

IV. The Mineralogical Composition of the Banded Ironstone

A microscopic study of core-samples obtained from a bore-hole drilled on Botha (M2), about 15 miles south of Danielskuil, and from one drilled on Pomfret (B4) was carried out and the results may be regarded as representative of the banded ironstone of the Northern Region. The bore-hole on the first farm penetrated almost the entire Banded Ironstone Substage and was abandoned after having reached the underlying Dolomite.

The main constituents of the banded ironstone of the Northern Region are chert, magnetite, carbonate (chiefly calcite, siderite and dolomite) accompanied by accessory amounts of minnesotaite and stilpnomelane. Minnesotaite is especially abundant in the lowermost portion of the Banded Ironstone Substage at Pomfret. Banding or lamination in the rock is extremely well developed and the contacts between successive laminae are sharp as a rule. The laminae generally have a monomineralic composition but laminae in which two or more of the constituent minerals figure are also found.

A petrological study of the rocks of Lower Griquatown Stage in the Southern Region was carried out on bore-hole cores obtained from a bore-hole (W2) on Westerberg (Q2). The banded ironstone from the Banded Ironstone Substage in this region corresponds largely with that of the Northern Region except that minnesotaite is more abundant.

In the following paragraphs the different minerals which constitute the crocidolite-bearing banded ironstone and their mode of occurrence are described in order of abundance.

(a) Chert

According to definition the term "chert" is applied to "cryptocrystalline varieties of silica regardless of colour, composed mainly of petrographically microscopic chalcedony and/or quartz particles whose outlines range from easily resolvable to nonresolvable with a binocular microscope at magnifications ordinarily used. Particles rarely exceed 0.5 mm. in diameter" (Howell, 1957, p. 49). Rice (1949, p. 71) defines chert as "a dense, cryptocrystalline rock, composed mineralogically of chalcedony and cryptocrystalline quartz".

Chalcedony is virtually absent in the banded ironstone and the associated rocks from the Lower Griquatown Stage and the free silica in these rocks could best be described as microcrystalline quartz. However, the term "chert" has

more or less a world-wide usage in the description of microcrystalline, non-calcitic quartz which typically forms laminae in the Precambrian banded ironstones. For this reason and because of its long usage in the description of similar rocks the term is retained in this paper.

The microforms of quartz may be divided into two classes, microcrystalline quartz and chalcedonic quartz (Keller, 1941). Microcrystalline quartz forms the bulk of the chert laminae in the banded ironstone and in addition may be accompanied by quartz of larger grain size. During the microscopical investigation of many sections of the banded ironstone from the Lower Griquatown Stage chalcedony was found in two thin sections only (HH285). Under the microscope it is composed of radiating or sheaf-like bundles of fibres radiating from a core of microcrystalline quartz or carbonate. This mode of occurrence, where the chalcedony is clustered around a core of crystalline material, suggests that the chalcedony grew in geodes formed during late-stage solution and reprecipitation of microcrystalline quartz.

The late solution of quartz could have taken place during the intrusion of mafic dykes and sills from which alkaline solutions were derived. Real evidence to prove this was not found. Where trapped these solutions could have caused the solution of quartz and the more or less simultaneous precipitation of carbonate. Under very sluggish conditions of migration of the dissolved silica the solutions then became gradually acid with the result that the silica in solution was precipitated to form chalcedony. This line of thought is strengthened by the common association of chalcedony with a core of carbonate.

The microcrystalline quartz is present as closely interlocking xenoblastic grains with random optical orientation. Grain-sizes range from less than 0.012 mm to 0.122 mm in diameter, the average being around 0.025 mm. Microcrystalline quartz is also present as distinctly elongated crystals (HH279).

A single lamina of microcrystalline quartz in the banded ironstone could be composed of grains having approximately the same size or the grain-sizes may decrease or increase gradually in a direction perpendicular to the bedding. Gradual variation in grain-size from 0.012 to 0.06 mm in diameter in a single thin chert lamina is often observed. In the case of relatively thick chert laminae the laminae are often composed of several thin layers of different grain-size. This type of phenomenon of grading and alternative layers of

various grain-size could possibly be explained by the rate of precipitation of silica during the deposition of the rocks. Sharp changes in grain-size from one lamina to an adjacent one apparently indicates that conditions controlling the precipitation of silica, whether chemically or as colloids, must have changed quite abruptly.

Elongated crystals of quartz are most common in late fractures which often traverse the bedding at right or at steep angles. In these fractures the quartz grains, often accompanied by carbonate, are orientated with their major axes perpendicular to the walls of the fracture. Growth took place from both walls or in rare cases from one wall only. The lengths of the elongated crystals naturally depends on the width of the fractures; the maximum length of microcrystalline crystals observed in these fractures is 0.48 mm (Sp. HH271).

In many specimens chert laminae separating magnetite laminae are chiefly composed of tiny acicular grains of microcrystalline quartz orientated perpendicular to the bedding. In some specimens the acicular crystals of microcrystalline quartz were observed to separate cross-fibre crocidolite from adjacent magnetite laminae. In these specimens the microcrystalline crystals of quartz could be remarkably elongated and intergrown with the crocidolite (Plate IX).

Laminae composed solely of chert or with chert as the chief constituent alternate with either magnetite, carbonate, minnesotaite or stilpnomelane in the banded ironstone. The borders between chert and adjacent laminae of magnetite are sharply defined as a rule, but in some specimens idioblastic crystals of magnetite are sparingly distributed within laminae of chert in close proximity with laminae of magnetite. The same applies to contacts between laminae of chert and stilpnomelane.

Where laminae of chert are intercalated with carbonate-bearing laminae, the contacts between the different laminae may be sharp, but more often the chert is accompanied by interstitial carbonate and "vice versa". Chert laminae often alternate with minnesotaite-bearing laminae (HH324). Although the contacts between such laminae are generally well defined, acicular crystals of minnesotaite generally tend to grow perpendicular to the bedding into the neighbouring laminae of chert (Plate X).

The banded ironstone from Precambrian ironstones is generally regarded as a rock which consists chiefly of alternating laminae of chert and magnetite. If this was

the case it would have contributed considerably to the solution of the problem of the origin of these rocks. However, the sequence of precipitation of the different materials constituting these rocks is more complex. Alternating laminae may be composed of either silica, magnetite, carbonate, stilpnomelane or minnesotaite or of a mixture of two or more of these minerals and the sequence of the laminae varies from specimen to specimen. In addition one can distinguish between chert laminae of different grain-size in the same thin section. This variation in grain-size may be gradual but it is more often sharp and well defined.

The variation in the grain-size of microcrystalline quartz in adjacent chert laminae or in separate laminae can hardly be explained by metamorphic influences because of the very sharp contacts often seen between laminae of fine-grained and coarse-grained microcrystalline quartz. It is believed that the difference in grain-size is a primary character and was caused by changes in the rate of precipitation of silica owing to rapidly changing physico-chemical conditions and/or rapid changes in the concentration of silica.

A fast rate of precipitation would result in the accumulation of fine-grained material in contrast with larger grains caused by slow precipitation. A lamina of carbonate is often bordered on both sides by fine-grained microcrystalline quartz or by fine-grained quartz on one side and coarse-grained quartz on the other side. Such anomalous features apparently indicate that the difference in grain-size was not only governed by a variation in the concentration of the silica but was also controlled by rapid variations in physico-chemical conditions during the process of precipitation.

(b) Magnetite

Magnetite, like chert, represents one of the essential constituents of the banded ironstone. The mineral is generally present as discrete idioblastic crystals or as closely interlocking crystal aggregates arranged in thin laminae parallel to the bedding. The mineral is also found as disseminated crystals in laminae of chert, carbonate, stilpnomelane and riebeckite (HH263).

Individual laminae of magnetite in the rock vary in thickness from 0.05 mm to about 4.0 mm. The thicker laminae of magnetite are actually composed of very thin laminae of magnetite varying in thickness from 0.05 to 0.45 mm. The latter are separated from one another by equally thin partings of chert and/or carbonate, arranged parallel to the bedding.

Under the ore-microscope the magnetite displays a distinct white-grey colour which is attributed to its very low titanium content. Analyses of bulk samples of the banded ironstone show that the TiO_2 -content of the rock does not exceed 0.2 per cent (Table 32). Crystal faces are well developed as a rule and no definite signs of martitization could be traced in specimens of fresh rock taken from below the zone of oxidation. In specimens from outcrops, almost all of the magnetite has been altered to hematite (martite). The martite again is largely altered to goethite. Martite lamellae parallel to the (111) crystallographic direction of magnetite are common. The original crystal faces of the magnetite are retained even where the mineral is almost completely replaced by martite and goethite.

The magnetite laminae are built of dense aggregates of crystals or are composed of single crystals or clusters of magnetite separated by chert and/or carbonate. Although the individual laminae are generally continuous in thin section they frequently pinch out within very short distances. In this way discontinuous streaks, often not longer than 3.0 mm are formed parallel to the bedding. Under low magnification the linear orientation of a series of short streaks of densely packed magnetite crystals appears as one continuous lamina. In the plane of the bedding the discontinuous streaks of magnetite are separated from one another by chert or carbonate or both. Where magnetite, chert and carbonate occupy the same microstratigraphical position in the rock this feature points to a possible rearrangement of material during or immediately after deposition. The rearrangement could have resulted under the influence of weak currents or waves.

Although chert may separate the individual laminae of magnetite from each other, this position is dominantly occupied by carbonate. It was further observed that laminae which are chiefly composed of carbonate and which are found on the edges of magnetite laminae may have an appreciable amount of magnetite distributed through the carbonate matrix. The amount of magnetite in them decreases gradually away from the magnetite laminae. Where chert is found adjacent to a magnetite lamina the contact is commonly sharper and magnetite is seldom conspicuous amongst the chert grains.

In some sections the magnetite laminae are not only discontinuous, but also very irregular in thickness. In other specimens the laminae are intricately curved or folded in the same manner as in the conical structures observed in crocidolite seams. It is important to note that small-scale

folding or warping of magnetite laminae is not restricted to crocidolite seams only where, according to Cilliers (1961), such structures were caused by the growth of the crocidolite fibre. Evidence of similar structures are found in many localities where crocidolite is completely absent and where the laminae on both sides of the curved laminae of magnetite are composed of only or mainly chert. Figure 9, which is a camera lucida sketch of a warped lamina of magnetite between two laminae of chert, illustrates the considerable variation in thickness of one chert lamina and the concomitant folding of the magnetite lamina. These structures are attributed to pressure phenomena and the pinch-and-swell character of the chert lamina indicates that some lateral transfer of material had taken place. In many specimens where such folding of the laminae in the rock took place small-scale faulting, illustrated by minute fractures which run perpendicular or at steep angles to the bedding or stratification of the rock, can be observed. These fractures are filled with relatively coarse-grained chert and carbonate. The laminae of magnetite also display drag-structures on the opposite sides of such fractures. These features would imply that the rock was completely consolidated during the time of the formation of the small fold structures. Very closely spaced laminae of magnetite often diverge so as to form two separate laminae over short distances before converging again. The elongated lenses between such divergent magnetite laminae could be occupied by either chert or carbonate or both.

(c) Carbonate

Although subordinate to chert and magnetite, carbonate is a common mineral in the banded ironstone. The total amount of carbonate differs quite considerably from one specimen to another and from one layer of banded ironstone to another. It is generally more abundant in the intercalations of banded chert in the banded ironstone and is most abundant towards the base of the Banded Ironstone Substage. Laminae of carbonate usually alternate with laminae of magnetite in the banded ironstone, but it is far more abundant in the intercalations of banded chert in the banded ironstone. This suggests that during the deposition of the chert or jasper there must have been a deficiency in iron in the basin of deposition. The absence of magnetite laminae in intercalations of jasper can therefore not be attributed to changing physico-chemical conditions only. If there was a deficiency in iron hydroxides during the deposition of the banded jasper then silica and ferruginous material could hardly have been derived simultaneously from the selective weathering of

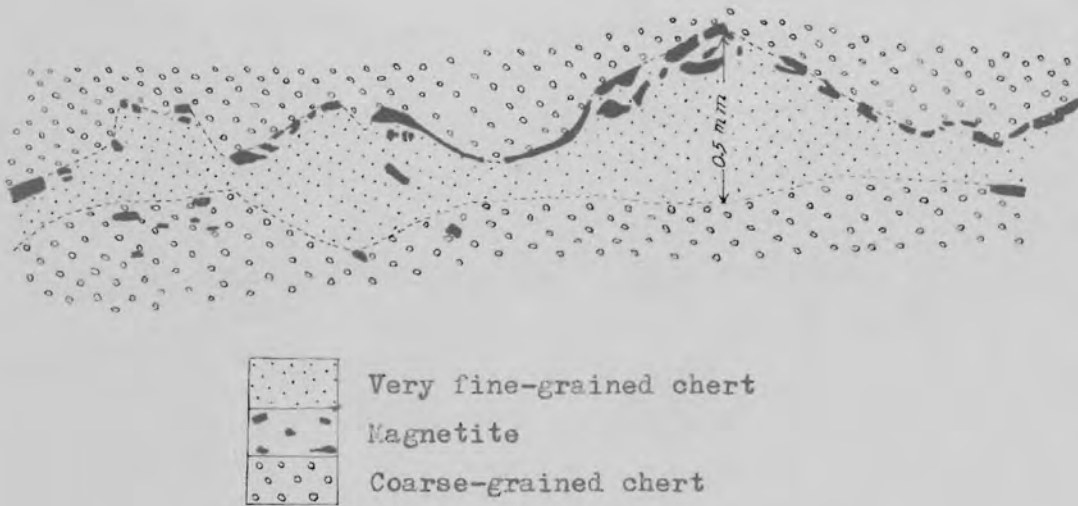


FIGURE 9

Camera lucida sketch of the microfolding of a lamina of magnetite adjacent to a lamina of fine-grained chert which displays distinct pinch-and-swell characteristics.

continental rocks.

The carbonate is present as:

- (i) Interlocking xenoblastic to ididioblastic grains and crystals constituting individual laminae.
- (ii) as isolated ididioblastic rhombs in chert laminae where it originated through the replacement of chert, and
- (iii) as ididioblastic to xenoblastic crystals in minute fractures which transgress the bedding of the banded ironstone.

(i) Laminae which are composed chiefly of carbonate range in thickness from around 1.4 mm to a maximum of 4.7 mm. The laminae often contain scattered crystals of magnetite and also interstitial grains of microcrystalline quartz. The chert which accompanies the carbonate is generally concentrated in irregular patches throughout the carbonate laminae. Carbonate constitutes individual laminae, but more commonly is found interstitial to magnetite.

(ii) Relatively large ididioblasts of carbonate, scattered in a matrix of chert, are quite common in most of the chert laminae. In this mode of occurrence various stages of replacement of chert by carbonate may be observed (Plates XI and XII). The growth of these rhombs commences with the replacement of a single grain of microcrystalline quartz. Adjacent grains are subsequently attacked in such a manner that a group of quartz grains are completely enveloped by carbonate. In thin section the progressing mantle of carbonate appears as two prongs which extend from the original point of replacement. After enveloping the grains of microcrystalline quartz and while the carbonate crystal is assuming a rhombohedral outline the enclosed grains of microcrystalline quartz are gradually replaced (Plate XI). Many of the ididioblasts of carbonate show various stages in the replacement of the chert in the cores in which any number, from one to several, grains of quartz are still intact. The carbonate rhombs display a frosty appearance and under strong magnification they appear to be built up of many small crystals, each having the approximate size of the original quartz grains which were replaced (HH273). The carbonate which replaces chert includes magnetite poikiloblastically (Plate XII) and crystallised obviously later than most of the carbonate which constitute separate laminae in the rock. Such replacement could be attributed to the percolation of low-temperature, alkaline solutions.

(iii) Carbonate, like chert, occupies late fractures in the banded ironstone, either alone or accompanied by chert. Where accompanied by chert the carbonate crystals are generally found along the walls of the fractures and micro-crystalline quartz occupies the centre or core of the fractures. In these fractures the carbonate crystals are orientated with their major axes perpendicular to the walls of the fracture similar to the orientation of the micro-crystalline quartz.

Carbonate staining techniques on specimens of banded ironstone from the Northern Region proved the carbonate to be chiefly calcite, dolomite and siderite. This was confirmed by X-ray analysis. The staining techniques applied involved the following chemicals and were administered according to the methods described by Friedman (1959, p. 88) and Warner (1962, p. 35-37).

- (i) Alizarine Red S, cold; Calcite - deep red.
- (ii) Alizarine Red S plus a 30% NaOH solution and boiled; Dolomite - purple; siderite - dark brown to black.
- (iii) A 2% HCl solution plus a 2% potassium ferrocyanide solution, heated; Siderite and dolomite - dark blue.
- (iv) A hot, concentrated solution of caustic potash plus hydrogen peroxide; siderite - brown.

The different staining methods indicated the general presence of siderite in the dominantly carbonate-bearing laminae of the banded ironstone. It further shows that siderite usually occupies all of a specific carbonate-bearing lamina whereas calcite is commonly the chief constituent in other laminae and is often accompanied by dolomite. Associations of dolomite and calcite are especially abundant in the lower portion of the Banded Ironstone Substage.

(d) Minnesotaite

Minnesotaite is an important constituent of the banded ironstone in both the Northern and the Southern Region although it is generally only abundant in certain layers in the succession. In the Northern Region minnesotaite is found chiefly in the lower portion of the banded Ironstone Substage whereas in the Southern Region the mineral is one of the chief constituents in the uppermost portion of this substage. It is especially abundant in rocks belonging to the Westerberg Beds where the mineral is usually the principal constituent.

In the lower portion of the Banded Ironstone Substage of the Southern Region minnesotaite is virtually absent, its place being taken by stilpnomelane.

The mineral appears as microscopic plates and more often as needles, which are orientated radially or arranged in sheaths. Individual slender needles reach a maximum length of 0.19 mm, but generally they do not exceed a length of 0.05 mm. In the Westerbeg Beds, where minnesotaite forms the bulk of the rock, and in the lower portion of the banded ironstone succession in the Northern Region where it is conspicuous, the mineral is present as very fine needles and flakes so closely packed together that it forms yellowish-green, felt-like masses (HH204; HH324).

In the Northern Region minnesotaite is commonly restricted to separate laminae which are intercalated with chert laminae or the mineral is found interstitial with chert. Where it constitutes separate laminae the needles attain their maximum length on the outer edges of the laminae. All needles on the contact with a lamina of chert project well in to the latter at about right angles to the bedding (Plate X). The larger needles are pale yellow to pale green and exhibit faint pleochroism. On account of the small sizes of the needles optical properties are difficult to determine accurately. The following optical properties were determined on the large needles in sodium and ordinary light.

$2V_a = 0$ to small	δ/c	= Zero
$n_c = 1.580 (\pm .005)$	Blongation	= positive
$n^A = 1.620 (\pm .005)$	Birefringence	= moderate to high

An X-ray diffraction pattern of a specimen in which minnesotaite is the most important mineral is given in Table 21. After grinding the specimen the material was suspended in water and all the magnetite was removed with a hand magnet while the suspension was continuously stirred.

(e) Stilpnomelane

Two kinds of stilpnomelane are present in the banded ironstone of the Lower Griquatown Stage, viz. ferri- and ferrostilpnomelane. Ferristilpnomelane is the most abundant and is one of the essential minerals in the lower portion of the Banded Ironstone Substage in the Southern Region. This mineral is comparatively scarce in the banded ironstone of the Northern Region and where present it forms tiny flakes associated with magnetite. In the Southern Region the mineral commonly constitutes separate laminae or fills interstitial cavities in magnetite laminae almost completely.

Table 21. - X-RAY DIFFRACTION PATTERN OF MINNESOTAITE-
BEARING ROCK

(Cobalt radiation. Magnetite removed. Philips X-ray diffractometer. Spc. HH318.)

Minnesotaite-bearing banded ironstone from approx. 400 ft. above Dolomite, Pomfret		Minnesotaite Gruner (1944a) p. 366		Quartz A.S.T.M. Card 5-0490		Chlorite A.S.T.M. Card 11 - 11	
dÅ	I/I ₀	dÅ	I/I ₁	dÅ	I/I ₁	dÅ	I/I ₁
15.31	10					15.0	100
9.50	90	9.53	100				
7.14	5					7.10	100
4.77	20	4.77	10				
4.58	5					4.70	100
4.25	40			4.26	35		
3.57	5					3.59	100
3.50	5	3.50	10				
3.35	100			3.34	100		
3.19	80	3.18	50				
2.74	10	2.75	5				
2.65	5	2.65	5				
2.52	20	2.52	20				
2.45	10			2.46	12		
2.40	5	2.40	10				
2.31	5	2.31	5				
2.28	10			2.28	12		
2.24	5			2.24	6		
2.20	5	2.22	10				
2.12	5	2.11	5				
2.05	5	2.01	10				
1.98	5			1.98	6		
1.90	5	1.92	10				
1.82	20			1.82	17		
1.67	5	1.66	10				

Much of the ferristilpnomelane, especially where it is present as relatively large flakes, in the banded ironstone may readily be mistaken for biotite. Like biotite, the mineral is optically negative and has a small optic axial angle - often approaching zero. It shows strong absorption with the basal cleavage parallel to the vibration direction of the polarizer and parallel extinction. The pleochroism is similar to that of biotite. The mineral displays no mottled effect on extinction.

Ferristilpnomelane is often accompanied by ferrostilpnomelane which is pleochroic in yellow and green. The following optical properties were determined in sodium light and ordinary light on ferrostilpnomelane in a core-sample obtained from about 350 feet above the Dolomite at Pomfret (B4):

$$2V\alpha = 0 - +5^\circ \quad \text{Pleochroism} =$$

$$\alpha = + 1.584(\pm 0.005) \quad \alpha = \text{yellow-brown}$$

$$\gamma = \beta = 1.612(\pm 0.005) \quad \gamma = \beta = \text{Deep brown}$$

An X-ray diffraction pattern of a stilpnomelane-bearing rock which contains mainly ferristilpnomelane is given in Table 22.

Table 22. - X-ray diffraction pattern of stilpnomelane-bearing rock from Botha, Postmasburg District (Cobalt radiation, Phillips X-ray diffractometer. Spc. HH 263)

Stilpnomelane-bearing rock		Stilpnomelane Gruner, 1937, p. 919		Chlorite Brindley 1951, p. 320	
$d\text{\AA}$	I/I ₀	$d\text{\AA}$	I/I ₁	$d\text{\AA}$	I/I ₁
12.01	100	11.90	100		
7.09	10			7.0-7.2	90
4.74	10	4.74	5		
4.12	5	4.14	5		
4.04	10	4.05	50		
3.55	10			3.52-3.58	100
3.34	5	3.35	10		
3.04	5	3.04	40		
2.72	20	2.69	20		
2.57	40	2.55	40		
2.36	10	2.34	30		
2.11	10	2.11	20		
1.57	20	1.58	30		
1.56	20	1.56	30		

(f) Chlorite

Most of the specimens on which X-ray determinations were carried out contained subordinate amounts of chlorite. Where concentrated in appreciable amounts the mineral is present as a fine-grained greenish mass with low birefringence, mottled extinction and often shows abnormal interference colours characteristic of the interference colours of penninite.

A X-ray diffraction pattern of a specimen in which appreciable amounts of chlorite accompanies minnesotaite is given in Table 23.

(g) Tremolite-Richterite

Small prismatic to acicular crystals of a mineral which strongly resembles tremolite were observed in specimens of banded ironstone obtained from localities away from intrusive dykes and sills of diabase. These small crystals show positive elongation, inclined extinction ($\gamma_{Ac} = +14^\circ$) and are biaxial negative ($2V\alpha = \text{large}$).

Table 23. - X-ray diffraction pattern of Chlorite-bearing

<u>Rock</u>							
(Co-radiation. Only strongest lines given. Magnetite removed. Philips X-ray diffractometer. Spc. HH307.)							
Chlorite-bearing banded ironstone from approx. 350 feet above Dolomite Pomfret		Chlorite (Penninite) A.S.T.M. Card II - 153		Minnesotaite Gruner (1944a, p. 366)		Quartz A.S.T.M. Card 3-0407	
$d\text{\AA}$	I/I ₀	$d\text{\AA}$	I/I ₁	$d\text{\AA}$	I/I ₁	$d\text{\AA}$	I/I ₁
9.56	100			9.53	100		
7.16	30	7.20	100				
4.77	10			4.77	10		
4.58	5	4.60	100				
4.25	50					4.26	80
3.58	20	3.54	100				
3.34	100					3.35	100
3.19	40			3.18	50		
2.79	5	2.81	70				
2.74	10			2.75	5		
2.64	10			2.65	5		
2.52	40	2.54	100	2.52	20		
2.45	30					2.46	60
2.40	10	2.40	70	2.40	10		
2.28	20					2.28	60
2.24	15	2.23	70				
2.20	25			2.22	10		
2.13	20					2.13	50
2.10	10			2.11	10		
1.98	15	1.99	100	2.01	10		
1.91	5			1.92	10		
1.82	40	1.81	50				
1.67	15	1.68	60	1.66	10		
1.60	15			1.60	10		
1.54	30	1.52	100			1.54	70

A specimen in which many of these crystals are found was analysed by means of X-rays after the extraction of the magnetite. The X-ray analysis is given in Table 24 (HH318).

A similar mineral is often found in the banded ironstone near to or on the contacts of diabase sills and dykes. It is found, for example, in relatively large amounts in specimens of banded ironstone from about one foot above the top of the uppermost diabase sill in the Pomfret area. The maximum observed dimensions of basal sections of the amphibole average 0.09 mm. The crystals are present as irregular prisms.

In a specimen obtained from near the contact of the diabase sill the mineral is optically negative, $2V_c$ is about 85 degrees and the extinction angle γ^*c varies between 12 and 28 degrees (HH311). The mineral is weakly pleochroic from colourless to pale green. The γ^* index of

the amphibole, determined in sodium light, is 1.677. These optical properties correspond best with that of richterite (Deer, et al 1963, p. 352). Why actinolite did not crystallize in these iron-rich rocks is not clear.

Table 24. - X-ray Analysis of a Tremolite-bearing Rock

(Cobalt radiation. Magnetite removed. Philips X-ray diffractometer. Spc. HH318.)

Tremolite-bearing rock from about 500 feet above the Dolomite Fomfret		Grammatite (Tremolite) Johansson (1930, p. 43)		Tremolite Comefero and Eitel (1951, p. 369)		Quartz A.S.T.M. Card 3-0407	
$d\text{\AA}$	I/I ₀	$d\text{\AA}$	I/I ₁	$d\text{\AA}$	I/I ₁	$d\text{\AA}$	I/I ₁
8.53	100	8.49	50	8.41	30		
4.55	10	4.52	50	4.50	20		
4.24	10	4.21	20			4.26	80
3.40	10	3.38	50	3.38	20		
3.34	20					3.35	100
3.16	80	3.13	100	3.13	40		
2.96	20	2.94	50	2.94	20		
2.84	20	2.80	50	2.80	20		
2.73	40	2.71	100	2.71	40		
2.62	20	2.59	50	2.59	20		
2.54	20	2.53	80	2.52	30		
2.36	20	2.33	50	2.33	20		
2.28	10	2.27	20	2.26	10	2.28	60
2.18	20	2.16	50	2.16	20		
2.05	10	2.04	20				
2.03	10	2.01	50	2.01	20		
1.72	10	1.74	20				
1.67	20	1.65	50	1.65	20		
1.62	20	1.62	20				
1.59	10	1.58	50	1.57	20		
1.52	20	1.51	50	1.51	20	1.54	70
1.49	30	1.50	50	1.50	20		
1.46	40	1.44	80	1.44	30		

(h) Acmite

Minerals of the pyroxene group are seldom found in the banded ironstone where the rock has not been affected by thermal metamorphism caused by subsequent emplacement of intrusive rocks. Although the mineral is rarely found in these rocks, pyroxene was observed in specimens which originated from localities outside thermal-metamorphic aureoles. Rogers and Du Toit (1908, p. 87) described a faint pleochroic pyroxene from near Prieska Poort in the Southern Region and noted that the mineral has an extinction angle as high as 12 degrees. They concluded that the mineral belongs to the aegirine-augite group.

Du Toit (1945, p. 175) described acmite in the banded ironstone from Westerberg as well as from near Buisvlei and remarked that this mineral is now known from at least

six localities in the Lower Griquatown Beds and that its occurrence is apparently restricted to a thin stratum in the banded ironstone - such as might possibly mark a stratigraphical horizon.

Cilliers (1951, p. 30) did not observe pyroxene in rocks outside metamorphic aureoles of intrusive dykes and concluded that minerals such as biotite, grunerite, garnet and pyroxene are not present in the rocks of the Koegas—Prieska area.

During the present investigation acmite was found in thin sections made of bore-hole cores from Westerberg (Q2) as well as in rocks which outcrop on Orange View (R3) in the Southern Region. In both localities the specimens investigated are found well below the thick intrusive sill in this area and not near to diabase dykes. From observations made on Orange View the pyroxene-bearing rocks are restricted to thin intercalations in the banded ironstone (HH511).

In the thin sections investigated acmite is generally restricted to separate laminae in which the acmite crystals are distinctly elongated parallel to the bedding. In thin section cut perpendicular to the bedding of the rock, basal sections of the pyroxene display good cleavages approximately at right angles to each other.

The grain size is variable, but grains measuring as much as 8 mm x 1.5 mm are not uncommon. In specimens from Orange View (Q3) the crystals are even larger, some measuring about one inch in length. Du Toit (1945, p. 175) described prisms, up to 1½ inch long, in outcrops on the boundary between Buisvlei and Geduld (Portion of Middelwater, R3).

The acmite shows no or very faint pleochroism and high birefringence. The optic axial angle, measured with the Fedorow Stage, varies from 54 to 62 degrees and the extinction angle α/c is about 4 degrees. Du Toit (1945) reported the acmite from Westerberg to have the following refractive indices:

$$\begin{aligned} \alpha &= 1.765 - 1.778 \\ \beta &= 1.82 \\ \beta - \alpha &= 0.055 \end{aligned}$$

The α index was determined in sodium light and a value of 1.780 was obtained.

The mineral includes idioblastic magnetite crystals and appears to be replaced by both fibrous riebeckite and ferristilpnomelane. The replacement took place parallel to the prismatic cleavage of the acmite (HH511). Distorted

cleavage faces, observed by Du Toit in acmite from near Goduld, led him to conclude that the mineral crystallised during the late stages of folding. He also found "needle" crocidolite to penetrate acmite crystalloblasts and concluded that the crystallization of riebeckite took place at a higher grade of metamorphism than the acmite. The present investigation did not supply sufficient evidence to support this view.

(j) Muscovite

A mineral distinctly pleochroic in bright-green and yellow was observed in thin sections made of bore-core from Westerberg (HH262). Crystals of the mineral have an average length of 0.93 mm and are seldom wider than 0.11 mm. They are haphazardly distributed in a matrix of ferri-stilpnomelane, accompanied by accessory magnetite and chert.

The mineral is biaxial negative with $2V$ practically zero, it shows parallel extinction and has negative elongation. It has an excellent basal cleavage. In view of the extraordinary optical properties of the mineral an X-ray analysis, which proved the mineral to be muscovite, was carried out (Table 25).

(k) Riebeckite

Riebeckite is found as slender needles up to 0.25 mm long, but seldom more than 0.005 mm wide. Where separate layers in the banded ironstone are composed chiefly of riebeckite the riebeckite needles are orientated at random and intimately interlocked (Plate XIII). In many thin sections the riebeckite needles radiate from a core of magnetite or from a core composed of a dense mass of riebeckite with the same optical orientation. In the latter case the crystals are present as sheaf-like aggregates (Plate XX). The magnetite from which needles of riebeckite radiate represents the centre of initial crystallization of the riebeckite because the iron necessary for the formation of riebeckite, was derived from iron-rich centres. Several stages in the formation of riebeckite, in which magnetite participated, can be observed. In many places the needles of riebeckite radiate from idio-blastic crystals of magnetite whereas in other places irregular grains of magnetite, often only a very minute grain, remained in a dense mass of riebeckite which displays optical continuity (Plate XX, HH273).

Riebeckite is also found as prominent lath-shaped crystals which display a strong cleavage parallel to the length of the laths. The lath-shaped crystals of riebeckite

Table 25. - X-ray Analysis of Bright-green Muscovite in
Banded Ironstone from Westerberg

(Cobalt radiation. 114 mm. X-ray camera. HH262)

Bright green muscovite		Muscovite Nagelschmidt (1937, p. 516)		Muscovite Smith and Yoder (1956, p. 230)	
dÅ	I/I ₀	dÅ	I/I ₁	dÅ	I/I ₁
10.06	100	9.98	s	10.08	vs
4.96	5	5.0	s	5.04	w
4.51	80	4.49	s	4.49	s
4.36	5			4.35	w
4.16	5			4.12	vw
3.90	10	3.91	w		
3.76	10	3.73	w	3.66	m
3.49	15	3.50	m		
3.33	100	3.33	vs	3.36	vs
3.23	20	3.20	m		
2.99	20	3.00	m	3.07	m
2.88	20	2.88	m	2.93	vw
2.79	20	2.80	m	2.69	vw
2.58	80	2.57	vs	2.56	s
2.53	5	2.475	wb	2.55	w
2.40	80	2.385	m		
2.25	5	2.28	m		
		2.19			
2.13	25	2.134	s		
1.995	15	1.995	vs		
1.66	25	1.65	wb		
1.62	5				
1.56	5	1.546	vw		
1.52	25	1.523	vw		

s = strong
 vs = very strong
 w = weak
 vw = very weak
 m = medium strong
 wb = weak and broad
 | = edges of broad

are commonly arranged parallel to the bedding of the banded ironstone and is almost invariably found within or next to magnetite laminae. These lath-shaped crystals are frequently inclined to the bedding, the crystals, if of large dimensions, then grew across one or more laminae of different mineral composition. The lath-shaped riebeckite crystals commonly include grains and crystals of magnetite in poikiloblastic manner, which suggests that the riebeckite crystallised later than the magnetite (Plate XIV, HH295).

Crystals of needle-like riebeckite or mass-fibre crocidolite which are found adjacent to magnetite laminae are often curved around magnetite crystals in such a way as to suggest simultaneous crystallization of magnetite and massive riebeckite and the continuation of the crystallization of magnetite after the crystallization of the riebeckite (Plate XV). The larger crystals of riebeckite are strongly pleochroic in dark-blue and green-blue, but in the very slender needles pleochroism is seldom distinct.

Separation of individual crystals of riebeckite for optical determinations is extremely difficult. Markedly elongated lath-shaped crystals occurring in a cross-fibre

habit like crocidolite, were obtained from the Cairn Brae asbestos mine (S4), Southern Region and these crystals show the following optical properties, determined in sodium and ordinary light:

$2V_{\gamma}$ = Large ($\pm 80^{\circ}$)	Extinction: Parallel
$\alpha = 1.688$	Elongation: Negative (Length fast)

Pleochroism: α : Blue

γ : Pale-blue to colourless

(1) Crocidolite

The crocidolite is a typical cross-fibre asbestos with a composition similar to that of riebeckite (Table 36). The hair-like fibres are extremely fine and even under very high magnification the fibres still appear as bundles.

The crocidolite fibres are elongated parallel to the c-crystallographic axes and are commonly orientated in fibre bundles. Owing to the fineness of the fibres it is not possible to determine the optical properties of individual crystals, but because small bundles of crocidolite are commonly in optical continuity certain optical measurements on such bundles are possible.

The elongation of crocidolite fibres not affected by thermal metamorphism caused by the intrusion of diabase dykes and sills or by weathering is invariably negative. Crocidolite which is even slightly affected by the heat from later intrusions is positively elongated, and the colour of the fibres is changed to pale green or black (HH501, 53). Even slight alteration under thermal-metamorphic conditions therefore affects the optical properties of the crocidolite.

Frankel (1953, p. 78) reported positively elongated fibres amongst negatively elongated crocidolite fibres in specimens from a number of localities in the Northern Region and pointed out that the positively elongated fibres have paler and different pleochroic colours. He concluded that unless these fibres are amosite they must have undergone alteration sufficient to induce different optical properties. He does not elaborate on the conditions which are responsible for the alteration.

In the present study it was found that at various localities and from different asbestos-bearing zones the optical properties of crocidolite from below the level of oxidation differ quite appreciably with regard to the refractive indices, but the hair-like crystals are always

negatively elongated (length fast). The optic axial angle is always large, in excess of 70 degrees and the pleochroism is very similar in the different specimens investigated.

The value for ω differs only slightly in eleven of the twelve specimens of crocidolite on which determinations of the refractive indices were carried out (Table 26). In a specimen from Pomfret (Table 26, I) the value for ω is relatively low compared with the values obtained on the remaining specimens. A chemical analysis of the same material from which this specimen was taken shows that the crocidolite contains 6.23% Al_2O_3 (Analysis I) compared with an Al_2O_3 -content which ranges from 0.38 to 4.75 per cent in the remaining eleven samples (Table 36). The value for ω in Specimen No. I leans towards that of crossite which has an intermediate composition in the glaucophane-riebeckite series (Winchell, 1956, p. 441). The difference in refractive indices could therefore probably be attributed to the amount of aluminium present in the composition of the crocidolite.

Table 26. - The optical properties of crocidolite from the Northern Cape Province

(Refractive indices determined in sodium light)

Specimen number*	2V ω (± 0.005)	ω	Δn_c	Pleochroism	
				X	Z
I	Large	1.669	0°-7°	Blue-green	Purple-black
III	"	1.686	0°-2°	Blue-green	Blue-grey
IV	"	1.696	0°	Blue-green	Yellow-green
V	"	1.693	0°-2°	Blue-green	Blue-violet
VI	"	1.690	0°	Blue-green	Blue-black
VII	"	1.691	0°	Blue-green	Yellow-green
VIII	"	1.694	0°-2°	Blue-green	Yellow-green
IX	"	1.689	0°-3°	Blue-green	Yellow-green
X	"	1.680	0°	Blue-green	Yellow-green
XI	"	1.694	0°	Blue	Yellow-green
XII	"	1.693	0°	Blue	Yellow-green
XIV	"	1.688	0°	Blue-green	Green-yellow

*The specimen numbers in Table 26 correspond with the numbers of samples analysed chemically (Table 35).

(m) Pyrite

Well-formed crystals of pyrite are found chiefly in the lower portion of the Banded Ironstone Substage in the Southern Area and are associated with thin intercalations of pyroclastic material in the banded ironstone. The pyrite not associated with the pyroclastic material is commonly intimately associated with mass-fibre riebeckite.

In the latter association the crystals of pyrite are disseminated in the seams of massive riebeckite but more commonly they are arranged along definite streaks parallel to the bedding of the banded ironstone. Segregations of large pyrite crystals, measuring up to a $\frac{3}{4}$ inch in diameter, are especially concentrated along the lowermost portions of seams of massive riebeckite in the banded ironstone (HH518).

V. Thermal Metamorphic Effects on Crocidolite

Crocidolite fibres affected by thermal metamorphism caused by the intrusion of mafic dykes and diabase sills generally exhibit different colours than ordinary crocidolite. It would appear that under very slight metamorphic alteration the fibre bundles turn to an ash-grey to slightly silvery colour, but still retain some of their flexibility (HH53). Under more intense thermal metamorphic conditions the fibre changes to a greenish to almost black colour and becomes progressively more brittle as the colour becomes darker (HH164, HH501). Such thermally altered crocidolite fibres are characterised by being elongated positively and by the fact that they become optically negative (Table 27).

Table 27. - Optical properties of thermally altered crocidolite

(Refractive indices determined in sodium light)

Specimen number	2V _d	n_{ω} ($\epsilon_{0.005}$)	n_{ϵ}	X	Pleochroism	Z
I	Large	1.676	0'	Yellow-green	Blue-green	
II	"	1.667	0	Pale-yellow	Yellow	
III	"	1.674	0	Pale-yellow	Pale-blue	
IV	"	1.662	0	Pale-green	Yellow-green	

I Altered crocidolite from closely above an intrusive diabase sill on Erfrust, Hay District.

II Crocidolite from same asbestos zone as specimen I on Kameelfontein, Hay District.

III Crocidolite from contact of dolerite dyke in the Riries, Kuruman District (analysis, Table 28).

IV Crocidolite from immediately above a diabase sill on Groenwater, Postmasburg District.

The refractive indices of the thermally altered crocidolite are as a rule smaller than for ordinary crocidolite (Table 27), and fall within the range given for anthophyllite (Winchell, 1955, p. 439; Deer, et al, 1963, p. 211). The size of the optic axial angle and the parallel extinction correspond with that of anthophyllite and riebeckite. The optic sign is negative and therefore differs from that of ordinary riebeckite, but corresponds with that of magnesium-rich varieties of anthophyllite. However, the mineral is pleochroic and according to the chemical composition of the thermally altered crocidolite (Table 28) it contains only a small amount of magnesium and is rich in both ferrous and ferric iron.

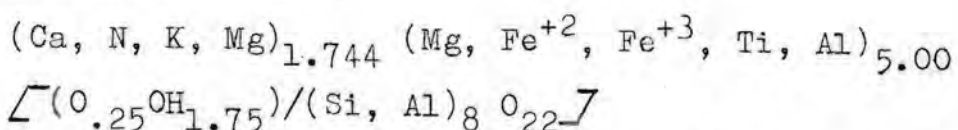
The chief difference between the chemical composition of thermally altered crocidolite and ordinary crocidolite is displayed in the amount of molecular water. In the altered crocidolite the H_2O^+ -content is 1.69 per cent (Table 28) compared with an average of 3.03 per cent and a range of 2.02 to 3.84 per cent for crocidolite (Table 35, Analyses I to VIII and X).

Table 28. - The Chemical composition and the unit-cell formula of thermally metamorphosed crocidolite on contact of diabase dyke, Riries (II) Kuruman District

Oxide	Weight per cent	Number of anions on the basis 24(O+OH)
SiO ₂	48.00	Si ⁺⁴ = 7.438
Al ₂ O ₃	3.60	Al ⁺³ = 0.657
Fe ₂ O ₃	18.28	Fe ⁺³ = 2.131
FeO	18.14	Fe ⁺² = 2.351
MgO	4.51	Mg ⁺² = 1.042
CaO	1.36	Ca ⁺² = 0.100
Na ₂ O	3.13	Na = 0.940
K ₂ O	0.21	K = 0.042
TiO ₂	0.37	Ti ⁺⁴ = 0.043
P ₂ O ₅	0.00	Total 14.744
MnO	0.00	
CO ₂	0.60	
H ₂ O ⁺	1.69	O ⁻² = 22.253
H ₂ O ⁻	0.26	OH ⁻¹ = 1.747
Total	100.15	24.00

Si = 7.438	8.00	Mg = 0.662	
Al = 0.562		Ca = 0.100	1.744
Al = 0.095		Na = 0.940	
Fe ⁺³ = 2.131		K = 0.042	
Mg = 1.038	5.00	O = 22.253	24
Ti = 0.043		OH = 1.747	
Fe ⁺² = 2.351			

Unit-cell formula.



Hodgson, et al (1964, p. 8) found that when crocidolite is heated in a neutral atmosphere the fibres lose their lustre and the tensile strength decreases at about 200°C; at about 500°C the colour fades to a greyish blue. When heated in an oxidizing atmosphere to 300°C the colour and the tensile strength undergo similar changes. At 330°C to 550°C the colour changes to a deep brown (Hodgson, et al, p. 18). X-ray analyses carried out by Hodgson and co-workers on the decomposition products of crocidolite heated in an oxidizing atmosphere showed that an oxyamphibole (referred to as oxyriebeckite) is formed at temperatures between 400°C and 790°C. Above 840°C the crystalline decomposition products include pyroxene, spinel, cristobalite and hematite. The X-ray results showed that the oxyamphibole has slightly smaller parameters than the unoxidized material, but otherwise the pattern obtained is closely similar to that of crocidolite.

The change in the colour of crocidolite when heated in neutral or oxidizing atmospheres is much the same as the colour changes found in crocidolite which is thermally metamorphosed to different degrees. The colour changes from ash-grey through greenish-brown to almost black, with an increase in the degree of thermal metamorphism. The chemical composition of brown-black metamorphosed crocidolite is given in Table 28. The calculated cation proportions according to the chemical composition of the mineral are provided (Table 28) and show that the composition of the mineral fits fairly well into the amphibole formula.

Intensely altered crocidolite was found directly on the contact of a transgressive diabase sill in the Owendale No. 1 Mine, Owendale (M2), Postmasburg District. The crocidolite in this specimen is almost completely altered to quartz (HH526). Other minerals associated with the intensely altered crocidolite are magnetite, micro-crystalline quartz, secondary calcite and probably talc.

An X-ray diffraction pattern of the thermally altered crocidolite is provided in Table 29 where the d-values are compared with those of ordinary crocidolite from the immediate vicinity and of anthophyllite. The X-ray pattern of the altered crocidolite is hardly different from that of crocidolite, but differs appreciably from that of anthophyllite.

Table 29. - X-ray diffraction pattern of Crocidolite

Metamorphosed by dolerite dyke Riries, Kuruman District

(Cobalt radiation, 57.3 mm. X-ray camera. HH164)

Thermally metamorphosed crocidolite Riries Kuruman District		Anthophyllite Salotti, Amer. Min. 47, 1962, p. 1061		Crocidolite Eldoret, Kuruman District	
$d\text{\AA}$	I/I ₀	$d\text{\AA}$	I/I ₁	$d\text{\AA}$	I/I ₀
8.43	100	8.24	100	8.42	100
4.90	10	5.01	16	4.89	20
4.53	20	4.45	21	4.52	40
4.27	10	4.12	12		
3.89	10	3.88	14	3.89	10
3.42	20	3.64	21	3.42	40
3.35	80	3.35	35		
3.27	10	3.22	70	3.26	20
3.11	40	3.05	85	3.11	70
2.97	20	2.87	18	2.97	10
2.79	10	2.75	20	2.79	10
2.73	60	2.67	37	2.72	80
2.60	20	2.57	40	2.60	40
2.53	60	2.55	38	2.53	60
2.45	10	2.50	36		
2.32	10	2.32	17	2.32	20
2.26	10	2.28	17	2.26	20
2.17	10	2.16	17	2.17	20
2.10	10	2.14	19	2.08	10

VI. Layers of pyroclastic material

Reference has been made to the numerous thin dark layers which are found intercalated in the banded ironstone and which are especially numerous in the asbestos-bearing zones of the area. These layers, generally referred in the mines to as "mudstone" or "siltstone" and in drill-core often as "shale", are commonly used as Marker-beds in the mines. Because of their unique mineralogical composition and their characteristic textural features these layers are given special attention.

The layers range in thickness from less than one quarter inch to a maximum observed thickness of two feet. In bore-holes more than 10 feet of "shale" were intersected.

The rock in unaltered layers is lightly greenish in colour, for example the "Khakiband" in the Whitebank mine, but most of the layers have a green-black to pitch-black colour and are extremely fine-grained. Where exposed in mine-workings the unweathered material is present as massive, brittle layers with a subconchoidal fracture. Near the surface they are weathered to reddish limonitic clay or to a pale-green nontronitic claystone.

The dark colour of the unweathered material is mainly due to the fine-grained texture of the material which is commonly composed almost entirely of yellow-green ferrostilpnomelane or brown to black ferristilpnomelane. Accessory minerals are magnetite, riebeckite, chert and carbonate (HH 167).

Well preserved shards have been observed in thin sections made of some of the layers. The shards are actually represented by stilpnomelane pseudomorphs after shards. Only the larger shard structures appear to be preserved; the smaller shards had their identity destroyed by the spherulitic crystallisation chiefly of stilpnomelane. The shards are often accompanied by elongated axiolitic structures. The axiolites consist chiefly of minute needles of ferrostilpnomelane oriented at right angles to a median axis (Plate XVI; HH 148).

Several specimens from the dark bands were investigated by means of X-ray diffraction in the Geological Survey laboratory and most of them are composed chiefly of stilpnomelane. Many of the specimens treated show only a few peaks on the X-ray diffraction pattern which suggest that much of the minute isotropic grains in the matrix of stilpnomelane represent volcanic glass.

The needles of stilpnomelane have an average length of 0.02 mm and are generally less than 0.003 mm wide. They grow in minute spherulites and form a yellowish-green felt-like matrix.

Stilpnomelane in association with pyrite, chlorite, albite and quartz has been reported in metamorphosed schistose pyroclastic rocks from the Omi District, Japan (Banno 1958). Seki (1958) reported stilpnomelane in metamorphosed rocks derived from mafic pyroclastic and igneous rocks from the Kanto Mountains, Central Japan. Interesting about the latter occurrence of stilpnomelane is its association with soda-amphiboles, chiefly glaucophane but also magnesioriebeckite and riebeckite. (Seki 1958, p. 240).

In a recent publication by La Berge (1966a) attention is drawn to the occurrence of altered pyroclastic rocks, now composed almost entirely of ferrostilpnomelane, and which are found intercalated in the banded ironstone of the Hamersley Range, Western Australia. According to La Berge the pyroclastic layers (locally called black shales) are common in at least the lower 1,000 feet of the Brockman Iron-Formation in the Wittenoom Gorge area of the Hamersley Range. Their presence indicate that several periods of volcanism took place during the sedimentary history of the iron-formation (p. 149).

The same author (1966b) subsequently published a paper on pyroclastic rocks in South African ironstones. In this paper attention is also drawn to pyroclastic layers, composed dominantly of stilpnomelane, which are present in the Pretoria Series.*

The chemical composition of one of the tuffaceous bands intercalated in the banded ironstone is provided in Table 30 (Analysis I). A study of thin sections of this particular specimen show that the stilpnomelane in the rock is almost entirely the ferrostilpnomelane variety. Minute grains of chert, riebeckite and magnetite and tiny flakes of chlorite were also observed in thin sections of the specimen.

* The investigation of which the results are presented in this paper was carried out and the present paper was prepared before the appearance of Dr. La Berge's paper.

Table 30. - Chemical composition of stilpnomelane-bearing tuffaceous rock from Riries Asbestos Mine and of ferrostilpnomelane

	I	II	III	IV	V
SiO ₂	41.65	40.4	44.40	44.77	48.03
Al ₂ O ₃	6.92	9.9	6.2	6.32	6.48
Fe ₂ O ₃	8.93	3.9	3.2	20.79	4.12
FeO	22.37	26.9	23.6	12.83	22.88
MgO	7.36	7.8	7.60	4.01	4.94
CaO	0.55	0.1	n.d.	0.10	0.83
Na ₂ O	0.51	0.1	0.56	0.07	Nil
K ₂ O	3.47	7.20	3.3	3.31	0.83
H ₂ O ⁺	5.97	4.68	6.15	5.64	6.90
H ₂ O ⁻	1.10	0.46	3.0	1.96	2.64
CO ₂	0.00	n.d.	n.d.	n.d.	n.d.
TiO ₂	0.41	0.1	n.d.	0.04	0.23
P ₂ O ₅	0.72	0.03	n.d.	n.d.	n.d.
MnO	0.05	0.40	0.04	0.21	2.67
Total	100.01	101.97	97.65	100.05	100.55

- I. Ferrostilpnomelane from tuffaceous layer, Second Lower Asbestos Zone, Second Cut, Hanging-wall Marker, Riries Mine, Kuruman District. Analysts E.C. Hauman and J.F. Dry, Soils Research Institute.
- II. Ferrostilpnomelane from near Koegas, South Africa. Reported by La Berge (1966b, p. 580).
- III. Ferrostilpnomelane from the Hamersley Range, Australia. Reported by La Berge (1966, p. 159).
- IV. Stilpnomelane from the Mesabi Range, Minnesota. Reported by Gruner (1937, p. 913).
- V. Ferrostilpnomelane from Western Otago, New Zealand. Reported by Hutton (1938, p. 184).

At the Koretsi South Mine (H1) a number of closely spaced seams of pyroclastic material yielded on recrystallization a peculiar fibrous mineral. The seams range from less than a quarter inch to just over one inch in thickness and recrystallization of the material in them is restricted to local areas of intense folding. The fibrous crystals are shiny and dark when fresh and grow at right angles to the bedding of the enclosing rock (Plate XXII). On exposure they rapidly change colour

from black to yellow-green (HH 356).

The fibrous crystals often attain lengths exceeding 1 mm and are arranged parallel to one another. In thin section the mineral appears fibrous in some portions of the section and flaky in other spots. In crushed powders of the material the crystals are commonly well elongated parallel to β .

In thin section the fibres are yellow-green to dark green and fairly strongly pleochroic in yellow-green (β) and olive-green (β'). Optical figures are as a rule very poor, but in some of the larger flaky portions the mineral is biaxial negative with $2V\alpha$ large. The mineral displays strong birefringence and has apparently only one strong cleavage direction with α approximately perpendicular thereto. The following refractive indices of the mineral were determined in sodium light:

$$\begin{aligned}\beta &= 1.612(+0.005) \\ \beta' &= 1.646(+0.005)\end{aligned}$$

The chemical composition of the fibrous material obtained from a seam approximately $1\frac{1}{2}$ inch thick is given in Table 31. Microscopic studies of thin sections showed that the material analysed is probably contaminated with interstitial goethite and silica.

The chemical composition of the material given in Table 31 and the optical properties determined on the fibres correspond closely with that of morencite, a fibrous hydrated ferric silicate first described by Lindgren and Hillebrand (1904) and later by Larsen and Steiger (1928).

An X-ray analysis of the fibrous material was carried out by the Ceramic Unit of the C.S.I.R. and the following results were obtained:

- (i) An X-ray diffraction pattern of an orientated section shows two peaks; a strong one at 9.75 \AA and a weaker one at 10.77 \AA .
- (ii) The orientated section treated with glycol gave a strong peak at 17.31 \AA and a second order peak at 8.73 \AA .
- (iii) An orientated section heated for one hour at 500°C gave a sharp peak at 9.688 \AA .
- (iv) The orientated section treated with MgCl_2 and then with glycerol gave a well-defined peak at 18.39 \AA and a second order peak at 9.32 \AA .

Table 31. - Chemical composition of a fibrous mineral

	I	II
SiO ₂	48.55	45.74
Al ₂ O ₃	0.00	1.98
Fe ₂ O ₃	31.02	29.68
FeO	2.50	0.83
MgO	2.90	3.99
CaO	1.29	1.61
Na ₂ O	0.16	0.10
K ₂ O	0.03	0.20
H ₂ O ⁺	7.06	5.08
H ₂ O ⁻	5.76	8.84
CO ₂	0.00	n.d.
TiO ₂	0.14	tr.
P ₂ O ₅	0.14	0.18
Cl	0.22	n.d.
F	0.00	n.d.
FeS ₂	n.d.	0.62
MnO	0.61	tr.

- I. Fibrous material in seam of recrystallized pyroclastic material from Koretsi South Mine, Kuruman District.
 Analysts, E.C. Haumann & J.F. Dry, Soil Research Institute.
- II. Morencite from Arizona. Analyst W.F. Hillebrand (Larsen and Steiger, 1928, p. 6).

(v) An orientated section treated with MgCl₂ only gave a fairly well-defined peak at 9.86 Å.

According to the X-ray analysis the mineral is not nontronite as anticipated before the X-ray and the microscopical investigation were carried out. According to the chemical composition provided in Table 31 and the refractive indices of the mineral it is tentatively determined as morencite until further detailed work is done in collaboration with the Ceramic Unit of the C.S.I.R. From the chemical analysis it is clear that the mineral is a hydrated ferric silicate and probably closely related to nontronite.

The mode of occurrence of the morencite is of importance. It is found as fibres perpendicular to the bedding of the host-rock, similar to the mode of occurrence of cross-fibre crocidolite and, more important, it is found only in restricted localities which were intensely folded.

Table 32. - Chemical Analyses and Niggli Values of Banded Ironstone and Related Rocks from the Northern Cape Asbestos Field

Chemical Analysis

Sample number	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	H ₂ O ⁺	H ₂ O ⁻	CO ₂	TiO ₂	P ₂ O ₅	Cl	F	S	MnO	Minus O=Cl, S	Total oxides	Total Fe
I	44.20	0.14	3.00	33.82	2.83	2.22	0.93	0.47	3.42	0.05	7.80	0.18	0.18	0.07	0.00	0.05	0.18	0.04	99.50	28.39
II	39.63	0.19	6.51	32.49	2.80	2.31	1.83	0.45	3.19	0.04	8.82	0.18	0.28	0.15	0.00	0.00	0.36	0.03	99.70	29.80
III	37.66	0.00	15.97	28.10	3.10	3.48	0.70	0.52	3.07	0.02	5.67	0.20	0.23	0.16	0.00	0.00	0.79	0.04	99.63	33.01
IV	43.07	8.08	14.31	20.20	2.45	2.43	0.63	1.22	0.66	0.00	6.10	0.18	0.35	0.15	0.00	0.00	0.40	0.03	100.20	25.71
V	43.54	1.32	21.93	19.66	2.36	2.29	0.97	1.08	0.71	0.02	5.35	0.18	0.27	0.15	0.00	0.00	0.13	0.03	99.93	30.62
VI	43.30	5.93	23.78	17.24	2.10	1.24	2.15	0.51	0.76	0.00	2.22	0.18	0.18	0.16	0.00	0.00	0.06	0.04	99.77	30.03
VII	36.00	0.17	23.48	23.04	2.60	2.76	1.57	1.31	0.80	0.00	7.50	0.19	0.37	0.06	0.00	0.00	0.33	0.01	100.17	34.33
VIII	45.95	2.21	20.67	13.08	4.50	4.11	0.46	0.16	0.60	0.00	7.45	0.05	0.14	0.19	0.00	0.00	0.14	0.04	99.67	24.63
IX	37.18	0.70	28.88	16.61	3.30	3.35	1.19	0.19	0.50	0.00	7.14	0.09	0.23	0.18	0.00	0.00	0.26	0.04	99.76	33.11
X	44.50	2.18	21.15	18.46	4.50	1.10	0.37	0.09	0.74	0.05	6.13	0.09	0.14	0.09	0.00	0.00	0.08	0.02	99.65	29.14
XI	35.20	0.27	19.21	24.52	3.47	2.93	1.41	0.05	0.39	0.50	11.44	0.12	0.32	0.17	0.00	0.05	0.43	0.06	100.42	32.50
XII	47.95	1.01	1.45	17.58	3.00	8.00	0.74	0.12	0.38	0.02	19.33	0.04	0.09	0.23	0.00	0.00	0.26	0.05	100.15	14.68
XIII	55.06	2.93	1.90	16.80	3.78	3.85	0.16	0.32	0.80	0.05	13.95	0.14	0.18	0.17	0.00	0.05	0.32	0.06	100.40	14.39
XIV	16.30	2.80	3.53	1.75	3.46	36.87	0.20	1.87	0.64	0.02	30.63	0.09	0.09	0.23	0.00	1.00	0.66	0.55	99.59	3.83
Average III - XI	40.71	2.32	21.04	20.10	3.15	2.63	1.05	0.57	0.91	0.07	6.56	0.12	0.25	0.15	0.00	0.00	0.29	0.03	99.89	30.34

Niggli Values

Sample number	si	al	fm	c	alk	mg	k	p	ti	co ₂	h+	cl ₂	s	qz
I	114.5	0.5	90.5	6.0	3.0	0.12	0.25	0.2	0.3	27.5	29.5	0.3	0.2	+10.9
II	94.0	0.5	87.5	7.0	5.0	0.11	0.14	0.3	0.3	29.0	25.5	0.7	-	-11.4
III	83.0	0.0	90.0	8.0	2.0	0.11	0.33	0.3	0.3	17.0	22.5	0.6	-	-19.5
IV	107.0	12.0	78.0	6.5	3.5	0.12	0.56	0.4	0.3	20.5	5.5	0.6	-	-8.2
V	105.0	2.0	88.0	6.0	4.0	0.10	0.42	0.3	0.3	17.5	5.5	0.6	-	-4.4
VI	101.5	8.0	83.0	3.0	6.0	0.09	0.14	0.2	0.3	7.0	6.0	0.6	-	-21.2
VII	77.5	0.5	88.0	6.5	5.0	0.09	0.35	0.3	0.3	22.0	5.5	0.2	-	-28.3
VIII	116.0	3.5	84.0	11.0	1.5	0.20	0.19	0.0	0.1	25.5	5.0	0.8	-	+10.6
IX	81.0	1.0	88.0	8.0	3.0	0.12	0.09	0.2	0.1	21.0	3.5	0.7	-	-24.7
X	108.5	3.0	93.0	3.0	1.0	0.28	0.14	0.2	0.2	20.5	6.0	0.4	-	+4.5
XI	78.0	0.5	89.5	7.0	3.0	0.13	0.02	0.3	0.2	34.5	3.0	0.6	-	-26.2
XII	157.5	2.0	67.5	28.0	2.5	0.22	0.09	0.1	0.1	87.0	4.0	1.3	-	+49.2
XIII	199.5	6.0	77.5	15.0	1.5	0.26	0.57	0.3	0.4	69.0	9.5	1.0	0.3	+94.5
XIV	31.0	3.0	19.0	75.5	2.5	0.52	0.86	0.1	0.1	80.0	4.0	0.7	3.6	-79.5
Average III - XI	94.5	3.0	87.0	6.5	3.5	0.13	0.25	0.2	0.2	21.5	7.0	0.6	-	-8.0

Analysts: E.C. Haumann and J.F. Dry
Soil Research Institute

- I Siliceous and ferruginous minnesotaite-slate between 100 and 240 feet above Inner Reef, Bore-hole W2, Westerberg.
- II Siliceous and ferruginous minnesotaite slate between 120 and 280 feet above Inner Reef, Bore-hole W2, Westerberg.
- III Banded ironstone between 30 and 180 feet below Visser Reef, Bore-hole W2, Westerberg (Upper Banded Ironstone Beds).
- IV Banded ironstone between 1400 and 1740 feet below lowermost asbestos reef in Westerberg Asbestos Zone, Bore-hole W2, Westerberg.
- V Banded ironstone between 1800 and 2020 feet below lowermost asbestos reef in Westerberg Asbestos Zone, Bore-hole W2, Westerberg.
- VI Banded ironstone between 2090 and 2190 feet below lowermost asbestos reef in Westerberg Asbestos Zone, Bore-hole W2, Westerberg.
- VII Banded ironstone between 2240 and 2350 feet below lowermost asbestos reef in Westerberg Asbestos Zone, Bore-hole W2, Westerberg.
- VIII Banded ironstone between 325 and 435 feet above Dolomite, Bore-hole DM 12A, Warrendale Mine.
- IX Banded ironstone between 180 and 300 feet above Dolomite, Bore-hole DM 12A, Warrendale Mine.
- X Banded ironstone between 70 and 320 feet above Dolomite, Bore-hole DW 19A, Pomfret.
- XI Banded ironstone between 325 and 650 feet above Dolomite, Bore-hole DW 19A, Pomfret.
- XII Banded chert (jasper) intercalated in banded ironstone between 20 and 155 feet above Dolomite, Bore-hole DM 12A, Warrendale Mine.
- XIII Banded chert (jasper) intercalated in banded ironstone between 10 and 380 feet above Dolomite, Bore-hole DW 19A, Pomfret.
- XIV Dolomitic limestone intercalated with banded ironstone at 25 feet above Dolomite, Bore-hole DM 12A, Warrendale Mine.

VII. Chemistry of the Banded Ironstone

Chemical analyses were carried out on bore-cores of banded ironstone and associated rock-types from three different localities in the Northern Cape. All the specimens from which samples for this purpose were taken came from rocks below the zone of oxidation. Bulk samples were obtained by cutting thin slices from each piece of available core at right angles to the bedding. Cores with approximately similar mineralogical composition and similar in texture were ground together and the pulps were thoroughly mixed. The results of 14 new chemical analyses of banded ironstone, iron-poor jasper, minnesotaite slate and carbonate-rich intercalations in the banded ironstone are given in Table 32.

(a) Tenor of Iron

On the average the tenor of iron in nine samples of banded ironstone obtained from different localities in the region is 30.34 per cent, and ranges from 24.63 to 34.33 per cent (Table 32, Analyses III to IX). The content of metallic iron in minnesotaite-slate of the Westerberg Beds which succeed the Banded Ironstone Sub-stage in the Southern Region is nearly the same, viz. 29.09 per cent (Table 32, Analyses I and II). Although the tenor of iron in the slate corresponds remarkably well with that in the banded ironstone there is a conspicuous difference in the mineralogical composition of the two rock-types. Magnetite is sporadically found in the slate in contrast with the general presence and abundance of the mineral in the banded ironstone. The amount of magnetite in the two rock-types is therefore not related to the amount of iron in them. It is also of interest to note that the magnetite-poor slate has an average FeO-content of 33.65 per cent (Table 32, Analyses I and II) compared with an average of only 20.10 per cent in the magnetite-rich banded ironstone (Table 32, Analyses III to XI). The amount of FeO in the composition of the banded ironstone and the related rocks is therefore not an indication of the amount of magnetite in these rocks.

The tenor of iron in the banded ironstone proper is remarkably constant in samples derived from different stratigraphical heights at the same locality and also in samples obtained from widely distributed localities. The content of metallic iron in the banded ironstone over a

vertical thickness of some 2,000 feet at Westerberg (Table 32, Analyses III to VIII) ranges from 25.71 to 34.33 per cent. The average iron-content of the banded ironstone at this locality is 30.74 per cent compared with 28.87 per cent at Botha (M2) and 30.82 per cent at Pomfret (B4) located approximately 100 and 260 miles respectively from Westerberg.

Two samples of magnetite-poor banded chert obtained from intercalations in the banded ironstone from between 15 and 370 feet above the Dolomite (Table 32, Analyses XII and XIII), have an average content of metallic iron of 14.53 per cent and an average Fe_2O_3 -content of only 1.68 per cent. The conspicuous difference in the iron-content of banded jasper and banded ironstone in which the jasper is intercalated has an important bearing on the origin of the rocks concerned. The iron-poor layers of chert reach a thickness of up to 40 feet and differ chiefly from the banded ironstone in mineralogical composition with respect to a deficiency or total absence of magnetite. If the thinly bedded nature of the enclosing banded ironstone is taken into consideration, there must have been quite long periods during which either no iron was available or the conditions for the deposition of iron were not favourable. Neither of these possibilities is strictly reconcilable with the hypotheses that the iron was derived from the weathering of continental rocks.

The distribution of iron in the banded ironstone and the associated rocks from different localities is shown in Figure 10. The graph shows that the tenor of iron in the banded ironstone as well as in the minnesotaite-slate (lower portion of the Westerberg Beds) is remarkably uniform; (Analyses I to XI); in the intercalations of banded jasper it is much lower (Analyses XII and XIII) than in the banded ironstone and in intercalations of calcareous rocks the iron-content is still very much lower (Analysis XIV).

(b) Silicon

The banded ironstone and most of the related rock-types are characterized by a high content of silica. Silica is contained in chert and/or microcrystalline quartz in the banded ironstone and in the intercalated layers of banded jasper. The average content of SiO_2 in the banded ironstone is 40.70 per cent, and it ranges from 35.20 to 45.95 per cent (Table 32, Analyses III to XI). The content of SiO_2 in the intercalations of

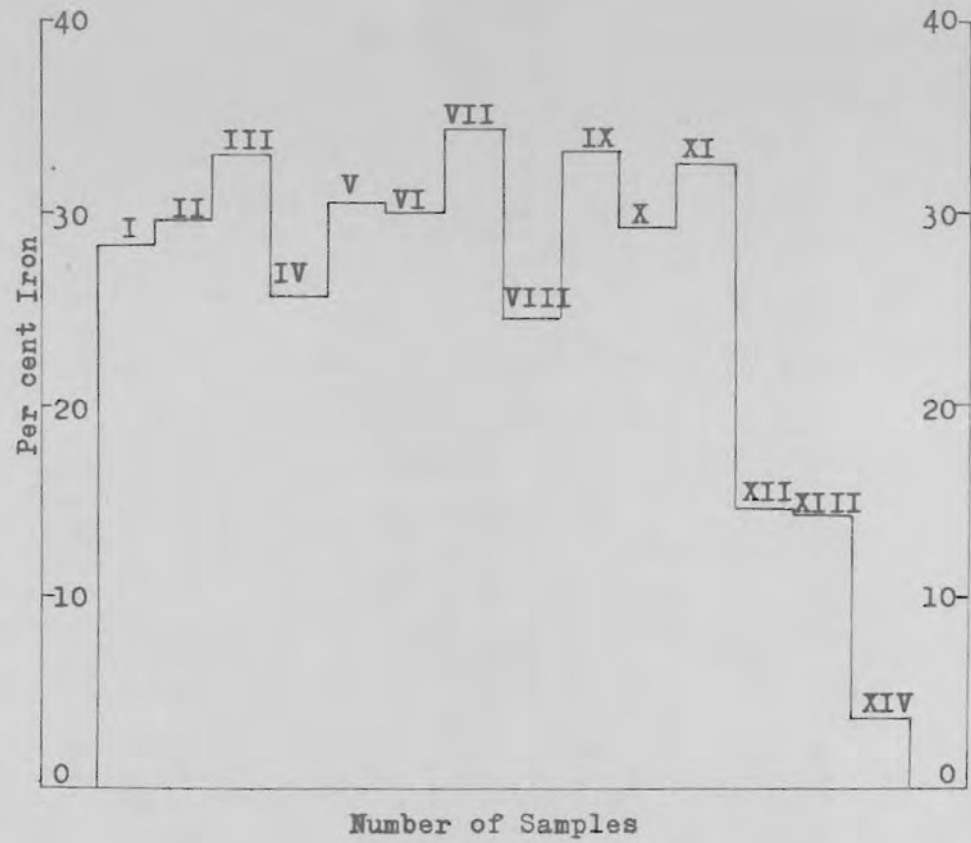


FIGURE 10

DISTRIBUTION OF IRON IN 14 ANALYSES OF SAMPLES
 DERIVED FROM ROCK TYPES IN THE BANDED IRONSTONE
 SUBSTAGE, LOWER GRIQUATOWN STAGE. (Compare Table 32).

banded chert in the banded ironstone is appreciably higher than for the banded ironstone viz. a range of 47.95 to 55.06 per cent and an average of 51.50 per cent. The SiO_2 -content in the minnesotaite-slate (Westerberg Beds) immediately overlying the banded ironstone ranges from 39.63 to 44.20 per cent with an average of 41.91 (Table 32, Analyses I and II). . . This average is slightly higher than that for banded ironstone although a microscopical investigation of the rock-types showed that free silica is far less common in the minnesotaite slate than in banded ironstone. The total amount of silica in these rocks can therefore not be taken as an indication of the amount of free silica, present as chert or as microcrystalline quartz, in the rocks.

(c) Calcium and Magnesium

The CaO -content of the banded ironstone ranges from 1.10 to 4.11 per cent, and averages 2.63 per cent (Table 32 Analyses III to XI). If the analyses of banded chert and minnesotaite-slate are included (Table 32, Analyses I to XIII) the average is raised slightly to 3.90 per cent. Lepp and Goldich (1964, p. 1039) pointed out that Precambrian ironstones are chemically characterised by a low CaO content (on the average 1.5 per cent) in contrast with Cambrian and later iron-formations which contain an average of 14 per cent CaO .

The MgO content of the banded ironstone ranges from 2.10 to 4.50 per cent, and averages 3.15 per cent. If the associated rock-types, banded chert and minnesotaite slate are included, the MgO content is raised to 3.19 per cent (Table 32, Analyses I to XIII). The average contents of CaO and MgO of the banded ironstone, 2.63 and 3.15 per cent, respectively, are slightly less than for intercalations of iron-poor banded chert for which the corresponding values are 5.92 and 3.39 per cent respectively. The average value for CaO/MgO in these rocks is 1.22.

Magnesium is tied up chiefly in silicate minerals like riebeckite and perhaps minnesotaite and the calcium chiefly as carbonates. Only a small amount of the magnesium is tied up in dolomite. Most of the magnesium probably represents part of the original precipitates, but much of the calcium could be secondary, having been derived from the solution of calcium by meteoric water from the underlying Dolomite. This statement is based on the general occurrence of calcite rhombs which replace chert in the banded ironstone. The magnesium derived from the dolomite presumably remained in solution.

(d) Manganese

The content of MnO in the banded ironstone is low. It varies from 0.06 to 0.79 per cent, and averages 0.29 per cent. The average for the banded ironstone and the associated banded chert and minnesotaite slate is 0.27 per cent. The Mn/Fe ratio for the banded ironstone, computed from seven analyses (Table 32, Analyses III to XI) is 0.0072 and for intercalations of banded chert (Table 32, Analyses XII and XIII) it is 0.015.

(e) Aluminium

The average content of Al_2O_3 of the banded ironstone and the banded chert is 2.32 per cent, but the values for individual samples vary considerably. The distribution of Al_2O_3 in the samples is shown in Figure 11. In one sample no aluminium was recorded, in five others the content of Al_2O_3 is less than one per cent; in the remaining seven samples it varies from 1.01 to 8.08 per cent. Most of the aluminium is probably tied up in riebeckite which enjoys a wide distribution in the banded ironstone.

VIII. The origin of the banded ironstone and the related rocks of the Lower Griquatown Stage

A. Introduction

The origin of banded ironstones of Precambrian age is an old problem. Various authors have offered solutions, but up to the present time different opinions are still held. In the hypotheses advanced two main views are advocated. One school of thought advocates a hypothesis of selective weathering combined with changing climatic conditions as well as a continuous change of physico-chemical conditions in the basin of deposition. A second school supports the direct contribution of material by magmas, volcanoes or fumaroles or the reaction of sea water with hot or cold lava.

In previous publications on the rocks of the Lower Griquatown Stage most authors have accepted the idea that these rocks are the products of chemical precipitation. Only a few authors have gone into the details of the contribution of material and the conditions under which the material was deposited.

Wagner (1928, pp. 64 & 65) discussed the origin of the banded ironstone of the Lower Griquatown Stage briefly and pointed to the regular transition between the banded ironstone and the underlying dolomite. This transition coupled with the complete absence of any recognisable sediment of mechanical origin in the more typical ironstone led Wagner to conclude that the sideritic cherts represent marine deposits. He considered them to be chemical precipitates and assumed that the silica was in colloidal solution in the sea water and that the iron was directly contributed to the ocean by magmatic waters.

Peacock (1928, p. 266) dismissed the possibility that the ironstone of the Lower Griquatown Stage could represent ancient deep-sea deposits, similar to the red clays now accumulating in the deepest part of the ocean, owing to the complex composition of the clays compared with that of the ironstones. He suggested that most of the chlorides now in solution in sea water was emitted by subaerial and submarine volcanoes and fumaroles (p. 269). He claimed that chlorides of iron, aluminium, calcium, magnesium and the alkalis were formed through the reaction between water vapour at a high temperature and deep-seated rocks which resulted in the concurrent liberation of silica.

Du Toit (1945, p. 163) suggests that the material was precipitated chiefly as colloidal silicates and double carbonates in a shallow, probably cold sea. He concluded that the precipitates were largely chemical and probably bacteriological and that precipitation took place under peculiar physiographical conditions. He does not give details of the sources of the material nor the mechanics of transportation and deposition.

Cilliers (1961) went into details concerning the transportation and deposition of the material constituting the rocks of the Lower Griquatown Stage and gives a good review of the changing physico-chemical conditions in the basin of deposition. He concludes (p. 74, & 78) that silica and compounds of iron were derived from the chemical, probably aided by bacteriological, weathering of exposed continental rocks. He suggests that the material was carried to the basin of deposition of the Lower Griquatown Stage both in true ionic and in colloidal solution. The silica and some of the silicates were precipitated in a colloidal condition (p. 82) whereas the precipitation of iron oxide, iron carbonate and iron sulphide was governed by purely chemical conditions (p. 84).

B. The Source on the Material and the Cause of the Laminated Structure

One of the major problems with regard to the origin of Precambrian ironstones is the source of the iron and the silica which constitute the bulk of these rocks.

In discussing the origin of sedimentary iron formations with special reference to the Biwabik Formation of the Mesabi Range, Gruner (1922, p. 459) concluded that iron and silica were derived from the weathering of the crustal rocks under humid and probably tropical or subtropical climatic conditions. Precipitation of the silica and iron and part of the organic colloids in the basin of deposition is chiefly attributed to the influence of algae and bacteria, but Gruner also recognises the influence of inorganic reactions in the process of precipitation.

Woolnough (1941, p. 465) expressed the opinion that the banded iron deposits of the older Precambrian throughout the entire world represent epicontinental sediments precipitated chemically from cold, natural solutions in isolated, restricted basins located on peneplaned land-surfaces.

Dunn (1941, p. 362) held that banded hematite quartzite might have various modes of origin, but pointed out that in the case of some iron ores of the Singhbhum District of India, the banded hematite quartzites have been formed by the silicification of ferruginous sediment which originated partly as tuffs.

Du Preez (1944) suggested that the banded ironstone of the Dolomite Series in the Thabazimbi area, Transvaal, formed through chemical precipitation of silica and iron oxide hydrosols, derived from the weathering of adjacent low-lying land and transported to the sea by rivers.

Alexandrov (1955, p. 459) suggests that the intermittent banding of silica and hematite in Precambrian banded iron ores is principally caused by selective weathering of the Precambrian soil. The selective weathering is attributed to seasonal changes of temperature, the amount of precipitation and the alternately higher and lower pH range of the leaching solutions. A similar view is held by Sakamoto (1950, p. 449) who inferred that shallow lakes in a paralic basin and a monsoon-like climate were responsible for the unique environmental condition.

Goodwin (1956, p. 588) is of opinion that weathering of continental rocks as a source of iron and silica does not apply to the Gunflint Iron Formation, Ontario. He points out that a direct and sympathetic relationship between volcanism and the iron formation is indicated by the presence of pyroclastics and lava on certain horizons in the iron-bearing rocks, and in a wider and more significant sense by the cyclical co-ordination of volcanism and sedimentation. He accordingly suggests that the main mass of iron oxide and silica in the Gunflint Formations has been contributed to the basin of deposition by volcanic activity, possibly through the activity of hot springs and mineral alteration of extruded volcanic material.

In the latest discussion of the origin of the rocks of the Lower Griquatown Stage, Cilliers (1961, p. 74) claims that there is no longer any doubt that sufficient material for the formation of these rocks could have been obtained from the weathering of the rocks exposed on the continent at the time this Stage was deposited. He suggests that the Ventersdorp lava was one of the sources from which the iron was derived.

Lepp and Goldich (1964, p. 1,026) assume that the Precambrian atmosphere had a relatively low oxygen content and that lateritic weathering under this condition permitted the transport of iron and manganese together with silica while the weathered mantle effectively retained aluminium, titanium, phosphorous and colloidal clay. They note that graphitic material of biogenic origin is closely associated with banded ironstone of Precambrian age and state that although it is uncertain whether iron oxide hydrosols were precipitated directly through biogenic processes, the removal of CO_2 and the liberation of oxygen to the sea water through photosynthesis of primitive plants undoubtedly influenced the energy relationships among the iron minerals.

Gross (1965, p. 1063) believes that the composition of the atmosphere was not a significant factor in the formation of the sedimentary banded ironstone. He stresses the significant fact that in whatever environment cherty banded ironstones were deposited evidence of volcanic activity contemporaneous with the deposition of these rocks is consistently encountered. He accordingly expressed the opinion that silica and iron oxides could have been contributed by processes that were related to this volcanic activity.

In considering the origin of the material necessary for the formation of banded ironstone Trendall (1965, pp. 1066 & 1067) refers to the Brockman Iron Formation of Western Australia which has a thickness of about 2,000 feet over some 25,000 square miles and calculated that even if the source-area of the material now constituting this formation was eight times as large as the area of deposition all the iron from a total thickness of 2,000 feet of Precambrian soil would have been needed for the deposition of the formation. He pointed out further that, after the extraction of iron, silica, alumina and other materials required for the deposition of the formation about 600 feet of aluminous laterite would have been left behind, assuming that most of the silica was removed in solution. He concluded that the preferential extraction of some constituents by weathering to these depths is difficult to reconcile with a relief sufficiently slight to prevent significant transportation of clastic debris towards the basin of deposition.

Reference has already been made (p.148) to the presence of thin, persistent layers of pyroclastic material which are intercalated with the banded iron-

stone over the entire area. These layers vary much in thickness from less than half an inch to several feet and although they are found at irregular intervals in the entire succession, the layers are commonly more abundant in the crocidolite-bearing zones, or in zones where many seams of massive riebeckite are present. It is accordingly suggested that the layers of tuff represent falls of volcanic ash during the time of deposition of the banded ironstone.

Many of the layers of tuff contain appreciable amounts of pyrite, generally in the form of irregular concretions elongated parallel to the bedding of the banded ironstone. The pyrite is commonly accompanied by black carbonaceous material and it is believed that both the organic material and the pyrite were derived from primitive micro-organisms which flourished in the basin of deposition (HH 284). A sudden fall of volcanic ash could result in the burial and the preservation of organic matter, followed by the development of pyrite.

That primitive life existed during the time of deposition of the Lower Griquatown Stage was suggested by Harrington and Cilliers (1963) after Harrington (1962) had proved the presence of certain primitive oils and waxes in the banded ironstone and associated crocidolite.

Pyrite is also found in abundance in seams of massive riebeckite especially towards the base of the Banded Ironstone Substage in the Southern Region. Should the pyrite in the seams of massive riebeckite have had the same mode of origin as the pyrite in the layers of tuff it is quite possible that the origin of the massive riebeckite and for that matter the origin of the material constituting the seams of asbestos is largely similar to that of the layers of tuff.

In discussing the origin of the material constituting the rocks of the Lower Griquatown Stage it is of importance to bear in mind that the banded ironstone in the succession is not always built of alternating layers of chert and magnetite only. The upper portion of the Banded Ironstone Substage in the Southern Region, for example the Westerberg Beds, contains actually very little free silica in the form of chert, but is chiefly composed of alternating laminae of minnesotaite, and riebeckite. Magnetite is abundant in certain zones only.

The Riebeckite Slate Zones of the Middelwater and Kwakwas Substages in the Southern Region also contain

several hundred feet of massive riebeckite accompanied by stilpnomelane and siliceous material. If selective weathering of continental rocks near the basin of deposition was the primary factor in the contribution of material for deposition then atmospheric conditions must have changed considerably to cause the leaching of such small amounts of iron compounds compared with silica for the formation of these amphibolitic rocks. Conditions controlling the precipitation of iron hydrosols and silica in the basin of deposition must also have changed remarkably to prevent the alternating precipitation of these elements as is found in the banded ironstone. Not only that, but assuming that little ionic migration took place during the low-grade metamorphism of the rocks, then almost the exact quantities of iron oxide, silica, alumina, magnesia and soda must have been precipitated simultaneously to cause the formation of the thick succession of strata in which riebeckite is by far the most important constituent. The question naturally arises whether the parent-material for the formation of the seams of massive riebeckite did not originate from a source capable of supplying identical material at regular intervals, such as volcanic ash from active volcanoes or from fissure-eruptions. When material of this kind is carried to the basin of deposition, either by wind or by rivers, it will be precipitated under the influence of gravity only and cause the formation of layers which have more or less the same chemical constituents evenly distributed through them. Subsequent metamorphism could then cause the formation of layers which have the same mineralogical composition.

Sodium which represents an essential element in the chemical composition of the riebeckite is largely restricted to seams of massive riebeckite and crocidolite. If the parent-material for the formation of these minerals is also mainly of sedimentary origin, as some investigators believe why should sodium, which is highly soluble, be associated only with those elements necessary for the formation of riebeckite. The present writer is of the opinion that none or very few of the elements which constitute the riebeckite was precipitated chemically, but that all material was derived from extraneous sources of volcanic activity, most probably in the form of volcanic ash which settled on the floor of deposition under the influence of gravity.

Another phenomenon which is contradictory to the

suggestion that the material for the formation of the banded ironstone was obtained through the selective weathering of continental rocks is the common occurrence of thick intercalations of iron-poor banded chert in the thinly laminated banded ironstone. These intercalations in places measure more than 40 feet in thickness. If seasonal changes were responsible for the selective leaching of iron and silica then the presence of intercalations of the iron-poor banded chert could probably be explained as follows: A long period during which only a little iron was leached out ensued and only silica was carried to the basin of deposition. Should this be the case one would expect layers of the banded chert to be followed by bands exceptionally rich in iron or bands in which the amount of iron gradually increases until a normal banded ironstone is present once more. In the layers of banded ironstone immediately below such an iron-poor chert one would expect a gradual decrease in the iron-content as a result of progressive seasonal changes. This is not the case; the amount of iron in the banded ironstone below and above the iron-poor layers of chert is the same as in the beds away from the banded chert. The presence of the layers of chert can also not be due to changing physico-chemical conditions in the basin of deposition because, if ferruginous material is steadily contributed to the basin during the entire deposition of the banded chert, a tremendous volume of iron hydrosols must remain in solution over a long period. Again if conditions became favourable for the precipitation of it one would expect layers in which the iron-content will be abnormally high owing to rapid precipitation partially influenced by the high concentration of iron.

Peacock (1928) suggested that the material for the formation of the banded ironstone could have been contributed by the reaction between vapours from submarine fumaroles, water and the walls of the fissures along which water vapour charged with hydrochloric acid ascended. He suggests that the selective precipitation of silica, iron and subordinate aluminium hydrates was caused by subsequent emission of alkaline reagents either in the form of ammoniacal vapours or soluble alkaline silicates in the manner proposed by Van Hise and Leith (1911, pp. 499-529). With subsequent denudation of the basin of deposition this heterogeneous precipitate would become largely dehydrated and indurated to form a rock corresponding substantially with a ferruginous chert.

One of the chief objections against the contribution of iron and silica by fumaroles is that insufficient material is supplied (Gruner, 1922, p. 488-449 & Cilliers, 1961, p. 72). Cilliers maintained that it would be difficult to find an adequate source of iron in the older formations which could have been mobilised through fumarole activity in sufficient quantities to satisfy the requirements of the Lower Griquatown Stage. At the same time he suggests that the weathering of the Ventersdorp lava underlying the Transvaal System probably supplied the bulk of the iron in the Lower Griquatown Stage. He also claimed that no trace of intense chemical activity associated with fumaroles are preserved in the formations older than the Pretoria Series.

With regard to the absence of exceptionally iron-rich formations underlying the Lower Griquatown Stage, from which iron could have been dissolved the question arises whether it is necessary for the iron to be derived from the solution thereof from underlying formations, or whether the bulk of the iron now present in these rocks is not truly magmatic in origin. As shown on Folder 1, the Lower Griquatown Stage is traversed by a profusion of diabase dykes. Many of these dykes probably represent the original fissures along which the Ongeluk Lava, which directly succeeds the Lower Griquatown Stage, was intruded. Some of the material now occupying the dykes is of Karroo Age. Composite dykes, occupied by material very similar to the Ongeluk Lava and by material similar to dolerite dykes of Karroo Age, also exist. If only a fair percentage of the dykes represent the channels along which the Ongeluk Lava was intruded these linear structures could well have been the passages for juvenile gases prior to the extrusion of the lava.

After having investigated the iron ores of the Pretoria Series in South Africa, Schweigart (1956) came to the conclusion that because of prevailing oxidizing conditions during Early Precambrian times and the absence of a continental flora, adequate amounts of iron for the formation of large deposits of iron ore in the Pretoria Series could not have been dissolved and transported to the Precambrian oceans in solution. He accordingly suggests that the iron in these sediments was formed from acid submarine volcanic exhalations, contributed to the sedimentary basins from time to time during the precursory phases of the Bushveld Igneous Complex. He pointed out that intermittent volcanic activity took place during various stages of the deposition of the Transvaal System.

of which the Pretoria Series (Lower Griquatown Stage) forms an integral part.

The well-known periods of volcanic activity during the history of deposition of the Transvaal System is marked firstly by restricted flows of andesitic lava in the basal part of the Black Reef Series. This was followed by intermittent volcanic episodes during the Daspoort Stage represented by andesitic lava, tuff and agglomerate (Ongeluk lava), tuff and agglomerate during the Magaliesberg Stage and amygdaloidal andesite and tuff during the Smelterskop Stage of the Pretoria Series. Less well-known is the occurrence of pyroclastic material intercalated in the basal portion of the Dolomite Series. The occurrence of this pyroclastic material is described by Young and Mendelssohn (1948). They noted (p. 57) that near Schmidt's Drift in the Northern Cape Province pyroclastic material is embedded in a calcareous matrix and the rock-type which resulted can best be described as tuffaceous limestone. Despite the appreciable alteration of the pyroclastic material the authors could recognise fragments of volcanic origin which are characterized by curvatures attributable to conchoidal fracture or reminiscent of a vitroclastic texture.

Including the presence of layers of tuff intercalated in the banded ironstone, which has not been realised to date, there is therefore at least six cycles of volcanic activity associated with the deposition of the Transvaal System. Two of these cycles preceded the deposition of the Lower Griquatown Stage, one was contemporaneous with and three cycles succeeded its deposition. One of the last three periods, the andesitic lava flows of the Daspoort Stage (Ongeluk lava), followed immediately upon the deposition of the Lower Griquatown Stage.

Present-day contribution of iron, silicon, aluminium and other elements to the ocean by fumaroles is demonstrated at Karuchatka, Russia, and the Kurile island. Studies during the last decade have shown that distinctively acid water (pH - 1 to 2) and unusual cation associations are characteristic of these two areas (Strakhov, 1964, p. 842). The major elements in this water are Al and Fe (ferrous and ferric) whereas Na, K, Ca, and Mg are not among the major elements.

Strakhov (p. 842) maintains that it is now clear that the acid waters originate from the circulation of water in unconsolidated tuffs which are exposed to gases derived from fumaroles (HCl, HF, H₂S, SO₂, CO₂). These gases

dissolve in the water and give rise to fairly strong sulphuric and hydrochloric acid solutions which attack the surrounding rock and extract Na, K, Ca, Mg, SiO₂ and also Fe and Al from it. With the extraction of these elements the rocks are greatly altered, giving rise in some places to pure opalolites (virtually only SiO₂) and in others to clays.

Where these acid solutions mix with sea water there is a very prominent chemical differentiation for Al and Ti, which become spatially separated during migration, and also for silica and iron oxides. As the pH of the water rises gradually several components are deposited. Results given by Strakhov (p. 843) of a detailed study of the conditions at Lake Tikhoye from where a stream leads to the North Chirip River which enters the sea of Okhotsk, reveal the following:

A ferruginous sludge forms in the lake itself and much ferruginous material is deposited in the stream. Where the river enters the sea three turbid zones can be distinguished due to variations in colour of the water and the coloured precipitate suspended in the water. An analysis of the suspended matter at different points from the lake to the sea is shown in Table 33.

Table 33. - Composition of Suspensate
 (After Strakhov 1964, p. 843)

Site	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	SO ₃
Lake Tikhoye	1.61	0.06	1.29	56.73	10.75	7.78
In the North Chirrip River	0.36	Trace	0.55	72.92	4.06	5.14
At the mouth of the North Chirrip River	4.42	None	9.82	52.43	1.32	1.32
In sea 50 m from mouth of river	None	None	24.17	31.61	-	-
In sea 250 m from mouth of river	None	None	32.40	17.18	-	-

Judging by the analyses of the suspensates given in Table 33 it is apparent that the chemical differentiation caused by pH changes of the water is conspicuous. A rhythmic or constant influx of thermal exhalations into a basin of deposition, especially if the channels for these exhalations are widely strung out along fissures, would not only cause subsequent addition of material, but will also cause almost regular changes in the pH conditions

which will result in the differential precipitation of silica, iron oxide and alumina.

From the chemical composition of two representative chlorite-sulphate springs given by Strakhov (p. 842) it is evident that the concentration of ferrous iron in the water is about twice to about seven times that of ferric iron; 0.7742 g/L and 0.2194 g/L compared with 0.4170 g/L and 0.0312 g/L. If the amount of ferric iron in the suspensate which formed in the Lake Tikhoye is compared with the amount of ferrous iron in the same suspensate (Table 33) it is clear that oxidation of the ferrous iron takes place rapidly and on a large scale.

The chemical composition of the water from the thermal springs also indicates that the ratio of iron oxide to alumina varies appreciably from one spring to another and the conclusion is accordingly drawn that alumina could be completely subordinate in the composition of waters from thermal springs.

Much stress has been laid on the role played by varying pH and oxidation-reduction potential (Eh) in the deposition of chemical sediments (Krumbein and Garrels, 1952; Huber and Garrels, 1953; etc.).

Cilliers (1961) gave a detailed description of how these physico-chemical influences could have contributed in the selective precipitation of iron oxide and silica during the deposition of the Lower Griquatown Stage. It is not necessary to repeat the details of the processes because the effect of changing physico-chemical conditions on the precipitation of these elements would have been much the same whether the silica and iron oxide were contributed by processes of selective weathering or by successive periods of submarine fumarolic exhalations.

However, it is necessary to remark on a few anomalous features concerning the contact between the Dolomite and the succeeding Banded Ironstone Substage. It has been pointed out by Krumbein and Garrels (1952) that the precipitation of calcium carbonate is chiefly controlled by pH conditions - an increase in pH creates alkaline conditions which will cause the precipitation of calcium carbonate. Truly acid environments are therefore unfavourable to limestone deposition. In this respect the contact between the Dolomite and the banded ironstone of the Lower Griquatown Stage needs special consideration.

It has been pointed out repeatedly in previous pages that the contact between the two rock-types is transitional

in the sense that separate siliceous and also ferruginous intercalations become abundant towards the top of the dolomite, and that intercalations of dolomite again are found in the basal portion of the banded ironstone. The mineralogical composition of the siliceous intercalations in the dolomite and the different layers in the lower portion of the Banded Ironstone Substage vary considerably from one locality to another. In many localities they are almost completely devoid of iron, as shown by borehole sections on Whitebank, Kuruman District (Table 3). In the environment of Danielskuil the very first siliceous intercalations towards the top of the dolomite contain considerable amounts of magnetite and represent a banded ironstone. In the Kauningsvlei area again the basal portion of the Banded Ironstone Substage is characterized by the frequent occurrence of layers of black shale largely similar to the tuff bands described earlier. Although many, if not most of the siliceous intercalations in the upper portion of the Dolomite contain accessory amounts of carbonate there is, however, a distinct and sharp contact between calcareous and siliceous bands. This would indicate that the conditions which influenced the change in pH conditions must have changed abruptly because almost no gradation from carbonaceous to siliceous beds exists.

Taking into consideration the frequent occurrence of pyroclastic material intercalated in the banded ironstone the writer draws the conclusion that volcanic activity played a major role in the formation of the Lower Griquatown Stage. The sharp contacts between laminae of different mineral composition may indicate that the conditions which controlled the alternating precipitation of iron hydrosols and silica and also carbonate in places were rhythmically repeated and must have had a remarkably instantaneous effect. The pH of the water in the basin of deposition was most probably the chief controlling factor in the precipitation of silica, iron oxide and carbonate. If submarine exhalations took place over wide areas in the basin of deposition the intermittent influx of exhalations rich in iron oxide and silica, would have had a remarkable and abrupt influence on the pH of the water. This would result in the rapidly alternating precipitation of chemically different elements.