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THE BUSHVELD HYDROTHERMAL SYSTEM: FIELD AND PETROLOGIC EVIDENCE

CRAIG M. SCHIFFRIES* and BRIAN J. SKINNER**

ABSTRACT. A systematic search for evidence of fluid-rock interactions in the Bushveld Complex has resulted in the recognition of extensive networks of hydrothermal veins that crosscut the intrusion. The veins were produced by fracture-controlled flow of aqueous fluids during the subsolidus cooling history of the intrusion. Hydrothermal veins occur in each of the major stratigraphic units that comprise the Bushveld Complex. We have observed veins across a stratigraphic thickness of 9 km and along a strike length of at least 100 km. The Bushveld hydrothermal system has affected tens of thousands of cubic kilometers of rock, and it is probably among the largest paleo-hydrothermal systems in the world.

Petrologic studies of the veins indicate that fluid-rock interactions occurred over a wide range of physical and chemical conditions, producing a wide variety of hydrothermal veins. The mineral assemblages in the veins are similar to those found in mafic igneous rocks that were subjected to metamorphism under conditions ranging from upper amphibolite facies to zeolite facies. Many veins that formed at intermediate or high temperatures contain hydrothermal pyroxene that is texturally and chemically distinct from its magmatic counterpart. Hydrothermal clinopyroxene, where present, is enriched in Ca but depleted in non-quadrilateral components relative to magmatic clinopyroxene. The occurrence of hydrous minerals with high concentrations of chlorine (up to 5 wt percent in amphibole) and the presence of fluid inclusions with a halite daughter crystal indicate that many of the fluids that interacted with the Bushveld Complex were chloride-rich brines. Fluids of this type are effective agents for mass transfer, and this raises the possibility that mineralogical, chemical, and isotopic features of the Bushveld Complex may not be due solely to primary magmatic processes but may instead have been influenced by the effects of hydrothermal processes.

INTRODUCTION

The Bushveld Complex is the world's largest layered igneous intrusion and the greatest repository of magmatic mineral resources ever discovered. There is a vast literature about this extraordinary intrusion, but the roles that hydrothermal processes may have played and the influences they may have had in producing the features we see today have been largely neglected. This paper reports results and conclusions arising from portions of a systematic study of fluid-rock interactions in the Bushveld Complex. The major goals of the study are to establish that: (1) the Bushveld Complex was influenced by a hydrothermal circulation system as evidenced by crosscutting networks of hydrothermal veins; (2) fluid-

* Department of Earth and Planetary Sciences, Harvard University, Cambridge, Massachusetts 02138

** Department of Geology and Geophysics, Yale University, New Haven, Connecticut 06511

rock interactions occurred over a wide range of physical and chemical conditions as indicated by the mineralogy of the veins; (3) the fluids responsible for the formation of many veins were chloride-rich brines; and (4) fluid-rock interactions modified many igneous features, such as the chemical and isotopic compositions of minerals, and may have played a significant role in producing some of the macroscopic features that have been generally presumed to be of primary magmatic origin. This is the first paper arising from our studies, and it presents field and petrologic data that indicate that a major hydrothermal system was active in the Bushveld Complex as it cooled. There was probably a continuous sequence of hydrothermal influences on the Bushveld Complex, starting with the intrusive-magmatic stage, and continuing through the entire post-magmatic cooling history. This paper is concerned only with the post-magmatic hydrothermal stage.

Geological setting.—The Bushveld Complex has a stratigraphic thickness ranging between 7 and 9 km and is exposed as three major lobes that enclose an area of 65,000 km^2 (fig. 1). It is 100 to 1000 times larger than the Skaergaard intrusion, the Stillwater Complex, and the Cuillin Complex (Willemse, 1969). Size alone draws attention to the Bushveld Complex, but it is also famous for the remarkable persistence and regularity of its igneous layering. Individual layers and cyclic units have been traced along strike for more than 100 km in both the eastern and western lobes of the intrusion. Some of the layers contain immensely valuable mineral resources, and these deposits produce a major fraction of the world's output of chromium, vanadium, and the platinum-group elements.

The Bushveld Complex contains a broad spectrum of plutonic rock types. The magmatic stratigraphy is divided into four major zones (fig.1). The Lower Zone consists of ultramafic rocks, including dunite, orthopyroxenite, and harzburgite. Above the Lower Zone lies the Critical Zone which contains abundant norite, anorthosite, and orthopyroxenite, as well as approx 20 chromitite layers which individually are up to 2 m thick. The Main Zone is dominated by gabbro and contains subordinate amounts of anorthosite. The Upper Zone is composed of magnetite gabbro, ferrodiorite, anorthosite, and approx 25 magnetitite layers.

The age of the Bushveld Complex is 2050 ± 25 Ma ($\lambda_{Rb} = 0.0142$ Ga^{-1}; Hamilton, 1977). It is intruded into a thick pile of early Proterozoic sedimentary and volcanic rocks that belong to the Transvaal Supergroup. The emplacement of the Bushveld was preceded by an extensive period of volcanism as recorded in the Rooiberg Felsite and Dullstroom Volcanics. Intrusion of the mafic rocks was postdated by a period of granitic plutonism recorded in the Bushveld Granite. Additional details about the geology of the Bushveld Complex, and the current thinking about processes involved in its petrogenesis, are summarized in numerous papers (for example, Sharpe, 1985; Irvine, Keith, and Todd, 1983; Campbell, Naldrett, and Barnes, 1983; Tankard and others, 1982; von Gruenewaldt, 1979; Willemse, 1969; and Wager and Brown, 1968).

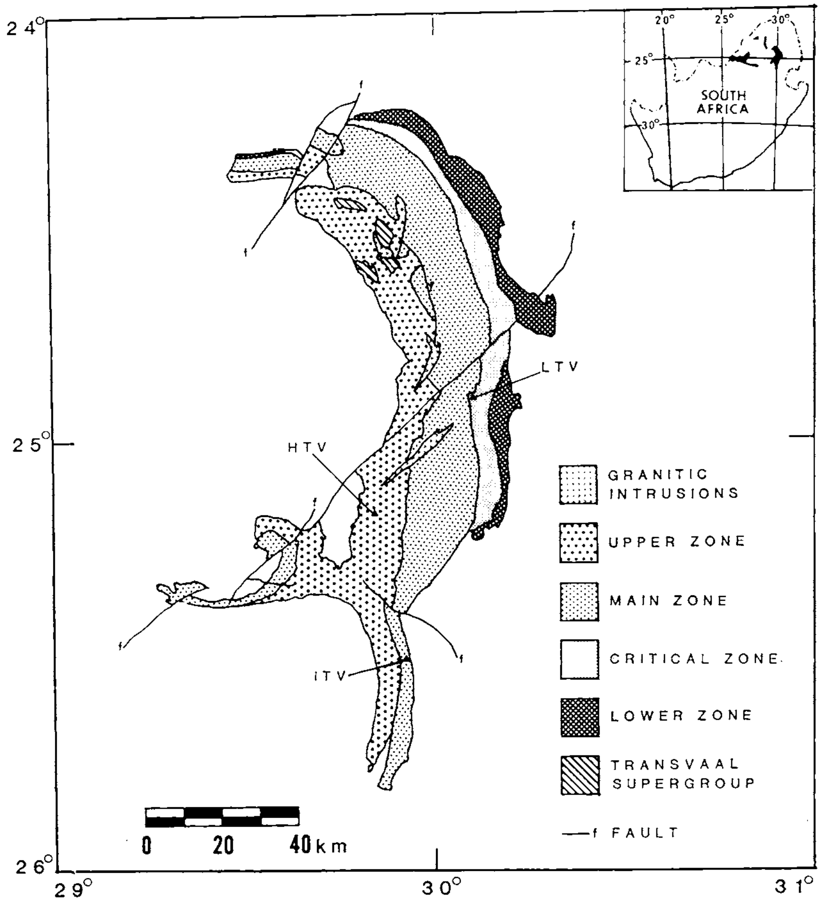


Fig. 1. Generalized geologic map of the eastern lobe of the Bushveld Complex. The type localities of the early-stage, high temperature veins (HTV), middle-stage, intermediate temperature veins (ITV), and late-stage, low temperature veins (LTV) are indicated by triangles.

FIELD RELATIONS AND PETROLOGIC DATA

Overview and classification of hydrothermal veins.—The most obvious manifestations of hydrothermal activity are mineralized veins deposited by the hydrothermal solutions and the zones of secondary alteration that flank the flow channels. Hydrothermal veins were first recognized by us in 1981, and a systematic search was conducted in the eastern lobe of the Bushveld during 1985 and 1986. Veins were discovered in each of the major stratigraphic units that comprise its 9 km thickness and over a distance of 100 km along the strike length (Schiffries, 1985b). More than 1000 kg of veined samples were collected for laboratory studies; petrologic investigations of this material have resulted in the recognition of a wide variety of mineral assemblages.

The frequency of veins is quite variable, but careful mapping reveals that veins are everywhere present. Most veins are narrow — a few millimeters in diameter is a wide vein — and most are difficult to see, but even so, extensive networks of hydrothermal veins constitute a prominent, first-order structural feature of the outcrops in some areas (pl. 1). Larger veins are uncommon and appear to be virtually absent in many areas. Appearances can be misleading however. The ease with which veins of any width can be seen in the field is strongly dependent on the mineralogy of the vein, the type of host rock, and the extent of weathering; it is likely that many veins were not detected by us during our field work.

We have grouped the post-magmatic hydrothermal veins into three temporal families on the basis of their relative chronology as determined by crosscutting relations. The three groups are referred to as early, middle, and late veins. This classification system is introduced for organizational convenience and is not meant to imply that temporal discontinuities occurred in the vein forming processes. Nor does this classification imply that only three mineralogically distinct types of veins are present in the Bushveld Complex. To the contrary, mineral assemblages in the veins are strongly influenced by the composition of the host rocks. Despite compositional differences, there appears to be a commonality in each vein group. The commonality reflects the conditions of formation. The mineral assemblages in the early, middle, and late veins are similar to those found in mafic igneous rocks that have equilibrated under metamorphic conditions corresponding to amphibolite, greenschist, and zeolite to prehnite–pumpellyite facies, respectively (table 1). The sequence of formation suggests that the veins may have formed as the Bushveld Complex cooled.

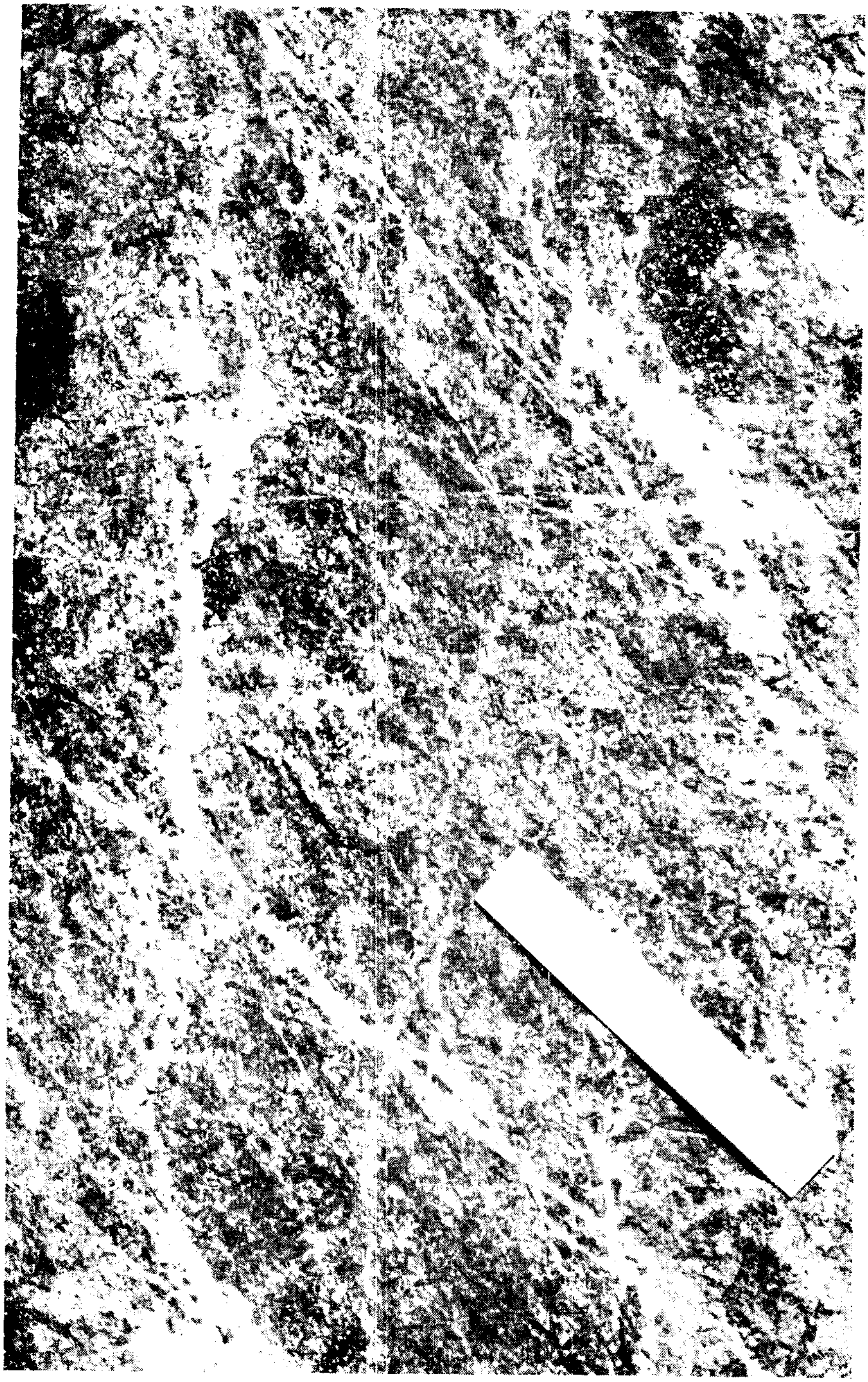
TABLE 1

Summary of vein characteristics

	Early Stage	Middle Stage	Late Stage
Major Vein Minerals	Hornblende Plagioclase Fe–Ti oxides	Actinolite Clinopyroxene Quartz Sphene	Prehnite Pumpellyite Quartz
Additional Alteration Minerals	Clinopyroxene Biotite Symplectite	Albite Pumpellyite Epidote Microcline Chlorite Biotite Clinozoisite Carbonate	Chlorite Carbonate
Facies equivalent	Upper Amphibolite	Greenschist	Prehnite–Pumpellyite
Temperature	>600°C	300-600°C	<300°C
Frequency	0.2 to 2 m^{-1}	0.2 to 1 m^{-1}	<10 to >100 m^{-1}
Vein width	0.5 to 5 cm	0.5 to 5 cm	1 mm
Type locality	Onverwacht 152 JS Upper Zone	Kwaggskop 359 JS Main Zone	Dwars River 372 KT Critical Zone
Host rock at type locality	Magnetite Gabbro	Gabbro	Mottled Anorthosite

(C) Middle-stage vein surrounded by a prominent alteration halo, in gabbro of the Main Zone, at the type locality on the farm Kwaggskop 359 JS.

PLATE 1 (continued)



(D) Dense network of late-stage veins in mottled anorthosite of the Upper Zone, on the farm Zwartkop 142 JS.

outcrops. They are difficult to see on weathered surfaces. At the type locality the veins occur along three distinct sets of steeply dipping fractures (fig. 2) and have a frequency of 0.2 to 2 m^{-1} (vein frequency is reported as total vein length per unit area and hence has units of m^{-1}). Individual veins are commonly continuous for 10 m or more along strike. They are nearly planar throughout their length, but near their terminations they are commonly curved and can be observed to break up in an *en echelon* fashion (pl. 3). When viewed on a macroscopic scale, wall-rock alteration is not readily apparent adjacent to early-stage veins (pls. 1-A, 3). On a microscope scale, however, vein contacts are irregular, and large changes in mineral composition can be observed at the boundary between a vein and its host rock.

Early-stage hydrothermal veins from the type locality are composed predominantly of calcic amphibole and calcic plagioclase, with subordinate amounts of clinopyroxene and Fe–Ti oxides, ± biotite ± orthopyroxene ± sulfides (Schiffries, 1986). In mafic igneous rocks subjected to regional metamorphism, this mineral assemblage is characteristic of upper amphibolite facies. We conclude that the early-stage veins were formed under similar physicochemical conditions to those characteristic of the upper amphibolite facies. Our Bushveld observations are not unique. A very similar assemblage has been reported for hydrothermal veins that crosscut the Skaergaard intrusion (Bird, Rogers, and Manning, 1986).

Calcic amphibole is the most abundant mineral in the early-stage veins. It ranges in composition from actinolite to a ferroan pargasite; pargasitic hornblende is the most common amphibole variety. For amphibole normalizations calculated on the assumption that the number of total cations minus the sum of Ca, Na, and K is equal to 13, the concentration of Al^{IV} ranges from 0.1 to 2.1 atoms per formula unit, and $Fe^{2+}/(Fe^{2+} + Mg)$ varies from 0.46 to 0.79. The absence of a distinct miscibility gap in the measured compositions of the amphiboles is possibly significant. It suggests the amphiboles formed above 600°C, under supersolvus conditions (for example, Kimball, Spear, and Dick, 1985). Some amphiboles contain extremely high concentrations of chlorine (up to 5 wt percent). There is a strong positive correlation between chlorine content and $Fe^{2+}/(Fe^{2+} + Mg)$ (fig. 3). This relationship is consistent with structural controls on the chlorine content of amphiboles (Volfinger and others, 1985). The presence of chlorine-rich amphiboles indicates that the fluids that formed the early-stage veins were chloride-rich brines (Vanko, 1986).

Plagioclase in the early-stage veins is invariably fresh, with little or no textural evidence of incipient alteration to hydrous minerals. Many veins contain plagioclase that is substantially more calcic than that in the surrounding rock; differences up to 25 mole percent anorthite occur across the margins of many veins (fig. 4).

Hydrothermal clinopyroxene is common at or near vein margins, where it occurs as fine-grained pseudomorphs after igneous clinopyroxene

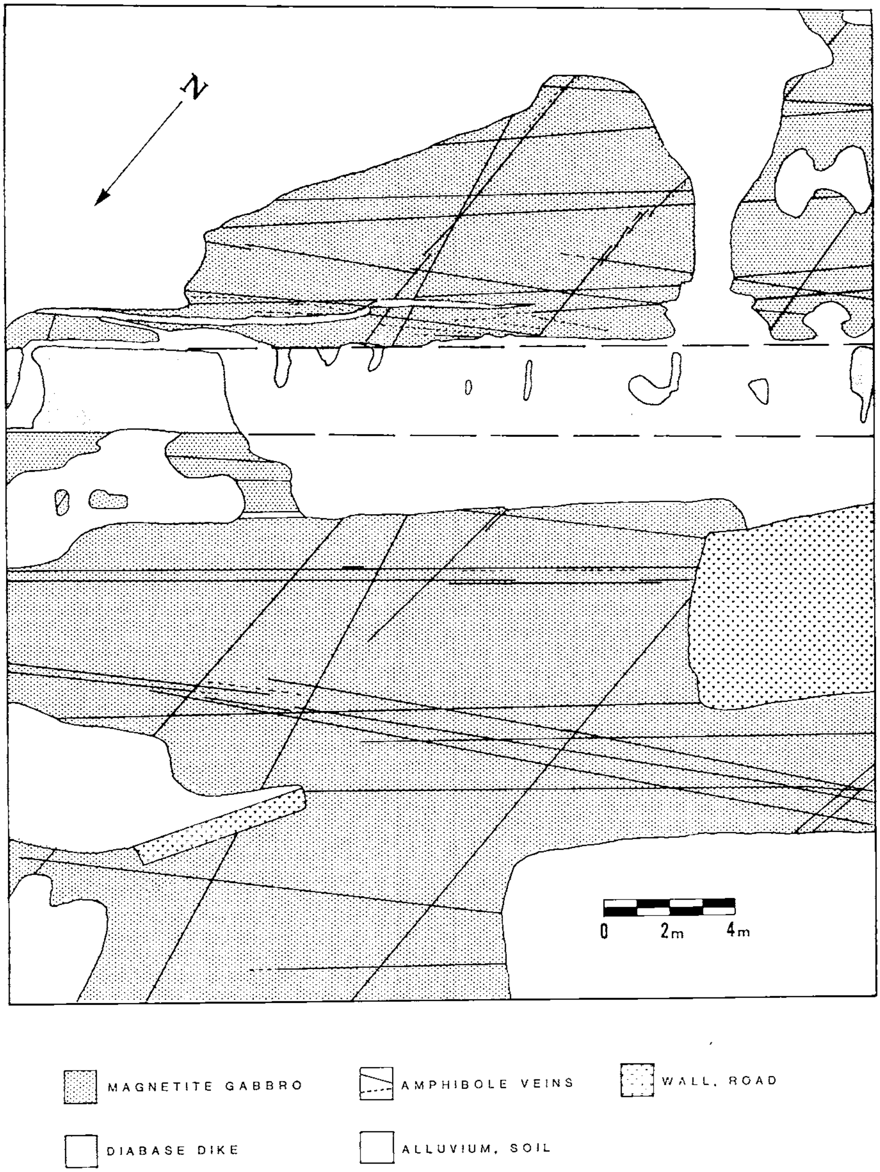
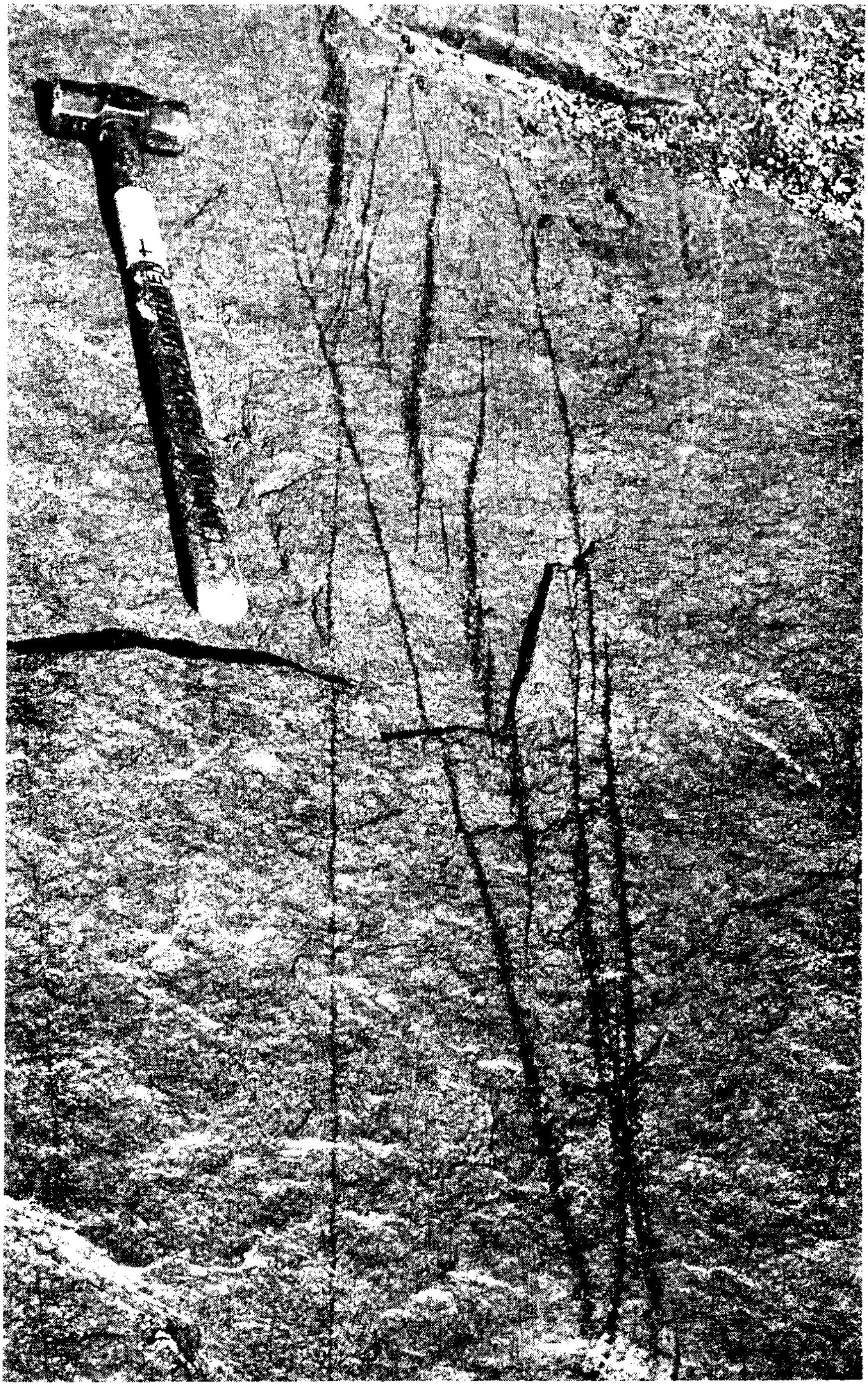


Fig. 2. Detailed outcrop map (1:20) at the type locality showing the distribution of early-stage amphibole veins in magnetite gabbro of the Upper Zone. The veins are cross-cut by later diabase dikes that were emplaced along dilational fractures. The veins occur along three distinct sets of fractures and generally have dips greater than 60°.



Field photograph of the *en echelon* structure that is commonly observed near the termination of early-stage veins at the type locality in the Upper Zone.

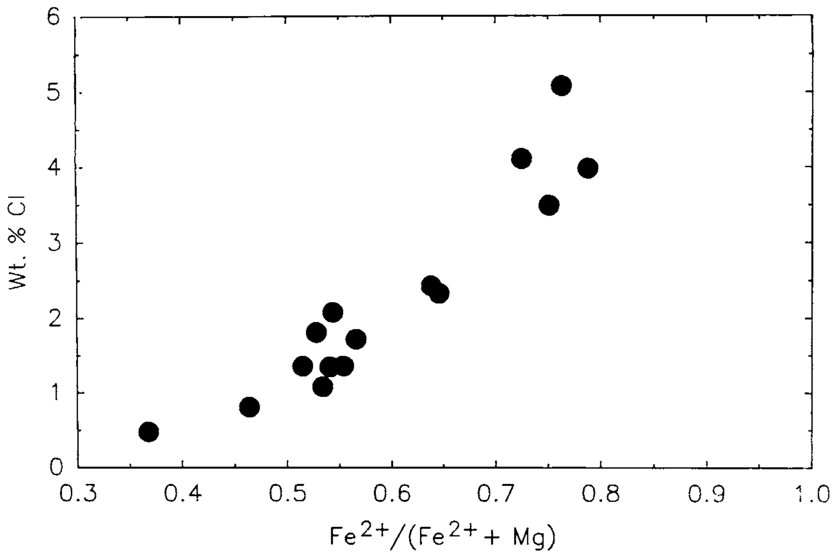


Fig. 3. Chlorine content versus $Fe^{2+}/(Fe^{2+} + Mg)$ of calcic amphibole in an early-stage vein at the type locality in the Upper Zone. Chemical analyses were obtained with a Cameca MBX electron microprobe equipped with 3 wavelength dispersive spectrometers and a Tracor-Northern automation system. The instrument was operated with an accelerating potential of 15 kV, a beam current of 15 nA, and a maximum counting time of 30 sec.

(pl. 4). The replacement process is accompanied by progressive chemical and textural changes that are essentially identical to those described in the following section for middle-stage veins.

Early-stage hydrothermal veins are common in all parts of the Bushveld Complex we have examined, but there are significant variations from the characteristics described above. In some areas the mineralogy of the early-stage veins is very similar to that of the type locality, but the field relations are quite different. For example, many outcrops of Main Zone gabbro are crosscut by a reticulate network of closely spaced (10-100 m^{-1}), narrow (approx 1 mm wide) veins (pl. 1-B). On a microscopic scale, the widths of these veins can be seen to be strongly dependent upon the local mineralogy (pl. 4). Where a fracture passes from a grain of pyroxene to an adjacent grain of plagioclase, the width of the vein decreases from 1 mm to 0.2 mm.

The mineralogy of early-stage veins in some areas is rather different from that in the type locality. For example, orthopyroxenite layers of the Lower Zone in the Oliphants River trough are crosscut by early-stage veins predominantly composed of an orthopyroxene texturally and chemically distinct from the orthopyroxene in the host rock. Relative to the magmatic orthopyroxene, orthopyroxene in the veins is enriched in Ca but depleted in non-quadrilateral components, particularly Al and Cr (fig. 5).

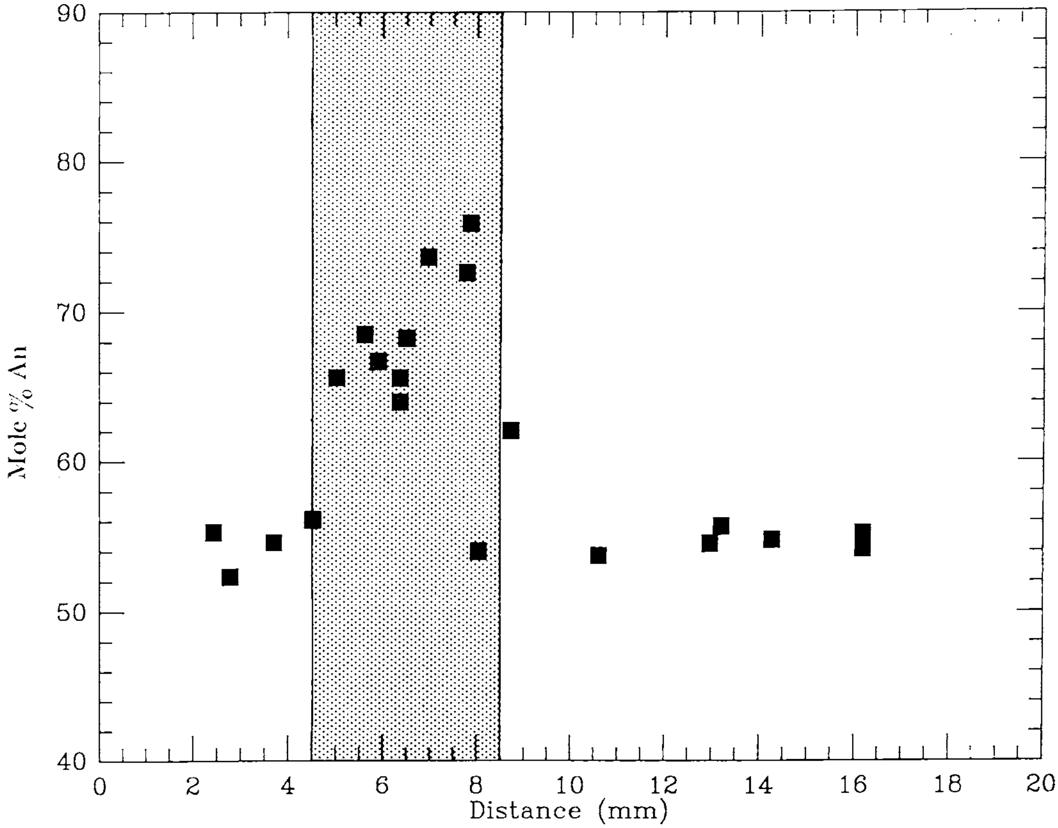


Fig. 4. Variation in plagioclase composition across an early-stage vein at the type locality in the Upper Zone. Plagioclase in the vein (shaded) is strongly enriched in anorthite relative to plagioclase in the host rock. Instrumental data related to conditions for the chemical analyses are summarized in the caption of figure 3.

Early-stage veins also occur in the contact aureole beneath the Bushveld Complex. The frequency in the aureole of the veins is strongly dependent upon the type of host rock, but there is a distinct increase in abundance as the intrusion is approached. A particularly interesting example of early-stage veins in the contact aureole occurs on the farm Middelkraal 221 JS, where volcanic rocks of the Dullstroom Formation are intruded by the Main Zone rocks of the Bushveld Complex. The volcanic rocks are crosscut by a network of narrow (1 mm) veins that are primarily composed of hornblende. Intense contact metamorphism resulted in migmatization of the volcanic rocks immediately adjacent to the intrusion. Igneous segregations produced by partial melting of the volcanic rocks appear to crosscut the earliest hydrothermal veins (pl. 5), indicating that a hydrothermal circulation system had been initiated prior to the onset of partial melting in the country rocks.

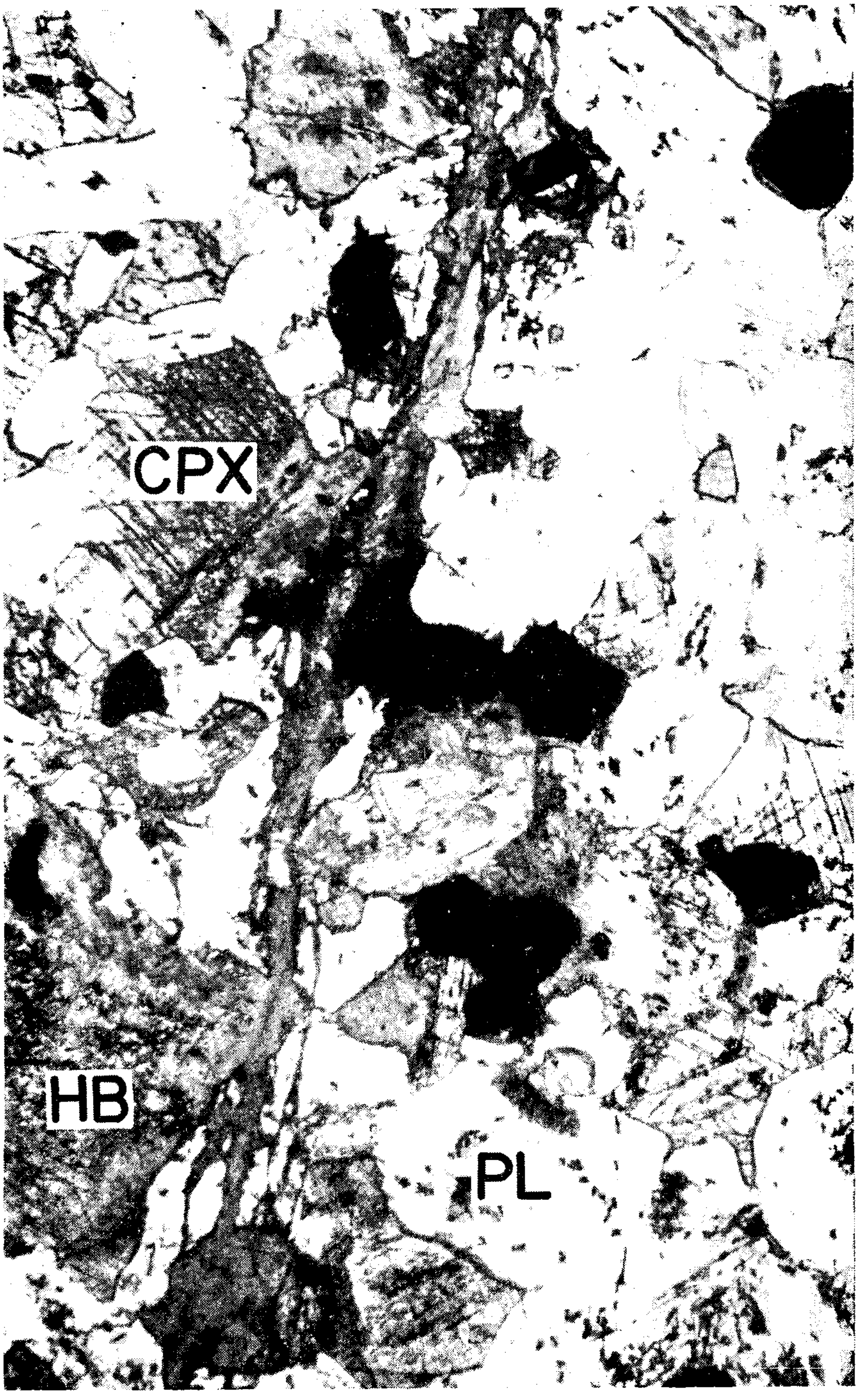
Middle-stage hydrothermal veins.—A regionally extensive swarm of veins has been traced for more than 15 km between Stoffberg and Belfast (fig. 1), and it is the type example selected for middle-stage veins. The veins are found in a two-pyroxene, Main Zone gabbro that has an unusually dark color and is locally called "black granite." They are well exposed in a series of dimension-stone quarries. Middle-stage veins in this region are bright green in color and are surrounded by a distinct, whitish-colored zone of intense wall-rock alteration. The strong color contrast between the white alteration halo and the dark gray host rock makes the veins very conspicuous (pls. 1-C and 6).

The middle-stage veins are generally between 0.5 and 5 cm in width (pl. 1-C), but occasional veins reach a thickness of 10 cm (pl. 6). The white alteration zone is commonly about one-third the width of the vein. Individual veins are commonly continuous for 20 m or more along strike. Most of the veins occur in steeply dipping sets of conjugate fractures, although a single orientation is dominant in most localities. The frequency of veins in the quarries is between 0.2 and 1 m^{-1}; however, the average frequency for the region may be greater than this because the miners avoid rocks that are highly veined.

The mineralogy of the green, middle-stage veins is variable. In many instances they are composed largely of clinopyroxene and quartz, with subordinate amounts of sphene and actinolite. In some instances the most abundant mineral is actinolite or actinolitic hornblende. Chalcopyrite is locally abundant in a few veins. Some veins have been affected by multiple episodes of fracturing and mineral growth, and early clinopyroxene can be seen to be partially replaced by later actinolite.

The mineralogy of the white wall-rock alteration is dominated by clinopyroxene, albite, and pumpellyite, with subordinate amounts of sphene, actinolite, quartz, microcline, carbonate, and epidote. Plagioclase in the alteration halo is completely replaced by an extremely fine-grained intergrowth of albite and pumpellyite, plus trace amounts of microcline and epidote. In some cases, clinozoisite occurs near the interface between the vein and the white alteration zone. In addition to the prominent

PLATE 4



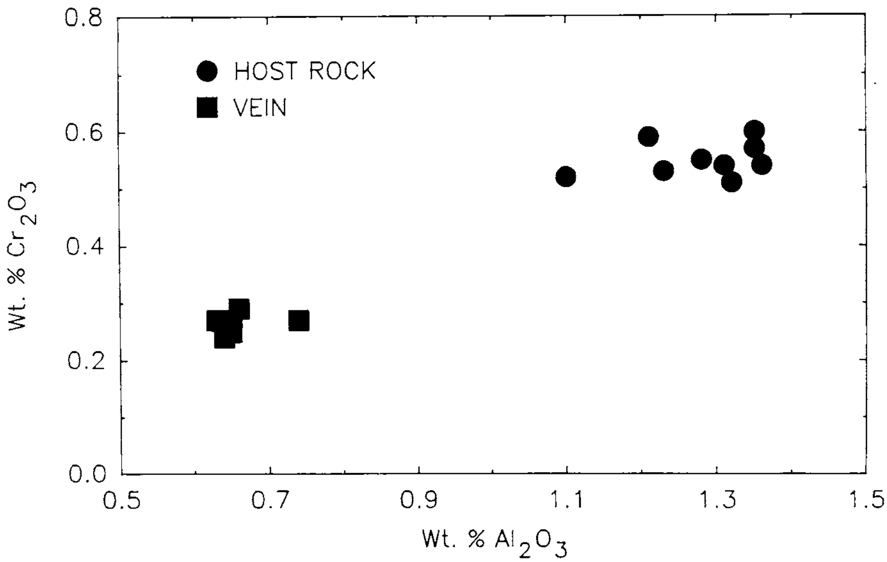


Fig. 5. Comparison of hydrothermal orthopyroxene from an early-stage vein with magmatic orthopyroxene from the adjacent host rock. Hydrothermal orthopyroxene (squares) is depleted in non-quadrilateral components relative to orthopyroxene in the host rock (circles). Analytical conditions are summarized in the caption of figure 3.

alteration zone there is a more subtle alteration zone that surrounds the white sheath. This zone, which is enriched in ferromagnesian minerals, particularly biotite, actinolite, and chlorite, is approximately half the width of the white alteration zone.

The clinopyroxene that formed through hydrothermal processes in the middle-stage veins occurs as equigranular, euhedral to subhedral grains which are substantially coarser than the grains of igneous clinopyroxene in the adjacent host rock. Exsolution bodies and twin lamellae are common in the igneous clinopyroxene grains but are relatively uncommon in hydrothermal clinopyroxene. Relative to its igneous counterpart, hydrothermal clinopyroxene is enriched in Ca but strongly depleted in non-quadrilateral components, particularly Al, Ti, and Cr (fig. 6). Clinopyroxene in the alteration halo has a chemical composition intermediate between those of the hydrothermal and magmatic clinopyroxene. Igneous clinopyroxenes exhibit a series of progressive textural changes as a vein is approached. These include a coarsening of the exsolution lamellae, development of granular exsolution, and, immediately adjacent to the vein, the formation of relatively homogeneous grains that display little twinning or exsolution. The progressive chemical and textural

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← Photomicrograph of an early-stage hornblende vein (vertical) in gabbro from the Main Zone (Mapochsgronde 500 JS). Plagioclase (PL) near the vein is fresh. One grain of clinopyroxene near the vein has been completely replaced by hornblende (HB). Another grain of igneous clinopyroxene (CPX) has been partially replaced by fine-grained hydrothermal clinopyroxene (top edge of the grain) and hornblende (right edge of the grain). The field of view is 3 mm wide.

(A) Contact metamorphism produced a partial melt (light color) that crosscuts earlier hydrothermal veins.

Formation immediately adjacent to the Bushveld Complex

(B) Close-up of the crosscutting relations.

PLATE 6

Photograph of a 10-cm wide middle-stage vein from the type locality in the Main Zone. The width of the alteration halo is about one-fifth the width of the vein. The original igneous texture is still visible in the white alteration halo.

changes that accompany the replacement of magmatic clinopyroxene by hydrothermal clinopyroxene in the Bushveld Complex are similar to the sequence of changes that occur adjacent to hydrothermal veins in the Skaergaard intrusion (Manning and Bird, 1986).

Quartz occurs in the middle-stage veins as anhedral grains that are commonly interstitial to the clinopyroxene grains. The quartz contains abundant fluid inclusions, many of which contain a halite daughter crystal. The presence of fluid inclusions saturated with respect to halite indicates that at least some of the fluids that flowed along these fractures were highly saline brines.

Late-stage hydrothermal veins.—The type examples of late-stage veins occur in anorthosites of the Critical Zone on the farm Dwars River 372 KT. The veins are white to light green in color, they tend to be

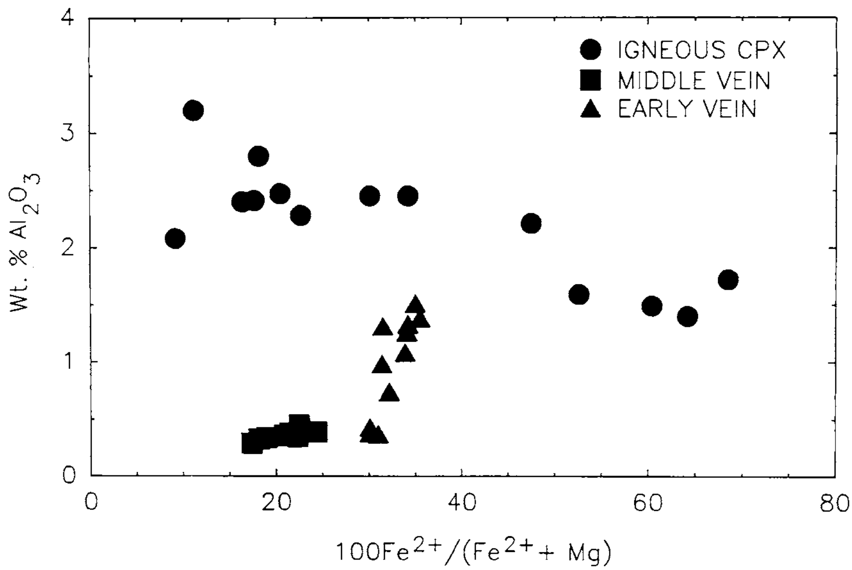

Fig. 6. Comparison of the composition of magmatic clinopyroxene from throughout the Bushveld Complex (circles), with hydrothermal clinopyroxene from a middle-stage vein (squares). The large, euhedral grains of hydrothermal clinopyroxene are strongly depleted in non-quadrilateral components relative to all magmatic clinopyroxene. Partially replaced clinopyroxene grains from an early-stage vein (triangles) have compositions that span the range between magmatic and hydrothermal clinopyroxene. Analytical conditions are summarized in the caption of figure 3. Data for magmatic clinopyroxene are from Atkins (1969).

small, and they occur in great abundance. Most of the veins are approx 1 mm wide, although a few are locally more than 1 cm wide. Late-stage veins in the type area are commonly associated with a set of subparallel shear zones. The distribution of macroscopic veins is highly irregular; vein frequencies of greater than 100 m^{-1} are attained within 1 m of a shear, but the frequency falls off to less than 10 m^{-1} in intervals between the shears. Although the vein systems exhibit a preferred orientation parallel to the shears on the scale of an outcrop, individual veins are highly irregular on the scale of a thin section, with abundant branching and crosscutting relations.

Most of the late-stage veins in the type area consist largely of prehnite and quartz, with subordinate amounts of pumpellyite. Other minerals, including zeolites, calcite, and sulfides are locally abundant.

Some of the late-stage veins are strongly zoned and exhibit unambiguous textural evidence of open-space filling. A typical quartz + prehnite vein has the following sequence of zones: a central core of anhedral quartz, a pair of symmetric bands of cockade-textured prehnite, then a sequence of crustiform layers of anhedral prehnite adjacent to the wall rock (pl. 7). The veins are surrounded by a broad halo of wall-rock alteration. The quartz core contains rare fragments of cockade-textured

Photomicrograph of a late-stage vein in mottled anorthosite of the Critical Zone, at the type locality on the farm Dwars River 372 KT. The central core consists of quartz (dark) and is bounded by bands of cockade-textured prehnite. The cockade-textured prehnite is strongly zoned, and crystals on opposite sides of the quartz core have symmetric zoning profiles. The field of view is 1 mm wide. Crossed-polarized light.

prehnite that broke away from the vein walls. Fluid inclusions are common in the quartz, and many of the inclusions contain a halite daughter crystal. The cockade-textured prehnite has strong internal growth-zoning developed parallel to crystal faces that project into the quartz core, and crystals on opposite sides of the core have symmetric zoning profiles. Compositional variations in crystals of prehnite expressed in terms of $Fe^{3+}/(Fe^{3+} + Al)$ range from 0.005 to 0.14. The zoning is very striking in thin section because the compositional changes produce large variations in the indicies of refraction. The crustiform layers of prehnite also have large variations in composition, but the variations are highly irregular due to the superimposed effects of repeated fracturing and fluid flow.

Some late-stage veins are surrounded by a zone of wall-rock alteration several times as wide as the vein. The alteration halos of adjacent veins may merge together, and in some cases the overlapping alteration halos permeate through the entire rock and produce the effect of pervasive alteration.

Late-stage hydrothermal veins occur in other parts of the Bushveld Complex. They are particularly abundant in anorthosite layers that lie immediately below magnetitite layers of the Upper Zone. There, the late-stage veins have frequencies that exceed 100 m^{-1}, and they occur as irregular, three-dimensional networks that penetrate entire outcrops (pl. 1-D).

DISCUSSION AND CONCLUSIONS

The fracture-controlled flow of aqueous fluids that produced the extensive networks of hydrothermal veins here described from the eastern lobe of the Bushveld Complex are also present in the other lobes of the Bushveld Complex we have examined. To date, however, we have only had an opportunity to map carefully in the eastern lobe of the complex. Our observations establish that extensive networks of hydrothermal veins occur across a stratigraphic distance of 9 km and a strike length of 100 km. The Bushveld hydrothermal system has affected tens of thousands of cubic kilometers of rock in and around the eastern lobe alone, and it is probably among the largest paleo-hydrothermal systems in the world. The Bushveld hydrothermal system is probably much larger than the largest hydrothermal system that has been documented around a single granitic pluton (Criss, Ekren, and Hardyman, 1984).

Fluid-rock interactions occurred over a wide range of physical and chemical conditions, producing a wide variety of hydrothermal veins. Broad constraints on the conditions under which the various veins formed can be inferred from the mineral assemblages in the veins and from the alteration halos. More exact estimates of intensive thermodynamic parameters associated with the formation of individual veins will be presented in a subsequent paper that includes a detailed discussion of mineral chemistry and element partitioning among coexisting phases.

Early-stage veins are predominantly composed of calcic amphibole and calcic plagioclase, with subordinate amounts of clinopyroxene and Fe–Ti oxides, ± biotite ± orthopyroxene ± sulfides. These mineral assemblages are characteristic of metamorphosed mafic igneous rocks that have been equilibrated at upper amphibolite facies conditions and are indicative of temperatures greater than 600°C.

The mineral assemblages of the middle-stage veins do not always provide tight constraints on the conditions under which they formed. However, the alteration halos around middle-stage veins contain actinolite, albite, chlorite, quartz, and sphene. This mineral assemblage is typical of metamorphosed mafic igneous rocks that have been equilibrated at greenschist facies conditions and is indicative of temperatures between 300° and 600°C.

Late-stage veins composed of prehnite, quartz, and pumpellyite are representative of prehnite–pumpellyite facies conditions. These veins probably formed at temperatures below 300°C.

The mineral assemblages in the three families of veins exhibit a decreasing temperature sequence with decreasing relative age (table 1). The temperature sequence represents an alternative basis for classifying the veins: high, intermediate, and low temperature veins correspond to early-stage, middle-stage, and late-stage veins, respectively. A classification system based on relative temperature, rather than relative timing, can be useful in areas where only one generation of veins is present, or where the relative chronology of the veins has not been determined. A classification based on temperature may be helpful in order to make comparisons with vein systems in other mafic intrusions.

Textural evidence indicates that some of the hydrothermal minerals formed by the replacement of magmatic precursors, whereas others precipitated directly from hydrothermal fluids. The development of alteration minerals has progressed further adjacent to low and intermediate temperature veins (pls. 1-C, D, and 6) than high temperature veins (pls. 1-A, 3, and 4). Take the example of plagioclase. Adjacent to high temperature veins, plagioclase is texturally fresh, but adjacent to intermediate temperature veins it is replaced by an intergrowth of albite and hydrous minerals, and adjacent to low temperature veins it has been completely replaced by hydrous minerals, such as pumpellyite and prehnite.

Fluid inclusions are widespread in quartz that occurs in low and intermediate temperature veins. A microthermometric study is in progress that will provide quantitative constraints on the compositions and concentrations of the fluids, as well as the temperatures of the fluid-rock interactions. Initial results of this study indicate some of the fluids were highly saline, because the fluid inclusions contain a halite daughter crystal, although lower salinity inclusions are also present.

Fluid inclusions are rare in most of the early-stage, high temperature veins we have examined. However, evidence that at least some of the early, high temperature fluids were highly saline is indicated by the occurrence of amphibole that contains high concentrations of chlorine (up to 5 wt percent). The partitioning of chlorine between amphibole and a coexisting aqueous fluid is poorly understood, but similar high-chlorine amphiboles from oceanic gabbros are believed to have formed as the result of interactions with highly-saline hydrothermal fluids (Vanko, 1986). Most amphiboles in low and intermediate temperature veins have relatively low concentrations of chlorine, but this does not imply that these amphiboles formed as the result of interactions with low-salinity fluids. Amphiboles in the low and intermediate temperature veins we have investigated tend to have low $Fe^{2+}/(Fe^{2+} + Mg)$ ratios, and the structures of these amphiboles do not favor the fixation of chlorine due to the small size of the anionic (OH, Cl) site (Volfinger and others, 1985).

Saline fluid inclusions and chlorine-rich hydrous minerals are not restricted to hydrothermal veins in the Bushveld Complex. These features have been reported recently from the Merensky Reef and other stratiform layers (Schiffries, 1984; 1985a; Johan and Watkinson, 1985; Ballhaus and Stumpfl, 1986; Boudreau, Mathez, and McCallum, 1986), as well as from mafic pegmatoids that are transgressive to the layering (Schiffries, 1982). Although we have not yet determined whether the fluids that produced the fluid inclusions and chlorine-rich hydrous minerals in the stratiform layers and the transgressive pegmatoids are genetically related to the fluids that produced the networks of hydrothermal veins described in this paper, it is clear that different flow regimes, pervasive as opposed to channelized, would be necessary to explain the origin of these features.

The timing of the onset of the post-magmatic, channelized circulation system can be constrained by field evidence. Early-stage, high temperature veins are present in both the Bushveld Complex and the pre-Bushveld rocks of the contact aureole. Within the Dullstroom Volcanics, and immediately adjacent to the intrusion, there are numerous pods and segregations of migmatite. A few hundred meters away from the intrusive contact the migmatitic segregations are absent. The early-stage hydrothermal veins are present in both the intrusion and the contact aureole. They pre-date and are cut by the migmatitic segregations (pl. 5). The relative ages of the early-stage veins and the migmatitic segregations indicate that the hydrothermal activity responsible for the early-stage veins must have commenced before the peak of migmatitic melting was reached and, possibly therefore, before the process of magmatic intrusion was complete.

The timing of the middle-stage veins can also be related to the cooling history of the Bushveld Complex, although less convincingly than for the early-stage veins. Middle-stage veins are widespread in the Complex, and they are the vein set we have observed most commonly. We have observed them in the three major lobes of the Bushveld, but not in any of the post-Bushveld rocks. The wide geographic distribution of middle-stage veins in the intrusion and their absence in post-Bushveld rocks are strong evidence in favor of a Bushveld-controlled origin.

Evidence concerning the time of formation of late-stage, lower temperature veins is less constrained. We strongly suspect that there is a continuum in the vein-forming processes, and that the formation of late-stage veins was the last hydrothermal manifestation of the Bushveld's cooling history. We have been unable, so far, to prove the point. It is possible that late-stage veining was a consequence of post-Bushveld magmatism or tectonism. We do not believe this is a likely possibility for the following two reasons. First, while there are low temperature hydrothermal veins of questionable origin that cut the Bushveld, they are restricted geographically, and they have distinct compositions. Examples of such veins can be seen in the region around the Pilanesberg intrusion. This is an alkaline intrusion that postdates the Bushveld Complex by 500 Ma. Dikes and veins associated with the Pilanesberg intrusion are structurally and geographically constrained. The veins have quite different compositions from the early-stage hydrothermal veins we have described. For example, the Pilanesberg-associated hydrothermal veins have been observed to contain fluorite (observations in the Rustenberg area made by Skinner in 1976). Such observations suggest that even though post-Bushveld veins may be present, they are compositionally distinct and geographically restricted to the region around the intrusions from which they were derived.

A second reason for our conclusion that late-stage veins are artifacts of the cooling of the Bushveld Complex concerns the Bushveld Granite. The Bushveld Granite is the only post-Bushveld igneous body large enough to generate a hydrothermal system that could produce Bushveld-wide, late-stage hydrothermal veins. There are certainly veins in, and associated with, the Bushveld Granite and also veins associated with other post-Bushveld rocks, but none of these veins resembles the early-, middle-, or late-stage veins we have described. The mineralogies are very different, the veins appear to be restricted to the granitic intrusion and the immediately surrounding rocks, and so far as we know there is no good evidence that the Bushveld veins we describe cut any of the younger rocks. Whilst not proven, therefore, the evidence strongly suggests that the processes we describe had ceased by the time the Bushveld Complex had cooled and before the next major igneous intrusion or tectonic event in the region.

Insights into the types of processes that may have operated in the Bushveld Complex can be gained from studies of other mafic intrusions. It has been proven that hydrothermal systems were generated by several epizonal layered mafic intrusions as they crystallized and cooled. Examples of gabbro-hosted hydrothermal systems that occur in continental crust include the Skaergaard intrusion (Taylor and Forester, 1979, Norton and Taylor, 1979; Norton, Taylor, and Bird, 1984; Bird, Rogers, and Manning, 1986) as well as the intrusions at Skye, Mull, and Ardnamurchan (Taylor and Forester, 1971; Forester and Taylor, 1977; Ferry, 1985). An example of a gabbro-hosted hydrothermal system that occurs in oceanic crust is present in the Samail ophiolite in Oman (Gregory and Taylor, 1981).

The most instructive and complete set of observations of a gabbro-hosted hydrothermal system has been made on the Skaergaard intrusion.

Several superimposed networks of hydrothermal veins are present in the Skaergaard intrusion, and as in the case of the Bushveld Complex, the mineral assemblages in the veins exhibit a decreasing temperature sequence with decreasing age. The fracture patterns and mineral assemblages of high temperature veins in the Skaergaard intrusion are very similar to those in the Bushveld Complex, and it is probable that the processes responsible for their formation were similar in both cases. The hydrothermal system was established early in the cooling history of the intrusion, and it appears to have ceased by the time the Skaergaard had cooled. Stable isotope studies have established that the Skaergaard intrusion underwent a significant and widespread depletion in ^{18}O due to subsolidus isotopic exchange with hydrothermal fluids derived from meteoric waters (Taylor and Forester, 1979). The hydrothermal veins are interpreted as flow channels for the fluids that altered the isotopic composition of the gabbro as the intrusion cooled. The lack of pervasive hydrothermal alteration throughout the Skaergaard intrusion implies that much of the isotopic exchange took place under conditions that were within the stability limits of the original minerals. Numerical modelling of the Skaergaard hydrothermal system indicates that at least 75 to 90 percent of the fluid-rock exchange occurred at temperatures above 450°C (Norton and Taylor, 1979; Norton, Taylor, and Bird, 1984).

There are important chemical differences between the hydrothermal fluids that interacted with the Bushveld Complex and those that interacted with the Skaergaard intrusion. The fluids that circulated through the Skaergaard intrusion were very dilute brines, and there was little chemical advection along the hydrothermal veins (Norton, Taylor, and Bird, 1984; Bird, Rogers, and Manning, 1986). In contrast, many of the fluids that circulated through the Bushveld Complex were highly saline brines, and such fluids have the capacity to transport appreciable quantities of dissolved metals (Skinner, 1979). Chalcopyrite and other sulfide minerals are locally abundant in some hydrothermal veins. This indicates that copper and other base metals were transported by hydrothermal fluids and then deposited in the veins. In addition to sulfide minerals in hydrothermal veins, there are high platinum values in transgressive dunite pipes that have been interpreted as hydrothermal replacement bodies (Cameron and Desborough, 1964; Schiffries, 1982; Stumpfl and Rucklidge, 1982). These observations indicate that fluid-rock interactions may have played a more extensive role in determining the distribution of sulfide and platinum mineralization than has been previously recognized. Further consideration should be given to the possibility that some of the mineralogical, chemical, and isotopic features of the Bushveld Complex may not be due solely to primary magmatic processes but may instead have been influenced by the effects of subsequent fluid-rock interactions.

It is becoming increasingly clear that questions concerning the relative importance of crystal fractionation, magma mixing, crustal assimilation, and mantle heterogeneity cannot be resolved adequately until the effects of magmatic processes can be distinguished from those of hydrothermal processes. The recognition of hydrothermal processes in gabbroic intrusions raises many new questions and challenges the conventional answers to a plethora of longstanding ones.

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