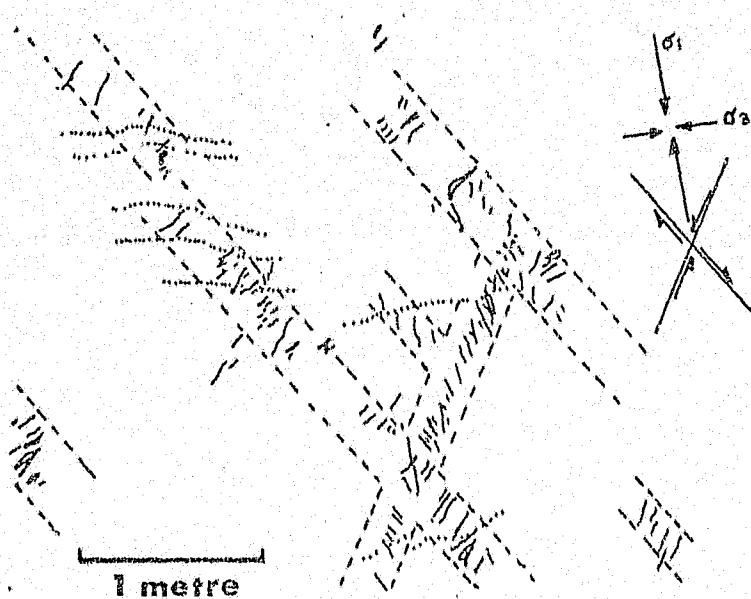
The basic 3-D form of these veins is a central quartz filled extension joint, and two symmetrically arranged conjugate quartz filled veins. The three veins tend to intersect along a common line which is horizontal east-west. The conjugate quartz veins are only planar near

**DEVELOPMENT OF QUARTZ
VEINS IN SHEAR ZONES
(AFTER VARIOUS WORKERS)**

Figure 6.9



After Durney and Ramsay
(1973)



After Roering (1968)

the intersection with the extension joint, but away from the intersection they tend to parallelism with the extension joint. This basic form is repeated a number of times, in whole or part, to account for most of the features seen in Figures 6.8 and 6.7C.

6.4.2 Origin

The significance of quartz veins with respect to the stress field operating at the time of their formation, has been discussed by several workers in terms of both dynamic and static conditions. Durney and Ramsay (1973) discuss the formation of a complex array of extension joints within a shear zone. In this mechanism, the central part of the joint rotates, but the tips continue to propagate in the same orientation until dilation of the joint ceases, after which new joints form, as illustrated in Figure 6.9. Roering (1968) also discusses the formation of extension joints associated with shear zones; he suggests that the complex joint pattern forms, not as a direct result of rotation, but as a result of various second order stresses associated with the shear zones, as illustrated in Figure 6.9. The propagation of veins out of the shear zone adds to the complexity of the system.

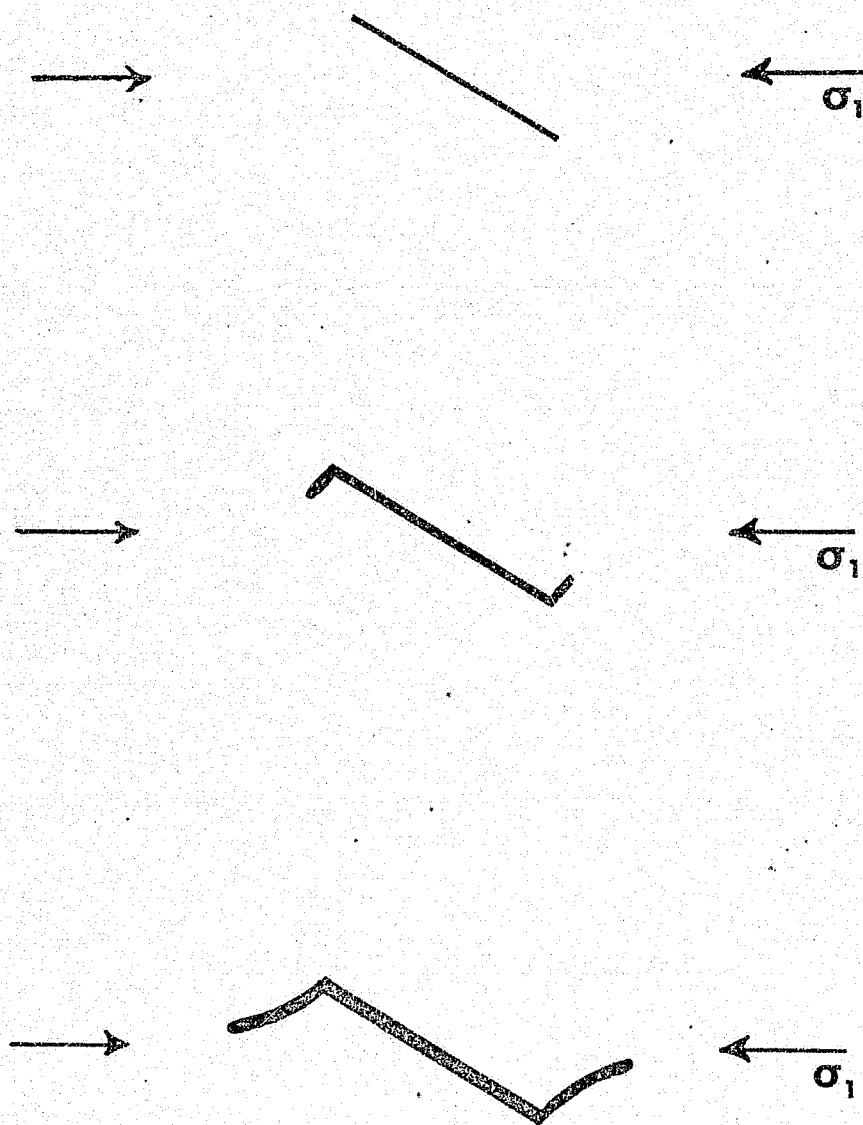
Shainin (1950) described veins also developed in shear zones which have an overall similarity to those described by Roering (1968), but have some aspects similar to those described by Durney and Ramsay (1973).

The explanation of the above workers hinges on the initial development of a shear zone; there is no suggestion of any such shear zone associated with the quartz veins in the aplite sills, although in the two well exposed cases, the veins are developed in and around the intersection of the sill with vertical mafic dykes. Moreover, neither the postulate of Roering (1968) or Durney and Ramsay (1973) explains the bilateral symmetry of the veins.

PROPAGATION OF A CRACK IN UNIAXIAL COMPRESSION

Figure 6.10

After Brace and Bombolakis(1963)

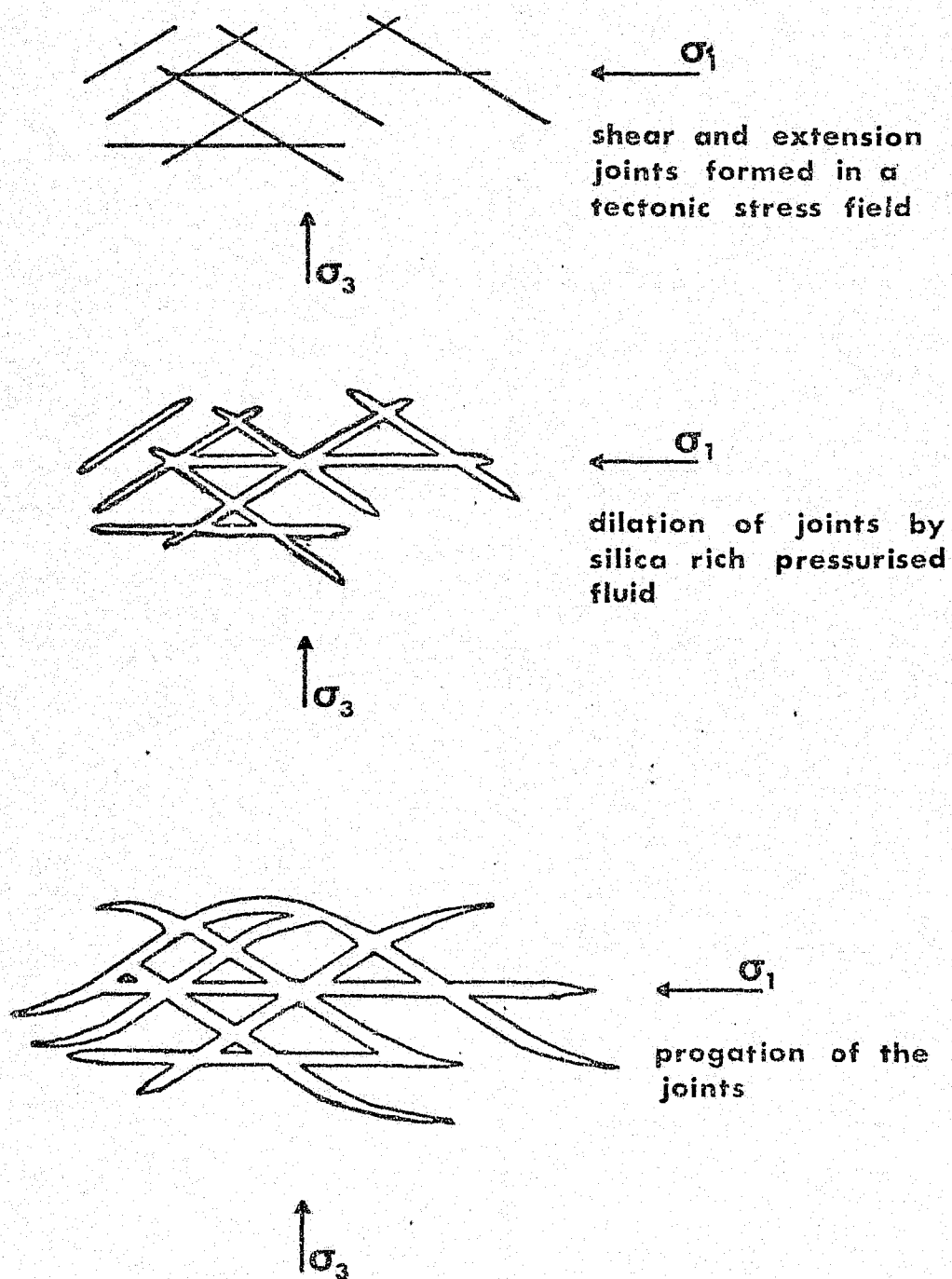


More promising is the work of Brace and Bombolakis (1963) who studied, experimentally, the propagation of a fracture initially inclined at an angle to a uniaxial compressive stress. In their system, fracture propagation does not start at the tip of a pre-existing fracture but at a point near it where the induced tensile stress is maximum. The orientation of the new crack rotates as it grows from a position normal to the tangent to the crack at the point of initiation, towards the maximum principal stress where the crack becomes 'stable' (Figure 6.10).

Instead of a dry crack subjected to increasing stresses as studied by Brace and Bombolakis (1963) and found as quartz veins in the field by Roering (1968), consider a crack filled with fluid under pressure in a block of material subjected to external static stresses. If the pressure in the fluid were sufficient to cause the crack to dilate, propagation would occur from the tip as in a normal Griffith crack in tension. Such a mechanism is advanced for the emplacement of intrusive dykes and sills (Jaeger and Cook, 1969, p.419; Pollard, 1973). Also, dyke swarms at Spanish Peak, Colorado, (mapped by Knopf, 1936) have a similarity in morphology to the quartz veins in E.R.P.M., in that dykes which are radial at the main volcanic vent, become parallel to the regional dyke trend at a distance. The classical analysis done by Odé (1957) on the dykes at Spanish Peak showed that they are parallel to the principal stress trajectories of a stress system containing the regional stress superimposed on the stresses around two volcanic pipes in which the pressure is greater than the maximum regional stress. However, the analogy between these dykes and the quartz veins is not exact because there is a greater density of dykes radiating from the stock at Spanish Peak, sub-parallel to the maximum regional stress direction; this is not found in the quartz veins. Also the quartz veins do not show a focus

ENVISAGED FORMATION OF ASSOCIATED QUARTZ VEINS

Figure 6.11



of origin. These differences can be explained by the initial fractures being part of an early set of shear and extension joints, formed in the regional stress field, into which the silica rich fluids intruded and caused the propagation of the veins, (Figure 6.11), until they became parallel to the far field σ_1 , and so formed the observed, spectacular vein-fracture patterns.

6.5 Folding

Gentle folding of the strata appears to have taken place after the emplacement of the sills. This is indicated by the steeper dips (50°) of the sills in the shallower levels, as seen in the shaft sections (Figure 2.2), where the folding is about an E-W axis.

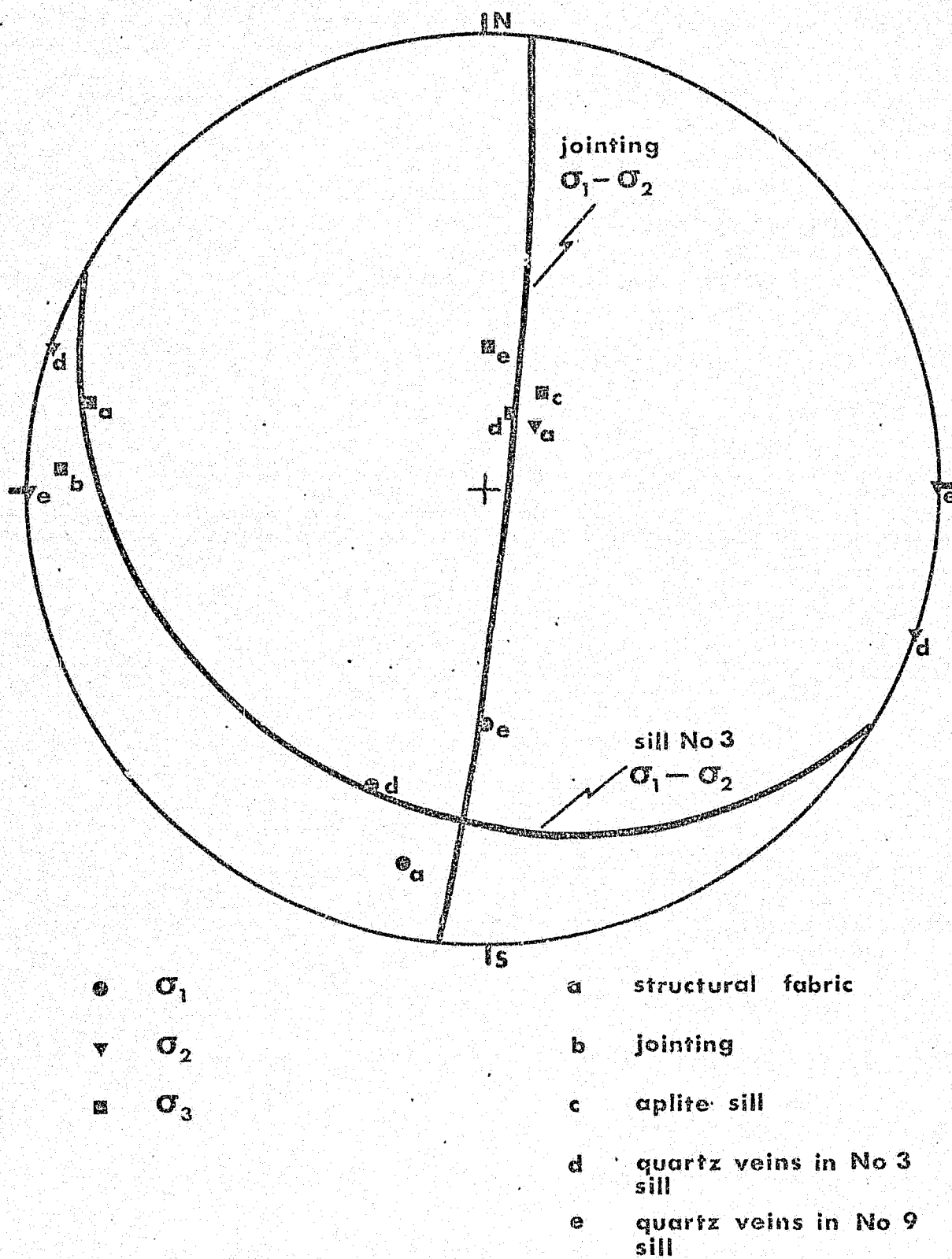
6.6 Evolution of the stress field

The tectonic features described in the preceding sections, plus the morphology of the sills and their age can be used to determine the orientation of the stress field and possible magnitudes of the principal stresses.

The features that can be used are the jointing, the structural fabric, (ie. ^{hides} Bands) the quartz veins and, of course, the attitude of the sills themselves. It is generally accepted that intrusions occur in a surface whose normal is the minimum principal stress (Anderson, 1938; Odé, 1957); which means the maximum and intermediate stresses are in the plane of the sill. The earliest feature recognised in the sills is the structural fabric which gives a good stress determination. According to the work of Friedman and Logan (1973) the intersection of the two planar sets corresponds to σ_2 , while σ_1 or σ_3 can bisect the acute angle, depending on the effective confining pressure. If the direction of σ_1 or σ_3 is known the effective confining pressure can be determined, and if the effective confining pressure is known, the orientation of σ_1 and σ_3 can be found. Jointing

ORIENTATION OF THE TECTONIC STRESS FIELD ASSOCIATED WITH THE EMPLACEMENT OF THE APLITE SILLS

Figure 6.12



is the next structural feature developed in the sills, but because the three dimensional shape of the shear joints are not fully known, the 'tectonic cross' described by Price (1966, p.110) cannot be used to determine the complete stress orientation, only the orientation of σ_3 normal to the extension joints is given. The last structural features known to be associated with the sills are the quartz veins, whose inferred development has already been discussed. The orientations deduced for the stress field from these features, are shown in Figure 6.12.

The attitude of the sills shows that the early σ_3 was approximately normal to the bedding and σ_1 and σ_2 were approximately in the plane of the bedding. Because the jointing bisects the acute angle of the structural fabric, σ_1 has been plotted bisecting the acute angle. As such, it is near horizontal in a north-south direction. σ_2 is approximately normal to the bedding and σ_3 is horizontal east-west. The orientation of σ_3 from the jointing is in good agreement with that determined from the structural fabric, and the intersection of the great circle for the jointing, with the great circle for the sill is near to the position of σ_1 determined from the structural fabric. The quartz veins show a greater scatter but σ_1 is approximately the same direction as the other features with σ_2 horizontal east-west and σ_3 normal to the bedding.

From this, the evolution of the stress orientation can be deduced, for a short time interval after intrusion. In all cases, σ_1 is near horizontal aligned north-south in the plane of the bedding, with only σ_2 and σ_3 changing. At the time of intrusion, σ_3 was approximately vertical and σ_2 horizontal east-west, but the structural fabric, which is thought to have formed shortly after the magma had crystallised, shows σ_2 approximately vertical and σ_3 horizontal; i.e. σ_2 and σ_3 changed positions during this short time interval. Similarly, the jointing shows the same stress

orientation as the Luders bands but the quartz veins indicate a state of stress like that which occurred at the time of intrusion of the sills. Therefore, for a short time span, some additional factor affected the stress field to cause the inversion of σ_2 and σ_3 . This could possibly be a tensile component caused by the thermal contraction of the rock. Such a component has a uniform affect in all directions, but the shape of the sills will tend to confine its main influence to within the plane of the sills (Jaegar, 1961).

As regards the magnitudes of the principal stresses, during intrusion of the sills σ_3 can be estimated from the weight of the overburden at the time. In Chapter 4 it was concluded that the age of the sills might correspond to the end of deposition in the Transvaal times. So at the time of emplacement a thickness of about 9 kilometres of sediments and volcanics of the Upper Witwatersrand, Ventersdorp and Transvaal systems would have overlain the main reef horizons (based on data from du Toit, 1954). The stress due to the weight of these strata would be approximately 235 MPa assuming a density of $2,67 \times 10^3 \text{ Kg/m}^3$. The difference between σ_2 and σ_3 in magnitude would have been less than the thermal contraction tensile stress, otherwise these two stress directions would not interchange; and the difference between σ_1 and σ_3 must be greater so that σ_1 does not change. A very rough estimate of the magnitude of this thermal contraction stress can be obtained by multiplying the Young's modulus by the linear thermal contraction. Using a value of $0,8 \times 10^5 \text{ MPa}$ for the Young's modulus and a volume contraction of 0,6% between 1000° and 800° C (which is possibly representative of the temperature drop prior to jointing; a third of this volume contraction, calculated from a value of the coefficient of volume expansion in Skinner, 1965, approximates the linear contraction,) a tensile stress of the order of 160 MPa is generated. Therefore, the magnitude of σ_2 at the time of emplacement is

Table 6.1

Magnitude and orientation of the principal stresses

	<u>Magnitude</u>	<u>Orientation (azimuth and plunge)</u>
1. <u>Intrusion</u>		
Ambient σ_1	395MPa	182°/27°
Ambient σ_2	235 - 395MPa	273°/10°
Ambient σ_3	235MPa	028°/70°
Magma pressure	265MPa	-
2. <u>Lüders' Bands and jointing</u>		
Ambient σ_1	395MPa	180°/20°
Ambient σ_2	235MPa	030°/77°
Ambient σ_3	-	278°/8°
Pore pressure	200MPa	-
3. <u>Quartz veins</u>		
Ambient σ_1	395MPa	006°/30°
Ambient σ_2	235 - 395MPa	280°/0°
Ambient σ_3	235MPa	008°/70°

approximately between 235 MPa and 395 MPa, and σ_1 is greater than 395 MPa. Since the effective confining pressure after solidification is between 30 - 40 MPa using Friedman and Logan's (1973) results on the effect of confining pressure of the development of Liller's bands, the hydraulic pressure was approximately $\sigma_3 - 35$ MPa, i.e. 200 MPa. The magma pressure must be greater than σ_3 plus the tensile strength of the quartzites and shales (Jaeger and Cook, 1969, p.419), (30 MPa Hoek, 1964). Therefore the magma pressure is in excess of 265 MPa. These results are summarised in Table 6.1.

7. ROCK MECHANICS

This study was carried out to try and correlate the mechanical properties of the aplite with geology and its mechanical behaviour as observed underground.

7.1 Behaviour of the aplite in underground workings

In addition to the geological joints described in section 6.1, the fracture pattern of the aplite in the deep level workings on the mine is strongly controlled by the magnitude of the mining induced stresses. In the footwall and hangingwall drives the aplite is relatively unfractured compared to the adjoining sediments, being cut only by a few mining induced 'type 1' (defined in Chapter 6) fractures. In the stopes the sills are either extremely fractured or essentially unfractured compared to the adjacent country rock; especially important is the fact that when badly fractured, the aplite is very difficult to support if it is in the hangingwall of the stopes. Sometimes closely spaced shear planes, parallel to contact of the sills, occur in the immediate country rock within the stopes, with a lot of comminuted material within the shear planes.

Whether the aplite is fractured or unfractured depends on where the lower contact of the sill is, in respect to the roof of the stope. If the contact coincides with the roof, the aplite is generally only cut by a few type 1 fractures, spaced tens of centimetres apart (Sill No. 8, F69W stope, December 1973, Sill No. 6, top of G76W stope, September 1973). If the roof falls within the contact zone, the intensity of fracturing is similar to that in the quartzite (Sill No. 8, F70W stope, December 1973) and can easily be supported. However, the fracturing of the hanging is most severe as in large sills, Nos. 1 and 6, when the roof of the stope coincides with the central zone. Closely spaced fractures with a type 1 orientation form and then allow the aplite to continually

TABLE 7.1

The rock mechanical properties of the aplite mined by other workers.

Sill No.	C _o	σ_p	E	ρ	D	L	Comments	Reference
3	472 (30)	0	0,82 (0,02)*	2630	25	25	Average of 6	N.M.E.R.I. (1959)
8	476 (33)	0	0,84 (0,02)	2630	25	25	Average of 6	N.M.E.R.I. (1959)
8	503 (14)	0	0,83 (0,02)	2630	25	25	Average of 6	N.M.E.R.I. (1959)
8	502 (3)	0	0,73 (0,01)	2610	25	25	Average of 3	N.M.E.R.I. (1959)
197		-13			13	25	Dog bone	Hoek (1965)
605		0			13	25	Dog bone	Hoek (1965)
614		0			13	25	Dog bone	Hoek (1965)
814		78			13	25	Dog bone	Hoek (1965)
814		78			13	25	Dog bone	Hoek (1965)
1550		145			13	25	Dog bone	Hoek (1965)
1820		241			13	25	Dog bone	Hoek (1965)
2300		287			13	25	Dog bone	Hoek (1965)
2430		325			13	25	Dog bone	Hoek (1965)
605		0		2614	13	25	Dog bone	Brace et al (1966)

595	0	2614	13	25	Dog bone	Brace et al (1966)
720	77	2614	13	25	Dog bone	Brace et al (1966)
1380	143	2614	13	25	Dog bone	Brace et al (1966)
1530	238	2614	13	25	Dog bone	Brace et al (1966)
1980	283	2614	13	25	Dog bone	Brace et al (1966)
2070	320	2614	13	25	Dog bone	Brace et al (1966)
572	0					Hoek (1964)
0	-35		13	25	Dog bone	Hoek (1965)
2	-22		13	25	Dog bone	Hoek (1965)
3	-30		13	25	Dog bone	Hoek (1965)
612	0		22	43	Cylinder	Hoek (1965)
567	0		22	43	Cylinder	Hoek (1965)
594	0		22	43	Cylinder	Hoek (1965)
558	0		22	43	Cylinder	Hoek (1965)
580	0		22	43	Cylinder	Hoek (1965)
582	0		22	43	Cylinder	Hoek (1965)
1082	58		22	43	Cylinder	Hoek (1965)
1165	63		22	43	Cylinder	Hoek (1965)
1235	66		22	43	Cylinder	Hoek (1965)
1280	68		22	43	Cylinder	Hoek (1965)

1290	63	22	43	Cylinder	Hoek (1965)
1340	71	22	43	Cylinder	Hoek (1965)

C_o Uniaxial compressive strength in MPa.

σ_p Confining pressure MPa.

E Young's modulus 10^5 MPa.

ν Poisson's ratio.

ρ Density kg/m^3

D Diameter of sample millimetres.

L Length of sample millimetres.

* Brackets encloses standard deviations

collapse as falls of hanging or in rock bursts of various magnitudes (H75E stope, March 1973, magnitude $2\frac{1}{2}$, McGarr et al., 1975). This unstable situation continues until the complete thickness of the sill has fallen out and more stable quartzite forms the hanging. When these type 1 fractures in the central zone approach geological joints, or encounter an inhomogeneity, they are refracted towards the normal of the plane of the inhomogeneity. Moreover, if the sill is thin (e.g. No. 8), jointing tends to be less severe, which also improves the stability of the hanging, possibly because the typical central zone material is not really developed.

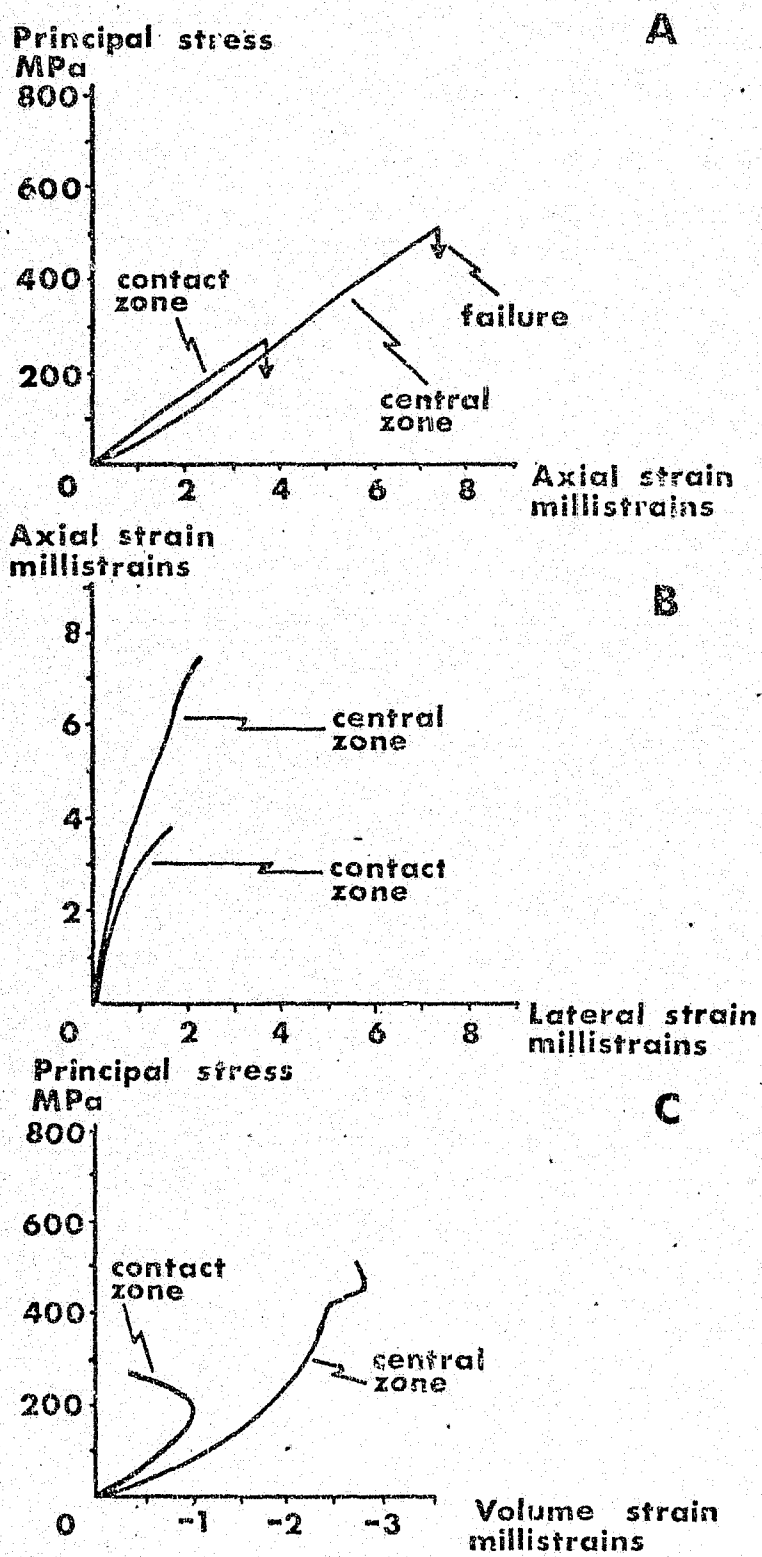
7.2 Experimental rock mechanics

The mechanical properties of the sills have been investigated in detail by a number of previous workers including the National Mechanical Engineering Research Institute (N.M.E.R.I.), (1959); Hoek, (1964, 1965); Brace et al. (1966) and Jaeger and Cook (1969, p.146) all of whom reported brittle behaviour of the aplite. The work of N.M.E.R.I. (1959) was carried out on four samples from Sill Nos. 3 and 8 but the localities of the material used by Hoek (1964, 1965) and Brace et al. (1966) is unknown. The results of these workers are listed in Table 7.1 (compressive stresses are positive). They are used here to complement the present work.

All samples were taken from boreholes drilled into the footwall from overstoped ground, except those cored in the laboratory from specimens F76C and G76C. The core from the three boreholes was of good quality in the quartzite but deteriorated in the aplite, where it was often broken into 5 - 10 centimetres long pieces or cut into discs 1 - 2 centimetres thick. The longer pieces of core were used to prepare test cylinders with length to diameter ratio 3 : 1. Wherever possible, samples were taken to cover the thickness of the sills. The blocks F67C and G76C were taken from the lower contact of the sills. The axial and circumferential strains were normally recorded independently using pairs of strain gauges

TYPICAL DEFORMATION CURVES FOR CENTRAL ZONE AND CONTACT ZONE MATERIAL

Figure 7.1



mounted parallel and perpendicular to the axis of the cylinder and wired in series. However, in one test, one axial and two lateral gauges were connected in series to record volume strain directly. The gauges were generally 10 millimetres long, but for the specimens taken from sample F69C the circumferential strain gauges were 1 millimetre in length. All tests were carried out in a soft compressive testing machine; strains were recorded manually every 20 kN of applied load on a Huggenburger tepic indicator, or else the load strain curve was plotted directly on an X-Y recorder.

7.3 Results of laboratory experiments

Figure 7.1A shows that the stress-strain curves for the contact and central zones have essentially the same slope. This slope, which gives the Young's modulus for the rock, is at the upper end of the range of static values for granitic rocks reported by Birch (1966) and has a large scatter. However, the contact zone samples do not show any indication of 'settling in' which corresponds to the closing of micro-fractures (Jaeger and Cook, 1969, p.319), thus indicating that the contact zone has fewer micro-fractures compared to the central zone.

The other curves in Figure 7.1 show that the contact zone has a proportionally larger lateral strain than the central zone, and after a certain point the axial strain is less than twice the lateral strain: therefore the volume strain tends towards a positive value. This means that the contact zone shows a certain degree of *dilatancy* which is not present in the central zone, but the values of Poisson's ratio for the samples from the contact zone, prior to this change, are similar to those for the central zone samples and to values for other rocks listed by Birch (1966).

The type of failure for all the specimens irrespective of their structural setting was violent, but micro-seismic activity (detected

Table 7.2

Mechanical properties of the aplite sills and adjacent Quartzites.

Sill No.	Sample	C _o	E	v	C _a	D	L	Ratio	Comments
3	BC5 0,73	268	0,813	0,245	122	32,3	100,2	3,10	Cylindrical specimen, failed along a joint.
3	BC5 0,88	485	0,767	0,278	339	32,4	100,2	3,09	Cylindrical specimen, violent failure.
3	BC5 0,97	430	0,785	0,268	215	30,8	91,7	2,98	Cylindrical specimen, violent failure.
3	BC5 1,80	533	0,754	0,274	316	32,4	98,9	3,05	Cylindrical specimen, violent failure.
3	BC5 3,00	482	0,798	0,278	199	30,0	90,5	3,02	Cylindrical specimen, violent failure.
3	BC5 3,68	286	0,920	0,188	*	29,7	99,9	3,06.	Cylindrical specimen, non-violent failure.
Thickness of sill in this borehole 3,68 metres.									
	BC7 +0,4	106	0,716	0,156	*	26,8	76,7	2,86	Quartzite, cylindrical specimen, non-violent failure.
	BC7 +0,3	62	0,812	0,748	*	26,8	79,0	2,95	Quartzite, cylindrical specimen, non-violent failure.
	BC7 +0,1	123	0,713	0,362	*	29,9	89,7	3,00	Quartzite, cylindrical specimen, non-violent failure.
1	BC7 0,15	478	0,904	0,220		31,8	71,5	2,25	Cylindrical specimen, violent failure.
1	BC7 2,16	460	0,748	0,332	218	32,4	96,8	2,99	Cylindrical specimen, violent failure.
1	BC7 2,35	376	0,762	0,289	291	32,4	97,9	3,02	Cylindrical specimen, violent failure.
1	BC7 2,50	433	0,778	0,287	357	31,6	98,4	3,12	Cylindrical specimen, violent failure.
	BC7 3,95	181	0,778	0,197	*	26,5	70,6	2,66	Quartzite, cylindrical specimen, non-violent failure.

Thickness of sill in this borehole 3,45 metres.

	BC8 +0,2	208	0,711	0,219	29,9	84,9	2,84	Quartzite, cylindrical specimen, violent failure.	
	BC8 +0,1	221	0,658	0,269	28,4	81,6	2,87	Quartzite, cylindrical specimen, violent failure.	
1	BC8 2,50	541	0,759	0,299	256	29,9	80,6	2,69	Cylindrical specimen, violent failure.
1	BC8 3,00	375	0,801	0,295	289	29,7	85,9	2,89	Cylindrical specimen, violent failure.
1	BC8 3,10	407	0,800	0,292	29,6	88,9	3,00	Cylindrical specimen, violent failure.	
1	BC8 3,60	478	0,749	0,260	345	31,0	92,9	3,00	Cylindrical specimen, violent failure after hysteresis.
	BC8 4,67	260	0,736	0,225	29,7	84,8	2,86	Quartzite, cylindrical specimen, violent failure.	
	BC8 4,72	216	0,680	0,261	29,7	87,8	2,96	Quartzite, cylindrical specimen, violent failure.	
	BC8 4,77	245	0,743	0,270	29,7	83,0	2,80	Quartzite, cylindrical specimen, violent failure.	
	BC8 4,87	239	0,740	0,268	31,0	90,9	2,93	Quartzite, cylindrical specimen, violent failure.	

Thickness of sill in this borehole 4,57 metres.

6	G76C	259	0,492	*	24,0	62,0	2,58	Irregular cylindrical specimen, produced Luder's bands	
8	F67C No2	268	0,714	0,568	*	25,8	74,7	2,90	Cylindrical specimen, produced Luder's bands.
8	F67C No3	274	-	-	*	25,8	75,1	2,91	Cylindrical specimen, produced Luder's bands, violent failure
8	F67C No4	251	0,708	0,298	*	25,7	76,4	2,97	Cylindrical specimen, produced Luder's bands in a hysteresis.
8	F67C No5	173	0,776	0,189	*	25,7	77,0	3,00	Cylindrical specimen, produced Luder's bands in a hysteresis.

C₀ Uniaxial compressive strength or maximum load in MPa.

E Young's modulus 10⁵ MPa.

ν Poisson's ratio.

C_a Stress at start of microseismic activity MPa, * indicates none detected.

D Diameter of the specimen millimetres.

L Length of the specimen in millimetres.

Ratio Ratio of length to diameter.

Average strength of the central zone specimens 456 MPa.

Average stress at fracture initiation 267 MPa.

Figure 7.2

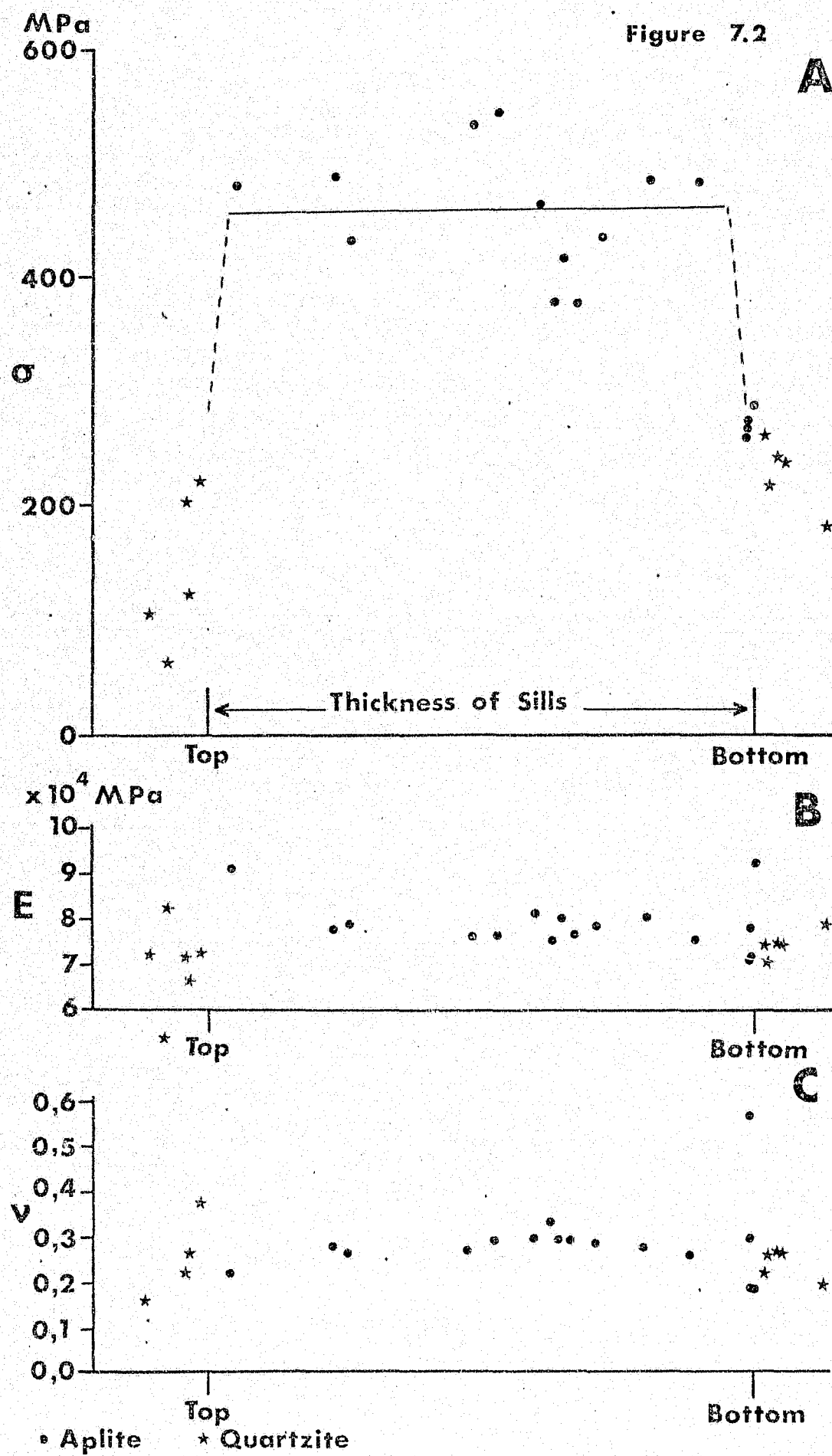


Figure 7.2

The variation of the rock mechanical properties across the sills.

- A. The uniaxial compressive strength. The solid line is the average central zone strength determined in this work.
- B. Young's Modulus
- C. Poisson's Ratio

audibly) always occurred in the central zone samples prior to failure but never in the contact zone specimens which suggests a marked difference in the respective behaviour of the specimens. In particular the uniaxial compressive strengths of the contact zone samples is very much less than the strengths of the central zone samples (c.f. Table 7.2 and Figure 7.2). The strength of the adjacent footwall quartzites is relatively low, ranging from 62 to 260 MPa. Similarly, the chert in the lower contact zone is not particularly strong, having an average strength of about 280 MPa. However, away from the contact zone, the strength increases remarkably and in the central zone the average value is 456 MPa with a range from 375 to 541 MPa. The large scatter in the results from the central zone may be due to several factors, including:-

- (i) Sampling difficulties : samples were taken from different localities and levels in the sills
- (ii) Fractures induced in the core by vibration of the compressed air diamond drill used for coring
- (iii) The presence of natural and mining induced joints in the core

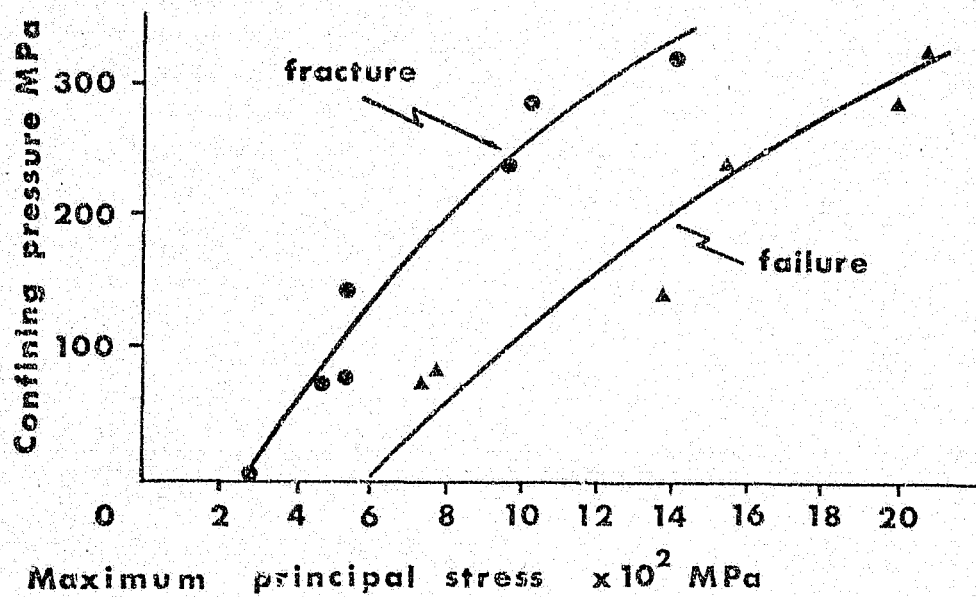
However, the average strength of the central zone agrees fairly well with the values determined by N.M.E.R.I. (1959), but are significantly lower than those obtained by Hoek (1965) and Brace et al. (1966). The Young's modulus and the Poisson's ratio are also in agreement with the results of N.M.E.R.I. (1959).

7.4 Discussion

The underground observations show that it is the very strong central zone aplite and not the weaker contact zone aplite or quartzite, which fractures badly and is difficult to support. Why it behaves in this way

**CHANGES IN THE STRESS AT
INITIATION OF FRACTURING, AND
FAILURE WITH CHANGING CONFINING
PRESSURE FOR THE CENTRAL ZONE
MATERIAL.**

Figure 7.3



Data from Brace et al (1966)

is not known; it is in contrast to what one would expect from experimental observations, but it does correlate with the fact that the geological joints are also best developed in the central zone.

The initiation of fracturing in the aplite has been shown by Brace et al. (1966) to begin very early in a test, with very little accompanying dilation compared to other rock types. Figure 7.3 shows the increase in the stress required for fracture initiation with confining pressure. The reported value of the uniaxial stress at which fracturing began agrees well with the average stress (267 MPa) at which microseismic activity started in the present tests on the central zone specimens. The volume strain-axial stress curves (e.g. Figure 7.1C) show that fracturing starts at a stress level of 100 - 200 MPa in the contact zone specimens. Similarly, McGarr (personal communication, 1974) reported that microseismic activity during testing of hangingwall quartzites started at about three-quarters of the failure load which corresponds to an average uniaxial compressive stress of 260 MPa.

Pallister et al. (1971) have shown that the virgin horizontal/vertical stress ratio on E.R.P.M. was 0,72, but the vertical stress was less than expected from the weight of the overburden. At 3,2 kilometres the stress due to the overburden is 85 MPa and so the maximum value of the virgin horizontal stress is approximately 60 MPa. Therefore, from Figure 7.3 a stress of marginally less than 400 MPa would be required to fracture the central zone rock in the virgin stress field. Towards the mined out area, where the minimum principal stress tends towards zero and the maximum principal stress increases (McGarr, 1971a), a stress of more than 250 MPa is required to fracture the rock in the central zone. According to McGarr (1971a) stresses of these magnitudes are easily realised in the vicinity of the stope, with fracturing of the rock possibly extending more than 10 metres in advance of the stope face. With respect to the aplite the contact zone should fracture before

the enveloping quartzites and central zone aplite. In fact, if one considers the extent of the induced stress field in front of the stope (McGarr et al., 1975) and the failure criterion postulated by Cook (1962), (failure stress = $\sigma_1 - 6\sigma_3$, where σ_1 is the maximum principal induced stress and σ_3 is the minimum principal induced stress) the rock of the contact zone should be fractured up to 30 metres in advance of the stope, and the central zone aplite and the quartzite should start to fracture at about 10 metres from the stope face. Therefore, considering the greater time available for deformation underground, compared to the experimental work, the contact zone does not deform elastically but in a ductile manner (see Figure 7.1c) and so has relatively few mining induced fractures. Therefore it forms a relatively stable sheath enclosing the stronger central zone, which despite geological jointing may have considerable strain energy stored in it which is released by mining stresses, resulting in intense fracturing. This correlates with the poor core recovery of aplite in this region.

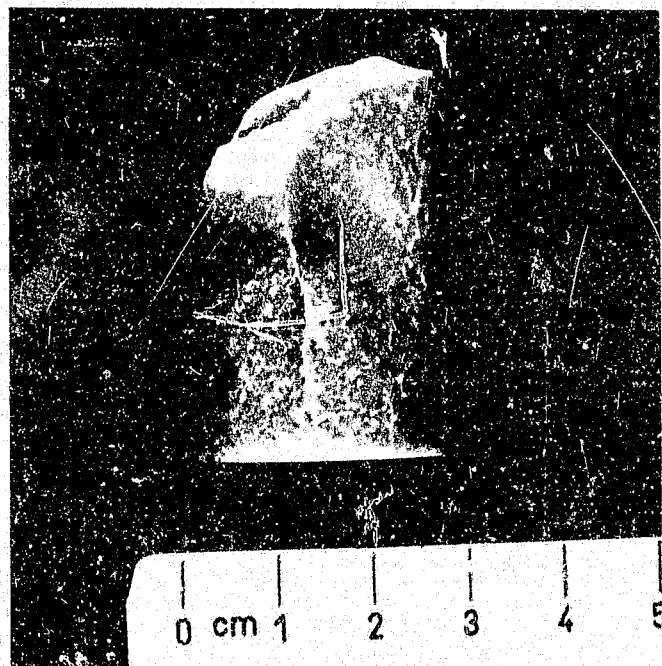
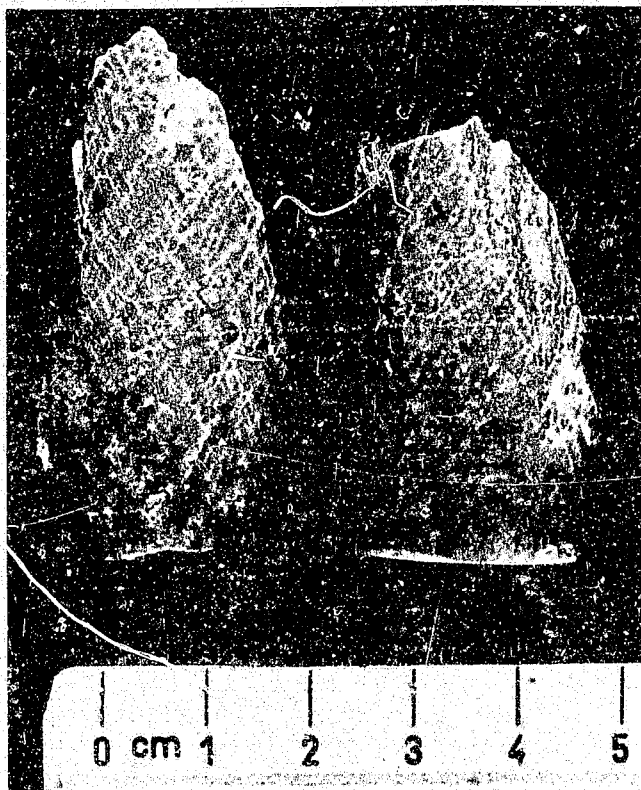


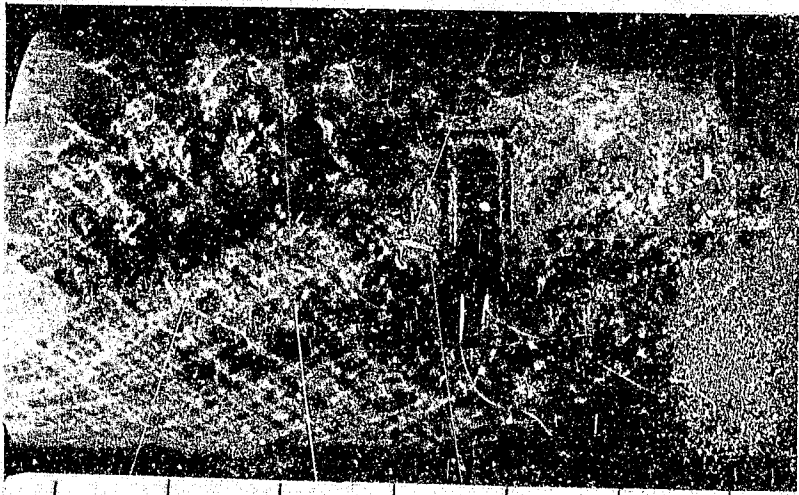
Plate 8.1

Lüder's bands formed during uniaxial compression tests; the bands are generally restricted to the central part of the specimens.

(Photograph by M. Hudson)

Plate 8.2

The lithologic control on the formation of Lüder's bands: the bands only occur in the mica rich zones originally parallel to the contact, but now at an angle to the core. (Photograph by M. Hudson)



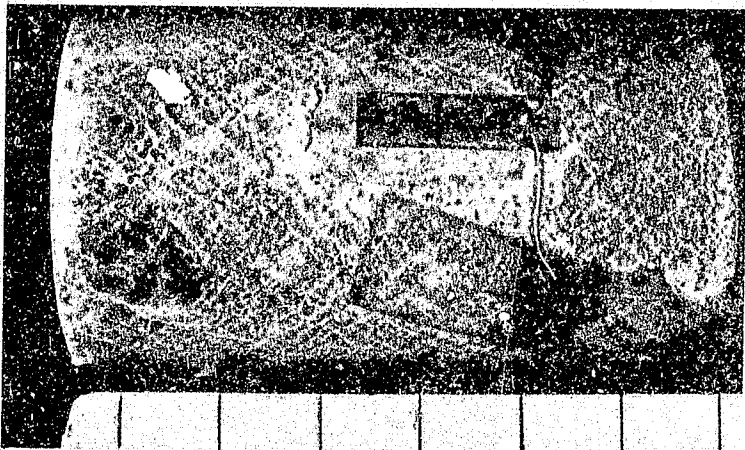
0 cm 1 2 3 4 5 6

A



0 cm 1 2 3 4 5 6

B



C

Plate 8.3 A, B, C

A more penetrative development of Lüder's bands formed during uniaxial compression with a small confining pressure.

Plate 8.3 A, B, C

A more penetrative development of Lüder's bands formed during uniaxial compression with a small confining pressure.

Table 8.1

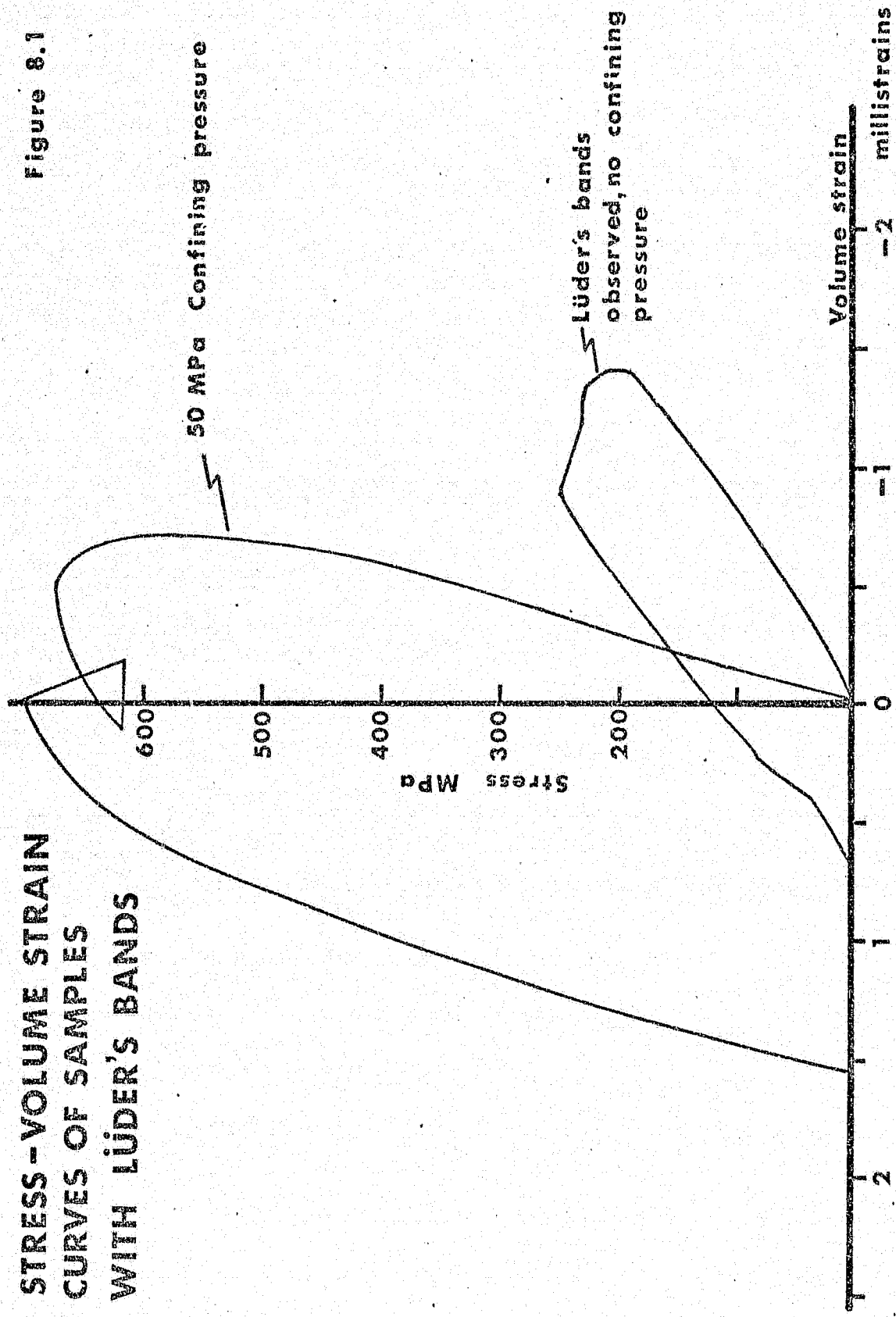
Experimental conditions and the angle between conjugate sets of Luder's bands

Sample	Stress at initiation MPa	Maximum stress MPa	* Mean angle degrees	Standard deviation degrees	Volume strain increase millistrains	No. of measurements	Comments
BC5/3,68	-	286	71	2,7	-	20	Bands only found at the end of a test specimen adjacent to a shear failure plane.
BC5/3,68	-	286	87	3,3	-	10	Bands only found at the end of a test specimen away from the shear failure plane.
G76C	199	259	58	5,2	-	10	Step loaded at 10 kN/5minutes; taken to failure.
F67C No. 2	134	268	57	6,0	-	10	Step loaded but unloaded prior to failure.
F67C No. 3	-	274	-	-	-	-	Continuously loaded until failure.
F67C No. 4	110	251	55	6,6	0,7	10	Step loaded but unloaded prior to failure.
F67C No. 5	96	173	59	10,9	0,1	10	Step loaded but unloaded prior to failure.
F67C No. 6	-	512	71	8,6	1,6	10	Step loaded with 50 MPa confining pressure; penetrative bands covering the whole specimen.

* Angle between band sets bisected by axial stress.

**STRESS-VOLUME STRAIN
CURVES OF SAMPLES
WITH LÜDER'S BANDS**

Figure 8.1



8. LÜDERS' BANDS

In addition to the main rock mechanical work, a limited amount of experimental work was done on the formation of Lüder's bands in samples from the contact zone.

8.1 Previous work

Lüders' bands are closely spaced shear planes that occur in one or two sets at an angle either side of the applied maximum stresses, during the deformation of metals in tensile, compressional and bending tests. Some theory has been developed and so far only applied to tensile tests (e.g. Nadai, 1960, p.319-339; Irwin, 1953; Reid, 1973, p.44-47) which suggests that the bands form parallel to the surfaces of no finite strain. This theory does not discriminate between tensile or compressional tests.

The occurrence of Lüders' bands in experimentally deformed rocks has only been reported by Prandtl and Rinne (1907), Heard (1960) and Friedman and Logan (1973); in addition, Jaeger and Cook (1969, p.87) describe a type of failure as 'multiple shear planes' which is presumably the same phenomenon. Naturally occurring samples have been described by Friedman and Logan (1973) and are reported here.

8.2 Practical results

In the present experiments Lüders' bands were found to form in several specimens from the contact zone, especially if the specimen was loaded carefully at the relatively slow rate of 10 kN/5 minutes. By polishing the surface of the specimens the progress of band initiation could be studied.

The detailed results of the tests are outlined in Table 8.1 and Figures 7.1 and 8.1 show the volume strain-axial stress curves of the samples that formed Lüders' bands; these curves show that there was considerable release of volume strain energy during Lüder band development.

The initial observed development of the bands coincides with an abrupt change in the slope of the axial stress-volume strain curve (Figure 7.1c, 8.1). These curves also show that band formation is accompanied by considerable permanent increase in volume.

Three styles of formation have been observed in these specimens. Firstly, the Lüders' bands start to form at between one and two-thirds of the failure load, being made visible by the change in appearance of small felspar phenocrysts in the aplite. In the undeformed rock these phenocrysts appear as 1 millimetre diameter, light green specks, set in a dark green matrix. During loading the specks suddenly become whiteish in colour, after which the Lüder's bands begin to propagate away from the felspar phenocrysts and form around the quartz phenocrysts. Two sets of bands form, resulting in a rhombohedral pattern, at which stage, the original texture of the specimen is lost. Further increases in the load cause more and more bands to form and the surface area covered by the bands to increase.

The second style of formation occurs in specimens poor in phenocrysts and loaded at very slow rates. The first sign of formation of the bands is a general lightening of the colour of the central area of the specimen. On closer inspection a series of very finely spaced (0.1 millimetre) bands can be seen. The intensity of the bands increase as the load is increased, with the bands propagating away from the central area. The direction of propagation is in the plane of the bands and not normal to the plane of the bands as in metals under tension (Reid, 1971). The third style has not been observed during formation but two distinct sets of bands form; one parallel to an adjacent shear failure plane and the other set inclined to and some distance from the shear plane.

For all three styles, the Lüders' bands formed in the specimens tested by uniaxial compression, are mainly confined to the surface of the specimen but in the specimen tested triaxially they are more penetrative (Plate 8.3).

Figure 8.2

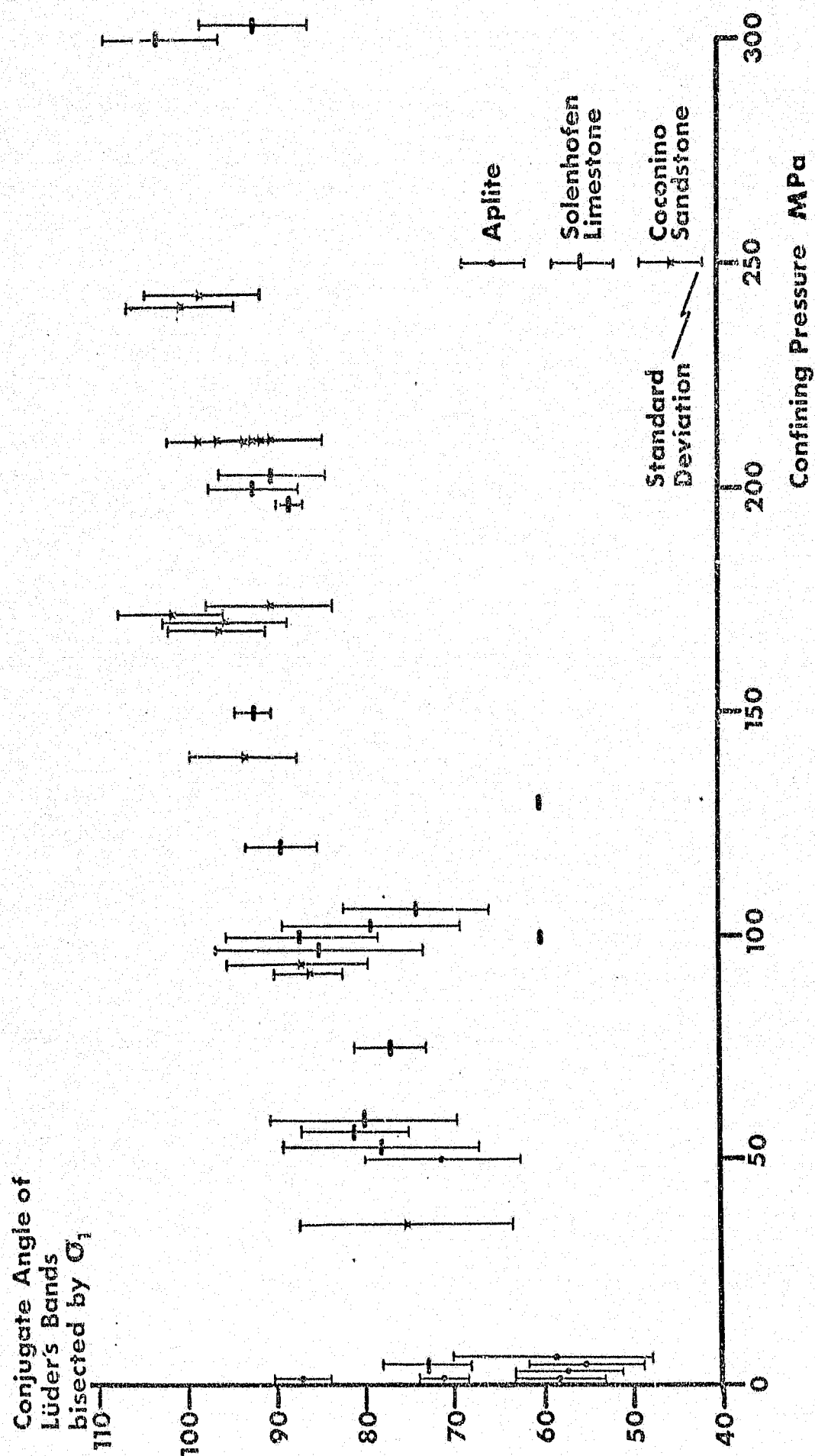


Figure 8.2

A synthesis of the available data on the change in the conjugate angle of the Lüder's bands with confining pressure.

The aplite results were obtained in this work

The Coconino sandstone results are from Friedman and Logan (1973)

The Solenhofen limestone results are from Friedman and Logan

(1973) and Heard (1960). The latter results do not have error bars.

There is a strong lithologic control on their formation and they occur preferentially in the areas of high mica concentration (Plate 8.2). This correlates with the fact that no bands were formed experimentally in samples from the central zone.

8.3 Discussion

In tensile tests on steel, the specimens deform elastically until an upper yield point is reached, when there is a sudden stress drop to a lower yield point. At this point Lüder's bands begin to form at a constant stress, and when they have formed throughout the test piece, plastic deformation begins (Nadai, 1948; Reid, 1973). Similar, though less well defined stress-strain curves have been obtained in compression tests on Solenhofen limestone by Heard (1960) and Carrara marble by Von Karman (1917). However, Lüder's bands were not always produced in these tests, and those that formed did so when the rock was deforming in the transitional zone between brittle and ductile behaviour, a result also obtained by Friedman and Logan (1973). This is in contrast to the results reported here for the aplite, in which the Lüder's bands formed during brittle deformation of the rock, as defined by Jaeger and Cook (1969, p.77)

Friedman and Logan (1973) studied Lüder's bands formed in Solenhofen limestone and Coconino sandstone tested under various confining pressures. They found a definite correlation between the confining pressure and the angle bisected by the maximum principal stress. Their results are plotted in Figure 8.2 with those reported by Heard (1960) for Solenhofen limestone and those obtained in this work for aplite. All three rock types for which data are now available plot on a curve which shows an increase in the angle from 55° to 105° as confining pressure increases to 300 MPa. The errors are relatively large particularly at low confining pressures, but they are reduced at slower strain rates. However, Heard's (1960)

results do not lie on this curve which suggests that parameters other than confining pressure and strain rate influence the formation of Lüder's bands and may affect the angular relationship between pairs of Lüder's bands.

9. MECHANISMS OF INTRUSION

From the isotopic work and the chemistry, it is obvious that the postulated origins of the 'chert dyke' put forward by Pretorius (1964) i.e. mylonitisation of the sediments, and Shipway (1972) i.e. low temperature lateral siliceous secretion, are incorrect. In this chapter the emplacement of the sills as a hypabyssal igneous intrusion is discussed.

Some of the features that have to be explained include (1) the non-laminar flow textures and (2) the limited distribution of the sills within a particular horizon.

9.1 Speed of emplacement

A minimum average speed can be calculated from the time required to crystallise the complete thickness of the sill, during which the sill must have covered the maximum known distance for the sills, which is greater than six kilometres. From Jaeger's (1959, 1961) curves, the time required for a four metre thick sill to crystallise, using the assumption that the melting point is 900°C (Chapter 4), is 20 days. This means that the larger sills must have had a minimum average speed of 0,35 cm/sec; the actual value would have been slightly greater because little or no movement would have occurred during the final phases of crystallisation.

9.2 Viscosity of the magma

Bottinga and Weill (1972) and Shaw (1972) have recently shown that the approximate viscosity of a magma can be derived from the major element chemistry of the igneous rock. The numbers obtained using Shaw's (1972) technique are in good agreement with viscosities determined experimentally for magmas of similar composition. Shaw's (1972) method gives a viscosity for the aplites of $8,2 \times 10^5$ poise at 1000°C , a value which agrees with those for obsidian listed by Clark (1966). This calculated viscosity

would be reduced by the presence of volatiles (Shaw, 1972; Kogarko, 1974).

The importance of the viscosity is that most models for intrusion assume laminar flow, and can therefore be used to estimate the pressure gradient required to move the magma. In the aplite sills non-laminar textures were seen which may indicate localised disruptions of the laminar flow field by turbulent flow. Prandtl and Tietjens (1957) give a value for the critical Reynold's number of a thousand for unstable turbulent flow. The Reynold's number of the aplite magma, using estimated and determined parameters most favourable to generate a high Reynold's number can be calculated as follows:-

$$\begin{aligned} R &= \rho \cdot u \cdot d / \eta \\ &= \frac{2,4 \times 0,5 \times 600}{8,2 \times 10^5} \\ &= 8,7 \times 10^{-4} \end{aligned}$$

ρ density of the magma (g/cm^3)
 u speed of intrusion of the magma (cm/sec)
 d thickness of the sill (cm)
 η viscosity of the magma (poise)

This number is considerably smaller than required for turbulent flow, and only by reducing the viscosity to the unrealistically low value of $\pm 1,0$ poise will the Reynold's number equal the critical value.

The implication of this is that the non-laminar texture cannot be explained by classical fluid dynamics with the presently known or estimated parameters, though the texture looks very like some turbulent flow structures. In fact it suggests that during intrusion patches of turbulent flow occur in the centre of the sills, particularly near protruberences of the wall, with a boundary layer of laminar flow at the contacts.

A result of this low Reynold's number and the slow speed, is that the pressure gradient along the sills will be due to fractional drag and elevation changes and not due to turbulent drag as well.

9.3 Stratigraphic level

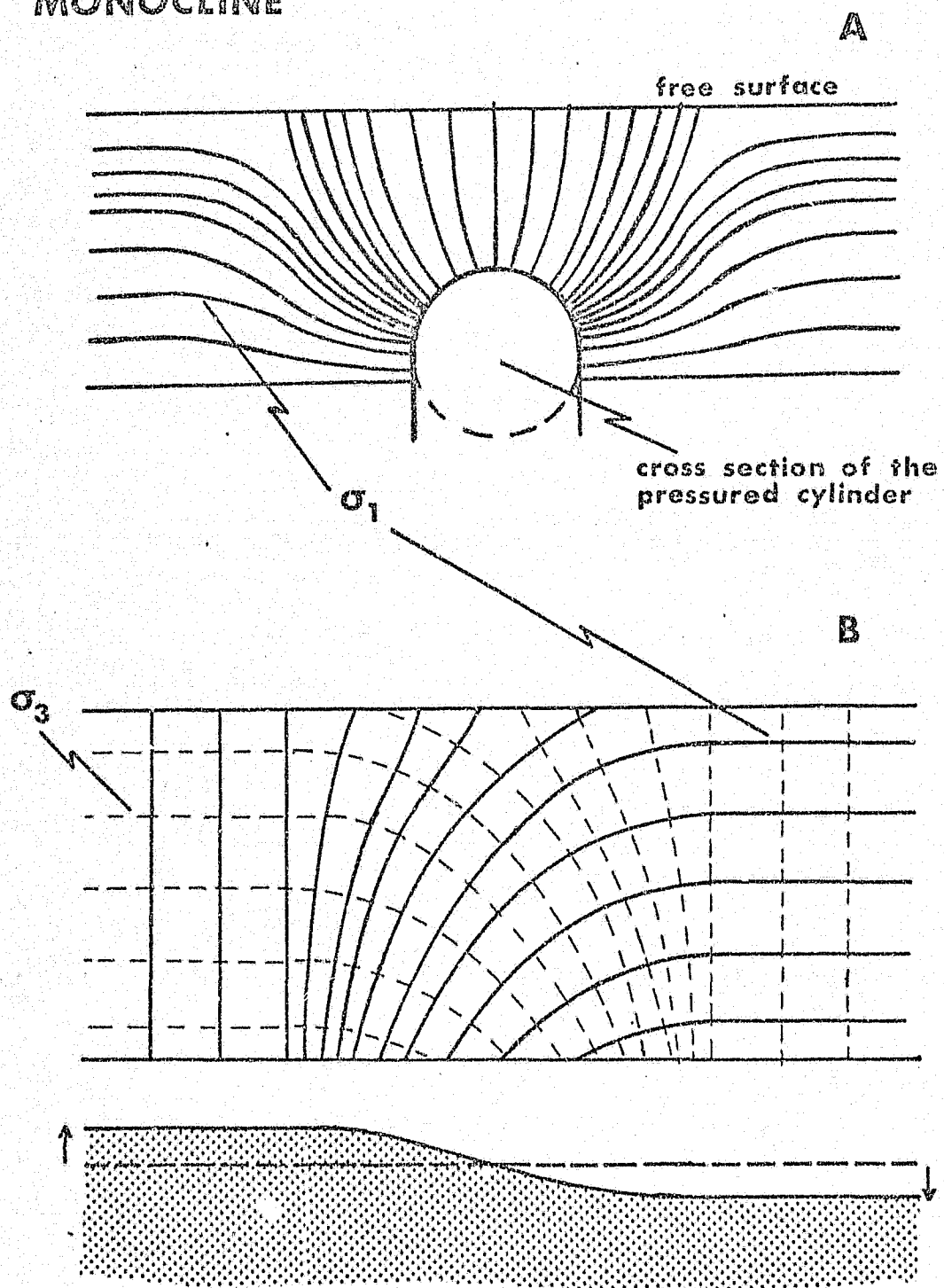
At the time of emplacement of the sills, given by the geochronology, it was estimated that there was nine kilometres of overburden (Chapter 6). This is in disagreement with the results of Mudge (1968), who found that the maximum depth for the emplacement of sills was about two kilometres. Therefore the depth of emplacement and the transgressive nature of the aplite sills must be explained.

A large number of workers including Anderson (1938), Gretner (1969), Roberts (1970), Pollard (1973) and Phillips (1974) have tried to explain why sills occur at a particular level. Gretner (1969) and Roberts (1970) have postulates that depend on the weight of the overburden and the rheological properties of a particular horizon. Phillips (1974) has a postulate that a dyke will change into a sill when the density of the magma equals that of the host rock. Pollard (1973) explains the occurrence by the physical nature of a pre-existing discontinuity such as bedding. Roberts (1970) also has an explanation which is solely dependent on the pore pressure of the host rock being nearly equal to the lithostatic pressure, so that the particular level at which the sills intrude is essentially equivalent to the topographical surface.

In Chapters 1 and 2 it was noted that the projected outcrops of the aplite sills have a linear expression which cuts across the folds and faults of the Witwatersrand sediments, especially in the West Rand. This suggests that the above explanations are not applicable to the emplacement of the aplite sills on a gross scale as they cut across rocks with radically different rheological properties (shale, quartzite) and across major faults. Also the inferred level of emplacement (9 kilometres) is far in excess of the usual level of sill emplacement of less than two kilometres (Mudge, 1968) and at an unfavourable level for the above postulates.

**STRESS TRAJECTORIES
AROUND A PRESSURISED CYLINDER,
AND ADJACENT TO A CRUSTAL
MONOCLINE**

Figure 9.1



after Roberts(1970)

The depth and the transgressive nature suggest that the gross distribution of the sills may be controlled by the principal stress orientation. The geometrical form of the feeders to the sills is unknown, but since the known geometry of the sills suggest they lie in the plane containing the maximum and intermediate principal stresses, the feeders may follow continuations of these planes to depth. The form of the principal stress trajectories surrounding possible magma sources has been considered by Sanford (1959), Roberts (1970) and Phillips (1974); the last two extended the original theory of Anderson (1938) by considering a pressurised cylinder or sphere representing the magma chamber in a solid near a free surface. The analysis predicts that the sills would emanate from the sides of the magma chamber (see Figure 9.1a) and then ascend through the crust towards the free surface asymptotically. The form of the stress trajectories will be modified by any additional stress components acting at the same instant. In the case of the aplite sills there are two additional stress components; a gravitational stress system with the maximum principal stress oriented vertically, and a tectonic stress system with the maximum principal stress oriented horizontally north-south. There is a possibility that this tectonic stress might be a result of the pressure in the magma chamber. In such a model, the shape of the magma chamber will have the largest single effect on the distribution of the stress trajectories. In Phillips' (1974) postulate, the absence of dykes at the surface is explained by the magma chamber rupturing on the sides due to a sudden expansion in the volume of the magma. As a result, magma would extrude laterally and not reach the surface.

The planes containing the maximum and intermediate principal stress can also change in orientation from the horizontal to the vertical at a crustal monoclinial flexure, (Sanford, 1959 and Roberts, 1970). In such a system (Figure 9.1b) sills will occur in the crustal block that is

depressed and dykes in the crustal block that is displaced upwards.

The density of a granitic magma is approximately $2,4 \times 10^3 \text{ Kg/m}^3$ at 800°C (Harris et al., 1970) and will be less at higher temperatures so a marked density contrast with all rocks (except water saturated, unconsolidated sediments) should have existed during the intrusion of the aplite. This density contrast would introduce a gravity potential which, with the magma pressure and bouyancy affect would provide most or even all of the force required for the emplacement of the sills.

The sills were emplaced at nine kilometres below surface as a result of the magma following a stress trajectory that was near horizontal at nine kilometres below surface, in the Central Rand Basin.

9.4 Conclusions

From the chemistry and isotopic work it was inferred that the magma originated at 35 kilometres below surface, and was emplaced at 9 kilometres below surface during the preliminary activity associated with the Bushveld Igneous Complex. From the preceding sections in this chapter it was inferred that sills intruded along flat lying planes containing the maximum and intermediate principal stresses at approximately $0,4 \text{ cm/sec}$. Finally, from the geometry given in Chapter 2 the sills appear to have been emplaced from the south east.

The crustal flexure model would appear to be a favourable one, for there would be crustal uplift resulting from the melting of the mantle, to form the Bushveld magma. This would result in transgressive sills being injected into stable block, and gently folding it which correlates with the folding of the sills. On E.R.P.M. the direction of emplacement would be north to south. However, there is evidence from the morphology of the sills that the direction of emplacement was south to north.

There are two many unknowns to assess whether the other model is a more favourable one.

10. CONCLUSION

The aplite sills on E.R.P.M. are part of a much larger set of associated sills occurring within the Witwatersrand sediments. The exposed aerial extent has extreme variations, but the thickness of the sill is relatively constant and the maximum thickness is noticeably smaller than other occurrences to the east and west of E.R.P.M.

Petrographically the sills appear to be similar to a large number of other felsic sills with a two stage cooling history. However, chemically the aplite is rather unique as a result of the peralkaline composition of the source rock which was anatectically melted at 30 - 35 kilometres below the surface. From the geochronology this source rock appears to be associated with the leucotonalitic gneisses of the Ancient Gneiss Complex.

The emplacement of the sills appears to be controlled by the tectonic principal stresses operating at that time with only a minor influence from different rock types and structures. The magnitude and orientation of these stresses as deduced from tectonic features are: σ_1 400 MPa, $182^\circ/27^\circ$; σ_2 230 - 400 MPa, $274^\circ/10^\circ$; and σ_3 230 MPa, $028^\circ/70^\circ$, associated with the aplite sills.

In as far as the mining activity is concerned, the aplite sills appear to present a structural problem only when the sills are greater than three metres thick. Such problems are only expected to be encountered in the Angelo and Driefontein sections in the long term future; in all other sections mining is taking place above the aplite sills and so is not influenced by its presence.

APPENDIX A

List of samples analysed with crushing procedure and brief description. Unless stated all samples are in excess of 2 kilograms and all samples except the very small ones were split into aliquots after the initial crushing in the jaw crusher.

Abbreviations: JC = jaw crusher, S = sieved, A = agate concentric ring mill, WC = tungsten carbide concentric ring mill, NiCr = nickel-chrome concentric ring mill.

Sample names: F64A = collected in F shaft on 64 level, BC2/1,2 = sample from No. 2 borehole 1,2 metres from top contact of the sill.

<u>Sample</u>	<u>Sill No.</u>	<u>Crushing</u>	<u>Description</u>
F64F	10	JC, A	High mafic content in ellipsoidal concentrations
F64G	9	JC, WC	Small euhedral pyrite crystals in a sample otherwise similar to F64F
F67B	8	JC, WC	One metre from bottom contact with no noticeable differences from other central zone rock
F69C		JC, WC	Felsic intrusion in Sill No. 8, medium grained granular rock
F70E	8	JC, A	Laminar and non-laminar texture 30 - 40 centimetres from bottom contact
G41	1	JC, WC	Strong laminar texture from centre of sill composed of parallel ellipsoids of mafic minerals
G74A	3	JC, WC	Large pyrite phenocrysts (0,5 centimetre) 0,5 metre from bottom contact in an off-white, fine grained matrix, in contrast to aphanitic matrix.
G74B	6	JC, A	Large pyrite phenocrysts 0,3 metres from bottom contact. In a pale green aphanitic matrix.
G75B	6	JC, WC	Pink Coloured sample with laminar texture. about 15 centimetres from the top contact

<u>Sample</u>	<u>Sill No.</u>	<u>Crushing</u>	<u>Description</u>
G76B	6	JC, WC	Normal central zone grey specimen from the central part of the sill
G76D 0,1	6	JC, A	Normal contact zone specimen forming part of a traverse 0,1 metre from upper contact in the stope
G76D 0,2	6	JC, WC	Normal central zone specimen forming part of a traverse 0,2 metre from upper contact in the stope
G76D 0,75	6	JC, WC	Normal central zone specimen forming part of a traverse 0,75 metre from upper contact in the stope
G76D 1,2	6	JC, WC	Normal central zone specimen forming part of a traverse 1,2 metres from upper contact in the stope
G77A		JC, WC	Quartz-chlorite-calcite sill intruding Sill No. 6
H70A	1	JC, WC	Normal grey central zone specimen from the central part of the sill, and the furthest removed accessible sample from others for Sill No. 1
H73A	3	JC, WC	Specimen has a L��der's band texture, defined by linear concentration of mica, 1,5 metres from bottom contact
H75F 2,6	3	JC, WC	Specimen has a L��der's band texture, defined by linear concentration of mica, 1,5 metres from bottom contact
H75Z	3	JC, A	From the central part of the sill with a large scale colour banding of light-dark, green, grey colouring
H75II	3	JC, S, A	A bright but off-white sample with magnetite phenocrysts collected in the stope in the central part of the sill
H75IID	3	JC, NiCr	Duplicate sample of H75II
H76T	4	JC, WC	Near a post sill, quartz vein, in the central part of the sill that has no noticeable differences from other central zone material
H77D	4	JC, WC	From the central part of the sill, but 10 centimetres from a mineralised joint, containing mica, calcite and pyrite

Author Fumerton S L

Name of thesis The Felsic intrusives in E.R.P.M. Boksburg 1975

PUBLISHER:

University of the Witwatersrand, Johannesburg

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