A COMPARATIVE GEOLOGICAL STUDY OF SOME MAJOR KIMBERLITE PIPES IN THE NORTHERN CAPE AND ORANGE FREE STATE

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DECLARATION

I hereby declare that all the work presented in this thesis is my own, except where otherwise stated in the text.

Signed by candidate

Signature removed

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"And that, in seeking to undo
One riddle, and to find the true,
I knit a hundred others new:"

Tennyson

ABSTRACT

In the first part of this thesis the petrological status of 'kimberlite' is defined and mineralogical and textural classifications which allow systematic documentation and collation of these complex rocks are evaluated. Quantitative mineralogical classification based on the modal abundances of the matrix minerals is advocated. A textural classification is adopted within which different textures are related to different modes and conditions of kimberlite emplacement.

In the subsequent and major part of the thesis the results of a comparative geological study of six diamondiferous kimberlite pipes are reported. The pipes concerned are the De Beers, Wesselton, Dutoitspan, Bultfontein, Finsch and Koffiefontein occurrences (collectively termed the KIMFIK pipes). Prior to post-kimberlite erosion these pipes were characterized by morphologically distinctive crater, diatreme and root zones but only the lower parts of the diatreme zones and the underlying root zones are preserved.

The root zones are irregular in form and are characterized locally by prominent contact breccias. Most contact features of the root zones indicate that the morphology of these zones has been strongly influenced by pre-existing structures in the country rocks. The diatreme zones are more regularly shaped than the root zones and structural control of the form of these zones is less pronounced.

Evidence of the former presence of craters at the tops of the KIMFIK pipes is provided by down-rafted blocks of epiclastic kimberlite. These inclusions also indicate that the development of the pipes took place intermittently, over prolonged periods of time. The surface areas of the craters probably exceeded 60ha in all cases.

Four groups of kimberlite dykes and rare sills are associated with the pipes. These minor intrusions display diverse petrographic features and more than one period of intrusion is recognized within each group. The injection of precursor dykes and sills may have been a prerequisite for the subsequent formation of the pipes.

The root zones of the KIMFIK pipes commonly contain several discrete intrusions (6-18 major intrusions have been recognized in individual pipes). These intrusions display varied mineralogical and textural features. The diatreme zones of the pipes are mainly occupied by pelletal-textured tuffisitic kimberlite breccias (TKB's) and relatively few major intrusions (1-3) are present.

Two periods of crystallization of primary minerals in the KIMFIK intrusions are recognized; pre-intrusion (i.e. upper mantle) and post-intrusion (i.e. near-surface) crystallization has occurred. The general crystallization sequence (which ignores substantial crystallization overlaps and thus in effect refers to the onset of crystallization of each mineral) is:

Pre-intrusion: ilmenite/spinels/perovskite/rutile, olivine, diopside, phlogopite.

Post-intrusion: ilmenite/spinels/perovskite, diopside, monticellite, apatite, phlogopite, calcite and serpentine.

The foregoing sequence is not applicable to diatreme-facies kimberlites. In these rocks the crystallization sequence is affected by loss of volatiles, and, probably, by the incorporation of contaminants. The main effect is delayed or prolonged crystallization of groundmass diopside.

It is suggested, mainly on the basis of their variable compositions and limited compositional zoning, that the ubiquitous olivine phenocrysts in the KIMFIK intrusions crystallized in the upper mantle (in many instances from discrete magma batches that were subsequently mixed) prior to the incorporation of upper mantle peridotite-derived olivine xenocrysts.

Whole-rock major and trace element analyses suggest that bulk composition variations in the Kimberley pipes may be due in the first instance to mixing of magma batches which were subjected to varying degrees of olivine and orthopyroxene enrichment by the addition of 'average' upper mantle. Further olivine enrichment (involving early-crystallized olivine phenocrysts) probably reflects batch-mixing and/or degassing.

It is concluded that the diamonds in the KIMFIK pipes are xenocrysts and that the grade variations within and among discrete intrusions in the pipes may partly be related to differential sampling of diamondiferous upper mantle sources. Diamond distributions and abundances also reflect varied degrees of mixing during intrusion. Such mixing may involve more than one batch of mixing or may relate to mixing, involving turbulence or convection, within individual batches.

A general theory of kimberlite pipe formation is proposed which takes into account the geological complexity of the KIMFIK pipes:

To account for root zone features emphasis is placed on the importance of a variety of interrelated subsurface processes which result in the initial formation of embryonic pipes. These embryonic pipes develop (migrate) upwards and culminate in explosive breaching of the surface from depths of less than 0,5km (the explosions may, at least in part, be phreatomagmatic in character). The pre-breakthrough subsurface processes involved in embryonic pipe formation include hydraulic fracturing and wedging, magmatic stoping, brecciation by intrusion of magma along discontinuities in the wallrocks, intermittent explosive and/or implosive brecciation of wallrocks associated with vapour differentiates and/or ground-water intersection, spalling, slumping and, possibly, rock bursting from temporary free faces.

The subsequent development of kimberlite pipes is activated by explosive breaching of the surface with the consequent formation of explosion craters. Explosive breakthrough is probably accompanied by authigenic brecciation at levels below the base of the crater zones.

The development of diatreme zones is ascribed to post-breakthrough modification of the basal parts of the crater zones and considerable parts of the extended root zones, i.e. the embryonic pipe columns. The rapid depressurization accompanying explosive breakthrough would result in an upsurge of precursor gas phases and partly degassed magma and the nature of the TKB's in the diatreme zones of the KIMFIK pipes indicates that vapour-solid-liquid fluidized systems developed at this stage. As a consequence of vaporization and rapid adiabatic expansion these systems are considered to have evolved backwards down the embryonic pipes. During this process diatreme zones would be formed by the incorporation of previously brecciated sidewall material into the fluidized systems. Fluidized intrusion is considered to have been short-lived due to rapid quenching of the systems.

Following the cessation of fluidization and concomitant deflation and consolidation, the final stage of the formation of kimberlite pipes is the deposition of epiclastic kimberlite (derived from the erosion of tuff rings) in the crater zones.

The KIMFIK pipes reflect multiple intrusive events hence the cycle of formation summarized above may have been repeated several (in some instances many) times. However, intrusion parameters are likely to have varied considerably in respect of each cycle of events hence the relative importance and duration of each of the processes involved is likely to have varied and individual cycles may not always have continued to completion. Variations in intrusion parameters may result in diatreme and root zone characteristics being developed at the same level in different parts of a pipe (as is the case in several KIMFIK pipes).

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I am grateful to many people for help in various forms during the long and often unsteady course of the research which is reported in this thesis:

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Prof. A M Reid took over the supervision of the project after the departure of Dr A C Moore from U C T and also edited a draft manuscript. I am grateful for his expert help and advice and for prompting me into 'getting a move on' towards the end.

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Most writers of theses receive much assistance and encouragement from their families and I am no exception in this respect. The enthusiastic support which I received from my mother and late father is much appreciated and I am especially grateful to my wife for helping in innumerable ways and for 'standing back' so often. Joan and our children, Vivienne, Adrienne and Roderick probably felt it would never end. I thank them for their patience.

FOREWORD

Several papers embodying some of the results obtained from work carried out for this thesis have been published in the scientific literature. These papers, some of which involve co-authors, are listed below.

1. Clement C.R. (1973) Kimberlites from the Kao pipe, Lesotho. In: Lesotho Kimberlites (editor P.H. Nixon), 110-121. Lesotho Natl. Dev. Corp., Maseru.
2. Clement C.R. (1975) The emplacement of some diatreme-facies kimberlites. In: Physics and Chemistry of the Earth, 9 (editors L.H. Ahrens, J.B. Dawson, A.R. Duncan and A.J. Erlank), 51-59. Pergamon Press, Oxford.
3. Clement C.R. (1979) The origin and infilling of kimberlite pipes. In: Extended abstracts, Kimberlite Symposium II, Cambridge.
4. Clement C.R., Gurney J.J. and Skinner E.M.W. (1975) Monticellite - an abundant groundmass component of some kimberlites. In: Extended abstracts, Kimberlite Symposium I, Cambridge.
5. Clement C.R., Skinner E.M.W. and Scott B.H. (1977) Kimberlite redefined. In: Extended abstracts. Second International Kimberlite Conference, Sante Fé, New Mexico.
6. Clement C.R., Skinner E.M.W., Hawthorne J.B., Kleinjan L. and Allsopp H.L. (1979) Precambrian ultramafic dykes with kimberlite affinities in the Kimberley area. In: Kimberlites, Diatremes and Diamonds: Their Geology, Petrology and Geochemistry (editors F.R. Boyd and H.O.A. Meyer), 101-110. Am. Geophys. Union, Washington.
7. Clement C.R. and Skinner E.M.W. (1979) A textural-genetic classification of kimberlitic rocks. In: Extended Abstracts, Kimberlite Symposium II, Cambridge.

8. Clement C.R., Harris J.W., Robinson D.N. and Hawthorne J.B. (In Press) The De Beers kimberlite pipe - A historic South African diamond mine. Geol. Soc. S. Afr., Johannesburg.

9. Skinner E.M.W. and Clement C.R. (1979) Mineralogical classification of southern African kimberlites. In: Kimberlites, Diatremes and Diamonds: Their Geology, Petrology and Geochemistry (editors F.R. Boyd and H.O.A. Meyer), 129-139. Am. Geophys. Union, Washington.

10. Hawthorne J.B., Carrington A.J., Clement C.R. and Skinner E.M.W. (1979) Geology of the Dokolwayo kimberlite and associated paleo-alluvial diamond deposits. In: Kimberlites, Diatremes and Diamonds: Their Geology, Petrology and Geochemistry (editors F.R. Boyd and H.O.A. Meyer), 59-70. Am. Geophys. Union, Washington.

The author's responsibilities in respect of the papers which involve joint authorship are noted below.

Paper 4: Written by CRC. Petrography by CRC and EMWS. Chemical analyses by JJG.

Paper 5: Joint responsibility of the authors. Written by CRC.

Paper 6: Petrography by CRC and EMWS. All other aspects of paper except section dealing with age dating by CRC.

Paper 7: Classification devised jointly by CRC and EMWS. Written by CRC.

Paper 8: All aspects of paper by CRC except sections dealing with diamonds.

Paper 9: Classification devised jointly by EMWS and CRC. Paper written by EMWS.

Paper 10: Author responsible for petrography and petrographic interpretation only. Paper written by JBH.

As a result of additional work since the papers were written some of the conclusions recorded in Papers 1, 2, 3, 5, 7 and 9 have been revised (in some instances considerably) in this thesis.

CONTENTS

VOLUME 1

VOLUME 2

CHAPTER 1

INTRODUCTION

1.1 SCOPE AND PURPOSE OF THE INVESTIGATION

This study is primarily concerned with the geology of the six major kimberlite pipes that are mined for diamonds by the Kimberley Division of De Beers Consolidated Mines Limited. Four of these pipes (De Beers, Wesselton, Bultfontein and Dutoitspan) lie within, or immediately adjacent to, the City of Kimberley. The City owes its existence to the discovery of these diamondiferous pipes (together with the famous Kimberley pipe or 'Big Hole') in the latter half of the 19th Century. The Kimberley group of occurrences, which include 5 major pipes and a number of small neighbouring bodies, constitute the type locality for kimberlites.

The remaining two pipes studied are the Koffiefontein occurrence which lies approximately 80km SSE of Kimberley and the much more recently discovered (1961) Finsch pipe which is located 150km west of Kimberley. All six localities are shown in Figure 1.1, together with the positions of several other kimberlite occurrences to which references are made in the text. Such references have generally been made to support observations or interpretations relating directly to the six major pipes studied. For brevity and convenience these six pipes are henceforth collectively referred to as the KIMFIK pipes (KIM=Kimberley group, FI=Finsch, K=Koffiefontein).

The nature of the work carried out and the extent of the investigations has varied considerably from pipe to pipe. This variation has been largely governed by the degree of accessibility and extent of exposure within the individual mines during the course of the study. The geological complexity of individual pipes has also strongly influenced the amount of work carried out. Relatively little attention has been paid to Bultfontein as, at current levels of exposure, it is geologically the simplest of the KIMFIK group. Furthermore, mining operations at Bultfontein were suspended for three years during the duration of this study.

A fortunate aspect of the investigation is that mining operations have advanced to different degrees at the individual mines. As a result exposures at different levels (in relation to the original surfaces) can be examined. Consequently it has been possible to establish a composite picture of pipe characteristics over a vertical distance in excess of 1000m.

The present study has largely been restricted to investigations relating to the nature of the pipes and their infilling material. It is only indirectly concerned with the ultimate origin of kimberlite magmas in the upper mantle or with their evolutionary history prior to reaching near-surface zones in the earth's crust. The investigation was initially prompted by a realization that a better geological understanding of these major mines might be of considerable economic as well as academic interest. The work was also prompted by the realization that these mines, like all others, are waning assests and therefore much useful information would be lost if it was not formally documented. It is, after all, only very rarely that one has an opportunity to investigate in considerable detail working mines that have been in existence for more than 100 years.

1.2 PREVIOUS WORK

The scientific literature dealing with kimberlites and/or the upper mantle-derived xenoliths which they contain is vast and varied and many publications exist which deal with specific aspects of one or other, or all, of the KIMFIK pipes. There are, however, relatively few contributions which deal comprehensively with the geology of the pipes. The last major works to be published are those of Wagner (1914), whose 'Diamond Fields of Southern Africa' is a classic contribution to kimberlite literature and Williams (1932). The latter author's monumental two-volume treatise contains a wealth of information but many of his views on the nature of kimberlite and the origin of the pipes have not been generally accepted (see Section 9.5).

The lack of more modern comprehensive investigations probably stems mainly from the limited access (to the older mines) which has, until relatively recently, been available to potential researchers. Research into kimberlites has, however, also been hampered by longheld views (now fortunately disregarded) that such research would be relatively unrewarding. This idea reflected beliefs that kimberlites are always highly altered, fine-grained, contaminated rocks with hybrid antecedents.

1.3 OBJECTIVES

The main aim of the present study is to document and compare the nature of the KIMFIK pipes and their infilling material and, based on the results obtained, interpret the modes of origin of the occurrences. Subsidiary aims within this overall objective are:

(i) To establish a detailed structural model of the KIMFIK pipes and determine to what extent this model can be generally applied.

(ii) To identify discrete intrusions within the KIMFIK pipes and establish their spatial and temporal relationships.

(iii) To establish whether relationships exist between mineralogical and textural features of the kimberlites and the relative ages of the intrusions and/or the distribution of the intrusions in the pipes.

(iv) To describe the main features of the major silicate phases and some oxide minerals in the kimberlites and evaluate their parageneses.

(v) To determine the nature and the sequences of deuteric alteration in the kimberlites.

(vi) To establish, compare and contrast the modes of formation of diatreme- and hypabyssal-facies kimberlites in the pipes and describe and interpret some intermediate textural varieties.

(vii) To indicate the variable whole rock geochemistry of intrusions within the KIMFIK pipes, compare new analyses with published results and evaluate certain geochemical trends in relation to sequences of intrusions.

(viii) To evaluate some aspects of the distribution of diamonds in the intrusions.

(ix) To establish whether certain minerals in the intrusions are xenocrysts or phenocrysts.

(x) To establish petrographic criteria whereby different textural and mineralogical classes of kimberlite can be identified and grouped. This aspect of the study forms part of on-going investigations being carried out by the author and his colleagues in the Petrological Research Unit of the Geological Department of De Beers Consolidated Mines and in the Kimberlite Section of the Anglo American Research Laboratories.

1.4 NOMENCLATURE

Several of the terms used in this thesis, particularly with reference to various mineralogical and textural features, have only recently been devised. The derivation of some of these terms has been indicated in previous publications (Clement et al., 1977; Skinner and Clement, 1979; Clement and Skinner, 1979); others are explained in Chapters 2 and 7.

1.5 GEOLOGICAL SETTING

Dawson (1970) has stated that the inland plateau of South Africa is "....the region of the most concentrated kimberlite intrusion in the world and is also the most thoroughly investigated kimberlite province." The former conclusion may largely be an artefact of the latter but there is, nevertheless, no doubt that kimberlites have been widely intruded, albeit in discrete clusters, over much of the vast interior plateau (Dawson, 1970, Fig. 2). The KIMFIK pipes represent three of these clusters; the Kimberley and much smaller Koffiefontein and Finsch groups.

Following Clifford (1966), Dawson (op. cit.) has also noted that kimberlites tend to be strongly concentrated in cratonic areas and that economically diamondiferous kimberlites occur within old cratons, i.e. areas that have been stable for at least 1500 m.y. The South African situation is consistent with these views. The KIMFIK pipes are located within the Kaapvaal Craton and the Archaean Basement of the area has been dated at $\sim$2840 m.y. (recalculated from Barret and Berg's 1975 results using $\lambda Rb = 1{,}42 \times 10^{-11} \times yr.^{-1}$, see Clement et al., 1979). The Kimberley kimberlites are Cretaceous in age and have been dated at 84 $\pm$3 m.y. by the Rb-Sr method (using $\lambda Rb = 1{,}42 \times 10^{-11} \times yr.^{-1}$; see Clement et al., 1979) and at $\sim$90 m.y. by U-Pb dating of zircons (Davis, 1977). The Finsch pipe has recently been dated by Rb-Sr at 120 m.y. (C. Smith, personal communication).

In the Kimberley area the upper 90-120m of the geological succession consists of horizontally bedded, black and grey, carbonaceous shales with minor interbedded sandstone and siltstone layers. At the base of these Dwyka Formation (Karoo Sequence) rocks a tillite horizon (Dwyka tillite) is erratically developed and seldom exceeds 3m in thickness. The Dwyka shales have been extensively intruded by generally concordant dolerite sills which reach 30m in thickness. Locally these sills have acted as barriers to kimberlite intrusion with the resultant formation of kimberlite sills at or near the base of the dolerites (Hawthorne, 1968).

The Karoo rocks are underlain by a variable thickness of Ventersdorp Supergroup strata. These strata consist mainly of amygdaloidal and porphyritic lavas (andesitic to basaltic in character) with intercalated quartzite horizons. Quartzite is particularly prominent in the vicinity of the De Beers and Kimberley pipes and at the latter locality exceeds 250m in thickness. A conglomerate is locally present at the base of the Ventersdorp column and has been exposed in the lower levels of most of the Kimberley mines.

The depth from surface to the unconformable contact between the Ventersdorp Supergroup and the underlying Basement Complex is highly variable. It ranges from about 350m near the Dutoitspan pipe to in excess of 1000m on the south side of the Wesselton pipe. These differences reflect the highly undulating nature of the pre-Ventersdorp topography. In the No. 1 shaft at De Beers mine the Ventersdorp/Basement contact lies at a depth of 650m (Fig. 1.2).

The Archaean Basement rocks consist mainly of granitic and amphibolitic gneisses with locally extensive development of talc schist. Numerous pegmatite veins cut these rocks.

As noted previously the only Karoo sediments preserved in the area belong to the Dwyka Formation. However, a wide variety of younger members of the Sequence (including Drakensberg Group lavas) have been preserved as xenoliths in the kimberlite pipes (Hawthorne, 1975). The presence of these xenoliths indicates that extensive erosion of the Karoo cover has post-dated and not preceded the intrusion of the Cretaceous kimberlites. Hawthorne (1975) estimated post-kimberlite erosion in the Kimberley area to total between 900 and 1900m.

At Koffiefontein the preserved sequence of Karoo stata is considerably thicker than in Kimberley. Approximately 300m of shales with concordant and discordant dolerite sills which reach 20-30m in thickness are present. The shales probably represent both the Dwyka and Ecca Formations. The Karoo rocks rest directly on Archaean granites and gneisses.

The geological setting of the Finsch pipe is entirely different from the other KIMFIK occurrences. The upper 130-160m of the pipe traverses more-or-less flat-dipping (locally contorted) banded ironstones which form part of the Asbesheuwels subgroup of the Griqualand West Sequence. Approximately 30m of finely interbedded, extremely fine-grained siliceous (cherty), dolomitic and ferruginous layers form passage beds between the banded ironstones and underlying dolomite which almost certainly persists downwards beyond the limits of the pipe.

CHAPTER 2

THE NATURE AND CLASSIFICATION OF KIMBERLITE

2.1 INTRODUCTION

The term kimberlite was originally introduced into petrological literature by Lewis (1887) to describe the host rock of diamond at the type locality, Kimberley, South Africa.

Following the early South African discoveries and the subsequent location of kimberlites in many other areas numerous attempts have been made to establish sound petrological criteria for the recognition and classification of kimberlites (e.g. du Toit, 1906; Wagner, 1914; Williams, 1932; Dawson, 1960, 1962, 1967a, 1967b, 1971; Kovalsky, 1963; Edwards and Howkins, 1966; Kennedy and Nordlie, 1968; Mitchell, 1970, 1979a). It is however, apparent from the literature that there is considerable lack of agreement regarding the nature of the kimberlite. This has led to some unfortunate consequences:

(i) Inaccurate, incomplete and/or misleading definitions of kimberlite (see discussion by Mitchell, 1970).

(ii) Terminological confusion arising from the introduction of poorly defined and imprecise terms such as 'central complex kimberlites' (Dawson, 1967a), 'kimberlitic' (Frantsesson, 1970), 'kimberlitic-carbonatitic' (Gittins et al., 1975) and carbonatite-kimberlite (Zhabin, 1967).

(iii) Unwarranted implications with respect to possible petrogenetic associations between kimberlite and several other rock types. Such inferred associations have been severely criticised by Mitchell (1970, 1979a).

(iv) An extension of 'kimberlite' to include many rocks which should be categorized within the framework of other petrological definitions and classifications. For example, there is a tendency to classify many olivine-phlogopite-carbonate ± clinopyroxene ± feldspathoid ± spinel assemblages as kimberlites rather than placing them in the lamprophyre group where, according to their mineralogy, fabric and rock associations, they clearly belong (Mitchell, 1979a).

(v) Failure to recognize the full range of petrographic variation exhibited by kimberlites sensu stricto (Clement et al., 1975; Skinner and Clement, 1977, 1979; Clement et al., 1977; Clement and Skinner, 1979).

(vi) Overemphasis of the role played by certain minerals (e.g. magnesian garnet), which invariably only occur in accessory amounts and are almost certainly xenocrysts, in the identification of kimberlites (Chapter 6).

In view of these problems some general comment on the petrological status and classification of kimberlite seems worthwhile prior to the detailed descriptions and interpretation of the KIMFIK pipes which follow.

In addition to noting the views of earlier authors much of this chapter is devoted to the justification of a revised definition of kimberlite and to descriptions of new classifications within which petrographic variants of kimberlite are grouped. This revised definition and the classifications are partly based on the investigations carried out for this dissertation but also reflect detailed examination of other kimberlites by the author and co-workers. The classifications were devised jointly by the author and E.M.W. Skinner. The revised definition is the joint responsibility of the author, E.M.W. Skinner and B.H. Smith. The revised definition and various aspects of the classification systems have been published previously (Skinner and Clement, 1977, 1979; Clement et al., 1977; Clement and Skinner, 1979). A detailed discussion and justification of the revised definition is in preparation (Clement, Skinner and Smith) but, because of its relevance to this thesis, much of the material in an unpublished draft of this paper is included here.

2.2 THE NATURE OF KIMBERLITE

2.2.1 Previous definitions

Most early workers (e.g. Lewis, 1887; Wagner, 1914) regarded kimberlite as a variety of peridotite on the basis of its ultrabasic character; reflected mineralogically by the presence of abundant olivine and other mafic minerals. The occurrence of abundant phlogopite in some instances led to kimberlite being classed as a mica peridotite and Mitchell (1970) concluded that in North America the two terms had become virtually synonomous. In contrast to most other early researchers du Toit (1906) interpreted kimberlite as a variety of limburgite.

In the last two decades a number of definitions of kimberlite have been published and a representative cross-section of these definitions is given below:

(i) AGI Glossary of Geology. Gary et al, editors (1977)

"A porphyritic alkalic peridotite containing abundant phenocrysts of olivine (commonly serpentinized or carbonatized) and phlogopite (commonly chloritized) and possible geikielite and pyrope in a fine-grained groundmass of calcite and second-generation olivine and phlogopite and with accessory ilmenite, serpentine, chlorite, magnetite and perovskite."

(ii) Milashev (1963)

"A group of ultrabasic porphyritic rocks, frequently enriched in aluminium and alkalies containing phenocrysts of olivine, phlogopite and ilmenite and accessory grains of pyrope, chrome diopside, and diamond, set in a groundmass containing microlites of clinopyroxene and sometimes monticellite and melilite but which is occasionally holohyaline."

(iii) Kovalsky (1963)

"A massive ultrabasic rock with an alkaline trend and effusive aspect, characterized by a porphyritic texture and the presence of phenocrysts of olivine, phlogopite, ilmenite and pyrope set in a fine-grained or glassy matrix." Other minerals regarded by Kovalsky as primary constituents of kimberlite are diopside, perovskite, apatite and diamond.

(iv) Sarsadskikh et al. (1963) (Quoted by Frantsesson, 1970)

"A rock consisting of a varying amount of fragmental material, which must include representatives of deep-seated rocks of ultrabasic composition, associated with zones of ultra-high pressures, and also their minerals and magmatic matrix, being the heteromorphic equivalent of these ultrabasic rocks."

(v) Seminar on kimberlite terminology and classification, Moscow (1969)(Quoted by Frantsesson, 1970)

At this seminar Soviet geologists proposed that kimberlite should be defined as "...an alkaline - ultrabasic rock with porphyritic texture with dominant olivine, and having variable amounts of phlogopite, ilmenite, pyrope, pyroxene, apatite, perovskite and some other minerals. Kimberlite is a parent medium of diamond."

(vi) Dawson (1962, 1967a)

"A serpentinized and carbonated mica-peridotite of porphyritic texture, containing nodules of ultrabasic rock-types characterized by such high-pressure minerals as pyrope and jadeitic diopside, it may, or may not, contain diamond. It occurs in diatremes, dykes, veins and sills of very limited size."

(vii) Edwards and Howkins (1966)

"An intrusive, undersaturated, feldspar-deficient, igneous rock, usually observed in an altered and brecciated state. It contains essential magnesian ilmenite and chrome-rich pyralspite garnet with one or some of the following accessory minerals, chromiferous diopside, phlogopite, zircon, perovskite, diamond, magnetite, celestite, baryte and calcite. The essential olivine of the primary rock is commonly altered to serpentine."

(viii) Kennedy and Nordlie (1968)

"All rocks from diamond pipes and all diamond pipes are of similar character. Kimberlite is finely ground, broken up, and altered rock from the deep mantle of the earth diluted with some shallow crustal material. With it are associated angular fragments and rounded boulders of all formations traversed by the kimberlite on its way to the surface of the earth. The single character to be emphasized is that diamond pipes are filled with fragmented debris..."

(ix) Mitchell (1970)

"Kimberlite is a porphyritic, alkalic peridotite, containing rounded and corroded phenocrysts of olivine (serpentinized, carbonatized or fresh), phlogopite (fresh or chloritized), magnesian ilmenite, pyrope and chrome-rich pyrope set in a fine-grained groundmass composed of second generation olivine and phlogopite together with calcite (and/or dolomite), magnetite, perovskite and apatite."

(x) Dawson (1971)

"Kimberlite is a very rare, potassic, ultrabasic, hybrid, igneous rock that occurs in small diatremes or in dykes or in sills of limited extent. It has an inequigranular texture, the porphyritic aspect being due to megacrysts of olivine, enstatite, chrome diopside, pyrope, picroilmenite and phlogopite set in a finer matrix of which serpentine, carbonates, phlogopite, magnetite and perovskite form the major part. Many of the megacrysts are derived from fragmentation of mantle-derived garnet lherzolite (blocks of which are embedded in the kimberlite) and are in various stages of reaction with the kimberlite matrix. The matrix may or may not contain diamond; even in the most diamondiferous kimberlite diamond is a very rare and widely dispersed mineral."

(xi) Nixon (1973)

Referring specifically to Lesotho kimberlites Nixon defined these rocks as "...hydrated magnesium silicate rocks, with or without residual olivine, having variable amounts of perovskite, Fe-Mg-Ti spinel, calcite and phlogopite, and deep-seated inclusions or their disaggregated minerals comprising, picroilmenite, pyrope, Mg-rich pyroxenes and olivine. Textures range from basaltic (Wagner, 1914) to micaceous and a wide range of volcanic structures are represented."

(xii) Mitchell (1979a)

Kimberlites are "...inequigranular alkalic peridotites containing rounded and corroded megacrysts of olivine, phlogopite, magnesian ilmenite and pyrope set in a fine-grained groundmass of second generation euhedral olivine and phlogopite together with primary and secondary (after olivine) serpentine, perovskite, carbonate (calcite and/or dolomite) and spinels. The spinels range in composition from titaniferous-magnesian-aluminous chromite to magnesian ulvöspinel-ulvöspinel-magnetite. Accessory minerals include diopside, monticellite, apatite, rutile and nickeliferous sulphides. Some kimberlites can contain major modal amounts of diopside or monticellite."

The foregoing definitions highlight the existing disparity of views regarding the nature of kimberlite. Despite some points of general agreement (e.g. the presence of abundant olivine and the ultrabasic character of the rock) there is a remarkable diversity of opinion regarding the fundamental characteristics of kimberlite.

2.2.2 A revised definition of kimberlite

The revised definition given here stems partly from a review of relevant published literature but is based mainly on detailed investigations of kimberlites and petrographically similar rocks from several hundred occurrences. Most of the occurrences are located in Africa (south of the equator) but specimens from North America, South America, Russia, Greenland and Australia have also been examined (Subsequent work on kimberlite specimens from India has shown that these rocks are also consistent with the revised definition; personal communication, E. Colgan).

This definition was originally published, in slightly different form, in the abstracts volume of the Second International Kimberlite Conference held in Santa Fé, U.S.A. (Clement et al., 1977). In general format this definition broadly follows that proposed by Dawson (1971) but it differs from the latter in several important respects. The definition is given below:

> "Kimberlite is a volatile-rich, potassic, ultrabasic igneous rock which contains abundant olivine and generally has a distinctively inequigranular texture resulting from the presence of macrocrysts set in a fine-grained matrix. This matrix contains as prominent primary phenocrystal and/or groundmass constituents, olivine and several of the following minerals: phlogopite, calcite, serpentine, diopside, monticellite, apatite, spinels, perovskite and ilmenite. Other primary minerals may be present in accessory amounts. The macrocrysts belong almost exclusively to a suite of anhedral, cryptogenic, ferromagnesian minerals which include olivine, phlogopite, picroilmenite, magnesian garnet, chromian diopside and enstatite. Olivine is extremely abundant relative to the other macrocrysts all of which are not necessarily present. In addition to macrocrysts smaller grains belonging to the same suite of minerals occur. Kimberlite

may contain diamond but only as a very rare constituent. Kimberlite commonly contains inclusions of mantle-derived ultramafic rocks. Variable quantities of crustal xenoliths and xenocrysts may also be present. Kimberlite is often altered by deuteric processes mainly involving serpentinization and carbonatization."

2.2.3 Discussion

In spite of overwhelming evidence for the igneous nature of kimberlite (e.g. Wagner, 1914; Williams, 1932; Dawson, 1960, 1971; Mitchell, 1970; Dawson and Hawthorne, 1970, 1973; Clement, 1973, 1975; Emeleus and Andrews, 1975; Scott, 1979; Skinner and Clement, 1979) the existence of kimberlite magma is denied in two of the definitions quoted in section 2.2.1 (Sarsadskikh et al., 1963; Kennedy and Nordlie, 1968). Kimberlite is interpreted by these authors as a heterogeneous assemblage of rock and mineral fragments derived dominantly from the upper mantle. This assertion is based on the occurrence, in many kimberlite pipes, of textural varieties of kimberlite which are classed as tuffisitic kimberlite breccias in this thesis (section 2.3.5). These tuffisitic kimberlite breccias do, however, contain abundant pelletal lapilli and autoliths (section 7.2.2) which can unambiguously be interpreted as having crystallized from a kimberlite magma (e.g. Plates 7.2-7.8). Furthermore, in most of these rocks much of the interstitial cement, which binds pelletal lapilli, kimberlite fragments (autoliths), discrete kimberlite mineral grains and xenogenic components together, is of direct magmatic origin; although the minerals involved have crystallized from highly evolved residual magma fractions (section 7.2.3).

Kimberlites are characterized by a group of anhedral ferromagnesian minerals which occur as relatively large insets in a fine-grained (aphanitic) matrix. These insets impose an inequigranular texture on the rocks (e.g. Plates 6.1-6.4 and 7.16a). Olivine is the dominant member of the group which also includes phlogopite, picroilmenite, magnesian garnet, chrome diopside, enstatite and several different types of spinel (Mitchell, 1979a; Pasteris, 1980a). Typically these insets occur as rounded, often irregular, corroded grains (Plates 6.1-6.4) or as angular, broken fragments (Plates 7.9a-b) and many show some degree of reaction with the magma (Dawson, 1971).

The presence of these insets has resulted in kimberlites being termed porphyritic rocks in most published definitions. This designation adequately describes the macroscopic appearance of the rocks but implies that all, or at least many, of the insets are phenocrysts. Such an origin is favoured by some authors, particularly with respect to olivine (e.g. Mitchell, 1973a; Scott, 1977). There is, however, a wide body of opinion in favour of all, or most, of these insets being xenocrysts; derived from the same, or similar, upper mantle rocks that are the sources of the ultramafic nodules that commonly occur as rounded xenoliths in kimberlite (e.g. Dawson, 1971; Lawless, 1974, 1978; Gurney, 1974; Dawson and Stephens, 1975; Stephens and Dawson, 1977; Boyd and Clement, 1977). To avoid this genetic controversy these cryptogenic insets are referred to as macrocrysts in the revised definition (section 2.2.2). Small (<0,5mm) grains belonging to this group of minerals also occur in most kimberlites. It is, however, the abundance of insets (particularly of olivine or its pseudomorphic replacement products) which are macroscopically visible and are significantly larger than the surrounding matrix that imposes the characteristic inequigranular texture of kimberlites and justifies use of the term 'macrocryst'.

In several earlier kimberlite definitions a number of minerals belonging to the macrocryst assemblage (e.g. olivine, chromian garnet and picroilmenite) are regarded as essential minerals (Kovalsky, 1963;

Edwards and Howkins, 1966; Mitchell, 1970, 1979a). However, with the exception of olivine the macrocryst minerals very rarely occur in more than accessory amounts and the complete macrocryst assemblage is not always present. For example, no macrocrysts of picroilmenite are present in the Finsch, Koffiefontein and Letseng pipes, all of which have been extensively mined for diamonds.

Olivine macrocrysts are generally abundant and their presence, together with a second population of smaller euhedral or subhedral olivine crystals (Plates 5.1a-b and 6.1a-d) is a diagnostic feature of kimberlites. In rare instances olivine (and other) macrocrysts may, however, be entirely, or locally absent. This is often due to near-surface flowage differentiation and, possibly, to other crystal fractionation processes at depth (Chapter 7).

The matrix referred to in the revised definition is equivalent to kimberlite minus the macrocryst assemblage and any xenolithic material present. It has recently been demonstrated (Skinner and Clement, 1979) that the matrices of kimberlites exhibit a wider range of mineralogical variation than previously supposed. In addition to the second population of olivine referred to above, any one of five other minerals may occur as the most abundant mineral in the matrix. These five minerals are phlogopite, calcite, serpentine, diopside and monticellite (Skinner and Clement, 1979).

Several other primary matrix minerals have been recorded (e.g. Skinner and Scott, 1979; Dawson, 1971, 1980). Most of these occur only in accessory amounts. Attention has, however, been drawn to four of these minerals in the revised definition because of their almost ubiquitous occurrence in kimberlites or because they exhibit chemical compositions, or compositional trends, that are characteristic of kimberlites. These minerals are apatite, perovskite, ilmenite and a variety of spinels. The spinels are of interest because they display compositional trends which are unique to kimberlites (Haggerty, 1975; Mitchell, 1979a; Pasteris, 1980a). Matrix ilmenites are unusual because they frequently contain even more Mg than picroilmenite macrocrysts (S. Shee has analysed matrix ilmenites which approach 20 wt.% MgO, personal communication).

Some previous definitions (e.g. Milashev, 1963) indicate that melilite is a characteristic mineral of kimberlites. The occurence of melilite has, however, been seriously questioned by Mitchell (1970). The melilite problem has not been satisfactorily resolved. No melilite has been positively identified by the author or his colleagues in any of the kimberlites which they have examined. However, numerous probable melilite pseudomorphs (composed of serpentine and clay) are present in certain Koffiefontein and Ebenhaezer kimberlites (section 5.11) and E.M. Skinner (personal communication) has found similar pseudomorphs in several kimberlites outside the KIMFIK group.

Although the majority of kimberlites appear to be devoid of diamond, the occasional presence of this mineral is noted in the revised definition for historical and economic reasons and because of its genetic implications regarding the depth of origin of kimberlite magmas. The origin of diamonds in kimberlite is enigmatic. They have been interpreted mainly as xenocrysts (Robinson, 1978, 1980) or phenocrysts (e.g. Gurney et al., 1979a). In terms of identifying or defining kimberlite, diamonds are of little relevance because of their rarity.

A noticeable feature of the mineralogy of kimberlites is the common occurrence of two populations of several minerals, i.e. one population belonging to the macrocryst assemblage and a second which forms part of the kimberlite matrix. Olivine is the outstanding example (Plates 5.1a-b) but phlogopite, picroilmenite and spinel also occur in both modes. Some kimberlites are also characterized by the occurrence of two generations of certain matrix minerals, e.g. diopside and phlogopite (Plates 5.6b and 7.13c-e).

Three aspects of the chemistry of kimberlites are stressed in the revised definition; namely that kimberlites are ultrabasic, potassic and volatile-rich rocks.

The ultrabasic nature of kimberlite is well established (e.g. Dawson, 1967b) and is indicated in most published petrological definitions of kimberlite. The SiO_2 content of kimberlites is generally well below 40 wt.% although certain diopside and phlogopite-rich varieties may approach or, in rare instances, slightly exceed 40 wt.% (Clement et al., 1979). Very high SiO_2 content ($\leq$50 wt.%) reflects contamination by inclusions of SiO_2-rich crustal material (section 7.8), particularly when accompanied by abnormally high levels of Al_2O_3 and Na_2O (greater than $\sim$4 and 0,5 wt.% respectively). A considerable number of published analyses and some new analyses reveal such contamination (section 7.8).

Although seldom referred to in kimberlite definitions the volatile-rich nature of those rocks is a characteristic feature (Dawson, 1967b; Mitchell, 1970; Gurney, 1974). The H_2O^+ and CO_2 contents of kimberlites are frequently as much as 5 and 10 wt.% respectively (section 7.8). Higher concentrations occasionally occur in some fresh kimberlites (Table 7.7) but in other kimberlites extremely high H_2O^+ and CO_2 concentrations probably reflect substantial metasomatic alteration and/or weathering. A further indication of the volatile-rich nature of kimberlite is given by the generally high P_2O_5 level (Table 7.7) relative to other ultrabasic rocks.

Several of the definitions quoted in section 2.2.1 refer to kimberlites as alkaline or alkalic rocks. As indicated in Table 7.7 kimberlites are generally potassic but they are not enriched in soda. The main host of K in kimberlites is phlogopite which, in some instances, is extremely abundant. Consequently some kimberlites contain >4 wt.% K_2O. The high K_2O levels of many kimberlites distinguish them from most other ultrabasic rocks (although some highly potassic lamprophyres and leucite lamproites have considerably higher K_2O levels, Scott, 1977). The high K_2O/Na_2O ratios of most kimberlites are therefore a characteristic feature.

As noted previously, kimberlites that contain more than about 0,5 wt.% Na_2O are considered to have been contaminated. No primary mineral in kimberlite is a major host of Na but minor amounts can occur in phlogopite and diopside (Emeleus and Andrews, 1975; Dawson and Smith, 1977; Scott, 1979) and in perovskite (Mitchell, 1972; Scott, 1979). Some kimberlites are virtually devoid of phlogopite (Skinner and Clement, 1979) hence whole rock analyses may indicate higher Na_2O than K_2O. It is, however, unlikely that Na_2O will exceed or even approach 0,5 wt.% in such instances.

Because of the frequency with which they occur and their important geobarometric and geothermometric implications attention is drawn in the revised definition to the presence of upper mantle-derived ultramafic xenoliths in kimberlites. The origin of these inclusions and their relationships to kimberlite have been extensively discussed in the proceedings volumes of two International Kimberlite Conferences held in Cape Town, South Africa (1973) and Santa Fé, U.S.A. (1977).

Attention is also drawn in the definition to the fact that many of the early or relatively early crystallizing minerals have been subjected to extensive deuteric alteration. Deuteric alteration is another characteristic feature of kimberlites (Skinner and Clement, 1979) and the major processes involved are serpentinization and carbonatization. Many other types of deuteric alteration and replacement do, however, occur and are described in Chapters 5, 6 and 7. Deuteric alteration is unrelated to weathering or metasomatic alteration of kimberlites but among the products of deuteric alteration are a variety of hydrous minerals which are prone to rapid weathering. Consequently kimberlite is often extensively weathered and this weathering, which results in the development of "yellow ground", often persists to considerable depths below surface.

A feature of many kimberlites is the presence of scattered inclusions known as megacrysts or discrete nodules (Dawson, 1971, 1980; Nixon and Boyd, 1973a). These megacrysts include discrete single grains or monomineralic aggregates of olivine, phlogopite, garnet, clinopyroxene, orthopyroxene, ilmenite and zircon and regular to irregular intergrowths of ilmenite-clinopyroxene and ilmenite-orthopyroxene. In view of their large size (often >2cm in diameter) and distinctive mineral compositions (Dawson, 1980), megacrysts are unlikely to be xenocrysts derived from the varied ultrabasic, upper mantle rocks which commonly occur as xenoliths in kimberlites and their origin remains enigmatic. All members of the suite are very rarely present in individual intrusions and they are generally rare components of kimberlites. Consequently no reference to megacrysts is made in the revised definition of kimberlite.

2.3 CLASSIFICATION OF KIMBERLITES

2.3.1 General comments on classification

As indicated in the previous discussion (section 2.2) kimberlites exhibit extensive mineralogical and textural diversity. This has been recognized, to a greater or lesser degree, by many authors (e.g. Wagner, 1914; Dawson, 1967a, 1967b, 1971; Frantsesson, 1970; Mitchell, 1970, 1979a; Skinner and Clement, 1979; Clement and Skinner, 1979). The variation is so extreme that 'kimberlite' can be regarded as a 'sack' term; uniting a family of rocks which are linked by characteristic petrological and geochemical features.

Because of the extreme variability of kimberlite and because this variability has direct genetic implications there is considerable merit in formulating classification systems. They provide a frame-

fications have not been satisfactory for two main reasons:

(i) In some cases a kimberlite must satisfy several criteria, yet these criteria are largely or entirely independent of each other (e.g. the classifications of Rabhkin, 1962 and Artsybasheva et al., 1964).

(ii) In other cases entirely unrelated criteria are used to distinguish individual classes within a single classification system (e.g. Brobnievich et al., 1964).

were not quantitatively defined, were termed basaltic and lamprophyric (micaceous) kimberlites.

Basaltic kimberlites were defined as "...porphyritic peridotites of basaltic habit..." composed of olivine phenocrysts and occasional insets of phlogopite, ilmenite, enstatite, pyrope and diopside, set in a groundmass interpreted as consisting mainly of secondary minerals (dominantly serpentine and calcite). Wagner also noted considerable perovskite, apatite, chromite and other opaque minerals in the groundmass of basaltic kimberlites. He considered that olivine generally formed 50-70 vol.% of basaltic kimberlite and that the amount of phlogopite rarely exceeded 3-4 vol.%.

Wagner's second major mineralogical class of kimberlites, the lamprophyric varieties, were regarded as "...porphyritic mica peridotites of lamprophyric habit." They were said to contain essentially the same 'phenocryst' assemblage as basaltic kimberlites set in a groundmass dominated by phlogopite (up to 50 modal%). Other groundmass minerals recognized by Wagner included olivine, clinopyroxene, apatite, chromite, perovskite and opaque oxides. The presence or absence of clinopyroxene was used as a basis for recognizing two subtypes of lamprophyric kimberlite.

Little further development of mineralogical classifications took place until the mid-1950's although, in 1929, Wagner provided a more specific interpretation of basaltic kimberlites. On the basis of provisional identifications of monticellite in the groundmass of kimberlites from the De Beers and Kimberley pipes he suggested that basaltic kimberlites were "...porphyritic monticellite peridotites of basalt habit."

The discovery of the Siberian pipes stimulated a vast amount of research by Soviet investigators. One of the consequences of this work was the development of a series of kimberlite classifications (based mainly on mineralogical parameters) between 1954 and 1965.

In most Soviet classifications initial subdivision is based on the abundance of phlogopite, resulting in the recognition of mica-poor and mica-rich varieties equivalent to Wagner's basaltic and lamprophyric types. Further subdivision, of varying degrees of complexity, is based on a variety of other minerals.

Soviet classifications have been extensively reviewed by Frantsesson (1970) hence detailed discussion is not required here. However, as an example of the more detailed and complex classifications that proposed by Milashev (1965) is given in Table 2.1.

Since 1965 few new or modified mineralogical classifications have been forthcoming but in 1970 Mitchell published a simple, threefold classification of kimberlites in which he recognized:

(i) Kimberlite (olivine-rich kimberlite approximately equivalent to basaltic kimberlite).

(ii) Micaceous kimberlite (equivalent to lamprophyric kimberlite).

(iii) Calcareous kimberlite (kimberlite containing abundant carbonate minerals).

2.3.3 Inadequacies of published mineralogical classifications

The classifications referred to in section 2.3.2 have been criticized for two fundamental inadequacies (Skinner and Clement, 1979). They do not incorporate the complete range of mineralogical variation exhibited by kimberlites and they are essentially qualitative in character. Several reasons can be advanced to explain these shortcomings:

(i) Most classifications are based on limited kimberlite suites from restricted geographic areas.

(ii) Few detailed mineralogical studies of the groundmass of kimberlites have (until recently) been undertaken. This situation apparently reflects a longheld belief that such investigations would be unrewarding because of the extremely fine grain-size and altered nature of kimberlite groundmass minerals.

(iii) Confusion between which minerals in kimberlites are primary and which are secondary alteration products or are xenocrysts.

(iv) Limited access by most researchers to fresh material.

2.3.4 Mineralogical classification used in this thesis

The classification used in this thesis is that proposed by Skinner and Clement (1977, 1979). It was originally based on petrographic investigations of a large number of southern African kimberlites. This classification has since been applied on a routine basis, in the De Beers Petrographic Laboratory and the Anglo American Research Laboratories, to numerous additional African kimberlites and to kimberlites located in North America, South America, Australia, U.S.S.R., Greenland and India.

For details of the methodology and for justification of the classification concepts the reader is referred to the published paper (Skinner and Clement, 1979). A brief summary is, however, given below, together with some minor additional comments. The classification is also shown in Table 2.2 which should be read in conjunction with the following comments.

The classification is based on the modal proportions of certain minerals in kimberlites and it therefore provides a quantitative system of subdividing kimberlites. Only certain matrix minerals are taken into account. The macrocryst assemblage is ignored because many of these minerals are considered to be xenocrysts (Chapter 6) and most occur sporadically and only in accessory amounts. Although it is a ubiquitous and abundant constituent matrix (phenocrystal) olivine is also ignored. There are several reasons for this:

(i) Olivine is frequently severely altered; to the extent that the grains as a whole, or their boundaries, may be extremely difficult to recognize.

(ii) It is not always possible to distinguish fully between the two populations of olivine which are generally present in kimberlites. One of these populations is considered to consist almost exclusively of xenocrysts (e.g. Boyd and Clement, 1977; this thesis, Chapter 6) and the other of early formed phenocrysts (section 7.7.1).

(iii) It has been found (Skinner and Clement, 1979) that the abundance of olivine in kimberlites is largely independent of the relative abundances and types of other major components.

The studies on which this classification is based, revealed that, in addition to olivine, any one of five minerals may be present as the major primary matrix mineral in a kimberlite. These minerals are diopside, monticellite, phlogopite, calcite and serpentine. Accordingly, in the 1st stage of classification, five basic subdivisions of kimberlite are recognized depending upon (and named after) whichever one of these five minerals is volumetrically most abundant (Table 2.2). Further subdivision (2nd tier) can be made if one or more of these five minerals, or any other mineral, is present in sufficient abundance for it to qualify as a characterizing accessory. Minerals are accepted as characterizing accessories if they are present to the extent of, or exceed, two-thirds of the volumetric abundance of the dominant mineral. In addition to the five minerals listed only apatite has to date been found in sufficient abundance to qualify as a characterizing accessory. In some instances, however, the total opaque mineral content of the matrix may reach, or exceed, two-thirds of the abundance of the dominant mineral. In such cases the kimberlite is qualified as "opaque mineral-rich" (Table 2.2).

It is possible that some other mineral may be found in such abundance in kimberlite as to require the classification shown in Table 2.2 to be extended. Melilite (if its presence in kimberlite can be proven) or some other feldspathoid, are possible candidates. In addition, Skinner and Scott (1979) have recorded noteworthy amounts of an amphibole in a kimberlite from the Swartruggens group of dykes. E.M.W. Skinner (personal communication) has optically identified an extremely fine-grained ($\sim$0,02mm) groundmass generation (as opposed to phenocrysts) of olivine in one of the Newlands (Barkly West) occurrences. The abundance of this olivine is such that analytical confirmation of its presence would require the classification to be extended.

Examination of Table 2.2 may lead to the impression that the various mineralogical varieties of kimberlite are more or less equally abundant. This is certainly not the case. For example, phlogopite kimberlites appear to be much more common than serpentine kimberlites. Also, diopside and phlogopite are often both abundant minerals in individual kimberlites and there are numerous examples of phlogopite-diopside or diopside-phlogopite kimberlites (depending on the relative abundances; in the first case diopside exceeds phlogopite, vice versa in the second case). In contrast, although diopside and monticellite do occasionally occur in the same kimberlite, one mineral is always much more abundant than the other.

This classification has been criticised by Dawson (1980) for two reasons:

(i) That the mineralogical character of kimberlites (in terms of the five 'key' minerals) will often be set by emplacement conditions and the partial pressures of volatiles rather than being a reflection of intrinsically different types of kimberlite.

(ii) That the groundmass mineralogy often reflects extensive alteration and is not pristine.

The point made under (i) above is entirely accepted but is not seen as a flaw in the classification system. Indeed, the linking of groundmass mineralogy to varying emplacement conditions indicates that the classification has a genetic basis, even if this is not always directly related to differences in the original kimberlite melts.

Provision is made in the classification for alteration effects. In the original paper stress was laid on the fact that 'monticellite', 'diopside', 'phlogopite', 'calcite', or 'serpentine' kimberlites are recognized when the primary nature of these minerals is established. When severe alteration precludes such identification it is recommended that kimberlites be classified in terms of the dominant alteration processes. Thus 'carbonatized' or 'serpentinized' kimberlites may be recognized instead of 'calcite' or 'serpentine' varieties.

2.3.5 Textural-genetic classification of kimberlites

The wide textural variations exhibited by kimberlites have long been recognized and were discussed in considerable detail by Wagner (1914). The greatest range of textures occurs within kimberlite pipes as a result of complex, near-surface, pipe-forming and infilling processes (Chapter 10). Because of the relationship between texture and the evolution of kimberlite pipes there is merit in attempting to formulate logical classification systems which differentiate between varieties of kimberlite on a textural-genetic basis.

The classification adopted in this thesis is a modified version of that proposed by Clement and Skinner (1979). It is illustrated in Table 2.3.

In terms of this classification kimberlite rocks are initially subdivided into three broad groups: hypabyssal-facies, diatreme-facies and crater-facies kimberlites. These three groups are largely (but not entirely) confined to corresponding vertical zones of kimberlite pipes, namely, root zones, diatreme zones and crater zones (Chapter 3). This initial subdivision corresponds closely with the views of other authors (Dawson, 1971; Hawthorne, 1975) on the distribution of different varieties of kimberlite in kimberlite pipes. Although some overlap occurs in complex pipes (where multiple intrusion has taken place) this initial subdivision provides a useful indication of the general vertical zonation of different varieties of kimberlite in kimberlite pipes.

Two main subdivisions of hypabyssal-facies kimberlite are recognized. They are 'kimberlite' and 'kimberlite breccia' (Table 2.3). The term 'kimberlite' is used here in a strictly petrological rather than general geological sense to reflect kimberlite as defined in section 2.2.2. It is the 'normal' crystallization product of kimberlite magma and at no stage has vapour-solid fluidization (Chapters 7, 9 and 10) been involved in its development. A number of other terms have been used to describe hypabyssal-facies kimberlite. These terms include massive kimberlite, porphyritic kimberlite, clastic-porphyritic kimberlite and intrusive kimberlite (Frantsesson, 1970). At this level of classification these terms are considered superfluous and they may be misleading. For example, use of the adjective 'intrusive' implies that no other varieties of kimberlite were intruded into position.

The second major subdivision of hypabyssal-facies varieties is kimberlite breccia (e.g. Plates 7.1e-h). Kimberlite and kimberlite breccias are distinguished from one another only on the basis of the abundance of relatively large (+4mm) country rock xenoliths and/or fragments of earlier crystallized kimberlite (autoliths) that are present. Thus there is a complete gradation between the two rock types. Kimberlite breccias merely reflect increasing incorporation of rock fragments (accidental or cognate) by kimberlite. The dividing line has been set at 15 vol.% of +4mm fragments, an arbitrary but practical boundary.

At the next level of classification hypabyssal-facies kimberlites can be separated macroscopically into those that are aphanitic and those with a porphyritic aspect (Table 2.3 and Plates 3.3h, 5.10d and 7.16a).

Aphanitic kimberlite is relatively rare. Such kimberlites are devoid of the coarse-grained ferromagnesian macrocryst mineral assemblages which are usually a prominent feature of kimberlites. Possible reasons for the absence of macrocrysts in these rocks are noted in section 2.2.3. The more characteristic macroporphyritic kimberlites contain abundant macrocrysts (e.g. Plate 7.16a), mainly of olivine, which commonly exceed 1,0mm and may (rarely) exceed 1,0cm in size. These macrocrysts are set in an aphanitic matrix which corresponds to aphanitic kimberlite. Many, or most, macrocrysts may be xenocrysts (Chapter 6) hence, as discussed in section 2.2.3, the term 'macroporphyritic' may not be entirely correct. It implies that the macrocrysts are true phenocrysts. However, no other suitably descriptive word appears to be available. A transition exists between aphanitic kimberlite, through kimberlite containing few, scattered, small macrocrysts, to kimberlites with a prominent porphyritic texture.

Kimberlites sensu stricto (and the corresponding matrix of kimberlite breccias) can be further divided according to differences in the distribution and mutual relationships of their components. Two main, gradationally related, classes are distinguished (Table 2.3):

(i) Kimberlites in which the minerals are uniformly distributed throughout the rocks.

(ii) Kimberlites in which prominent segregation, particularly of the groundmass constituents, is evident (e.g. Plates 5.5f-h). Such segregation is, in some cases, developed to the extent that emulsion-like or globular textures are formed (e.g. Plates 5.10e-f). Several different forms of segregationary textures can occur. They are discussed and illustrated in Chapter 7.

In addition to the textures noted above a variety of subtextures are evident under the microscope. These textures, which are also described in Chapter 7, generally reflect relationships between specific minerals and may only be developed on a local scale. They are useful aids to petrographic description and are important in elucidating the paragenetic evolution of kimberlites. To avoid further elaboration they are not, however, incorporated in this classification.

Kimberlite breccias have also been described as 'intrusive kimberlite breccias' (Dawson, 1967a), 'eruptive kimberlite breccias' (Frantsesson, 1970) and 'brecciaform kimberlites' (Artsybasheva, et al., 1963). Some of these terms are misleading and etymologically unsound. In particular the term 'intrusive kimberlite breccia' should not be used to describe a specific type of breccia in kimberlite pipes. 'Intrusive breccia' is a term which implies only that the breccia in question displays cross-cutting relationships (Wright and Bowes, 1963). Other types of breccias in kimberlite pipes (e.g. tuffisitic kimberlite breccias) also display such relationships. In contrast some of the kimberlite breccias (as defined in this thesis) do not display cross-cutting (intrusive) relationships. For example, some can be described as 'kimberlite intrusion breccias'. They occur where kimberlite magma has intruded along discontinuities in the country rocks, disrupted the latter and then crystallized around and between the resulting blocks and fragments. Displacement of the breccia blocks is limited and they have not in effect been transported by intruding magma although brecciation is the direct result of magmatic intrusion. The distinction between 'intrusive' breccias and 'intrusion' breccias has been discussed in detail by Wright and Bowes (1963).

For similar reasons kimberlite breccias should not be termed 'eruptive' breccias; not all kimberlite breccias are 'eruptive'.

Also there is little justification in terming kimberlite breccias 'brecciaform kimberlites' thereby implying that they look like breccias when, in fact, they are breccias, albeit of a specific (igneous) kind.

Two major, gradationally related, rock types are also recognized among diatreme-facies kimberlites. They are termed tuffisitic kimberlite (TK) and tuffisitic kimberlite breccias (TKB's). As in the case of the hypabyssal-facies rocks they are arbitrarily separated according to the abundance of accidental and/or cognate inclusions which measure >4mm across. A cut-off limit of 15 vol.% is again applied.

The rock fragments and discrete mineral grains (of diverse origins) within these rocks are often cemented by extremely fine-grained minerals which are interpreted as quench products derived from the condensates of the vapour phases of short-lived, vapour-liquid-solid fluidized systems (Chapters 7 and 10). The solid material transported and mixed by fluidization is thus in many cases cemented by the crystallization products of a primary magma - albeit a highly evolved, probably contaminated, residual liquid (Chapter 7). In view of this cement, these diatreme-facies rocks might not be interpreted as true tuffisites (or tuffisite-breccias) which by definition (Gary et al., 1977) are entirely clastic (fragmental) rocks. However, many tuffisites (intrusive tuffs) are emplaced by gas-solid fluidization (Reynolds, 1954) and it is likely that in many such cases similar interfragmental, volatile-derived, quench products occur. It is therefore concluded that the existing definition of tuffisite is too restrictive.

TK's and TKB's can, in a further stage of subdivision, be divided into four classes according to prominent textural features. These textural varieties, which display gradational relationships, are termed pelletal, segregationary, uniform and crystallinoclastic TK's or TKB's (Table 2.3). In view of the gradational relationships no class limits are set and the terms are intended primarily as descriptive aids to indicate prominent chracteristics of the relevant types of intrusions (Plates 7.1-7.9). As indicated in Chapters 7 and 10 and to some extent below these subdivisions do, however, have certain genetic implications.

Pelletal-textured TK's and (more commonly) TKB's contain numerous pelletal lapilli which are ascribed to crystallization prior to and/or during explosively induced fluidization. Segregationary-textured TKB's reflect considerable post-fluidization crystallization of minerals from vapour condensates in patchy or segregated fashion. Uniformly-textured TKB's reflect similar crystallization in much less patchy or segregated, essentially interstitial, fashion. Note that deuteric alteration, contemporaneous with vapour condensate crystallization, may extensively modify pre-existing pelletal lapilli, hence segregationary textures may, to a greater or lesser degree, represent modified pelletal textures. Crystallinoclastic TK's or TKB's are those in which the amount of the condensate-derived interstitial material is relatively low (in rare cases it appears to be almost entirely absent) and the rocks most closely approach clastic assemblages. The various classes of TK's and TKB's are described in more detail and their modes of origin are evaluated in Chapter 7.

Crater-facies kimberlites are also divided into two groups; epiclastic and pyroclastic (volcaniclastic) kimberlites. Epiclastic kimberlite (Plates 7.15g-h) is derived from the erosion of ejected pyroclastic material (in tuff cones) and subsequent redepositon of the reworked, transported, material within the craters of kimberlite pipes, or in nearby topographic depressions. Pyroclastic material may accumulate and be preserved as back-fall deposits of ejectamenta within kimberlite craters.

Both epiclastic and pyroclastic kimberlite can be classified according to standard grade scales (e.g. kimberlite shale, sandstone, grit or conglomerate in the case of epiclastic kimberlite and kimberlite tuff, tuff-breccia or agglomerate in the case of pyroclastic kimberlite). Such classification is often best applied only on small scales (e.g. on a hand specimen basis). On a megascopic scale complex lithologies are often evident which are best described and interpreted in terms of lithological-genetic facies concepts, e.g. alluvial fan facies, fluviatile facies, talus deposits, etc.

CHAPTER 3

MAJOR GEOLOGICAL FEATURES OF THE PIPES

3.1 INTRODUCTION

In this chapter the megascopic geological features of the KIMFIK pipes are described. These descriptions mainly reflect information obtained from detailed geological mapping of the occurrences and use has been made of a number of illustrations (some of which are partly diagrammatic) to clarify specific features. Reference is also made to a number of other kimberlite pipes which provide good examples of some of the features evident in the KIMFIK occurrences.

Geological mapping has been facilitated by exposures resulting from underground and/or opencast mining operations at the six pipes. Some information relating to parts of the pipes which have been mined out, or are no longer accessible, has been obtained from old records in the Geological and Survey departments of De Beers Consolidated Mines Ltd.

Description of the major geological features has been arranged so as to relate to (and justify) the development of a generalized structural model of a kimberlite pipe. This model provides a basis for evaluating proposed mechanisms of pipe formation (Chapter 10). The structural model is broadly similar to that put forward by Hawthorne (1975) but differs from the latter in several respects and involves a more detailed assessment of the features which characterize specific parts of kimberlite pipes.

In the generalized model three successive, transitionally linked but otherwise broadly distinct, depth zones are recognized. They are referred to as crater, diatreme and root zones, representing respectively the upper, central and lower parts of kimberlite pipes.

Due to post-kimberlite erosion only the lower parts of the diatreme zones and the root zones of the KIMFIK pipes are preserved. In the descriptive sections which follow particular attention is given to geological features which substantiate the recognition and delineation of these two zones and clarify the differences between them.

Evidence is presented which indicates that crater zones must have been present at the top of the KIMFIK pipes prior to erosion. However, the nature of these zones has had to be inferred by analogy with pipes in other areas which have undergone little or no post-emplacement erosion (e.g. the Orapa pipe in Botswana and the Mwadui pipe in Tanzania).

3.2 SIZE AND SHAPE

3.2.1 Surface areas

The surface area of the Finsch pipe (17,9 ha) was accurately established by trigonometrical surveying during the early stages of the development of the mine. The surface areas of the other KIMFIK pipes are less accurately known and the author has been unable to locate any plans which reliably depict the surface outlines of these pipes. Wagner (1914) listed the areas of these pipes in terms of mining claims (30 x 30 feet) and his figures, converted to hectares, are given in Table 3.1 (Column B). These figures are, however, of dubious reliability as the areas defined by registered mining claims will not have corresponded exactly with the surface areas of the pipes and may include some "break-back" of the Karoo shale country rocks after the initial stages of mining.

The areas listed in Column A of Table 3.1 are based on upward projection to surface of the margins of the pipe from accurately demarcated positions on underground mining levels. 1:1000 Scale mine plans were used for this purpose.

The areas obtained by projection are considered reasonably reliable because:

(a) The projected distances are short relative to the known lengths of the pipe columns.

(b) The upper parts of the pipes have regular shapes (see section 3.2.3).

(c) The slopes of the walls of the upper parts of the pipes are unaffected by changes in the nature of the country rocks and no outward flaring of the margins near the present surface is evident (Hawthorne, 1975).

3.2.2 Changes in area with increasing depth

In Figure 3.1 the areas of the Kimberley pipes at various mining levels are plotted against depth below surface. A significant feature of these plots is that, with one exception (Bultfontein), they indicate the presence of two contrasting zones within the pipe columns. The upper zones, which extend from surface to depths of between 300 and 600m, are characterized by systematic decreases in area with increasing depth. These zones are correlated with the lower parts of the diatreme zones referred to in section 3.1. The deeper parts of the pipes (root zones) differ from the diatreme zones in that the average rate of decrease in area with depth is considerably reduced. Furthermore, variations in area within the root zones are commonly much more irregular than in the diatreme zones, although an overall trend towards smaller size with increasing depth is always evident. In several of the pipes the transition from diatreme to root zone is marked by sudden increases in area. For example the area of the De Beers pipe increases by 35% between the 455 and 500m levels (Fig. 3.1A). Local increases in area also occur entirely within the root zones (e.g. between the 800 and 1000m levels of the Wesselton pipe, see Fig. 3.1A).

The Bultfontein pipe is anomalous in that a systematic and relatively rapid decrease in area persists to the deepest mining level (825m below surface). The apparent outward flaring of the pipe margins above the 200m level, which is implied by the plotted area/depth relationship (Fig. 3.1B), in fact reflects rapid near-surface enlargement of a NNW-trending elongate (dyke-like) body of kimberlite, emplaced prior to the formation of the main body. This kimberlite is referred to as the NNW extension in Fig. 3.1B).

Probable reasons for the absence of a distinct root zone at Bultfontein are considered in the discussion on the origin of the KIMFIK pipes (Chapter 10). The depths at which the transitions from diatreme to root zones take place, vary considerably, even in neighbouring diatremes. These transition depths are not directly related to changes in the nature of the country rocks (Fig. 3.1). Furthermore, the irregularities within the root zones themselves cannot be related to changes in country rock stratigraphy (see discussion in Chapter 10).

The exposed parts of the Finsch and Koffiefontein pipes can be correlated with the diatreme zones of the Kimberley pipes but the nature of the deeper parts of these two pipes is not known as in both cases the depth of exposure is limited. The area of the Finsch pipe decreases regularly from surface to a depth of 350m (Fig. 3.2) and exploratory drilling indicates that this trend will probably persist to a minimum depth of 500m. The symmetry of the pipe is interrupted on the west side by the partial detachment and incorporation of a wedge-shaped mass of country rock. Available information indicates that the Koffiefontein pipe decreases steadily in area with increasing depth and becomes slightly elongated (Fig. 3.3). Intrusions of pre-pipe age form the East and West dyke extensions at Koffiefontein (Fig. 3.3). These bodies become somewhat irregular and wider at depths of $\sim$250m but their subsurface form has not been fully established.

Diatreme and root zones can be recognized in several other kimberlite pipes according to the nature of variations in area with depth. The Jagersfontein, Lace and New Elands pipes are good examples.

The upper part of the Jagersfontein pipe decreases steadily in size with depth until a distinct increase in area, between 570 and 670m levels, marks the start of the root zone (Fig. 3.1C). As in the Kimberley occurrences the root zone of this pipe is marked by a reduction in the average rate of decrease in area with increasing depth (compared with the diatreme zone). The transition from diatreme to root zone of the Lace pipe is indicated by a substantial increase in area (Fig. 3.4). Similarly, although the area of the New Elands pipe decreases steadily downwards over most of the exposed column, a noticeable increase in area is evident at a depth of approximately 180m (personal observation). The nature of the pipe below this level is not known but the increase in area may mark the start of the root zone.

3.2.3 Pipe configurations

The diatreme zones of the KIMFIK pipes vary considerably in cross-sectional shape and are approximately circular, imperfectly elliptical, kidney-shaped or irregularly elongate. Within the limits imposed by these variations the upper parts of the pipes are, however, chracterized by a general regularity of form (Figs. 3.2-3.3 and 3.5-3.8) which is reflected by the systematic decreases in area noted in section 3.2.2.

Typically the axes of the diatreme zones are vertical or nearly so. Although minor irregularities occur and there is a gradation towards slightly more complex cross-sections downwards, the margins generally dip inwards at consistently steep angles ($\sim 80°$). On a megascopic scale contours of the pipe margins are smoothly curved or, in respect of elongate pipes, may be approximately linear over considerable distances (see for example Fig. 3.5). Abrupt changes in the strike of the contacts are rare and where present usually indicate the junction between separate intrusions of kimberlite in individual pipes (section 3.5).

Hawthorne (1975) has pointed out that the deeper (root zone) parts of the pipes are characterized by a much greater degree of irregularity. This irregularity is reflected not only by more complex shapes but also by changes in shape over relatively short vertical distances (Figs. 3.5-3.8). In addition the root zones of the pipes or sections of them may be markedly inclined, in strong contrast to the typical vertical or near vertical orientation of the diatreme zones (Figs. 3.5-3.8). The irregularity of the root zones is emphasized in several cases by the splitting of pipes into more or less discrete columns or channels. The De Beers and Wesselton pipes are good examples (Figs. 3.5 and 3.6). These columns of kimberlite are usually linked by dyke-like connections but may be separated by intervening country rock over limited vertical distances (Figs. 3.5 and 3.6). The irregular nature of the root zones is also emphasized by the presence in some pipes of "blind" dome-like off-shoots or appendages of kimberlite roofed by country rock. Such features occur at Dutoitspan and Wesselton (Figs. 3.6 and 3.7) and an excellent example is provided by the Monastery pipe in the eastern Orange Free State (Fig. 3.9).

Figure 3.1 reveals an apparently abrupt change from diatreme to root zones in terms of variations in area of the pipes. In this diagram no cognizance is taken of the multiple stages of intrusion involved in the formation of the Kimberley pipes. As a result the diatreme-root zone relationships are to some extent oversimplified. If a more detailed assessment of diatreme/root zone relationships is made, taking the configurations of the pipes as well as areal variations into account, it is evident that:

(a) In specific parts of individual pipes the change from diatreme to root zone characteristics is abrupt. This is well illustrated by some vertical sections through the De Beers and Wesselton pipes (Fig. 3.5 and 3.6).

(b) Elsewhere the change is gradational.

(c) The transition from diatreme to root zone configurations may occur at different levels, or over different depth zones, in different parts of the same pipe. Consequently, over a limited vertical depth range, both root zone and diatreme zone features may be evident.

(d) Locally within some of the pipes diatreme zone characteristics persist to the limits of underground exposure (cf. sections AA^1, BB^1 and CC^1, Fig. 3.7). Bultfontein (Fig. 3.8) is an extreme case nowhere displaying the morphological irregularity characteristic of the root zones of other Kimberley occurrences. Below the bottom (825m) mining level borehole information indicates that this pipe grades into a dyke (Fig. 3.12).

In geographic areas which have undergone considerable post-kimberlite erosion individual pipes may display root zone or diatreme zone characteristics at the present land surface. The Kimberley area is a good example. Although the diatreme zones of the large pipes are partly preserved several small pipes have been eroded to root zone levels. For example the Loxtondal and Dieppút pipes, located within 15km of Kimberley, are characterized by pronounced morphological irregularity at surface (Figs. 3.10 and 3.11). Similarly the highly irregular shape of the Otto's Kopje pipe (Fig. 3.13) suggests an outcropping root zone. Further afield the "North Blow" on the Main Dyke at Bellsbank is an excellent example of a complexly shaped root zone and the Newlands occurrences near Barkly West are other good examples. The St. Augustine's pipe, situated less than 2km west of the Kimberley pipe, becomes highly irregular and is markedly inclined a short distance below surface (Fig. 3.13).

3.2.4 Satellite pipes

A feature of some major pipes is the occurrence of neighbouring but apparently discrete satellite bodies. For example, satellite pipes occur adjacent to the Lace (Fig. 3.4), Koffiefontein (Fig. 3.27), Letseng la Terae and Liqhobong pipes. The last two occurrences have been described by Bloomer and Nixon (1973) and Nixon and Boyd (1973b) respectively.

The diatreme-zone morphology of some of these satellite pipes (e.g. Letseng la Terae and Ebenhaezer) and the occurrence of tuffisitic kimberlite breccias within them indicates that they reached the contemporaneous land surface. Other satellites (e.g. the Lace satellite) are blind intrusions which do not reach the present day surface. A probable example of the latter type of occurrence is located near the Wesselton pipe where a major breccia zone has been intersected 1 060m below surface (Fig. 3.19). This breccia is identical in character to contact breccias (Section 3.3) which partly surround the root zone of the main pipe at higher levels (660-930m below surface). The breccia zone is thus interpreted as a breccia cap above a blind kimberlite intrusion.

Since kimberlite pipes decrease in size with increasing depth (Section 3.2.2) and are generally vertical bodies it is likely that many satellite pipes are not linked directly to the main pipes below exposed levels. They may, however, in some cases, be indirectly linked by some form of common feeder channel at depth. It is also likely that many adjacent pipes were originally linked above current erosion levels (Section 3.4.1).

3.3 NATURE OF THE PIPE CONTACTS

Further justification for the delineation of distinctive diatreme and root zones in the Kimberley pipes, or within parts of the pipes, is afforded by a number of contrasting contact features. These differences in the nature of the contacts in the two zones are considered of major significance with respect to the origin of the pipes (Chapter 10).

3.3.1 Contact features in the root zones

Pronounced and often abrupt changes in the dip and strike of pipe contacts occur in the root zones. On a large scale this irregularity is illustrated in Figures 3.5-3.8. Figure 3.14 indicates that major variations in dip and strike may occur over very short distances (1-2m).

Locally the contacts of the root zones dip outwards, often at angles of less than 30°. In such areas kimberlite is roofed by overhanging country rock and the widths of the overhangs exceed 30m in some instances (Figs. 3.5, 3.7 and 3.14).

Pre-existing structural discontinuities in the country rocks have exerted considerable control on the attitudes of the contacts and, consequently, on the shapes of the root zones. At many underground exposures the contacts are defined by planar joints which are parallel to prominent joint sets in the country rocks. On a large scale, joint control is revealed by the linearity of segments of the subsurface contours of the pipes and by pronounced elongation of the pipes, or parts of the pipes, parallel to prominent joint trends (Figs. 3.15 and 3.16). The pipes or parts of the pipes are also elongated parallel to pre-existing kimberlite dykes which have been emplaced along joints in the country rocks (Fig. 4.1).

In areas where the surrounding Ventersdorp lava or underlying early Precambrian gneiss is traversed by closely spaced joints, root zone contacts are often jagged or blocky (Fig. 3.17). Similar features occur where wallrock gneiss is strongly foliated (Fig. 3.17). Where two joint sets intersect more-or-less at right angles a "saw tooth" or serrated contact is often present (Fig. 3.17). For contact features of this nature to have developed blocks of country rock, bounded by joint planes, must have been detached from the wallrocks during the formation of the pipes (Fig. 3.18). Some detached blocks have irregular angular outlines which are clearly due to mechanical disruption along pre-existing discontinuities (Fig. 3.18). The detached blocks vary tremendously in size ranging from a few centimetres to more than 10m across (Fig. 3.18).

Extensive intrusion of kimberlite veins and stringers, extending from the pipes into the wallrocks, is not often evident but a good example occurs on the south side of the De Beers pipe at the 785m level. This area is poorly exposed in mine excavations but core drilling has indicated the presence of a stockwork of gneiss blocks and kimberlite veins in a zone which has a maximum width of 30m.

The width of the gneiss/kimberlite stockwork at De Beers is unusual. Generally veins of kimberlite extending from the pipes do not penetrate more than 2 to 3m into the country rocks (Fig. 3.17). The maximum thickness of the veins seldom exceeds 10cm.

A conspicuous feature of the root zones is the occurrence of extensive contact breccias which, with minor exceptions, are situated under the country rock overhangs referred to previously. These breccias are described in detail as they are considered of major significance with respect to the origin of the pipes (Chapter 10). The best examples are found at the De Beers pipe below the 500m level and at Wesselton below the 660m level (Figs. 3.20 and 3.21).

At the De Beers and Wesselton pipes the contact breccia zones in several instances exceed 30m in width and a maximum width of >50m has been mapped on the 739m level of the De Beers mine (Fig. 3.20). The vertical continuity of the breccia zones is variable but is commonly several tens of metres. The distribution of the breccia zones is shown in Figures 3.29-3.35.

The nature of the contact breccias varies but some consistent features are present. Among the most notable of these is the local derivation of the blocks and fragments in the breccias. At Wesselton the breccias occur at elevations where the country rock is Ventersdorp quartz porphyry and the fragments are restricted to this material. At the De Beers pipe some contact breccias straddle the Ventersdorp/ Basement contact. Although minor dislocation or slumping of the breccia masses as a whole may have occurred the stratigraphic succession is maintained within the breccias indicating little relative movement between breccia fragments.

The dominant characteristic of the fragments making up the breccias is a high degree of angularity. In some zones the fragments are densely packed, consistently small (less than 5cm in size) and the breccia has the appearance of country rock which has been subjected to intense in situ shattering (Plate 7.16g). Elsewhere the packing density of the fragments is lower, greater dislocation of the fragments is evident, the fragments display a greater range of size (Plates 3.1a-d and 3.2a-b) and prominent cavities occur, occasionally more than 10cm wide. These cavities are often lined by euhedral secondary minerals which have grown on the surfaces of surrounding fragments (Plate 3.1b). The most common secondary minerals are calcite and pyrite. Limited quartz occurs and zeolites are locally present.

The fragments in the breccias commonly range from millimetre-sized particles to blocks up to 50cm across although large blocks (>1m) may be present. In rare cases size zoning is evident. Usually larger fragments are concentrated close to the wallrock contacts and the average size of the fragments decreases inwards towards the centre of the pipe. At one exposure in a tunnel on the 585m level at De Beers Mine smaller fragments are concentrated near the wallrock contact and this zone grades rapidly into an inner area where the average size of the fragments is conspicuously larger.

As previously implied the contact breccias are often devoid of kimberlite (e.g. Plates 3.1a-d, 3.2a-b and 7.16g) but in some cases interstitial areas are filled by soft, earthy, decomposed kimberlite. This kimberlite is similar in appearance to "yellow ground" typical of kimberlite in the near-surface environs of pipes (Wagner, 1914). It has been strongly hydrated and now consists mainly of clay. The amount of kimberlite present in the breccias may exceed 50 vol.% (visual estimate) but is often much less.

The local derivation of fragments is particularly well illustrated by several exposures of contact breccia between the 500 and 620m levels at the De Beers pipe. At the extreme outer edges of the zones brecciation is reflected only by fracturing of the country rock and individual blocks have not been displaced (Fig. 3.22). Inwards from these narrow (<1 to 3m) fracture zones some displacement of angular blocks of country rock is apparent but separation of the blocks is limited. These outer marginal parts of the contact breccias contain no kimberlite or the amount of interstitial kimberlite is low (less than 20 vol.% - visual estimate). Inwards, towards the centre of the pipe, kimberlite becomes an increasingly abundant constituent and country rock fragments become increasingly dispersed. Finally at the inner margins these breccia zones grade into kimberlite containing numerous scattered xenoliths (Fig. 3.22). Minor mixing of fragments from different stratigraphic horizons occurs at the extreme inner margins of the breccias. This reflects increasing incorporation, dispersion and transportation of fragments by intruding kimberlite. In these inner marginal areas flow-induced parallel orientation of elongate fragments is locally evident but generally the breccia fragments do not display any orientation pattern.

The margins of contact breccias are not always gradational. Frequently both inner and outer contacts are sharply defined. In such cases the contacts between breccias and adjacent undisturbed wallrock, although abrupt, are often jaggedly irregular. This irregularity stems from the influence of jointing in the wallrocks (Plate

3.1d). Sharp internal contacts are common and occur where later intrusions of kimberlite have sheared off and incorporated the inner parts of contact breccias (e.g. Figs. 3.23-3.24 and Plate 3.2c).

The breccias are often characterized by remarkable local variations in the intensity of brecciation. Areas of finely fragmental, shattered material occur adjacent to relatively large blocks which have clearly remained more or less undisturbed (Fig. 3.23). Elsewhere large blocks measuring several metres in diameter are set in fine breccia (Fig. 3.23-3.24 and Plate 3.1c) and these blocks may themselves in part be highly brecciated i.e. pockets of fine breccia occur within otherwise competent blocks (Fig. 3.23).

Pronounced angularity of the fragments is a prominent feature of the contact breccias but rounded fragments occur in a few of the breccia zones. Good examples are found in exposures on the south-east side of the De Beers pipe on the 595 and 720m levels of the mine. At both levels the breccia consists of a mixture of abundant angular fragments with fewer subrounded to well rounded boulder-like blocks (Plate 3.1e).

Contact breccias such as those described are not restricted to the Kimberley pipes but have been observed by the author at other localities. For example, similar breccia is exposed in the lower workings of the New Elands Mine near Boshof (Plate 3.2d) and narrow zones of contact breccia occur locally at the 120m level of the Loxtondal pipe north-west of Kimberley. Whitelock (1973) has described a similar breccia at the Monastery pipe.

3.3.2 Contact features in the diatreme zones

Various aspects of the contacts of the diatreme zones of the KIMFIK pipes have been noted, or are by implication apparent, in the previous descriptions of the pipe configurations (Section 3.2.3) and the discussion on size variations at different depths (Section 3.2.2).

For the most part the present study has confirmed the detailed descriptions of the contacts given by previous authors (e.g. du Toit, 1906; Wagner, 1914; Williams, 1932). In these earlier contributions particular emphasis is placed on three specific contact features; the contacts are sharp, smooth and regular; the immediate country rocks have not been structurally deformed; and little or no thermal metamorphism of the wallrocks has resulted from kimberlite emplacement. Mention is also made of mechanical abrasion of the pipe walls (indicated by prominent grooves) and the common occurrence of slickensiding along contacts. Grooves or striae (Plate 7.16h) are commonly vertically oriented, less often inclined or even horizontal (du Toit, 1906). Hawthorne (1975) has shown that the average inward dips of the walls of the pipes are consistently steep (ranging between 79° and 84°) and that the major features of the diatreme zone contacts probably persisted upwards as far as the base of the former crater zones.

Locally, within the diatreme zones, features are evident which are anomalous in terms of the foregoing, generally accepted, characteristics of the upper parts of kimberlite pipes. For example, the contacts of the Finsch pipe are prominently stepped in localized areas at elevations where the country rock consists of banded ironstone. Plate 3.12 shows that this feature is a function of the prominently stratified and highly jointed nature of the wallrocks. Slumping or plucking of joint- and bedding plane-bounded blocks has occurred at the contacts during the formation of the pipe. Similar features have not been seen in the upper parts of the Kimberley pipes but may have been present where the country rock consists of strongly jointed prominently bedded Karoo shales. No contacts are preserved in these mined-out areas. Similar features are evident at other localities and an excellent example of the influence of joints on the form of a diatreme zone contact has recently been exposed by mining at the Letseng Satellite pipe in Lesotho. The influence of jointing on the contacts and on the general form of the pipes is, however, less pronounced in the diatreme zones than in the root zones. This is shown by the smoother character and less pronounced elongation of the contours of the upper parts of the pipes (Figs. 3.5-3.8).

Another unusual feature in respect of the generally accepted characteristics of diatreme zone contacts is the occurrence of a broad zone of brecciated Ventersdorp lava along part of the north contact of the Dutoitspan pipe between the 280 and 420m levels (Fig. 3.25). This breccia has in part been impregnated by kimberlite but elsewhere consists entirely of lava fragments. Similar high level (diatreme zone) contact breccias have also been noted locally at Letseng la Terae and Wesselton. The implications of these breccias within diatreme zones are discussed elsewhere (Chapter 10).

Wagner (1914) and Williams (1932) referred to the common presence of narrow selvedges of secondary serpentinous and calcite-rich material at pipe contacts. These selvedges rarely exceed 2-3cm in width and are usually only a few millimetres wide. In addition to selvedges of introduced secondary material severe alteration of the kimberlite is often visible adjacent to the contacts. The altered zones seldom exceed 1m and are frequently less than 25cm in width. Secondary selvedges and marginal alteration of kimberlite are features which are not entirely restricted to the diatreme zones. They are also found at deeper levels in the pipes.

Williams (1932) noted that shearing had often occurred on one or both sides of the secondary contact selvedges. Spectacular examples of similar shearing have been exposed at the Finsch pipe where shear planes, marked by prominent slickensiding, have developed along the inner contact of a narrow zone of highly oxidized kimberlite rimming the pipe (Plates 3.2f-g).

3.4 THE NATURE OF THE ERODED PARTS OF THE KIMFIK PIPES

3.4.1 Original surface areas

The largest known kimberlite occurrence is the Mwadui pipe in Tanzania which has a surface area of 146 ha. Other large pipes include Orapa (111 ha) and Jwaneng (56 ha) in Botswana. The Camafuca-Camazambo and Catoca pipes in Angola measure 68 and 66 ha respectively. The recently located Successo 1 pipe in Brazil has a surface area of 120 ha (pers. comm., D. du Toit). Larger pipes, recently discovered in Brazil, have been located but their kimberlite character has not been confirmed (pers. comm., D. du Toit).

The KIMFIK pipes are small compared with the foregoing examples but direct size comparisons neglect the effect of different degrees of erosion. The large pipes noted above have not been substantially eroded (Hawthorne, 1975) whereas estimates of the amount of denudation in the Kimberley area, since kimberlite emplacement, range from 900m (Wagner, 1914) to between 900 and 1900m (Hawthorne, op. cit.). Since many authors (e.g. Wagner, op. cit.; Dawson, 1971, Fig. 2; this thesis, Fig. 3.1) have noted that kimberlite pipes decrease in size with increasing depth a substantial increase in the areas of the KIMFIK pipes at the original surface can be postulated.

Rough estimates of the original surface areas of the KIMFIK pipes can be made provided certain assumptions are accepted. These assumptions, which are based on the kimberlite pipe model put forward by Hawthorne (1975) are:

(i) The systematic increase in area upwards, shown by the preserved parts of the diatreme zones of the pipes (Fig. 3.1), continued to within 200 to 300m of the original land surface.

(ii) Outward flaring of the pipe margins occurred from within 200 to 300m of the surface reflecting the original presence of craters within this zone.

(iii) The average inward dip of the margins of the craters was 60°. This slope angle and the estimated depth of the craters are consistent with known examples (Mannard, 1962; Edwards and Howkins, 1966; Hawthorne, 1975).

(iv) Circular cross-sections are assumed (to simplify the calculations).

(v) The amount of post-kimberlite erosion in the Kimberley area is 1400m (the average of the minimum and maximum estimates noted previously).

(vi) A similar amount of erosion occurred in the Finsch and Koffiefontein areas.

On the basis of these assumptions the original surface areas of the KIMFIK pipes are estimated to have ranged between 65 ha (Kimberley pipe) and 110 ha (Finsch pipe). If the minimum estimated amount of post-kimberlite erosion in the Kimberley region (900m) is accepted these areas are reduced to 43 and 78 ha respectively. These estimates serve to indicate that, prior to erosion, the KIMFIK pipes were probably comparable in size with some of the largest known kimberlite occurrences.

In making the foregoing estimates no account has been taken of possible coalescence of neighbouring diatremes above the present surface to form very large composite pipes. The large Jwaneng pipe in Botswana provides an excellent example of this situation and clearly represents the coalescence of three distinct diatremes which are separated by country rock a short distance below surface (Fig. 3.26). There is little doubt that similar coalescence occurred in the case of the Koffiefontein and neighbouring Ebenhaezer pipes (Fig. 3.27). These two pipes are only 165m apart at the present surface. Bultfontein and Dutoitspan are separated by approximately 600m of intervening country rock. It is therefore possible that these pipes also coalesced at some point within the eroded stratigraphic column. Similarly three occurrences of kimberlite termed the "Treatment Plant" pipe, the

"North Water Tunnel Fissure" and the "Pan Tunnel Kimberlite" are located close to Dutoitspan (Fig. 3.28) and may have coalesced above the present surface.

3.4.2 Morphology and contact features of the eroded columns

The most detailed assessment of the probable nature of the eroded parts of the Kimberley pipes is that made by Hawthorne (1975). This author concluded that the regular shape and upward increase in area of the preserved parts of the diatreme zones, continued through the eroded columns to within about 300m of the pre-erosion surface. From depths of this order the contacts are assumed to have flared outwards producing the calyx or tulip shape referred to by Kennedy and Nordlie (1968). The broad morphological aspects of this model are consistent with evidence obtained from kimberlite pipes elsewhere which have undergone little or no erosion.

During the course of the present study inclusions of bedded epiclastic kimberlite were found at the Wesselton and Finsch pipes (Plate 7.16d). These inclusions are direct evidence of the former presence of craters (wholly or partly filled by epiclastic kimberlite) at the original surfaces of these two pipes. Similar inclusions occur in other kimberlite pipes and have been observed by the author at several localities. The type example is a pipe near Beit Bridge in Zimbabwe which contains abundant inclusions of epiclastic kimberlite set in primary kimberlite. Approximately 50 vol.% the upper 100m of this pipe is occupied by randomly oriented, frequently large (several tens of metres across), blocks of stratified epiclastic kimberlite. Inclusions of epiclastic material in primary kimberlite are also present in the Letseng and Kao pipes in Lesotho and detailed examination of other pipes would probably reveal similar inclusions in some instances.

Kimberlite pipes with preserved or partly preserved crater zones have been discovered in Tanzania, Zambia, Botswana, Angola, Zaire and Mali (Tremblay, 1956; Mannard, 1962; Edwards and Howkins, 1966; Hawthorne, 1975). Similar occurrences are also found in Brazil (personal observation, 1977) and epiclastic kimberlite has been intersected in a borehole on the farm Outpost No. 1 in the Northern Cape (Plate 3.1f). This pipe and others in the vicinity are the nearest occurrences to the kimberlites of the type area which contain in situ epiclastic kimberlite.

Known crater zones are characterized by relatively flat marginal dips which generally range between 50° and 75° (Hawthorne, 1975). Extensive brecciation of the immediate country rocks is a common feature and partial or complete collars of brecciated wallrock rimming the craters, have been recorded at several localities in Tanzania and Botswana (Mannard, 1962; Edwards and Howkins, 1966; Hawthorne, 1975). A similar breccia (Plate 3.2h) has been observed by the author at the Camutue pipe in Angola. It is probable that such breccia jackets occur around many other kimberlites which have not been eroded to diatreme zone levels but these breccias are not exposed because of overlying surficial deposits.

The best documented example of breccia zones surrounding craters occurs at the Mwadui pipe in Tanzania. The marginal breccia at Mwadui has a maximum width of 55m (D.G. Fullerton, personal communication). According to Tremblay (1956) the breccia zone dips inwards parallel to the crater margins and individual blocks have undergone little displacement.

Many of the breccias contain interstitial kimberlitic material, together with comminuted rock and mineral fragments (Tremblay, 1956; Mannard, 1962). The matrix material is generally extremely altered and includes secondary minerals such as chlorite and chalcedony. Detailed descriptions of the nature of the kimberlitic material are not available.

3.5 MULTIPLE INTRUSIONS IN THE KIMFIK PIPES

Probably the earliest reference to multiple intrusive activity in kimberlite pipes was made by Moulle (1885) who claimed to recognize 15 distinct columns of kimberlite in the upper part of the Kimberley pipe. Later Wagner (1914) stated that the kimberlite in different parts of specific pipes differed markedly "...in appearance, properties and diamond content, and the diamonds found in different varieties of pipe rock may likewise vary considerably in character." He concluded that most of the larger pipes were the products of a number of successive eruptions and noted that sharp contacts occurred between adjacent columns of kimberlite. Wagner recognized two intrusions of kimberlite in the De Beers, Bultfontein, Kimberley and Koffiefontein pipes and considered that four stages of intrusion were represented at Dutoitspan. In contrast Williams (1932) while admitting variations in the character of the kimberlite in individual pipes, firmly rejected any interpretation which recognized separately intruded, discrete, columns of kimberlite. During the course of the present investigation detailed geological mapping has verified the correctness of Wagner's views. This investigation has also shown that the pipes are much more complex in respect of the number of kimberlites present than Wagner supposed.

In the last two decades frequent references have been made in the literature to multiple intrusions within kimberlite pipes (e.g. Dawson, 1960, 1962, 1967a, 1971; Rabhkin et al., 1962; Frick, 1970; Verwoerd, 1970; Harris and Middlemost, 1969; Frantsesson, 1970; Ilupin, 1972; Zuyev, 1972; Rolfe, 1973; Clement, 1972; Hawthorne, 1975; Mitchell, 1979b; Skinner and Clement, 1979; Pasteris, 1980a). Generally, however, few data are available bearing on the variation in character, distribution, contact relationships and relative ages of different kimberlites in individual pipes. Consequently these aspects are examined in some detail in this thesis.

3.5.1 Recognition and delineation of different intrusions

Separate intrusions of kimberlite within individual pipes of the KIMFIK group can be distinguished in one or more of several ways:

(i) Discrete intrusions may differ substantially from one another in colour, hardness and the nature and extent of alteration.

(ii) Differences in the content of crustal and/or mantle-derived xenoliths are commonly evident and provide easily visible, megascopic evidence of multiple intrusion. The differences are usually most marked in respect of high-level country rock inclusions which may differ in terms of: rock type, abundance, relative abundances of different types, shape, degree of rounding and extent and type of alteration that has occurred. These aspects of the inclusion content of different intrusions in the KIMFIK pipes are discussed in Chapter 7.

(iii) Mineralogical differences may be evident on a macroscopic scale in respect of the minerals which occur as macrocrysts (Clement et al., 1977; this thesis, Chapter 2) and megacrysts.

(iv) Major differences in the mineralogy of individual intrusions are often revealed under the microscope. An important result of this study is the recognition of extensive petrographic diversity between kimberlites in specific pipes and between the different KIMFIK pipes (Chapter 7).

(v) Various authors (e.g. Lewis, 1887; Wagner, 1914; Bobrievich et al., 1959; Dawson, 1960, 1971; Rabhkin et al., 1962; Milashev, 1965; Frantsesson, 1970; Mitchell, 1970; Rolfe, 1973; Clement, 1973; Hawthorne, 1975; Kostrovitsky, 1976; Clement and Skinner, 1979) have noted that several textural varieties of kimberlite occur. Textural differences often provide an easy means of distinguishing different intrusions in the KIMFIK pipes. Textural variations are described in

Chapters 2 and 7.

(vi) Mine production and sampling operations have clearly shown that the diamond content of different intrusions often varies substantially, in some cases by a factor of 10 or more. Differences in diamond content can be useful in distinguishing separate intrusions which are similar in other respects (see Chapter 9).

Commonly discrete intrusions in the pipes are separated by sharp contacts examples of which are shown in Plates 3.1g-h, 3.3a-c and 3.4a-b. Such contacts often have a welded appearance but a physical plane of separation may be present. Prominent slickensiding may occur along such contact planes.

A narrow zone of secondary calcite, ranging from less than 1mm to 2cm in width, may occur along sharp contacts between intrusions (Plate 3.3b). This calcite frequently has a fibrous habit, the fibres being oriented perpendicular to the contact. Similar calcite also occurs in joints or fractures traversing individual intrusions.

The positions of contacts between major intrusions in the pipes are sometimes indicated by the presence of kimberlite dykes. Pre-existing contacts appear to have provided passageways for intruding magma to be emplaced as dykes within the pipes. Dykes also occur along the contacts between kimberlite and country rock.

Flow structures frequently occur adjacent to contacts but are not always restricted to the marginal parts of intrusions. Where present marginal flow structures may be reflected by:

(i) Vague flow lines parallel to the contact in the younger of adjacent intrusions (e.g. Plates 3.1g-h).

(ii) More irregular patterns suggesting turbulent emplacement.

(iii) Distinct, narrow (usually <10cm wide), alternating bands of coarse (macrocryst-rich) and fine-grained (macrocryst-poor) material. The bands are parallel or subparallel to the contact and occur in the younger intrusion.

(iv) Parallel orientation of elongate minerals and xenoliths in the younger of adjacent intrusions. Again the direction of orientation is parallel or nearly parallel to the contact. Such orientation may be evident on macroscopic and/or microscopic scales.

(v) Pronounced concentrations of xenoliths in bands adjacent to, and parallel to, the contacts. In such cases the xenolith bands also occur in the younger of adjacent intrusions.

Although different kimberlites can usually be recognized on individual mining levels by one or more of the foregoing criteria, vertical correlation of the intrusions is hindered by several factors. Most of the difficulties are of a practical nature. They relate mainly to the extent to which the pipes have been mined and to the amount and quality of exposure in existing mine workings.

From surface to depths of approximately 250m the Kimberley pipes have been completely mined out and very little information is available on the nature and distribution of different intrusions in these zones. Systematic, on-going geological mapping was not carried out prior to the mid-1960's.

Much of the kimberlite between depths of 250 and 500m has also been extracted and access to the remaining in situ material is limited. Many of the older mine workings have collapsed or entry is precluded by unsafe conditions (e.g. lack of adequate ventilation). Some information on the distribution of separate intrusions has, however, been obtained by mapping of local areas where access is (or has recently been) possible. Supplementary information has been obtained from Company files.

In recent years mining of the Kimberley pipes has been concentrated at depths of greater than 500m. Consequently the bulk of the geological mapping undertaken during the course of this study was carried out in these deeper parts of the pipes. The extent of exposure in these zones varies from pipe to pipe and from time to time due to changes in mining methods and to the stage reached in the extraction of ore from specific mining blocks. The amount of exposure is also influenced by varying requirements for the development of sampling levels to assess the tenor of the ore.

For many years (dating back to the 19th Century) the Kimberley pipes were exploited by a chambering mining system (Gallagher and Loftus, 1960). This method involves the excavation of a closely spaced grid of tunnels across the pipes on successive levels only 12m apart. At the Wesselton and De Beers pipes, between depths of 500 and 600m, a number of chamber levels, providing almost continuous exposure, have been mapped. Elsewhere in these two pipes and in the other Kimberley mines the ore has for some time been extracted by block caving methods (Gallagher and Loftus, 1960; Cleasby et al., 1975). Where block caving is practised exposures of kimberlite are restricted as major mining levels are only required at vertical intervals in excess of 100m. Recent development of sublevel caving practices has resulted in improved exposures at some mines.

Excellent exposures of kimberlite in the two pipes mined by opencast methods (Finsch and Koffiefontein) have been available throughout the duration of this study. The upper 100m of the Koffiefontein pipe was, however, mined out prior to World War II and little information relating to this part of the pipe is available.

The recognition, delineation and vertical correlation of individual kimberlites is also complicated in some instances by the inhomogeneous nature of some intrusions. This inhomogeneity may be erratic and local (even on a handspecimen or thin section scale) but may involve more systematic and widespread variations in character. Depth-related changes in texture are particularly relevant in the latter respect and various aspects of such variations are discussed in later sections of this thesis (Chapters 7, 8 and 10).

3.5.2 Distribution of separate intrusions in the pipes

The distribution of kimberlite intrusions is illustrated by a series of geological plans of the pipes, at different levels below surface, and by a number of vertical sections (Figs. 3.29-3.37). These plans and sections reflect only those parts of the pipes where sufficient data are available to allow reasonably complete interpretation of the vertical and horizontal configuration of discrete intrusions. Consequently the near-surface parts of the pipes are not illustrated. Also excluded from the diagrams are most of the numerous kimberlite dykes adjacent to (and within) the pipes. These dykes are described in Chapter 4.

Careful interpretation of the vertical sections is necessary, as in some instances they convey a misleading impression of the form of individual intrusions. In complex pipes it is impossible to select any one line of section across each pipe which will intersect all internal contacts between intrusions more or less perpendicular to strike. As a result some extremely flat and/or irregular apparent contact dips are reflected in the sections.

Wesselton

The distribution of kimberlites in this pipe is shown by geological plans of seven mining levels (Figure 3.29). Information derived from these and other levels has been used to construct the vertical sections shown in Figure 3.30.

Excluding minor dykes within the pipes, 9 (possibly 10, see Figure 3.29) different kimberlite intrusions have been recognized by nine geological plans (Figures 3.31 and 3.32) and by 3 vertical sections (Figure 3.33). One of these kimberlites (D7) apparently occurs as a large inclusion within D2 and hence represents a remnant of an intrusion emplaced at a relatively early stage of pipe formation (see Section 3.5.4). D17 (which occurs only as a single small exposure, within D11, in a tunnel on the 870m level) may also be a large inclusion. This interpretation is suggested by the irregular nature of the body which is terminated by a flat-dipping upper contact. There is, however, insufficient exposure to allow this conclusion to be verified.

The central part of Dutoitspan is mainly occupied by D11 and is relatively simple compared to the eastern and western parts where complex sequences of intrusion are present. This complexity is most pronounced in the southwestern area of Dutoitspan which is known as the West Auxiliary pipe (Figure 3.32). Kimberlite D1 occurs in a subsurface southerly extension of West Auxiliary pipe. This extension has only been exposed on the 870m level (Figures 3.32 and 3.33).

As at Wesselton most of the contacts between separate intrusions are sharp but the contacts between D13 and D14, D11 and D15 and D11 and D16 are gradational over distances of a few metres (<5m). Away from the contact zones the neighbouring kimberlites can be differentiated according to several criteria, notably differences in matrix mineralogy, texture and differences in crustal xenolith and diamond content.

An interesting feature at Dutoitspan is that D14 was intruded at depth along two discrete channels which linked, to form a bi-lobate intrusion, above the 675m level (Figures 3.32 and 3.33).

De Beers

This pipe is geologically less complicated than the Wesselton and Dutoitspan occurrences. Six separate kimberlites (DB1-DB6, Figures 3.34 and 3.35) together with marginal breccias, probably of several ages, occur between the 245 and 785m levels.

As at Wesselton and, to a lesser extent, Dutoitspan the upper part of the pipe is relatively simple and only two kimberlites, now largely removed by mining, are present. The occurrence of these two kimberlites (DB3 and DB5) was recognised by Wagner (1914) who states that the contact between the two kimberlites dipped eastwards at a steep angle.

At depth DB3 and DB5 are no longer in contact as DB2 occupies an intervening position (Figures 3.34 and 3.35). The contact between DB5 and DB2 is always gradational whereas, locally, the contact between DB2 and DB3 is indicated by a pronounced concentration of small (<10cm) country rock xenoliths in a narrow (<1m), near-vertical band at the margin of DB3. The latter kimberlite is much richer in diamond than DB2 and the contact between these two intrusions is clearly revealed by an abrupt change in diamond content in areas where sampling operations have been carried out (see Chapter 8). Over the depth zone within which sampling to determine diamond tenor has taken place DB5 also has a higher diamond content than DB2. The contact between DB5 and DB6 has not been exposed.

Two additional varieties of kimberlite (DB1 and DB4) occupy small areas of the root zone in the southern part of the pipe. Sharp contacts separate these intrusions from each other and from other adjacent kimberlites. The contact between DB2 and DB6 is gradational.

Broad zones of contact breccia are prominent in the southern part of the pipe and occur to a lesser extent in the central area (Figures 3.34 and 3.35).

Between the 500 and 620m levels of the mine most (but not all) of the contact breccia zones consist of locally derived country rock fragments with an interfragmental matrix of DB2 kimberlite (although, as noted in Section 3.3.1, the outermost parts of the zones may be devoid of kimberlite). In contrast the large area of contact breccia, exposed below the 720m level on the south side of the pipe, is entirely free of kimberlite. Sharp contacts occur between this breccia and the neighbouring DB1 and DB4 kimberlites.

Bultfontein

The distribution of different kimberlites at Bultfontein has not been well established due to poor accessibility. No access to in situ kimberlite is currently possible on any of the existing block cave mining levels and no sampling levels have been developed through the pipe since the early 1960's.

It is evident from the limited amount of mapping carried out and from old records in the Geological Department of D.B.C.M. Ltd. that at least three major intrusions of kimberlite occur within the pipe (B1-B3, Figures 3.36). It is, however, possible that additional intrusions have not been recognised.

B1 occupies the NNW-trending extension of the pipe (Figure 3.36) and is separated from the main intrusion (B2) by a sharp contact. This contact was well exposed on a number of old chambering levels (above the 580m level). B3 lies wholly within B2 and mapping on the 700m level has shown that these two kimberlites are also separated (at least at this level) by a sharp contact.

Finsch and Koffiefontein

As noted previously (Section 3.2.2) only the diatreme zones of these two pipes have been exposed by mining operations. At Finsch eight discrete intrusions (plus radial dykes) have been recognised between the surface and the 348m level of the mine. The distribution of these intrusions (F1-F8) is shown in Figure 3.37. F1 is volumetrically by far the most important of the intrusions . The only other intrusions of significant size are F2 and F3. The F2 kimberlite is extremely varied in character and may reflect more than one intrusion. However, sharp internal contacts have not been noted.

At current levels of exposure the Koffiefontein pipe is relatively simple (Fig. 3.38). Most of the pipe is occupied by the KOF1 TKB and a second, very similar, TKB (differentiated only on diamond grade) occurs within the West Dyke Extension (KOF2). A third, mineralogically different, TKB has been described (E.M.W. Skinner, De Beers Geology Dept. Int. Report DBG/PI/81-34) but this material has only been located as blast fragments. It is not known whether it occurs as a discrete intrusion or a large inclusion in KOF1. At least one hypabyssal-facies intrusion occurs in the East Dyke Extension at depth but this intrusion (KOF3) has not been clearly delineated.

3.5.3 Relative ages of intrusion

The relative ages of emplacement of different intrusions in the pipes can in many cases be determined according to one or more of the following criteria:

(i) Where inclusions of one kimberlite occur in another the relative ages of the two intrusions are unequivocably established (e.g. Figure 3.33 and Plate 3.3a).

(ii) The spatial relationships of adjacent kimberlite may provide a reliable indication of relative ages of intrusion. For example, several intrusions in the KIMFIK pipes occur as regular circular or oval-shaped plugs cutting and wholly within other kimberlites (e.g. Figures 3.31, 3.32, 3.34, 3.36 and 3.37). The form of these plugs clearly indicates that they have cored out and partly replaced pre-existing kimberlites and the age relationships are usually confirmed by one or more of the other criteria listed here. Similarly the contact between adjacent intrusions may be smoothly curved and the line of contact convex, often markedly so, towards one of the kimberlites (e.g. Figures 3.29, 3.32 and 3.34). Convex embayments of this type are due to partial coring out and removal of an earlier kimberlite by a subsequent adjacent intrusion.

(iii) In rare instances veins or dyke-like protuberances extend from major intrusions into older, adjacent kimberlites thereby allowing recognition of the age relationships (e.g. Note the manner in which D12 intrudes D11; Fig. 3.31, 300m level).

(iv) When three discrete intrusions are mutually in contact the youngest intrusion can be recognised where it transgresses the contact between the two older kimberlites (e.g. Note the manner in which D18 cuts the contact between D2 and D16; Fig. 3.32, 745m level).

(v) The position of a kimberlite within a pipe, particularly in relation to the configuration of the pipe margin, may allow its relative age to be determined. In the root zones of the De Beers, Wesselton and Dutoitspan pipes some intrusions are restricted to areas underlying overhanging parts of the pipe contacts. They are interpreted as protected remnants of kimberlite cut and partly removed by adjacent later intrusions (e.g. Figures 3.33 and 3.35).

(vi) A later intrusion may cause a narrow zone of alteration to be developed in the marginal parts of an adjacent earlier kimberlite. Such alteration zones seldom exceed 0,5m in width and may only be a few millimetres wide.

(vii) Flow structures such as those described in Section 3.5.1 may be present in a kimberlite close to, and parallel to, the contact with an older intrusion.

It has not been possible to establish the complete sequence of emplacement in all the pipes according to the foregoing criteria.

In complex pipes such as Dutoitspan and Wesselton individual intrusions are not always in contact with each other, nor are they mutually in contact with other kimberlites, in a manner which allows complete elucidation of the age relationships. The problem is illustrated by reference to Figure 3.41 which shows diagrammatically the relative ages (as far as they can be determined) of separate intrusions at Dutoitspan.

In the central and western parts of the Dutoitspan pipe a sequence of eight stages of intrusion has been established. As shown on the left side of Figure 3.41 kimberlite D6 is the youngest and D16 is the oldest of these intrusions. The ages of the other intrusions at Dutoitspan, relative to this sequence and to each other, are less clear but upper or lower limits have been established in all cases (as indicated in Figure 3.41). Thus kimberlite D10 and D18 are younger than D2 but their ages of intrusion relative to each other and to all other kimberlites younger than D2 are not known. One or both of these intrusions could be younger than D6. Similarly kimberlites D1, D7, D8 and D9 are all older than D2. Kimberlite D9 is older than D8 but the ages of emplacement of these intrusions, relative to all other kimberlites which are older, or may be older, than D2 are not known. D14 is younger than D13 and both are younger than D11 but no other age relationships have been established. Kimberlites D15 and D17 are relatively old intrusions; both are older than D11 and either or both may be older than D16. One or both of these kimberlites may, however, have been intruded after D1, D7, D8 and D9 as well as after D16.

Similar diagrams have been constructed to illustrate, as far as can be established, the age relationships of the intrusions in the other KIMFIK pipes (Figs. 3.42-3.44).

The relative ages of four of the intrusions in the De Beers pipe can be established by reference to Figures 3.34-3.35. These age relationships are illustrated in the central column of Figure 3.42. DB3 occurs as a plug within DB2 and clearly post-dates the latter. On the basis of the observations made by Wagner (1914) DB2, in addition to having been intruded after DB1, also post-dates DB5. However, the age relationships between DB1 and DB5 have not been established.

The relative age of DB6 is problematic. This intrusion has been exposed on the 720 and 785m levels of the mine. DB5 has not been located at these levels and it could be argued that DB5 has been stoped out and replaced by DB6, i.e. DB6 is the later of the two intrusions. There is, however, no evidence of such stoping, no inclusions of DB5 in DB6 have been noted and no subhorizontal or domed contact between DB5 and DB6 has been intersected above the 720m level. An alternative and more favoured interpretation is that DB5 cuts DB6 but that the former is a very small intrusion within the 720-785m depth zone where it has not been intersected in underground workings. The probable position of the DB5 intrusion on the 720 and 785m levels, based on downward projection and taking the configuration of the pipe into account, is shown in Figure 3.34. If this interpretation is correct DB5 must have been expanded rapidly upwards to core out DB6 and occupy the whole of the northwestern area of the pipe at some depth above the 720m and below the 560m levels.

The age of DB4 is also uncertain. It is undoubtedly younger than DB2 but its age in relation to DB3 has not been established. Similarly the age of DB1 in relation to DB5 and DB6 is not known. The formation of the contact breccia designated DB/CB1 on Figure 3.35 preceded the intrusion of DB1 and may have preceded DB5 and DB6. Other contact breccias may have formed contemporaneously or, more likely, may represent more than one stage of pipe formation (see Chapter 10). All the breccias appear to be older than DB2.

Interpretation of the age relationships at Wesselton is handicapped by problems in correlating some of the intrusions between different mining levels. These difficulties reflect the presence of some petrographically similar, but apparently separately intruded, kimberlites. In addition correlation is hampered by gradational changes in character within individual intrusions. However, in the central part of Wesselton a sequence of four stages of intrusion is clearly apparent. Within this sequence W5 is the youngest kimberlite and a contact breccia (W/CB1 containing interstitial kimberlite is the oldest. The age relationships of the other kimberlites at Wesselton, to each other and to this sequence, are shown in Figure 3.43. This diagram reveals numerous uncertainties regarding relative ages of intrusion at Wesselton.

At Finsch the F6 kimberlite cuts F5 and both intrusions are restricted (in the open pit) to a small outward-extending embayment in the pipe margin (Fig. 3.37). F5 and F6 have been truncated by F1 and hence pre-date the latter. F3 also appears to be a marginal remnant of an earlier intrusion truncated by F1 but the age relationships between F5 and F6 and F3 have not been established (Fig. 3.44). F2, F7 and F8 cut F1 and hence are younger than the latter. Dyke extentions of F2 cut F7 thus the former post-dates the latter. F2 also post-dates F8 (Fig. 3.37). F4 is the youngest of all eight intrusions as it has partly cored out F2. The age relations between F7 and F8 are not known.

In view of the limited number of discrete intrusions the relative ages of the kimberlites at Koffiefontein and Bultfontein are easily determined (Fig. 3.44).

3.6 MAJOR INCLUSIONS OF COUNTRY ROCKS

Among the most spectacular features of kimberlite pipes are huge inclusions of country rock. In the early mining days these inclusions were sometimes mistaken for in situ wallrocks and the term "floating reef" was used to distinguish them from the "reef" or country rock proper. Many authors (e.g. Harger, 1905; du Toit, 1906; Voit, 1907; Merensky, 1909; Wagner, 1914; Williams, 1932; Dawson, 1960, 1971; Hawthorne, 1975) have drawn attention to these major inclusions and Williams in particular has described them in considerable detail. Probably the best known and certainly one of the largest floating reef inclusions is the enormous mass of Waterberg Group rocks which occurs in the Premier pipe near Pretoria. This body is composed mainly of quartzite with associated conglomerate and sandstone (Williams, 1932).

Numerous floating reef bodies have been noted in the Kimberley pipes and, although many of the larger masses have been removed by mining operations, several examples are exposed at current mining levels. In the more recently discovered Finsch pipe numerous floating reef bodies have been exposed in the opencast mine (Fig. 3.37 and Plates 3.3d and 3.4c) and floating reef also occurs in the Koffiefontein pipe (Fig. 3.38).

3.6.1 Nature of the floating reef in the KIMFIK pipes

Floating reef in the Kimberley and Koffiefontein pipes is composed exclusively of rocks derived from the Karoo Sequence. Investigations by earlier workers (Wagner, 1914; Williams, 1932) have shown that a more-or-less complete cross-section of Karoo strata is representa-ted (or was present prior to mining) in the pipes. These authors presented lithological and palaeontological evidence indicating the occurrene of Drakensberg, Beaufort, Ecca and Dwyka rocks as floating reef. The floating reef inclusions therefore testify to the presence in the northern Cape of the entire Karoo sequence at the time of kimberlite emplacement (see Hawthorne, 1975).

The rocks which occur as floating reef in the Finsch pipe consist almost entirely of greyish-green to red-brown amygdaloidal or (more commonly) massive lavas; dolerite; red, purple, yellow, green and black mudstones and shales and white to buff-coloured sandstones. Although no Karoo rocks are now preserved in the vicinity of the pipe, this floating reef assemblage must, as at Kimberley and Koffiefontein, be derived from the Karoo Sequence.

This conclusion is supported by the investigations of Visser (1972) who correlated mudstone from one of the floating reef bodies with the Beaufort Group. In addition Hart (Unpubl. rpt., De Beers Consolidated Mines) has identified mudstones of Dwyka or younger (probably Lower Ecca) age on the basis of microfloral remains.

Floating reef in the KIMFIK pipes occurs in two forms. Some inclusions occur as discrete competent masses showing little internal disruption (apart from minor fractures) and original structures are well preserved. In contrast other floating reef masses are extensively brecciated (Plates 3.3e-f) and form irregular columns within the pipes (e.g. Figs. 3.39-3.40). To some extent these two forms of floating reef reflect the nature of the rocks involved. Thus large undisrupted inclusions of soft, fissile shales are rare, whereas more competent rocks, such as Drakensberg lava and dolerite, occur much more frequently (although not exclusively, see Plates 3.3d and 7.16f) in a relatively undisturbed state.

Some floating reef breccias are composed entirely of a single rock type (Plate 3.3e) but in others a variety of fragments are present (Plate 3.3f). In the latter case different fragments may be randomly mixed, to produce a heterogeneous assemblage, or different types of fragments may be zonally distributed within the breccias (Williams, 1932).

In many floating reef breccias the extent of brecciation is considerable and the individual fragments are small. In most breccias of this type the majority of fragments measure less than 25cm across and many are less than 5cm in diameter (Plates 3.3e-f). Typically the fragments are extremely angular (Plates 3.3e-f) although rounded fragments occur in rare instances. Some of the fragments in floating reef breccias composed of soft rocks such as shales or mudstones have sheared and striated surfaces indicative of mutual abrasion during subsidence (Section 10.2.3).

Many floating reef breccias are entirely devoid of kimberlitic material and are separated from the surrounding host kimberlite by sharp contacts (Plate 3.3e). In such cases the fragments are often closely packed and may be only slightly displaced relative to one another. In similar breccias where greater dislocation of individual fragments has occurred, the interstitial areas may be occupied by secondary calcite, quartz and zeolites (Williams, 1932). In a few cases voids are locally present between breccia fragments.

In some instances the kimberlite in which the floating reef masses are incorporated has penetrated the breccias and forms a matrix between the foreign rock fragments. The amount of kimberlite matrix in such breccias varies considerably. The interior parts of some breccias contain no kimberlite but kimberlite becomes an increasingly abundant constituent towards the margins and the contacts between breccia columns and the host kimberlite may be entirely gradational. In other examples a kimberlite matrix is present throughout the floating reef breccias (e.g. at Koffiefontein, Fig. 3.38) and, by increasing dispersion of fragments, such breccias grade into kimberlite containing scattered country rock fragments.

3.6.2 Size of floating reef inclusions

Reference has already been made to the large size of many floating reef bodies and specifically to the huge inclusion of Waterberg rocks in the Premier pipe. A second example of the enormous size of some floating reef bodies is the huge mass of Drakensberg basalt which occurs in the Voorspoed pipe near Kroonstad. The surface area of this down-rafted block is 5,95 ha and it occupies almost 50% of the total area of the pipe at surface (Fig. 3.45). This solid inclusion (brecciation is only evident locally, mainly at the margins) extends vertically to a depth of $\sim$300m (Fig. 3.46). Wagner (1914) interpreted this basalt as the remnants of a Drakensberg (Stormberg) basalt plug which had been partly cored out by later kimberlite that followed the same intrusive path. This interpretation is, however, disproved by:

(i) The presence of prominent horizontal or near-horizontal lava flows, including alternating amygdaloidal and massive bands, which would not be expected in a vertical volcanic plug.

(ii) Recently drilled (1979) exploratory boreholes which penetrated through the lava into underlying kimberlite.

Some of the floating reef inclusions which, prior to mining, were present in the Kimberley pipes were also extremely large. Both Wagner (1914) and Williams (1932) refer to several rock masses which extended vertically for 100-250m and had cross-sectional areas of several hundred m². Williams (op. cit.) also noted the presence of a brecciated column of sandstone, shales and dolerite in the Bultfontein pipe which extended from surface to a depth of 410m. Similarly large floating reef masses occur in the Finsch pipe and, to a lesser extent, at Koffiefontein (Figs. 3.37-3.38). Floating reef bodies which occur at relatively deep levels in the KIMFIK pipes are generally considerably smaller than the examples noted above and rarely have longest dimensions greater than 30-40m. Some substantial inclusions are, however, present (e.g. Figs. 3.39-3.40).

3.6.3 Distribution of floating reef inclusions

Several authors (e.g. du Toit, 1906; Wagner, 1914; Hawthorne, 1975) have stated that floating reef inclusions are confined to the upper parts of kimberlite pipes. Such inclusions do, however, persist to considerable depths in some of the KIMFIK pipes. In the Wesselton pipe brecciated shale masses (Figs. 3.39-3.40 and Plates 3.3e-f) occur almost 800m below the present surface. If the probable depth of post-kimberlite denudation is taken into account it is evident that these inclusions lie approximately 1,5-2km below the original Cretaceous kimberlite surface. Similarly at the Finsch pipe core drilling has intersected substantial floating reef masses up to 600m below the present land surface and such inclusions probably persist to greater depths. In the Kimberley pipe floating reef was located at a depth of 660m (Wagner, 1914) but similar inclusions have not been found in excess of 400-500m below surface in the other KIMFIK pipes. Of more significance than the actual depths to which floating reef masses persist is the fact that they are confined to parts of the pipes exhibiting diatreme zone characteristics and they occur exclusively within textural varieties of kimberlite which are referred to as tuffisitic kimberlite breccias (Chapter 2).

Hawthorne (1975) concluded that floating reef bodies are peripherally located in kimberlite pipes and both du Toit (1906) and Wagner (1914) noted that some floating reef masses occurred in direct contact with the walls of pipes. Concentration of floating reef near the margins of a pipe is well displayed at the Finsch occurrence (Fig. 3.37) and is a prominent feature of some other pipes. Such peripheral distribution is not, however, always apparent, especially in the deeper parts of the diatreme zones of pipes.

Subsidence of the floating reef bodies in the KIMFIK pipes, relative to their original stratigraphic positions, is invariably apparent. In many instances this subsidence is considerable. For example in the Finsch pipe floating reef composed of Stormberg lavas occurs at depths of 500-600m below the present surface. Since a considerable thickness of Karoo sediments must have been present in the area of the pipe at the time of its formation (Visser, 1972) it is evident that these lava inclusions must have subsided at least a 1000m and, possibly, considerably greater distances. Similarly in the Kimberley pipes considerable subsidence is indicated by the presence of Karoo shale floating reef as much as 600-700m below the base of the Karoo Sequence.

Surprisingly, in view of the overwhelming evidence for considerable subsidence, Williams (1932) quotes two cases, one in the De Beers pipe and the other at Dutoitspan, where he concludes that major floating reef bodies were moved bodily upwards in the pipes during kimberlite emplacement. In both cases he apparently identified floating reef masses of Ventersdorp lava at levels in the pipes where the walls consist of Dwyka shales. In both cases, however, the identifications are suspect. Hawthorne (1975) concluded that the occurrence in the De Beers pipe (now mined out) was lava of Drakensberg (Karoo) age. Williams himself noted that this floating reef inclusion consisted of "amygdaloidal diabase" together with Beaufort sandstone and Ecca shale, an association of rock types which supports Hawthorne's conclusion. The Dutoitspan occurrence, known as Mt. Ararat, was considered by Wagner (1914) to be a plug of "hardebank" kimberlite.

In a crude sense the original stratigraphic succession of the now largely eroded Karoo rocks is reflected by the distribution of the floating reef masses in the Kimberley pipes. Floating reef bodies derived from Upper Karoo rocks (Drakensberg and Beaufort Groups) appear to have been restricted to relatively high levels in the pipes. At current and recent mining depths (more than 400m below surface) floating reef consists mainly of shale, much of which is identical in character to the Dwyka shales still preserved in the area. Dolerite inclusions are also prominent and were probably derived from the thick and extensive dolerite sheets which were intruded into the Dwyka and Ecca rocks of the Kimberley-Koffiefontein region.

3.7 INCLUSIONS OF KIMBERLITE IN KIMBERLITE

Both Wagner (1914) and Williams (1932) noted the occurrence of inclusions of kimberlite in kimberlite within the KIMFIK pipes and reference has already been made to the D7 and D17 kimberlites at Dutoitspan which have been interpreted as very large inclusions. D7 has a maximum dimension not less than 30m across. In general kimberlite inclusions are, however, relatively small, ranging from microscopic fragments (e.g. Plates 6.4d, 6.4h and 7.6g) to blocks measuring 1-2m across (e.g. Plate 3.3g).

Williams (op. cit.) ascribed all kimberlite inclusions to autobrecciation; an interpretation he had little option but to adopt as he did not recognise multiple intrusive events within the pipes. A minority of the inclusions may indeed represent early-crystallizing parts of their host magmas. Such an origin is, however, manifestly not possible for the majority of kimberlite inclusions. These inclusions often differ substantially, in terms of mineralogy and texture, from each other, from their host kimberlite and from other known intrusions in the same pipe. The presence of such inclusions (autoliths) indicates that all the KIMFIK pipes have undergone a more complex intrusive history than is implied by the distribution of major intrusions (within the explored depth zones of the occurrences) described in Section 3.5.2.

CHAPTER 4

DYKES AND SILLS ASSOCIATED WITH THE PIPES

4.1 INTRODUCTION

Several hundred kimberlite dykes and a number of sills have been located in the central South African plateau. These dykes rarely exceed 1-2m in width but several occurrences are remarkably persistent along strike and Wagner (1914) records two instances of dykes (possibly dyke systems rather than individual dykes) which are several tens of kilometres long. Some of these dykes occur singly while others occur in well defined groups and frequently both dykes and sills are closely associated in time and space with kimberlite pipes. In several instances linearly aligned pipes have been shown to lie along kimberlite dykes and elsewhere it has been assumed that such pipes lie along concealed dykes which did not penetrate to surface (Wagner, 1914; Williams, 1932).

Numerous dykes and a few kimberlite sills are associated with the KIMFIK pipes although the abundance of these associated minor intrusions varies considerably at individual localities. On the basis of field relationships and ages of intrusion relative to the formation of the pipes, four groups of dykes can be recognized:

(i) Dykes which occur entirely within the wallrocks of the pipes and were emplaced prior to the formation of the pipes. Wagner (1914) referred to these kimberlites as antecedent dykes but henceforth in this thesis they are termed precursor dykes.

(ii) Contemporaneous dykes which represent off-shoots or extensions of major pipe intrusions into the wallrocks and were intruded during the period of pipe formation. Also included in this category are dykes which have not been directly correlated with pipe intrusions but where the field relations clearly indicate that they must have been emplaced between the onset and final stages of pipe formation.

(iii) Dykes which cut the kimberlite of the pipes and are entirely confined to the pipes. These dykes are referred to as internal dykes and conform broadly (together with those noted under (iv) below) to the consequent dykes of Wagner (1914).

(iv) Rare examples of dykes which cut the kimberlite of the pipes and the surrounding country rocks. These occurrences are henceforth referred to as cross-cutting dykes.

4.2 PRECURSOR DYKES

Precursor dykes are best exposed in the vicinity of the major Kimberley pipes where they have been intersected at various levels below surface in many underground tunnels. Because exposures are limited to mine tunnels it is not possible to establish the true abundance of such dykes. In the wallrocks of some pipes, for example De Beers (Fig. 4.1), as many as 20 or more precursor dykes have been mapped but similar intrusions are rare at some other localities (e.g. Bultfontein and Finsch).

The dykes are vertical or steeply dipping (>80°) bodies which vary considerably in width. The widest known precursor dykes occur at the Kamfersdam and Koffiefontein pipes. The Kamfersdam dyke (Fig. 4.2) has a maximum width at surface of 40m (although it narrows to <2m at a depth of 225m below surface). The "west dyke" at Koffiefontein is locally even wider (Fig. 3.3). Generally, however, precursor dykes do not exceed 2m in width and are commonly much narrower, ranging from about 0,5m to thin dykelets less than one centimetre wide.

Precursor dykes often occur singly but in some instances closely spaced, parallel dykes are concentrated in narrow zones (Fig. 4.3). Many dykes are tabular, uniform bodies but others pinch and swell considerably, both vertically and horizontally, over distances of a few metres. Some dykes bifurcate or pass along strike into a series of parallel, narrow dykes, veins and stringers (Fig. 4.3). Rare examples of anastomosing dykes have been located and frequently precursor dykes have minor branching apophyses which extend into the country rocks for limited distances (generally from a few centimetres to about 3m but occasionally in excess of 10m).

Due to limited exposures outside the immediate environs of the pipes the continuity of most of the precursor dykes has not been well established. Some dykes have only been located at single exposures in mine tunnels and their extent is thus completely unknown. A number of dykes are undoubtedly short lenticular bodies which pinch out in all directions and have strike lengths of 10m or less (Fig. 4.1). In other cases aligned exposures on opposite sides of pipes indicate that, prior to pipe formation, these dykes persisted laterally for at least 0,5km. In some instances dykes are offset in en echelon fashion.

Obviously some dykes may persist for distances considerably in excess of those quoted above. It is, however, likely that many precursor dykes are restricted in length and that these dykes are concentrated in the immediate vicinity of the pipes. There is substantial supporting evidence favouring these conclusions:

(i) A number of dykes which occur in close proximity to the pipes are known from underground exposures to have very limited strike lengths.

(ii) The main shafts at the Kimberley mines are located in the country rocks. At surface these shafts lie between 150 and 370m from the margins of the pipes and, due to the decrease in size of the pipes with depth, these distances increase at lower levels. These shafts are connected by haulages to the pipe workings at frequent intervals (at different levels) and it is noticeable that relatively few dykes have been intersected in these tunnels close to the shafts, i.e. fewer dykes at increasing distances from the pipes.

(iii) Two pipes, Bultfontein and Dutoitspan are linked by haulages developed on the 585 and 760m levels. No dykes have been intersected by either haulage in the intermediate area between the pipes.

(iv) Near surface (within and just below the Karoo succession) ring tunnels have been developed around the pipes for drainage purposes. The outermost tunnels are located at distances of between 200 and 450m from the margins of the pipes. Many of the dykes intersected at lower levels, closer to the pipes, do not intersect these ring tunnels.

Several precursor dykes in the vicinity of the Kimberley pipes are blind intrusions which do not reach the present surface level. Thus the apparent concentration of precursor dykes in the immediate vicinity of the individual pipes may, at least in part, reflect local intrusion, to relatively high levels, of dykes which are more persistent at depth. A similar situation has been postulated in Lesotho (Dawson, 1960).

It is apparent from underground exposures that the precursor dykes were intruded along pre-existing, vertical or steeply inclined, planar discontinuities (mainly joints). This is well displayed by the clear parallelism of many dykes to prominent joint sets, by pinching out of dykes along joints and by complex intrusion of dykes and veins which form reticulate networks of kimberlite along intersecting joint sets (Figure 4.3).

Detailed surveys to determine country rock joint directions have not been carried out but limited surveys at Wesselton and De Beers mines indicate the presence of complex joint systems (Figs. 3.15 and 3.16). Thus, although the precursor dykes in the vicinity of individual pipes commonly tend to follow one or more preferred directions, considerable variations in strike are evident and the same trends are not always followed at each pipe (c f. Figs. 4.1 and 4.4).

Although some variation in strike is evident, many of the precursor dykes near the De Beers pipe trend in a south-easterly direction (Fig. 4.1). The pipe itself is elongated in the same direction and, on a larger scale, this trend is reflected by the distribution of many kimberlite intrusions in the immediate vicinity of Kimberley (Fig. 1.1). A subsidiary NE dyke trend is also reflected (although less convincingly) by the distribution of some of the kimberlites in the area. The prominent SE dyke trend at the De Beers pipe also occurs at Wesselton and at Bultfontein (where few dykes are present). This trend is not, however, apparent at Dutoitspan where, as at Wesselton, the main precursor dyke trend is approximately E-W (Fig. 4.4). Furthermore the subsidiary NE trend of the De Beers dykes is not duplicated at Wesselton (where a similar but more easterly secondary trend occurs) or at Bultfontein and Dutoitspan. At the latter pipe the main dyke trend (approximately E-W) is accompanied by a subsidiary NNE trend and at Bultfontein a NNW trending dyke system parallels the direction of pipe elongation. Few precursor dykes are located in the wallrocks of the Finsch and Koffiefontein pipes.

There is no doubt that the precursor dykes were emplaced prior to the formation of the pipes with which they are associated. In every case where the contact relationships have been exposed the dykes are truncated by the pipes and, as previously noted, in some instances matching, aligned, dyke exposures have been located in the country rocks on the opposite sides of pipes. Furthermore, none of the precursor dykes cut any of the individual intrusions within the pipes, irrespective of the relative ages of the latter. There is thus little chance that some, or all, of these dykes were emplaced at an intermediate stage of pipe formation and that truncation only occurred during the later stages of pipe formation. The pre-pipe ages of precursor dyke intrusions are also indicated by truncation of dykes at root zone levels of the pipes. These root zones reflect the early stages of pipe formation (see Chapter 10).

Examples of precursor dykes cutting one another have not been located and it might be assumed that they were all emplaced contemporaneously. The precursor dykes in the vicinity of individual KIMFIK pipes do, however, exhibit considerable mineralogical and textural variations. It is therefore likely that more than one period of early (pre-pipe formation) kimberlite intrusion produced precursor dykes of several ages. In some instances more than one period of intrusion is indicated by sharp internal contacts within dykes.

4.3 CONTEMPORANEOUS DYKES

Compared with the abundance of precursor dykes in the vicinity of the Kimberley pipes, dykes which represent more-or-less linear off-shoots into the wallrocks, of major pipe intrusions, are rare. Excellent examples are, however, present at Wesselton and Dutoitspan. Such contemporaneous dykes wedge out rapidly away from the pipes, an indication of rapid decreases in country rock dilation away from the foci of intrusion within the pipes. Since these dykes are directly linked to, and form part of, major intrusions within the pipes it is clear that they must have been emplaced at different times. As might therefore be expected they vary considerably in character, both in regard to mineralogy and texture.

Contemporaneous dykes persist along strike for distances of at least 50m in some instances and may be several metres wide near their junction with the pipes proper (e.g. Fig. 4.6). In other instances they are insignificant features representing no more than vein-like protuberances which extend along joints into the wallrocks for as little as a metre or less.

A second type of contemporaneous dyke (noted in the introduction to this section) are dykes which cannot be linked directly to specific pipe intrusions. However, the form and location of these dykes clearly indicates that intrusion must have occurred during the formation of the pipes with which they are associated.

Such dykes do not appear to be abundant but an excellent example occurs near the De Beers pipe. This occurrence, known as the 'Eyebrow' dyke, has been described by Donaldson and Reid (in press): it is particularly well exposed on the 500m level of the mine. As is evident from Fig. 4.1 the Eyebrow dyke is located, together with several smaller, ancillary dykes, close to the northern boundary of the pipe. A noteworthy feature of the dyke is the manner in which it closely parallels the curving boundary of the pipe. This suggests that:

(i) Formation of the pipe must have been initiated prior to intrusion of the dyke.

(ii) The processes of pipe formation must have imposed some sort of concentric structural feature (jointing or minor faulting) on the wallrocks around the pipe.

With reference to (i) there is no doubt that the Eyebrow dyke does not post-date the entire period of pipe formation and infilling. This is indicated by truncation of the dyke by the pipe at relatively high levels.

There is generally very little evidence of the imposition of major joint systems on the country rocks during the formation of kimberlite pipes (Dawson, 1960, 1962, 1971). Detailed studies involving the measurement of many thousands of joints have been carried out around the Koffiefontein and Finsch diatreme zones during the course of open pit slope stability investigations. In neither case are there any clear-cut indications of radial or concentric joint patterns which could be ascribed directly to the formation of the pipes. It appears therefore that structural controls such as jointing (or perhaps minor faulting) which controlled the Eyebrow dyke may only be imposed at deeper levels (at root zone depths) and even under these conditions are neither common nor well developed.

It has been suggested (e.g. Wagner, 1914; Williams, 1932; Dawson, 1967a) that subsurface dykes are the feeder channels through which the kimberlite filling the pipes has passed. Such dykes, if present, would form a third type of contemporaneous dyke. No dyke which is clearly a feeder dyke has been exposed in any of the Kimberley mines with the probable exception of the Kimberley pipe. At this locality a broad ENE-trending dyke was interpreted by Wagner (1914) as being the feeder channel for the western part of the Kimberley pipe and the nearby St. Augustine's pipe. The two pipes are approximately 500m apart (Fig. 4.5). None of the other KIMFIK pipes has been mined to sufficient depth for feeder dykes to have been exposed. A feeder may, however, have been intersected in a borehole below the present bottom mining level at Bultfontein (825m below surface). As indicated in Figure 3.12 an exploratory borehole appears to have drilled along the strike of a feeder dyke. This dyke cannot extend upwards much above the 825m level as it has not been intersected in mine tunnels which cross its projected position at higher levels. There is no change in the character of the kimberlite throughout the length of borehole.

4.4 INTERNAL DYKES

A common feature of all the KIMFIK pipes is the presence of internal kimberlite dykes, i.e. dykes which are emplaced:

(i) Transgressively within the major intrusions of the pipes

(ii) Along the contacts between discrete pipe intrusions

(iii) Along the contacts between pipe intrusions and the wallrocks but which do not penetrate the wallrock

In some of the KIMFIK pipes such dykes are common (e.g. Figs. 4.6-4.7). In most cases no strike orientation pattern is evident although in the Finsch pipe a rather vague radial internal dyke disposition is evident (Fig. 4.7). These dykes are often of very limited extent, commonly displaying pronounced lateral lenticularity and having little vertical continuity. Mining operations have clearly revealed that some of these dykes are rootless. Tunnels traversing across the downward-projected positions of known dykes have failed to intersect them. Similarly internal dykes may also pinch out upwards.

Only a few internal dykes have strike lengths exceeding a few tens of metres although some dykes have been traced over vertical distances of more than 100 metres. Such dykes are often extremely irregular being markedly sinuous along strike and displaying pronounced changes in dip over short vertical distances (e.g. Fig. 4.7). In addition the width of such dykes often varies considerably, pronounced pinching and swelling being evident.

In some instances internal dykes which cut one or more major intrusions are truncated at the contacts with other pipe intrusions. Thus, while internal dykes are found cutting some of the youngest pipe intrusions, they do not all post-date the last stages of major intrusion within the pipes. In fact the distribution of internal dykes in relation to other intrusions in the KIMFIK pipes points undoubtedly to more than one, and possibly many, ages of intrusion.

It is possible that some internal dykes represent the waning stages of major intrusive events within the pipes while others possibly reflect precursor injection prior to major intrusions (Chapter 10).

As is the case with respect to precursor and contemporaneous dykes the internal dykes display considerable mineralogical and textural variation.

4.5 CROSS-CUTTING DYKES

Dykes which occur within the pipes and extend across pipe contacts into the adjacent wallrocks are extremely rare. Only two minor examples have been located among the KIMFIK pipes, one at Dutoitspan and the other at Wesselton.

4.6 KIMBERLITE SILLS

Wagner (1914) recorded the occurrence of several kimberlite sills at various southern African localities and Hawthorne (1968) drew attention to a major sill complex on the farm Benfontein on the outskirts of Kimberley. Except in a general sense, in that they are almost certainly of similar age and relate to the same regional magmatic event, the Benfontein sills do not appear to be directly associated with any of the major Kimberley pipes. The area of maximum development of these sills suggests that they were emplaced via a discrete and as yet undiscovered feeder channel (or channels).

In addition to the Benfontein complex the occurrence of a second major sill in the Kimberley area, somewhat lopolithic in character (Hawthorne, op. cit.), has been known for some time. This occurrence, known as the Wesselton Floors sill, is located approximately 1,5km south-east of the Dutoitspan pipe and, as in the case of Benfontein, the nature and position of the feeder is not known.

More recently a major complex of kimberlite sills (and associated dykes) has been located in drainage galleries developed around the Wesselton pipe (Fig. 4.8).

The sills at Wesselton are of particular interest for several reasons:

(i) They are directly related to, and were fed by, precursor dykes in the immediate vicinity of the pipe.

(ii) The sills were cut by the pipe (before mining) and therefore predate the latter.

(iii) At least two and possibly more stages of sill emplacement occurred. This is indicated by field relationships (sills and associated dykes cutting other sills) and by the occurrence of two distinct petrological varieties (Hill, 1977).

(iv) The sill complex is extensive. Prior to the formation of the pipe and the consequent coring out and removal of parts of the complex the sills must have extended over a minimum area of at least 70 ha (Fig. 4.8). This minimum area is indicated by the distribution of sill exposures in the drainage tunnels around the pipe; the true lateral extent of the sills is almost certainly considerably greater.

(v) For the most part the kimberlite sills lie at or just below the base of a major Karoo dolerite sill and in a few areas kimberlite sills and veins occur within the lower part of the dolerite. It is clear that, as at other localities in the Kimberley area (Hawthorne, 1968), the emplacement of the kimberlite sills can be directly related to the occurrence of dolerite sills. The dolerite formed a barrier against upward penetration of kimberlite magma thereby imposing lateral spreading.

(vi) Locally individual sills reach 5m in thickness (Hill, 1977) but the vertical extent of sills is generally $<1{,}0$m. In some areas individual sills range between a few centimetres and 0,5m in thickness over horizontal distances of several hundred metres. Such features indicate the original presence of a highly mobile magma and baking of the adjacent shales testify to relatively high intrusion temperatures.

Except for minor occurrences of very limited extent in the Karoo rocks adjacent to De Beers Mine no other kimberlite sills have been located in the immediate vicinity of the KIMFIK pipes. It should, however, be borne in mind that the discovery of kimberlite sills requires somewhat fortuitous circumstances; either erosion precisely to the sill horizons or, in the mines, the development of tunnels (or shafts) at the right levels and in the right areas for sills to be intersected. It is therefore reasonable to suppose that kimberlite sills are more abundant in the Kimberley area than is indicated by the number of known occurrences. This supposition is supported by the recent (1979) discovery of another group of sills near Warrenton (the Mayeng sills).

4.7 INTERNAL FEATURES OF DYKES AND SILLS

The dykes and sills associated with the KIMFIK pipes display a variety of megascopic structural and textural features.

Generally contacts with the wallrocks are sharp and alteration of the intruded rocks is absent or limited. Some discoloration over distances of a few centimetres may be evident where the country rocks comprise Ventersdorp lava, basement gneisses or kimberlite. In some instances intrusion of the kimberlites has resulted in baking of Karoo shales. In these cases the extent of metamorphism is usually limited to a few centimetres from the contacts. Similar features have been recorded by Dawson and Hawthorne (1970).

A feature of many dykes (and some sills) is extensive alteration and decomposition of the margins of the intrusions. In other cases, where the intrusions are thin (0,5m or less), alteration may be complete. Decomposition of this nature is a fairly universal feature of intrusions in the near-surface zones where the country rocks are Karoo shales and dolerites. Such decomposition is, however, also evident at depths well in excess of 500m. It is ascribed to hydration resulting from meteoric water percolating along the contacts of the intrusions.

Kimberlite dykes are often banded, commonly parallel to the dyke contacts. This banding may due to multiple injections along the same channelway in which case sharp internal contacts are usually evident (Scott, 1977; Donaldson and Reid, in press).

In other instances banding within dykes is the result of flowage differentiation (Bhattacharji and Smith, 1964). An example of such differentiation is shown in Plate 5.10d.

In some instances, particularly in regard to some internal dykes, a banded appearance is imposed by the presence of abundant, parallel calcite veins (Plate 3.4d). These veins may be concentrated in the marginal or inner zones of the dykes and are usually parallel to the walls.

Parallel or sub-parallel orientation of elongate macrocrysts and, less frequently, xenoliths is a feature of some of the minor intrusions associated with the KIMFIK pipes. The direction of orientation is normally parallel or nearly parallel to the walls of the intrusions. Similarly distinct flow lines reflecting minor differences in texture, mineral proportions or grain size are sometimes evident parallel to the walls of intrusions. In some intrusions complex macrocryst (and xenolith) distribution and orientation patterns and complex flow lines occur. Such features are considered indicative of the turbulent emplacement of highly mobile magmas.

The various flow structures which have been noted in the minor intrusions should not be overemphasized since many of the dykes are devoid of such features. They are often macroporphyritic rocks without any banding or directional fabric or they occur (less frequently) as massive, featureless rocks devoid of the macrocrysts which are normally a characteristic feature of kimberlites.

A noteworthy feature of some minor intrusions is the occurrence of segregations in the form of:

(i) Regular (in some instances spherical) and irregular ocelli (Plate 5.10e). Some ocelli are elongate bodies oriented parallel to the contacts of the intrusion as a consequence of flow.

(ii) Fairly regular, often spherical segregations (Fig. 5.10f) of the type which have been termed "nucleated autoliths" (Danchin et al., 1975). These bodies are discussed further in Chapter 7.

(iii) The presence of elongate, in some cases branching, 'migration amygdales' and minor diapiric structures, such as those described by Donaldson and Reid (in press) from a dyke near the De Beers pipe and by Dawson and Hawthorne (1973) from the Benfontein sills. The Wesselton sills display many of the structures recorded from the Benfontein complex by Dawson and Hawthorne (1973). These include prominent layering and magmatic sedimentation features.

4.8 PRECAMBRIAN ULTRAMAFIC DYKES

During the course of this study some attention was paid to unusual ultramafic dykes with kimberlite affinities in the Kimberley area. Five of these dykes have been located. One is cut by the Wesselton pipe and the other four dykes are situated close to the De Beers pipe (Fig. 4.9). These dykes have been described in considerable detail by Clement et al., 1979. Some features of these occurrences are summarized below.

Of considerable interest is the occurrence within these dykes of:

(i) Magnesian ilmenite commonly containing >10 wt.% MgO and 1,5 wt.% Cr_2O_3.

(ii) A variety of spinels including high-chrome chromites (up to 64 wt.% Cr_2O_3).

(iii) Titanium-poor pyrope garnets generally containing moderate Cr_2O_3 (3,5-5 wt.%) and CaO (4,5-6 wt.%).

(iv) Very rare, very small diamonds.

The minerals listed above are typically found in kimberlites (or in ultramafic nodules which occur as inclusions in kimberlites). In all other respects these dykes do not display close affinities with kimberlite and they cannot be petrologically classified as kimberlites.

Field relations clearly indicate that the dykes pre-date the Cretaceous kimberlites of the area and radiometric dating by the Rb-Sr method on one whole rock and four fine-grained mica concentrates from the dyke near Wesselton indicate a Precambrian age (1910 ±60 m.y.).

The typical kimberlite minerals found in these intrusions particularly diamond, indicate that the dykes must have originated as a deep-seated (upper mantle) melting event which preceded the much later intrusion of Cretaceous kimberlites.

CHAPTER 5

MATRIX MINERALOGY

5.1 INTRODUCTION

Detailed reviews of the mineralogy of kimberlites have been presented by other authors, one of the most recent and comprehensive being that by Dawson (1980). In this chapter (and Chapter 6) attention is therefore restricted to additional or more specific aspects of a limited number of minerals, particularly minerals that have significant paragenetic implications. A broad split has been made between matrix minerals, macrocrysts and megacrysts, following the arguments presented in Chapter 2. Some overlap in discussion is, however, unavoidable since certain minerals (e.g. olivine, phlogopite and ilmenite) may occur in all three modes and it is not always possible to draw sharp distinctions between the different populations.

5.2 OLIVINE

5.2.1 The number of olivine populations in kimberlites

Olivine is a ubiquitous and abundant mineral in kimberlites. It is commonly altered to some degree and, particularly in highly micaceous kimberlites, may be entirely replaced by other minerals (e.g. Skinner and Clement, 1979). Typically kimberlitic olivine is highly magnesian (Sections 6.2.2 and 7.7.2) but it displays considerable chemical heterogeneity, even on the scale of a single thin section (Boyd and Clement, 1977). In addition the olivine grains in kimberlites vary considerably in form and size.

The chemical and morphological variability of olivine in kimberlites is such that more than one mode of origin has been proposed and it has been suggested that the occurrence of two olivine populations is a diagnostic feature of kimberlites (e.g. Wagner, 1914; Dawson, 1962, 1971; Mitchell, 1970; Clement et al., 1977; Skinner and Clement, 1979). One of these populations is composed of relatively large (usually 0,5-5mm) anhedral macrocrysts and the other of smaller, commonly euhedral and subhedral phenocrysts. The primary nature of the latter population is widely recognized and is indicated by the euhedral morphology, ubiquitous presence and general abundance of these grains. The provenance of the macrocrysts is discussed in section 6.2.

In addition to the foregoing two populations a third group of olivines may be present in some kimberlites. This population is composed of large (commonly 2-3cm or larger) megacrysts (Nixon and Boyd, 1973a).

Some authors (e.g. Williams, 1932; Mitchell, 1979a; Elthon and Ridley, 1979; and Pasteris, 1980a) have subdivided the idiomorphic olivine crystals into 'phenocryst' and 'groundmass' or 'matrix' populations. This distinction is arbitrary and is not justified (see section 5.2.2).

As noted in Chapter 2 E.M.W. Skinner (personal communication) has suggested that a true groundmass generation of olivine is present in some kimberlites. This suggestion is based mainly on the presence in one of the Bellsbank dyke kimberlites of extremely fine-grained ($\sim$0,02mm) altered grains with preserved olivine(?) morphology and on the optical identification of similar-sized but better preserved grains in one of the Newlands kimberlites (Barkly West District). Analytical data is, however, required to confirm the identification of these very small crystals.

5.2.2 Size and shape of olivine phenocrysts

Primary (euhedral or subhedral) olivines in the KIMFIK pipes display considerable variations in size but the vast majority do not exceed 0,5mm in diameter (e.g. Plates 5.1a and 5.1b). Their small size is well illustrated by the histograms in Figures 5.1 and 5.2 which reflect size analyses of olivines in single thin sections from the DB2 kimberlite in the De Beers pipe and the D2 intrusion at Dutoitspan. In both cases between 90 and 95% of the primary euhedral or subhedral olivines occur as microphenocrysts measuring 0,5mm or less. In the De Beers kimberlite 70% of these microphenocrysts do not exceed 0,2mm while the corresponding figure for D2 is somewhat lower at $\sim$50%. Relatively large phenocrysts do occur in the KIMFIK pipes and Boyd and Clement (1977) recorded a euhedral crystal measuring 5,5mm in one thin section from the DB2 kimberlite. However, the rarity of large phenocrysts is emphasized by the fact that, in another slide of DB2 kimberlite, only one of 283 measured phenocrysts exceeded 1mm in size and none exceeded 2mm. None out of 213 measured phenocrysts in a slide of D2 kimberlite exceeds 1mm in size.

The histograms in Figures 5.1 and 5.2 do not provide any indications of breaks in phenocryst size ranges at 0,5-0,6mm, as suggested by Pasteris (1980a) for De Beers Mine olivines, or at 1mm, the arbitrary cut-off used by Mitchell (1979a) to separate phenocryst and groundmass olivines. The histograms thus provide no evidence of discontinuous olivine precipitation as suggested by Pasteris (1980a). On the contrary the size distribution could be adequately explained by a continuous crystallization model in which initial slow growth at relatively few nucleation sites proceeded until, at the final stages of olivine crystallization, rapid nucleation at many sites resulted in the growth of abundant small crystals.

A considerable degree of idiomorphism is a consistent feature of the phenocrysts and in some instances many crystals are perfectly euhedral (e.g. Plate 5.1b). The morphology of the phenocrysts may, however, be obscured by varying degrees of alteration (e.g. Plates 5.2a-h) and corrosion. Extensive morphological modification occurs in certain phlogopite kimberlites but complete destruction of the original euhedral character of the phenocrysts is very largely restricted to kimberlites which have been strongly metasomatised or highly weathered. Where alteration is essentially deuteric in character the original olivine grain boundaries are usually reasonably well preserved.

5.2.3 Alteration of olivine

The frequently complex nature of the alteration of olivines in kimberlites has been noted by numerous authors (e.g. Wagner, 1914; Williams, 1932; Dawson, 1962, 1971, 1980; Frick, 1970; Mitchell, 1970, 1978; Kresten, 1973; Fairbairn and Robertson, 1966; Skinner and Clement, 1979; Skinner and Scott, 1979; Pasteris, 1980a). Among the many alteration and replacement products reported are: serpentine (including antigorite, lizardite, chrysotile and serpophite), brucite, bowlingite, goethite, iddingsite, magnetite, millerite, heazlewoodite, rutile, sphene, chlorite (several varieties), perovskite, talc , calcite, dolomite, phlogopite, monticellite, hydrogrossular, quartz, baryte, pyrite and a wide variety of clay minerals. Many of these minerals have not, however, been found pseudomorphing olivine in the KIMFIK pipes and the following descriptions and discussions are in any event limited to the more common of the alteration products.

The olivine phenocryst and macrocryst populations in the KIMFIK pipes are commonly altered to some degree and in several discrete intrusions no fresh olivine has been observed. In other (rare) instances many individual grains appear to be entirely unaltered and the general level of alteration is very limited (Plates 5.1a and 5.1b). Much of the alteration is deuteric in character; it has usually occurred in pseudomorphous fashion and reflects the reactive, volatile-rich nature of post-olivine residual kimberlite liquids.

In addition to variations among intrusions the extent of alteration is often extremely variable within a single kimberlite or even within a single thin section (Plate 5.1c). Commonly co-existing phenocrysts and macrocrysts in individual kimberlites are altered in similar fashion although, because of their smaller size, phenocrysts are generally more extensively altered. Thus in many kimberlites the cores of relatively large macrocrysts are preserved while smaller grains are entirely replaced by alteration products.

Examination of hundreds of thin sections from the KIMFIK pipes has shown that the alteration of olivine may be extremely complex. In some cases it involves the successive development of several new minerals within the margins of the original grains. In other cases several stages of alteration and replacement have occurred during the formation of specific deuteric minerals. In some instances additional complexities reflect the effects of weathering or metasomatism superimposed on deuteric alteration.

The mineralogical study of olivine carried out during the course of this investigation has been based mainly on optical methods of examination. Much more information is required, particularly with respect to mineral chemistry, before the nature of olivine alteration will be fully understood. The descriptions and discussion which follow serve, however, to indicate the complex nature of this alteration and are applicable to both phenocrysts and macrocrysts.

Serpentinization is by far the most common form of alteration of olivine in the KIMFIK pipes. It should, however, be noted, as Dawson (1980) has pointed out, that little systematic work has been carried out on the sepentine-group species in kimberlites and 'serpentine' is used in this thesis in a general sense to describe a variety of minerals considered to belong to the serpentine group.

Serpentinization of olivine in the KIMFIK pipes is in some instances accompanied by extensive development of Fe in the form of magnetite (Plate 5.1d). This is, however, not always the case and some serpentinized olivines in the KIMFIK pipes contain little or no visible magnetite (Plate 5.1c). It is possible that the serpentines in the latter examples are unusually rich in iron, such as those reported from West Greenland (Emeleus and Andrews, 1975) and the Elwin Bay kimberlite in Canada (Mitchell, 1978). Mitchell (op. cit.) considers that such serpentines may reflect serpentinization at relatively low oxygen fugacities.

In some instances, particularly where alteration is limited, only one serpentinization event is evident. Elsewhere several stages of serpentinization appear to have occurred. This is indicated by discrete zones of optically different serpentines, often separated by sharp contacts, within individual olivine pseudomorphs. These zones are distinguished by differences in colour, texture and birefringence (e.g. Plates 5.1e and 5.1f) and by the manner in which the individual zones have, themselves, been altered.

It is possible that some of the zonal alteration referred to above reflects progressive changes in serpentine chemistry and texture, as a function of changing conditions, during a single, continuous, serpentinization event. In many cases, however, the distinct differences in character of discrete serpentine zones, the sharp boundaries between them and the highly selective subsequent modification of individual zones, are features which suggest a hiatus between the development of each zone.

It might be argued that each zone of optically distinctive serpentine reflects continuous serpentinization of originally zoned olivine. Such an explanation is refuted by

(i) The absence, where fresh olivine is preserved, of any optical evidence of major zoning affecting large areas of the grains.

(ii) A study by Boyd and Clement (1977) of olivines from the DB2 kimberlite at De Beers Mine. This indicated that zoning is limited to extremely narrow marginal rims of the olivines and that it reflects reaction between crystals and liquid: it is not a primary growth feature. Apart from these very narrow rims individual olivines in the DB2 kimberlite are chemically homogeneous (although $\frac{Mg}{Mg+Fe}$ ratios vary considerably from grain to grain).

(iii) The occurrence of similar serpentine zones in co-existing olivines irrespective of whether they are anhedral macrocrysts or euhedral phenocrysts.

In many cases it appears that the different zones represent successive modifications of pre-existing serpentine rather than successive increases in the physical extent of serpentinization (leaving less and less relict olivine). However, the latter situation probably also occurs. Thus a later stage of serpentinization may replace or alter previous serpentines and partly or wholly replace previously preserved olivine kernels.

Several of the complex serpentinization features described above are illustrated in Figures 5.3-5.6 and Plates 5.1c-f. An incipient stage of serpentinization is illustrated in Figure 5.3a. In the olivine illustrated serpentine occurs only along cracks in the grain and as a narrow marginal rim. Only one variety of serpentine is present. Such serpentine commonly has a homogeneous, massive, pale to dark grey appearance between crossed nicols or it may be fibrous in character. The fibres are usually aligned at right angles to the grain margins or to the cracks.

In other cases, even when the total extent of serpentinization is limited, several zones of serpentine are evident. In Figure 5.3b part of a crack within an olivine phenocryst is illustrated. This crack contains three matching, optically distinct serpentine zones, on either side of a median line. The latter is marked by a concentration of minute opaque oxide granules. 'Ghost' outlines of former olivine occur as needle-like projections (from residual olivine) into the serpentines of the crack. The needles apparently reflect areas where serpentinization has not proceeded to the extent that all trace of the original olivine has been destroyed.

Figure 5.3c shows part of an olivine phenocryst which has been subjected to two discrete stages of serpentinization. The Sp1 serpentine (within cracks) has been partly replaced and is truncated, near the margin of the grain, by the serpentine annotated Sp2. The latter forms a marginal rim around the olivine phenocryst. The contact between Sp1 and Sp2 is sharp.

Fig. 5.3d is an example of a partially altered olivine grain in which 3 types of serpentine occur. Initial serpentine (Sp1) has been followed by the fairly extensive development of later Sp2 serpentine. Both Sp1 and Sp2 appear to have been partly replaced by a later massive variety (Sp3) which occurs as a narrow marginal rim.

Figure 5.3f illustrates a more advanced degree of serpentinization. Considerable olivine is still preserved but two fairly broad rims of optically contrasting serpentine (annotated Sp and Sr) are present.

A similar situation is illustrated in Plate 5.1e. In Figure 5.3f the inner serpentine (Sp) forms embayments in the residual olivine core. Most of these embayments are angular and some of their edges are oriented parallel to crystallographic directions. In this example it therefore appears that Sp replaces olivine directly, that Sp is hence the first serpentine formed and that serpentinization is strongly controlled by the olivine structure. The structural directions appear to be preserved within the generally platy Sp serpentine. This is indicated by pyramidal features within Sp which are parallel to some of the phenocryst faces. This feature is shown in Figure 5.3f and is also illustrated in other altered olivine phenocrysts (Figures 5.3j and 5.5f-i). The outer rim of serpentine (Sr) in Figure 5.3f apparently represents a later modification or replacement of the Sp variety.

Similar Sp/Sr-type serpentine relationships are illustrated in Figures 5.3e and 5.3g-i. The phenocryst illustrated in Figure 5.3e has been entirely serpentinized. Sp occupies the central part of the grain and Sr forms a broad surrounding rim. The boundaries of the inner serpentine (Sp) are parallel to crystal faces. It is likely that intially the entire phenocryst consist of Sp-type serpentine but that subsequently the formation of the Sr-type proceeded inwards, by replacement of Sp. The replacement process appears to be controlled by 'inherited' olivine structure preserved in Sp-type serpentine. Clearer evidence for this conclusion is presented in Figure 5.3g. In this case similar parallelism between the relict, central Sp area and the crystal faces is evident. However, an entirely symmetrical pattern is destroyed by a large embayment of the Sr rim-type serpentine in Sp. This embayment is itself angular in character and its edges are also parallel to crystal faces. This indicates that:

(i) Sr is a modification of, or replaces, Sp.

(ii) Despite initial Sp serpentinization 'inherited' internal olivine structural directions control the subsequent development of Sr-type serpentine.

Similar features are evident in Figures 5.3h and i.

Figures 5.3j and k and Plate 5.1f illustrate more complex examples of serpentinization. In Fig. 5.3j a small kernel of preserved olivine is surrounded by platy serpentine (annotated Sp). This serpentine grades into another variety (termed Spr) which has a finely speckled appearance between crossed nicols. Spr is in turn surrounded by a third zone of serpentine (Sr). The former is separated from the latter by a sharp contact. The Sr serpentine has relatively high birefringence and contrasts strongly with Sgr, a fourth serpentine zone at the edge of the phenocryst. Sgr is almost isotropic and is optically identical with serpentine (Sg) which occurs in the groundmass of the kimberlite (Figure 5.3k). Consequently the boundaries of phenocrysts with marginal Sg-type serpentine tend to be blurred, somewhat irregular and difficult to discern. They are, however, in part defined by surrounding groundmass minerals, e.g. Figure 5.3k and Plate 5.2h. In Figure 5.3j Spr serpentine is sharply demarcated from Sr but its relationship with Sp is gradational and Spr and Sp may relate to a single period of serpentinization. In Fig. 5.3k no Spr serpentine is evident but otherwise the situation is similar to that illustrated in 5.3j. In the former case a little calcite occurs irregularly along the interface between Sp and Sr serpentines (Figure 5.3k).

Figures 5.3e-k are based on phenocrysts in a single slide from the D14 intrusion at Dutoitspan (Figs. 3.31-3.32). These sketches indicate a probable serpentine paragenesis within D14 olivine of:

$$\mathrm{Ol} \rightarrow \mathrm{Sp} \rightarrow \mathrm{Spr} \rightarrow \mathrm{Sr} \rightarrow \genfrac{}{}{0pt}{}{\mathrm{Sgr}}{\mathrm{Sg}}$$

This sequence is consistent with changes in the character of serpentine pseudomorphs with decreasing size. The main changes are:

(i) Less and less and finally no relict olivine preserved.

(ii) A decrease in the volumetric amount of Sp serpentine relative to other varieties.

(iii) The disappearance of all Sp serpentine but the preservation of Spr in the central part of the pseudomorphs.

(iv) In still smaller grains neither Sp or Spr serpentines are preserved. Pseudomorphs consist only of Sr and Sgr serpentines.

(v) Extremely small pseudomorphs consist only of Sgr serpentine and are difficult to distinguish when set in areas consisting mainly of Sg (groundmass) serpentine.

As noted previously and as implied by (v) above Sgr and Sg serpentines are optically similar and it is likely that they are contemporaneous. This conclusion is reinforced by the absence of sharp contacts between them. They represent the final stage of serpentine formation in the D14 kimberlite, i.e. the crystallization of primary serpentine are preserved. Abundant opaque oxides, apparently resulting from the replacement of Spr by calcite, are present in the central area of the grain. In Figure 5.4h a similar situation is evident but here calcite has in part replaced both the Spr and Sp serpentine. Both serpentine zones have also been chloritized.

Although an initial stage of serpentinization appears almost invariably to be a prerequisite for subsequent carbonatization later stages of serpentinization may replace calcite in olivine pseudomorphs. This is illustrated in Plate 5.1h and in Figures 5.5a-e. The latter are sketches of altered olivines from the DB3 kimberlite in the De Beers pipe.

In Figure 5.5a an irregular calcite kernel, within a euhedral pseudomorph, is surrounded and penetrated by serpentine. The serpentine adjacent to the calcite has an unusual, extremely fine-grained, speckled appearance between crossed nicols. The speckled appearance is due to an intimate association of high and low birefringence particles which cannot be resolved under the microscope. It is inferred that the speckled area represents incompletely serpentinized calcite. A similar situation is illustrated in Figure 5.5b. In this case calcite appears to have been replaced by two stages of serpentine; marginal material and a later vein which cuts both calcite and earlier serpentine.

In Figure 5.5c two adjacent calcite grains (Cc1 and Cc2) have been rendered higly irregular by extensive replacement by serpentine (the boundary between the two calcite grains is indicated by a dashed line). Similar replacement of calcite is shown in Figure 5.5d. In Figure 5.5e the relative positions are shown of 5 irregular shreds of calcite within the serpentine of an altered olivine phenocryst. All 5 shreds extinguish simultaneously between crossed nicols. This strongly suggests that they originally formed part of a single calcite grain which was subsequently extensively serpentinized (see also Plate 5.1h).

Phlogopitization of serpentinized olivine is a common feature of some KIMFIK kimberlites and is well illustrated by olivines in the DB3 intrusion of the De Beers pipe (Figures 5.6a-f) and in Plates 5.2a-h.

Figure 5.6a shows a relatively large (>0,5mm) euhedral olivine phenocryt. In this phenocryst a central core of unaltered olivine is traversed by thin serpentine veins. In addition patchy, irregular, serpentine zones occur within the core and it is surrounded by a prominent rim of serpentine. Phlogopite occurs as tiny, almost colourless, laths near the margin of the crystal. These tiny laths are oriented parallel to the euhedral edges of the phenocryst. This suggests that the crystallographic structure of the olivine has been 'inherited' by the rimming serpentine and has influenced the orientation of the subsequent phlogopite laths. The phlogopite laths are never in contact directly with olivine and their distribution suggests that they are a reaction product of serpentine (after olivine) and enclosing magma. A similar relationship has been noted by Pasteris (1980a) in olivines from De Beers Mine.

Figure 5.6b illustrates a more advanced stage of phlogopitization. In this example alteration is very severe. Phlogopite, calcite and serpentine each occupy about one third of the area of the phenocryst. The edges of the grain are somewhat obscure but are reasonably defined by fringing opaque oxides, perovskite and other groundmass minerals (only partly illustrated - a similar situation is shown in Plate 5.2h). In Figure 5.6b the centrally located serpentine is clouded by opaque material.

Figures 5.6c-e show altered olivines which consist only of phlogopite and calcite. These phenocrysts are devoid of serpentine but the textural and mineralogical relationships evident in other grains suggest that in these examples alteration is further advanced and calcite and phlogopite have entirely replaced pre-existing serpentine

Replacement relationships between calcite and phlogopite are not evident in Figure 5.6c but the former mineral has partly replaced the latter in Figures 5.6d and e. Figure 5.6d also ilustrates the manner in which phlogopite may straddle the boundaries of phenocrysts. This feature is particularly well illustrated in Plates 5.2c and 5.2d. It is interesting to note that the portion of the lath projecting beyond the original olivine grain boundary in Plate 5.2d is crowded with inclusions of earlier-crystallizing groundmass components, as are phlogopite laths in the groundmass itself. It can therefore be concluded that deuteric phlogopitization of the altered olivine was contemporaneous or overlapped, with the crystallization of groundmass phlogopite in this kimberlite (the DB3 intrusion in the De Beers pipe).

Figure 5.6f illustrates an olivine phenocryst which contains a little residual fresh material but has been serpentinized, phlogopitized and carbonatized. Serpentine occurs adjacent to the residual olivine and forms small embayments within it. This is indicative of direct serpentinization of olivine. Locally (top-right) phlogopite straddles the crystal boundary and appears clearly to post-date the serpentine. Calcite forms embayments in (and post-dates) both serpentine and phlogopite. More extensive replacement of olivine by these three minerals is shown in Plates 5.2e-g.

A fairly common feature of the serpentine replacing olivine in the KIMFIK pipes is some degree of subsequent chloritization (e.g. Plate 5.3a). Generally the extent of the chlorite (pleochroic in shades of green, blue and yellow-green) is fairly limited (except in highly weathered kimberlites). Preferential chloritization of specific serpentines in zonally serpentinized olivines is often apparent (cf. Figures 5.4g-h and 5.5f-j and Plate 5.3a).

As indicated in the foregoing discussion many of the ramifications of olivine alteration are centred around varying degrees (and, in some instances, repetitive stages) of deuteric serpentinization, phlogopitization and carbonatization. The resulting pseudomorphs are, in addition, quite frequently modified by a degree of chloritization.

It is important to emphasize that the first stage of olivine alteration involves some degree of serpentinization. Other deuteric alteration products reflect reaction between serpentine (s) and evolving kimberlite liquids rather than reaction between olivine and such liquids. This conclusion is supported by optical observations of many hundreds of thin sections of kimberlites, from all over the world, by the author and/or his colleaque E.M.W. Skinner. The only possible exception known to the author is shown in Plate 5.1g but E.M.W. Skinner (personal communication) has suggested that in some highly micaceous kimberlites direct conversion of olivine to talc may have occurred. In neither of the examples quoted is the evidence unambiguous.

Dawson (1980) has noted that early serpentine pseudomorphs may be replaced by various combinations of relatively low temperature minerals. However, of particular interest in the KIMFIK pipes, are observations that serpentinized olivine may also be replaced by monticellite and diopside (Plates 5.3b-e). These observations are not entirely without precedent as Williams (1932) tentatively identified diopside after serpentinized olivine in kimberlite from the Premier pipe. In addition monticellite rims around serpentinized olivine were reported by Verhoogen (1938) from kimberlite in Zaire. In view of the paragenetic implications arising from these observations (Chapter 7) it is surprising that they have not received greater attention.

A number of the minerals listed in the introduction to this section are, for the most part, not regarded as deuteric alteration products of olivine. These minerals include brucite, bowlingite, goethite, iddingsite, quartz, baryte, pyrite and clay minerals such as montmorillonite, nontronite, kaolinite, illite, stevensite, vermiculite, sepiolite, halloysite and saponite. The presence of these minerals in pseudomorphs may be due to:

(i) Secondary processes (mainly hydration) related to the weathering of kimberlite

(ii) Metasomatism due to the introduction of fluids from 'outside' or abstracted from incorporated xenoliths.

(iii) Extensive contamination of residual kimberlite liquids by meteoric water.

It is difficult to draw sharp boundaries between deuteric alteration, metasomatism and weathering but evidence in favour of a non-deutric origin for the minerals listed above includes:

(i) They are generally only abundant in olivine pseudomorphs in kimberlite samples derived from surface or near-surface levels. Some of these minerals have only been recorded from samples which are known to have been extensively weathered (e.g. Fairbairn and Robertson, 1966; Ruotsala, 1975).

These minerals (apart from some chlorite) are very rare constituents of olivine pseudomorphs in samples taken at depth from hypabyssal-facies kimberlites in the KIMFIK pipes. Clay minerals are commonly only abundant in altered contact zones (Plate 5.3f). These zones are often soft and earthy and reflect the movement of groundwater along pipe contacts. The general paucity of the minerals listed (mainly phyllosilicates) at depth in the KIMFIK pipes cannot be ascribed to the kimberlites concerned being specific mineralogical varieties not amenable to such alteration. As noted in Chapter 7 these hypabyssal-facies kimberlites exhibit extensive petrographic diversity.

(ii) Clayey or extensively chloritized pseudomorphs of olivine are common in tuffisitic kimberlite breccias and such pseudomorphs persist to considerable depths (>1 000m). This is consistent with the secondary modes of origin suggested for these pseudomorphs as TKB's: (i) Are highly contaminated by country rock xenoliths (Plates 7.1a-d and 7.2a-d).

(ii) Are the textural varieties (pelletal, segregationary or crystallinoclastic - see Chapters 2 and 7) most prone to penetration by circulating groundwater.

(iii) Are the kimberlites most likely (in view of their mode of origin) to have incorporated abundant meteoric water during emplacement (Chapter 10).

With respect to some of the phyllosilicates listed, the conclusion of a generally non-deuteric origin is at odds with the views of some authors. For example, Dawson (1980), based on work by Kresten (1973), recognizes 3 stages of deuteric alteration affecting olivine, pyroxene and phlogopite. The 3 stages are the talc/serpentine stage, the chlorite/vermiculite stage and the saponite stage. This interpretation rests heavily on Kresten's conclusion that samples of Lesotho kimberlites on which he worked were 'altered' but not 'weathered'. Kresten (op. cit.) remarked specifically that such was the case with respect to samples of the K6 kimberlite from the Letseng-la-Terae pipe. However, Skinner (personal communication) has concluded, after a detailed study involving the examination of over 400 thin sections that the upper part of this pipe is substantially weathered. Furthermore the K6 kimberlite is a tuffisitic kimberlite breccia which contains abundant (up to 50 vol.%) country rock inclusions. The contaminated nature of the K6 kimberlite has also been noted by Kruger (1978) and Lock (1980).

It should be noted that processes such as serpentinization, and carbonatization of olivine are not always deuteric in character. This is clearly shown in Plate 5.1c where it is evident that extensive serpentinization of olivines is restricted to a small area surrounding a highly altered country rock xenolith. This distribution of highly serpentinized olivines clearly reflects a situation where metasomatism has occurred. 'Kimberlitization' (Chapter 7) of the xenolith appears to have been accompanied by the abstraction of material (mainly H_2O + SiO_2) from the inclusion which has resulted in serpentinization of nearby olivine. This situation is commonly evident in kimberlites of the KIMFIK pipes.

Although calcite is often a primary or deuteric mineral in kimberlites (e.g. Andrews and Emeleus, 1971; Dawson and Hawthorne, 1973; Clement, 1973, 1975; Mitchell, 1979a; Skinner and Clement, 1977, 1979; Dawson, 1980) it may also be an abundant secondary mineral. In some cases near-surface calcretization results in almost no relicts of the original kimberlite mineralogy (including olivine) being preserved.

5.2.4 Inclusions in olivine phenocrysts

A noticeable feature of many olivine phenocrysts is the presence of oriented rod-like or needle-like rutile inclusions generally less than about 0,02mm in length (Plate 5.10c). As noted by Pasteris (1980a) these inclusions are usually oriented parallel to crystal faces and tend to be concentrated towards the edges of the olivines. The abundance of phenocrysts bearing rutile inclusions varies considerably among kimberlites in the KIMFIK pipes and in many intrusions rutile-bearing olivines appear to be absent. Other inclusions present in olivine phenocrysts include ilmenite and spinel (Dawson, 1980) and Pasteris (1980a) has recorded the occurrence of numerous fluid inclusions in olivine phenocrysts from the De Beers pipe.

5.2.5 Unusual phenocrysts

The olivine phenocrysts in the KIMFIK pipes are not always restricted to simple, well-formed polyhedral crystals. In rare instances phenocrysts occur as multiple growth aggregates. Also in rare instances anhedral macrocrysts of olivine may have a degree of 'pseudo-idiomorphism' imposed on them by corrosion. This generally involves the development of some straight edges and the formation of angular re-entrants, usually located at the terminations of grain cracks at the edges of the grains. In some cases it is difficult to distinguish such corroded macrocrysts from multiple growth phenocrysts (except where obvious size differences are apparent).

5.3 PHLOGOPITE

5.3.1 Mode of occurrence

Wagner (1914) drew attention to the fact that phlogopite forms as much as 50 modal% of some kimberlites and phlogopite-rich kimberlites are present in all the KIMFIK pipes. In these pipes phlogopite occurs as:

(i) Anhedral macrocrysts of diverse origin (see Chapter 6).

(ii) Large megacrysts (Chapter 6).

(iii) Subhedral or euhedral, lath-shaped phenocrysts and microphenocrysts.

(iv) A primary groundmass mineral of diverse habit and variable grain size.

(v) Cryptocrystalline 'pools'.

(vi) Overgrowths around earlier-formed phenocrysts and xenocrysts of phlogopite.

(vii) A deuteric alteration product of earlier-formed minerals.

(viii) A metasomatic mineral replacing xenoliths and xenocrysts in kimberlite (e.g. Plate 7.10e).

In addition to occurring as a deuteric or metasomatic alteration product a noticeable feature of phlogopite in the KIMFIK pipes is its common occurrence in two (and occasionally three) distinct generations of primary material. Commonly an early generation of phenocrysts and a later-crystallizing groundmass phase are present. As noted elsewhere in this thesis (Chapter 2) the presence of two populations of several primary minerals is a characteristic feature of kimberlite.

5.3.2 Phlogopite phenocrysts

The problem of distinguishing between phlogopite phenocrysts and macrocrysts which may not be primary minerals is discussed in Chapter 6. However, in some instances the abundance and idiomorphic character of phlogopite insets clearly indicates a primary (phenocrystic) origin.

Phlogopite phenocrysts often occur as euhedral to subhedral laths ranging in length from $\sim$0,1 to 0,5mm. The laths have variable length/breadth ratios (2:1 to 10:1) and their idiomorphic character is often modified to varied degrees by corroded margins. Zoning is common and may be complex. Zoning is commonly revealed by distinct colour variations within grains and by differences in the extent and nature of pleochroism.

In many instances the cores of phlogopite phenocrysts display normal pleochroism while reverse pleochroism is exhibited by the rims. However, the reverse situation also occurs and in some kimberlites neighbouring phenocrysts display contrary pleochroic schemes while unzoned grains, displaying either normal or reverse pleochroism, may also occur.

In some cases, particularly in certain dykes associated with the KIMFIK pipes, elongate phlogopite phenocrysts display marked parallelism (Plates 5.3g-h). Such alignments must have been imposed by flow and indicate that the phenocrysts crystallized intratellurically. Crystallization prior to intrusion is also indicated by concentric arrangements of phlogopite laths in globular segregations (discussed in Chapter 7) in some kimberlites (Plate 7.13e). Since phlogopite may be stable to depths of 150-200km (Carswell, 1975) it is possible that intratelluric phenocrysts crystallized in the upper mantle.

Some phlogopite phenocrysts are randomly oriented but are also believed to have crystallized prior to intrusion because:

(i) Many laths are bent or otherwise deformed.

(ii) A later generation of post-intrusion groundmass phlogopite is also present.

5.3.3 Groundmass phlogopite

Phlogopite occurs as a late-crystallizing (post-intrusion) primary groundmass mineral in a wide array of guises, some of which are described in the following examples.

A striking form of occurrence is as conspicuous, usually anhedral, interstitial plates of relatively large size (up to $\sim$1,0mm in diameter). Good examples of such phlogopite are found in the DB3 kimberlite in the De Beers pipe (Plates 5.4a-b). Both these illustrations show the manner in which phlogopite grains poikilitically enclose (or in some instances are partly moulded around) earlier-crystallized components such as olivine, perovskite, opaque oxides and apatite. These textural relationships are unambiguous evidence of the late-crystallizing nature of the phlogopite.

The post-intrusion phlogopite grains described above are generally anhedral but subhedral and euhedral grains also occur in similar textural relationships. Post-intrusion crystallization is indicated by random orientation of subhedral and euhedral laths and by the presence of numerous inclusions of other groundmass minerals in the phlogopite. In extreme cases these phlogopite laths are so crowded with inclusions they are best described as having sieve textures (Plates 5.4c-g and Fig. 5.7a-b).

The late-crystallizing nature of sieve-textured phlogopite is particularly well illustrated in Figure 5.7b and Plate 5.4g. In both examples phlogopite laths project from 'normal' groundmass areas into calcite-rich segregations. The parts of the laths in the 'normal' areas are crowded with inclusions but portions within the segregations are virtually devoid of inclusions. The contrasting inclusion content of different parts of the same laths implies the following paragenesis:

(i) Intrusion of magma containing early crystallized components.

(ii) Crystallization of some groundmass (post-intrusion) minerals (e.g. monticellite) and the formation of carbonate-rich liquid segregations.

(iii) Essentially simultaneous crystallization of mica within the segregations and within areas of 'normal' groundmass. Many mineral inclusions are incorporated within mica crystallizing in the latter areas.

(iv) Crystallization of calcite within the segregations (and elsewhere if present) accompanied by carbonatization of previously formed phlogopite (Plate 5.4g).

To account for the textural relationships illustrated and the crystallization sequence described above it appears essential for phlogopite crystallization to have occurred under stagnant (post-emplacement) conditions.

In some of the KIMFIK kimberlites groundmass phlogopite is concentrated within late-crystallizing segregations. In such areas phlogopite commonly occurs as fine or coarse blades which have nucleated at the margins of the segregations and grown inwards (e.g. Plate 5.4h and Figure 5.7).

Groundmass phlogopite frequently exhibits considerable variation in character, often on an extremely local scale, e.g. in a single thin section. A detailed explanation of these variations is beyond the scope of this thesis but they appear to reflect varying combinations of reaction or growth zoning, or overgrowth development, with or without corrosion during crystallization gaps. The D18 kimberlite at Dutoitspan provides examples of groundmass phlogopite which has been subjected to some of these processes.

The groundmass of the D18 kimberlite contains abundant small (0,05-0,15mm long) subhedral to euhedral laths of phlogopite (Plate 5.5a). These laths are randomly oriented and are often intimately intergrown (Plate 5.5d) hence there is little doubt that they have crystallized in situ. These laths display complex zonation commonly shown by patchy, extremely irregular areas of different colour and degree of pleochroism. The shapes of these areas are commonly unrelated to the form of the grains; they may occur as irregular bands across grains, as discrete patches within grains and as impersistent marginal fringes (some grains are, however, surrounded by thin, fairly regular overgrowths). The irregularity of the zoning and the gradational internal contacts which the zones often display suggest that they are due to reactions with late liquids. The patchiness of the zoning also suggests that, at least in some cases, local micro-environments may have influenced the nature and extent of the reactions (see Fig. 5.7f). Impersistent marginal zones may be due to post-crystallization or post-metasomatic corrosion (Fig. 5.7e).

In addition to the zoning described above the groundmass phlogopite in the D18 kimberlite varies in character due to markedly different degrees of overgrowth development in different parts of the rock. In (and adjacent to) calcite-rich segregations the phlogopite is considerably coarser-grained than elsewhere (laths and basal plates generally measure between 0,1 and 0,2mm and reach 0,4mm in size). In addition, within the calcite-rich segregations (which have diffuse boundaries) phlogopite crystals generally display a greater degree of morphological perfection than elsewhere in the groundmass (Plates 5.5b-c). The foregoing features reflect extensive development of overgrowths around phlogopite in the segregations, a feature not shown to any extent by groundmass mica elsewhere in the rock. These overgrowths are illustrated in Plates 5.5b-c and in Figure 5.8. The boundaries of the cores (although often highly irregular) are commonly sharp and the cores contain inclusions which are generally absent in the rims. Some cores have a corroded appearance and considerable chemical differences between cores and rims are implied by the abrupt colour changes. Substantial chemical differences are also implied by the different degrees of susceptibility to alteration of cores and rims (Figure 5.8c). Considered as a whole the foregoing features strongly suggest that the rims are indeed overgrowths, rather than reaction borders or a function of zoning during continuous growth. Furthermore, within the segregated areas discrete euhedral grains, corresponding in character to the overgrowth rims, provide additional evidence of a second phlogopite crystallization event. Both the overgrowths and discrete crystals of the same material also display faint parallel growth lines which are indicative of compositional zoning during their formation.

The features noted above indicate a crystallization hiatus between cores and rims. Thus it might be concluded that phlogopite cores in the D18 kimberlite represent pre-intrusion crystallization while the overgrowths reflect post-intrusion precipitation. In fact, the cores (and therefore also the overgrowths) probably crystallized after intrusion. Post-intrusion crystallization is suggested by:

(i) The abundance of 'normal' (without or with very limited, overgrowths) phlogopite outside calcite-rich segregations.

(ii) The absence of any preferred orientation of these 'normal' laths.

(iii) The intergrown nature of 'normal' laths and basal plates.

(iv) The probable presence of groundmass diopside (or, possibly, apatite) inclusions in 'normal' phlogopite (carbonatization of the inclusions precludes positive identification).

(v) The occurrence of other inclusions (mainly opaque oxides and perovskite) in both dark and light (core and rim respectively) mica.

(vi) The generally fine-grained nature of the phlogopite in the D18 kimberlite (Plate 5.5a).

The occurrence of a hiatus in phlogopite crystallization after final emplacement is difficult to explain since substantial changes in fluid composition are required to stop and restart phlogopite crystallization. Such a crystallization gap may relate to competition for some elements among other phases (e.g. opaque oxides, perovskite and serpentine), following the crystallization of core-type phlogopite but, perhaps more likely, the overgrowths reflect an influx of volatiles into partly crystallized magma. As mentioned the cores of some composite phlogopite crystals are highly irregular and have a corroded appearance. It is therefore possible that resorption of earlier core-type phlogopite aided the later precipitation of the overgrowth variety. Confirmation of any of the foregoing speculations requires more detailed work involving mineral composition determinations.

Groundmass phlogopite in the KIMFIK kimberlites is often much finer-grained than descriptions in the literature indicate (e.g. Plate 5.5e). For instance, in some intrusions, phlogopite laths rarely approach 0,1mm in length, are commonly smaller than 0,02mm and many measure <0,01mm (grading down to microlitic particles). Very fine-grained phlogopite tends to occur as overlapping mats of tiny anhedral to subhedral platelets and laths which are often difficult to discern individually (even under high magnification). Such cryptocrystalline phlogopite is often evident in kimberlites which pronounced segregationary textures (Plates 5.5f-h) and is prone to severe deuteric alteration. Alteration commonly involves serpentinization (Fig. 5.11). The lower limits of the grain size of groundmass phlogopite in the

Commonly monticellite crystals range in size from about 0,005 to 0,05mm and average dimensions rarely exceed 0,02mm. In some instances relatively large grains (up to 0,15mm) occur just within, or concentrated around, late-crystallizing, calcite-rich, groundmass segregations (Plate 5.7b). Monticellite crystals in the KIMFIK kimberlites have slightly lower relief than olivine (R.I.γ n $\leq$1,66) and have low birefringence (the maximum interference colour approximates that of quartz). Euhedral crystals have cross-sections resembling olivine. The grains show parallel extinction and give positive biaxial interference figures with 2V $\sim$80°. Minute 'bubbles' are often present and probably represent gas or liquid-filled cavities. Monticellite grains also often display patchy, poorly defined zones of contrasting 'colour' and relief, probably due to incipient alteration. The latter feature has been noted by Snowden (1981).

Few compositional data are available for monticellite in kimberlites. Anaylses of monticellite from De beers and Dutoitspan kimberlites reveal little compositional variation (Table 5.2). Clement et al. (1975) noted that minor increases in Fe correspond with decreasing grain size, suggestive of Fe enrichment during monticellite crystallization. Such a trend was not found by Snowden (1981) and Mitchell (1978) reported that monticellite grains in the Elwin Bay kimberlite (Canada) display a reverse trend from Fe-rich cores to Fe-poor rims. Mitchell (op. cit.) also noted a greater degree of compositional variation among monticellite grains from the Elwin Bay kimberlite than has been found in the Kimberley pipes (Clement et al., 1975; Snowden, 1981).

The nature, abundance and distribution of monticellite grains in several of the KIMFIK intrusions leave no doubt that this mineral crystallized directly from kimberlite melts. Furthermore the abundance, fine-grained nature and haphazard orientation of monticellite crystals (particularly in kimberlite which otherwise displays well developed flow textures) indicate that they are late-crystallizing (post-emplacement) constituents of the rocks. Monticellite also occurs (infrequently) as a deuteric mineral replacing serpentinized olivine (Plates 5.3b-c) and as a metasomatic mineral in crustal inclusions which have been 'kimberlitized' (Plate 7.9e). Following the views of Warner and Luth (1973) it is possible that deuteric monticellite contains more dissolved $MgSiO_4$ than primary monticellite in the same kimberlite. No analyses from the KIMFIK pipes are, however, available to test this supposition.

A noticeable feature of monticellite in the KIMFIK intrusions is that it is abundant only in those kimberlites which are relatively impoverished in volatile components. The apparent abundance of monticellite in such kimberlites may, however, in part be an artefact of alteration. Monticellite is prone to extensive replacement by calcite and 'ghost' outlines of completely carbonatized crystals are evident in some calcite-rich intrusions. Monticellite is also replaced by serpentine. The apparent ease with which monticellite is replaced by calcite and, to a lesser extent, serpentine suggests that it may be a more abundant original groundmass mineral than is currently supposed.

5.5 DIOPSIDE

Dawson et al. (1977) and Clement et al. (1977) drew attention to the occurrence of primary diopside in kimberlites and Skinner and Clement (1979) and Scott and Skinner (1979) noted that diopside is the most abundant groundmass mineral in some kimberlites. Abundant diopside often occurs in kimberlites which are also rich in phlogopite (e.g. Plate 5.6a). Among the KIMFIK pipes diopside is particularly abundant at Finsch but it also occurs in Kimberley and Koffiefontein kimberlites.

In hypabyssal-facies kimberlites or kimberlite breccias, within the KIMFIK pipes, diopside occasionally occurs as scattered small phenocrysts (rarely >0,5mm in size). These phenocrysts are often subhedral or euhedral (Plate 5.7c) but in some instances they have clearly been out of equilibrium with the magma which post-dated their crystallization. Consequently some phenocrysts are anhedral, have been extensively corroded and have rough, 'pitted', partly resorbed or melted appearances (Plates 5.7c-d).

Crystallization of the diopside phenocrysts in some of the kimberlites prior to intrusion into the upper crust, is indicated by parallel orientation (flow-induced) of elongate crystals. In other examples early crystallization is implied by the occurrence of distinct overgrowths around phenocrysts (Plate 5.6b). These overgrowths correspond to a later generation of post-emplacement groundmass diopside.

Diopside occurs most abundantly in the KIMFIK pipes as a late-crystallizing post-emplacement, primary, groundmass mineral. Such diopside is generally very fine-grained. In hypabyssal-facies kimberlites it occurs as slender prismatic crystals rarely exceeding 0,2mm in length and often measuring <0,05mm (Plate 5.6a). In diatreme-facies kimberlites groundmass diopside is even finer-grained (Plates 7.8f-h, 7.9c-d and 7.10a-b). In many instances all the grains measure less than 0,1mm and they grade down to microlites measuring <0,01mm in length. Very small crystals often have high length/breadth ratios and frequently occur in the form of acicular needles (Plate 7.10b).

Diopside also occurs as a deuteric alteration product in some KIMFIK intrusions. It occasionally replaces serpentine in olivine pseudomorphs (Plates 5.3d-e) and, in diatreme-facies (tuffisitic) kimberlites, extremely fine-grained (frequently microlitic) diopside is often an abundant deuteric mineral replacing a variety of pre-existing phases. In such rocks deuteric diopside commonly occurs together with varying (often abundant) amounts of primary diopside (Section 7.2.2).

Metasomatic replacement of crustal xenoliths by diopside is also evident in some KIMFIK intrusions. Metasomatic replacement of this nature is most prevalent in hypabyssal-facies kimberlites and it reflects the reactive nature of relatively hot, comparatively slowly cooling, kimberlite magmas.

Groundmass diopside is often segregated; in some instances together with other groundmass phases. Such segregationary features are discussed in Chapter 7. Diopside segregations in the hypabyssal-facies D5 kimberlite at Dutoitspan are shown in Plates 5.6c and 5.7g-h. In Plates 5.6c and 5.7g the primary character of the diopside crystals is indicated by their euhedral nature, distribution and relationships with other minerals. In Plate 5.7g the irregular but relatively sharp boundary (bottom) between the diopside-rich patch and the enclosing material suggests that the former represents an inclusion of earlier generation kimberlite in the D5 intrusion. This interpretation is not considered likely, however, because other minerals within the patch are identical to those outside and because, elsewhere in the slide, there is a complete gradation in character between segregations such as those illustrated in Plates 5.6c and 5.7g. The D5 kimberlite also illustrates a further mode of occurrence of diopside; the occurrence of radially disposed, commonly elongate, crystals in apparent segregation vesicles, ocelli or geode-like structures. In some instances these structures (which commonly range from <0,5mm to 1-2mm in diameter) probably represent metasomatized micro-xenoliths of country rock (Plate 5.7h). They also occur, however, in hypabyssal-facies kimberlites which are essentially uncontaminated by crustal material. In such circumstances there is little doubt that these structures are primary features. E.M.W. Skinner (personal communication) has noted many excellent examples from various localities, notably the Jagersfontein pipe.

Few compositional data are available for primary kimberlitic clinopyroxenes and no KIMFIK analyses have been published. Some analyses from elsewhere are listed in Table 5.3. It is evident from Ca-Mg-Fe plots

(Figure 5.12) of available analyses that, with rare exceptions, primary clinopyroxenes fall well within the diopside field if the pyroxene divisions of Poldervaart and Hess (1951) are followed. However, two analyses of pyroxenes from the Schuller pipe contain upwards of 10 wt.% Fe and plot in the salite field. In some kimberlites groundmass diopsides show a calcium enrichment trend with slightly decreasing or virtually constant Mg/Mg+Fe (Figure 5.12). As noted by Robey (1981) this trend is evident in hypabyssal-facies kimberlites but it is not repeated by the only available analyses of diopsides from a kimberlite that has a tuffisitic character (Premier type 3 kimberlite, Scott and Skinner, 1979). This Ca-enrichment trend is also not repeated by clinopyroxenes in the Schuller pipe which as a group are more Fe-rich than those from other localities. Texturally the Schuller kimberlite is somewhat intermediate in character but more closely resembles hypabyssal-facies than diatreme-facies kimberlites.

Figure 5.12 also shows the pyroxene compositional field from a Swartruggens lamprophyre which is associated geographically with the Swartruggens kimberlites and displays distinct mineralogical and geochemical affinities with kimberlite. The pyroxenes from the two rocks are, however, compositionally different and are characterized by different variation trends. Primary pyroxene compositions may thus serve to distinguish kimberlites from rocks which resemble them.

Diopside is often a well preserved mineral in the matrices of KIMFIK kimberlites, particularly in phlogopite-rich varieties. It may, however, be deuterically replaced by calcite, serpentine or chlorite.

5.6 CALCITE

According to Dawson (1980) carbonates reported from kimberlites include calcite, magnesian calcite, siderite, dolomite, aragonite, strontianite and shortite. Only calcite is, however, considered an important constituent. Staining tests (using Alizarin Red S) on numerous thin sections have confirmed that calcite is overwhelmingly the most abundant carbonate in the KIMFIK pipes and associated minor intrusions. Dolomite does, however, also occur (e.g. Donaldson and Reid, in press). Calcite is present to some extent in all the kimberlites investigated although it is restricted to trace quantities in some intrusions. Calcite is particularly rare in most of the tuffisitic kimberlite breccias within the KIMFIK pipes and in some of these rocks it probably occurs only as a metasomatic mineral. Possible reasons for the paucity of calcite (or other carbonates) in tuffisitic kimberlite breccias are considered in Chapter 10. In several of the hypabyssal-facies kimberlites calcite is the dominant matrix phase (e.g. Plates 5.1a, 5.5h and 5.6d).

In hypabyssal-facies kimberlites calcite displays various habits which reflect diverse modes of origin. There is no doubt, however, that much of this calcite is primary; an aspect of calcite in kimberlites which has been stressed in recent years (e.g. Watson, 1967; Andrews and Emeleus, 1971; Mitchell and Fritz, 1973; Clement, 1973; Dawson and Hawthorne, 1973; Clement, 1975; Clarke and Mitchell, 1975; Emeleus and Andrews, 1975; Robinson, 1975; Clement et al., 1977; Mitchell, 1978; Skinner and Clement, 1979; Dawson, 1980). Dawson and Hawthorne (1973) and Sheppard and Dawson (1975) have pointed out, on the basis of trace element contents and isotopic characteristics, that kimberlitic calcites are similar to the calcite of carbonatites, although Mitchell (1979a) has pointed out that this does not imply genetic relationships between the two rock types. In the KIMFIK pipes primary calcite often occurs in rudely ovoid (Plates 5.10e and 7.11h) or irregular segregations (Plates 5.63-f, 7.11b and 7.12a-h) where it is commonly associated with phlogopite, apatite and serpentine. The extent to which such segregation occurs varies from the development of rare, scattered, small (<0,1mm) bodies to more or less continuous, irregularly lobate web-like aggregates (cf. Plates 7.11b and 7.12d-h).

Some segregations have well defined boundaries while others have diffuse margins. The origin of these segregations is discussed in Chapter 7 (Section 7.3.4).

Commonly the calcite in the segregations occurs as clear, anhedral grains which occasionally reach 1,5mm across but usually do not exceed 0,3-0,4mm in diameter. Where relatively large grains occur small segregations may be entirely occupied by optically continuous calcite. In some instances calcite grains within segregations are subhedral or polygonal with straight or slightly curved boundaries (Plates 5.6e-f). When serpentine occupies the central parts of segregations euhedral crystals of calcite often occur at the margins, having grown inwards from the boundaries (Plate 7.12g).

In some instances serpentine has partly replaced calcite within segregations (see discussion in Chapter 7). Such replacement results in some calcite grains having extremely irregular boundaries (Plates 5.6g-h and 5.8a-b) or individual relatively large grains are reduced to several, very small, irregular remnants (Plates 5.8c-d). Rounded calcite grains (Plates 5.8e-f) may also reflect replacement by serpentine. Examples of partial replacement of calcite by serpentine are also shown in Figure 5.10.

Primary calcite is not restricted to segregated areas such as those described above. Calcite also occurs frequently as fine-grained (microlitic to $\sim$0,3mm), late-crystallizing grains evenly distributed throughout the groundmass of the kimberlites. Such calcite has two habits. Larger grains (upwards of 0,05mm) are commonly irregularly anhedral and poikilitic (Fig. 5.9). Minerals commonly present as inclusions in poikilitic calcite include monticellite, perovskite, apatite, spinels and other opaque grains. Groundmass calcite grains also occur in an interstitial relationship with earlier crystallized minerals (Fig. 5.9).

In a number of the KIMFIK intrusions extremely fine-grained or microlitic groundmass calcite is intimately intergrown with fine-grained silicate phases. In such cases it is difficult to decide to what extent the calcite may be primary or whether it replaces pre-existing minerals. In many cases such calcite is probably a mixture of both primary and deuteric material.

Other forms of primary calcite have been recorded from kimberlites in the Kimberley area (Dawson and Hawthorne, 1973; Donaldson and Reid, in press) but are rare or absent in the KIMFIK pipes. They include: euhedral tabular crystals, dendrites, prismatic laths (Plates 5.8g-h), parallel growth skeletal crystals and columnar aggregates of platy crystals.

Many earlier formed minerals are extremely susceptible to carbonatization. Consequently calcite is a widespread deuteric component of many of the KIMFIK intrusions, wholly or partly replacing monticellite, diopside, phlogopite and relatively early-formed serpentine (notably in olivine pseudomorphs). Most of the fore-going replacement relationships are discussed in Chapter 7 and are illustrated in Figures 5.4-5.9 and 5.11 and Plates 3.4g-h, 5.1g-h, 5.2c-g, 5.4h and 5.9d.

Metasomatic calcite occurs in the KIMFIK kimberlites as cross-cutting, regular or irregular, veins and veinlets. These veins range in thickness from a few centimetres to a fraction of a millimetre. In many instances, particularly in narrow veins, the calcite is fibrous in character and the fibres are arranged parallel to one another and at right angles to the veins. Calcite also frequently occurs as a metasomatic mineral in altered country rock xenoliths in the kimberlites.

Except where severe weathering has occurred alteration of calcite is generally restricted to serpentinization. Replacement by serpentine is evident where relatively late serpentinization episodes have affected calcite in olivine pseudomorphs (e.g. Plate 5.1h and Figs. 5.4-5.6). Also, as noted previously, partial replacement of primary calcite by late-crystallizing serpentine is commonly evident in calcite-rich segregations (Plates 5.8a-f).

5.7 APATITE

Apatite is a constant, often abundant, accessory mineral in the KIMFIK kimberlites and makes up several modal percent of some intrusions (Table 7.4). It commonly occurs as well formed, euhedral, stubby to elongate prismatic crystals. These crystals display characteristic cross partings (Figs. 5.7 and 5.9) and hexagonal basal sections (Plates 5.4b, 5.5c, 5.6c, 5.9g and 6.3g). Euhedral crystals which are commonly scattered fairly evenly through the groundmass of the kimberlites vary considerably in size but are usually small. In most of the kimberlites they rarely exceed 0,1-0,2mm in length but they may be much smaller often measuring <0,01mm. In some intrusions apatite crystals are slender with length/breadth ratios of more than 10:1. In the most extreme examples the crystals are acicular and in rare instances apatite grains occur in radiating or sheaf-like clusters (Plate 5.9a). Such clusters have been ascribed to quench crystallization (Dawson and Hawthorne, 1973; Clement, 1975). Apatite also occurs as skeletal crystals (Plate 5.9b) such as those described by Wagner (1914).

As indicated in section 5.6 apatite is often a prominent constituent of late-crystallizing calcite- or serpentine-rich segregations in the KIMFIK kimberlites (Plate 5.10a). Apatite within these segregations usually occurs as subhedral or euhedral crystals but anhedral grains are also evident. Apatite within segregations is frequently coarser-grained than elsewhere. Individual grains measure up to 0,7mm in length and may stretch entirely across and beyond the confines of small segregations (Plate 5.10a and Figs. 5.9c-d). In other examples apatite occurs as stubby-elongate prisms growing inwards from the edges of segregations (Plate 5.9c). The shapes of anhedral apatites in segregations have been dictated by earlier-crystallized confining minerals.

Large apatites contain inclusions of previously crystallized minerals such as monticellite, perovskite and opaque oxides, or are partly moulded around phenocrysts such as olivine (Plate 5.10b and Fig. 5.9c).

Apatite is prone to carbonatization and much original apatite has been replaced by calcite (Plate 5.9d and Fig. 5.9d). The preserved 'ghost' outlines of euhedral, pseudomorphed grains are often recognizable in thin section but other, partly replaced, apatites have an irregular, corroded appearance.

As is the case with several kimberlite groundmass phases there is a dearth of compositional data for apatite. Wagner (1914) and Dawson (1980) have indicated that only fluor-apatite varieties have been identified. Dawson (op. cit.) noted that apatites from one Russian locality contain appreciable amounts of rare earths (mainly Ce and Nd) and SrO.

5.8 PEROVSKITE

Perovskite, like apatite, is a characteristic accessory mineral in kimberlites (Mitchell, 1972) and in some cases accounts for several modal percent of these rocks. In some of the KIMFIK kimberlites perovskite is more abundant than similarly distributed groundmass opaque oxides but such cases are unusual; the reverse situation commonly applies.

Yellowish-brown to dark brown perovskite occurs in several forms but is particularly abundant as small, discrete, angularly irregular or slightly rounded grains and as equant square or oblong-shaped euhedra, scattered evenly through the groundmass of the host kimberlites (Plates 5.2h, 5.3a, 5.4g, 5.6b-d, 5.9b, 5.9c and 7.15a). In some instances these grains are preferentially located (together with opaque oxides) at the margins of altered olivine phenocrysts and macrocrysts. Some discrete perovskite grains display prominent angular re-entrants which indicate the occurrence of complex twinning. Perovskite grains occasionally occur in clusters (Plate 5.9e).

Perovskite also occurs in reaction mantles around ilmenite macrocrysts (Plate 6.2a). In addition complex associations between perovskite and groundmass opaque oxides are often evident. These include granular intergrowths, rims of perovskite enclosing opaque euhedra, opaque material surrounding perovskite and complex aggregates in which perovskite is one of the minerals present. Pasteris (1979, 1980a) and Shee (1979) have shown that the common opaque phases involved are ilmenite, titanomagnetites and chromites. Dawson (1980) has drawn attention to various examples of perovskite as a member of a series of epitaxial zones overgrowing other oxides and silicate phases in kimberlites.

The complexity of perovskite paragenesis in kimberlite is consistent with Pasteris' conclusion (Pasteris, 1980a) that this mineral has a long crystallization history. As noted in Chapter 7 perovskite often occurs as inclusions in other groundmass minerals. Boctor and Boyd (1979) note that perovskite is known for its capacity to accomodate REE and other incompatible elements such as Nb. This is borne out by the analyses quoted in Table 5.4 for perovskites from a De Beers dyke. The generally lower totals of other analyses in Table 5.4 point to possibly even higher levels of REE and other incompatible elements such as Nb and Sr. Boctor and Boyd (1979) determined REE abundances of up to 13,4 wt.% in perovskites from Bellsbank kimberlites. The latter authors have also pointed out that:

(i) Perovskite in kimberlite shows a remarkable enrichment in LREE relative to HREE.

(ii) Perovskite from kimberlite sills is more enriched in REE than perovskite from kimberlites in pipes.

(iii) Perovskite from carbonate-poor kimberlites has lower REE abundances than perovskite from carbonate-rich kimberlites.

Perovskite is generally well preserved in the KIMFIK kimberlites but Pasteris (1980a) found that some perovskite from the De Beers pipe is partly or wholly replaced by anatase. She considers that this is probably due to the reaction CO_2 + Perovskite → TiO_2 + Calcite.

5.9 SERPENTINE

Serpentine is an abundant mineral in the KIMFIK kimberlites and it displays wide textural and colour variations. Serpentine occurs in two main forms in these rocks; as a deuteric alteration product of many earlier-formed minerals and as a primary groundmass phase. The occurrence of primary serpentine, representing the final stages of crystallization of kimberlite magmas, was not recognised for many years but has recently been identified at several localities (e.g. Emeleus and Andrews, 1975; Robinson, 1975; Mitchell, 1978, 1979a; Smith et al., 1979; Pasteris, 1980a). In some KIMFIK intrusions primary serpentine is abundant (e.g. Plate 5.9f) and Skinner and Clement (1979) noted that serpentine may be the dominant groundmass phase in some kimberlites. The most spectacular examples of deuteric replacement involve serpentinization of olivine and examples illustrating the complexity and episodic nature of this process were given in section 5.2.3. As noted above serpentine displays notable colour differences and textures vary (even in individual pseudomorphs) from massive, through finely felted, to relatively coarse fibrous, platy, or mesh varieties. Several analyses of serpentines pseudomorphing olivine in kimberlites from the De Beers pipe (Pasteris, 1980a) and from an associated dyke (Pasteris, 1980a; Donaldson and Reid, in press) indicate wide compositional ranges and document pronounced compositional differences between adjacent serpentine zones in individual pseudomorphs (Tables 5.5 and 5.6).

As indicated in previous sections of this Chapter deuteric serpentinization is not restricted to olivine but also affects many other minerals, notably phlogopite phenocrysts and macrocrysts and major groundmass phases such as phlogopite, monticellite and calcite.

Primary serpentine is most prominent and easily identifiable where it occurs (usually with calcite) in late-crystallizing, regular to irregular, globular segregations or in the more-or-less continuous web-like matrix segregations noted in section 5.6. Serpentine-calcite 'amygdales' occur in a dyke near the De Beers pipe and have been described by Donaldson and Reid (in press). Similar features occur in another dyke at the same locality and occur rarely in other intrusions associated with or in the Kimberley pipes (sections 7.3.3 and 7.3.4).

Serpentine in late-crystallizing segregations (Plates 5.6g-h and 5.8a-f) ranges in colour from almost colourless, through pale green or yellow, to grey and brown. In some pools it occurs as structureless, isotropic serpophite but in others it occurs as extremely fine-grained 'felted' or microlitic aggregates with very low birefingence. Serpentine in segregated zones or patches is also distinctly fibrous in some cases. The fibres often radiate inwards from (and perpendicular to) the edges of segregations whereas, in the central parts, a finely spherulitic habit is often prominent (Plates 7.11c and 7.13h).

Serpentine segregations display considerable ranges in composition, even within the same intrusion (Table 5.7). The serpentine analyses from the De Beers pipe quoted in Table 5.7 are generally similar to groundmass serpentine analyses from elsewhere (e.g. Mitchell, 1978; Robey, 1981). Noteworthy features of the analyses are the substantial amounts of FeO present (3.82-9,57 wt.%) and the occurrence of considerable Al_2O_3 (often over 0,75 and up to 5,57 wt.%). Two of the quoted analyses contain unusually high amounts of K_2O. Robey (1981) has ascribed a similar result to interference from neighbouring phlogopite during analyses. Pasteris (1980a) has indicated that a possible explanation of high Al (and K) is that late-stage fluids are enriched in these elements due to phlogopite breakdown. Abundant optical evidence of replacement of phlogopite by calcite and serpentine (Chapter 7) supports her view.

Serpentine also occurs as a fine-grained interstitial groundmass mineral in some of the KIMFIK kimberlites. Commonly such serpentine is intimately associated with fine-grained calcite and in some instances it is difficult to assess to what extent either mineral represents the primary crystallization products of residual liquids.

Serpentine which has formed at a relatively early stage (e.g. much of the serpentine within olivine pseudomorphs) is often replaced by other minerals, notably calcite, phlogopite and, to a lesser extent, chlorite. Unless breakdown as a function of near-surface weathering processes has occurred alteration of primary groundmass serpentine appears to be limited to some chloritization and, possibly, carbonatization.

It is possible that much of the 'serpentine' which occurs as a late-crystallizing, isotropic or near-isotropic, mesostasis (either within segregations or as interstitial material is, in fact, a serpentinous colloidal deposit or a mixture of serpentine (sensu stricto) and such a deposit. This supposition is prompted by the unlikelihood of the mesostasis commonly corresponding compositionally to a single mineral (serpentine). Other residual (incompatible) components could, however, be incorporated in the proposed serpentine-rich, colloidal end product of consolidation.

5.10 OPAQUE MINERALS

Opaque minerals are present in variable amounts (trace to >15 vol.%) in the KIMFIK kimberlites but the examination, evaluation and interpretation of these minerals is not a goal of this thesis. Virtually no investigations of these minerals were carried although some aspects of ilmenite macrocrysts are considered in Section 6.4.

Detailed investigation of microphenocrystal and groundmass opaque oxide phases of certain Kimberley kimberlites have been undertaken by Pasteris (1979, 1980a, 1980b) and Shee (1979). These authors have shown that the matrix opaque suites are dominated by spinels which often occur as complex zoned crystals (sometimes atoll-textured). These primary spinels, which usually occur as equant subhedra or euhedra ranging from minute grains (<0,01mm) to about 0,2mm in size (Plates 5.3a, 5.5f, 5.6b and 5.9g), show a normal magmatic trend of evolving from low-Ti, high-Cr cores to low-Cr, high-Ti and high-Fe^{3+} rims. Shee (op. cit.) considers this trend to be a response to increasing fO_2. Pasteris (op. cit.) has detected a second, apparently unique, trend in the De Beers pipe of titanomagnetite evolving to Mg-Al spinel (Mg-pleonaste). Both the foregoing authors note that, despite some overlap, compositionally distinctive spinel suites can be distinguished in neighbouring kimberlite intrusions. Pasteris (op. cit.) has stressed that spinels have a long crystallization history in kimberlites but that much spinel zonation represents incomplete interaction of spinel with late-stage volatile-rich kimberlite fluids.

Groundmass ilmenite also occurs in the KIMFIK pipes. Pasteris (op. cit.) has noted, however, that in the De Beers pipe this mineral is rare relative to spinel. Shee (op. cit.) has drawn attention to the remarkably high MgO content coupled with high Cr_2O_3 of groundmass ilmenite. The latter author has also indicated that at Wesselton groundmass ilmenites display reverse zoning; the grains have more magnesian rims than cores.

Pasteris (op. cit.) has indicated that sulphides are common but volumetrically insignificant in De Beers pipe kimberlites. The major occurrence is as nickel sulphide (heazlewoodite) needles in serpentinized olivines and similar needles are evident in rare instances in altered olivines from other Kimberley pipes.

Attention was drawn to the presence of oriented rutile inclusions within the olivine phenocrysts of some KIMFIK kimberlites in section 5.2.5 (Plate 5.10c). Rutile has also been noted as inclusions in coarse phlogopite blades (Pasteris, 1980a) and Shee (1979) has described rutile xenocrysts rimmed by groundmass ilmenite and spinel and occasionally containing ilmenite exsolution lamellae. Haggerty (1975) has described several other ilmenite parageneses in kimberlites.

Two very rare oxide phases, armalcolite and an unnamed Ba- and V-bearing titanate, have been described from kimberlites in the Kimberley pipes (Haggerty, 1975).

5.11 MELILITE

The possible presence of melilite in kimberlites has intrigued petrologists for some time (e.g. Shand, 1934; Dawson and Hawthorne, 1973). However, experimental work (Yoder, 1975) has indicated that melilite is an unlikely phase, unless extensive degassing of kimberlite liquids had occurred. Yoder's work in the system akermanite - Co_2 showed that under high partial pressures of CO_2 calcite + diopside would crystallize preferentially (see also arguments by Moore, 1979).

There are, nevertheless, at least three occurrences of kimberlite within the KIMFIK pipes which contain an altered mineral which may have originally been melilite. One of these kimberlites occurs in the Finsch pipe, the others at Koffiefontein and at its satellite pipe, Ebenhaezer. The Koffiefontein/Ebenhaezer examples are the most convincing. In both cases an altered, lath-shaped mineral occurs within pelletal lapilli (Chapter 7) in tuffisitic kimberlite breccias (Plate 5.9h). These laths now consist of complex very fine-grained intergrowths of serpentine, chlorite and a probable clay mineral.

The laths do not represent altered phlogopite as relatively unaltered laths of the latter mineral occur in the same lapilli. They are also not considered to have been calcite (or another carbonate) as no trace of such a mineral is present and calcite laths are rare in kimberlites (although they have been reported by Donaldson and Reid, in press and Dawson and Hawthorne, 1973). There is also no evidence to suggest that the laths were diopside, a mineral which occurs in a relatively well preserved state in these kimberlites. There is also nothing to suggest that they might have been an amphibole such as that reported by Skinner and Scott (1979) from the Swartruggens kimberlites. Furthermore, they do not resemble apatite pseudomorphs.

In view of the foregoing comments it appears that melilite is one of few possibilities and this interpretation is supported by the presence, in some laths, of relict median lines (or narrow central zones) which may reflect the peg structure often characteristic of melilite in other rocks (Plate 5.9h).

5.12 OTHER MINERALS

In addition to diamond (discussed in Chapter 8) and the megacrysts and macrocrysts described in Chapter 6 numerous other minerals have been reported from the KIMFIK pipes. Apart from various xenocrysts derived from the local wallrocks, these minerals include: native copper, platinum group minerals, graphite, pyrite, iron hydroxides and a variety of clay minerals, quartz, barytes, afwillite, bultfonteinite, apophyllite, natrolite, pectolite and chlorite. Many of these minerals are secondary weathering products and are not found in the deeper parts of the pipes. Others are metasomatic, often occurring within inclusions in the kimberlites and reflecting interactions between these inclusions and kimberlite fluids.

In some cases minerals such as natrolite and pectolite occur in situations which suggest that they are primary (Plate 7.4a). However, it is noticeable that they only occur in hypabyssal-facies kimberlite breccias which contain abundant country rock inclusions. Commonly the pectolite and natrolite clusters of primary appearance are accompanied by extensive development of both minerals in altered xenoliths. The association suggests that the 'primary' occurrences reflect locally contaminated residual kimberlite fluids; the contaminants having been abstracted from xenoliths. The fact that both minerals are sodic while kimberlites are calcic supports this conclusion. Pectolite in the KIMFIK pipes has recently been investigated by Scott-Smith et al. (in press). Previously this mineral was incorrectly identified as cebollite by Kruger (1978).

CHAPTER 6

MACROCRYSTS AND MEGACRYSTS

6.1 INTRODUCTION

Kimberlites characteristically contain members of a group of anhedral, dominantly ferromagnesian, oxides and silicate minerals which occur as relatively large insets in a fine-grained (aphanitic) matrix. Commonly these insets range between 0,2 and 5,0mm in diameter and such grains have been termed macrocrysts (Chapter 2). In some kimberlite intrusions these macrocrysts are accompanied by even larger (commonly >2-3cm) but much rarer megacrysts (Dawson, 1971, 1980).

The minerals which commonly occur (as macrocrysts or megacrysts) are olivine (the dominant macrocryst phase), phlogopite, picroilmenite and magnesian garnet (the diagnostic indicator minerals of the kimberlite prospector), clino- and orthopyroxene and several varieties of spinel (the latter as macrocrysts only, spinel megacrysts have not been reported, J.v.A. Robey, personal communication). Although they are very rare relative to the foregoing minerals and are not ferromagnesian in character both diamond (discussed in Chapter 8) and zircon logically form part of the macrocryst/megacryst assemblages in kimberlites.

Macrocrysts and megacrysts have been extensively studied compared with the matrix minerals of kimberlites (see, for example the multi-authored volumes edited by Nixon, 1973, Ahrens et al., 1975 and Boyd and Meyer, 1979). The emphasis placed on these minerals reflects the relative ease with which they can be obtained, their well preserved nature and their importance with respect to evaluation of the nature of the upper mantle and processes which have occurred therein.

A detailed study of macrocrysts and megacrysts in the KIMFIK pipes has not been carried out during the course of this study. Consequently in this Chapter descriptions of these minerals are largely limited to brief petrographic reviews. However, certain aspects of the origin of some macrocryst minerals and of the megacrysts are considered.

6.2 OLIVINE MACROCRYSTS

6.2.1 Nature and origin

The olivine macrocrysts in the KIMFIK intrusions are typically anhedral. They occur as rounded (Plates 5.1a-b and 6.1a) or irregular, commonly embayed (Plates 6.1b-d) grains. Their xenomorphic character contrasts strongly with the idiomorphic phenocrysts (Chapter 5): they are usually larger than the latter, commonly ranging between 0,5 and 5mm in diameter (Figures 5.1 and 5.2).

In addition to size and morphological differences olivine phenocrysts and macrocrysts in the KIMFIK pipes display other contrasting features. The macrocrysts are frequently characterized by features indicative of dynamic metamorphism. These features which include undulatory extinction, kink banding and partial recrystallization (Plates 6.1b-e) are common in the olivines of peridotitic nodules found as inclusions in kimberlites. Such features are not displayed by phenocrystic olivine although very rare, apparently euhedral, crystals show slightly undulose extinction indicative of strain. Such crystals may, however, be tablets derived from recrystallized, sheared peridotites similar to those described by Boullier and Nicolas (1973, 1975). The 'secondary' euhedral tablets described by these authors were not characterized by deformation textures (e.g. Plate 6.1e) but, if further stress occurred subsequent to recrystallization, perhaps during incorporation as individual grains into the kimberlite, minor strain effects might be imposed. Macrocrysts also differ from olivine phenocrysts in that they are commonly traversed by much more abundant cracks, partings and poorly developed cleavages (e.g. Plate 5.1a). Such features probably also reflect a degree of dynamic metamorphism.

Although often implied by indiscriminate use of the term 'phenocryst' little positive evidence favouring a primary origin for the macrocrysts has been put forward. The contrasting view, that macrocrysts are entirely xenocrystic has been criticised by Mitchell (1973a) and

and Snowden (1981). Mitchell argued that, if olivine macrocrysts were all peridotite-derived xenocrysts, much greater abundances of similarly derived minerals, such as enstatite, would be present in kimberlites. He therefore concluded that only about 40% of the olivine macrocrysts in the Wesselton kimberlite which he studied were xenocrysts and that the balance were rounded phenocrysts. Snowden (op. cit.) considered a major proportion of the olivine macrocrysts in Dutoitspan kimberlite which she studied to be phenocrysts because good compositional relationships did not exist between macrocrysts and the olivines in peridotite nodules (Snowden's results are discussed later in this section).

In contrast to the views of the above authors it is concluded here that the vast majority of olivine macrocrysts in the KIMFIK pipes are xenocrysts and the latter term is used synonomously in the remainder of this thesis. Arguments in favour of a xenocrystic origin for these olivines are listed below:

(i) There is direct evidence in many kimberlites in the KIMFIK pipes of the break-up of large (fist-sized or greater) ultramafic nodules to smaller aggregates and finally to discrete xenocrysts (e.g. Plate 6.1f). Such break-up of nodules to produce xenocrysts has also been noted by Dawson (1980).

(ii) Micro-xenoliths of peridotite - often only 3 or 4-grain aggregates evident in thin sections - are much more abundant than the generally rare, much larger, ultramafic nodules that have been exhaustively studied in recent years (Nixon, 1973; Ahrens et al., 1975; Boyd and Meyer, 1979). In some instances more than one such micro-xenolith is evident in individual thin sections. The abundance of these micro-xenoliths relative to the larger nodules clearly implies that extensive disruption of upper mantle-derived peridotitic material has occurred and is consistent with interpretation of the macrocrysts as xenocrysts.

(iii) The deformation features displayed by many olivine macrocrysts cannot be adequately explained if these deformed grains crystallized directly from kimberlite magmas. Such features are, however, mirrored by the olivines in many peridotite nodules hence strained or recrystallized macrocrysts can confidently be interpreted as xenocrysts. Deformed crystals usually constitute >50% and in rare cases form >90% of the olivine macrocrysts visible in individual thin sections of KIMFIK kimberlites.

(iv) Mitchell (1973a, 1978) has suggested that macrocryst populations include xenocrysts and substantial quantities of large phenocrysts that have been rounded by abrasion during fluidized intrusion from the upper mantle. In this thesis argument is presented (section 7.7.2) to show that the euhedral phenocrysts probably crystallized in the upper mantle prior to relatively rapid intrusion to near-surface levels. In terms of this viewpoint Mitchell's 'abrasion theory' is untenable since it would require selective abrasion of xenocrysts and some phenocrysts while other phenocrysts maintained a high degree of idiomorphism. Furthermore, extensive abrasion of any olivine that crystallized at depth is unlikely as the transporting systems would be essentially 'lean-phase' liquid-solid fluidized intrusions (Chapter 9), i.e. systems that would not favour extensive abrasion. Considerable abrasion might be expected in near-surface environments under prolonged gas-solid fluidization conditions. However, since olivine macrocrysts are morphologically similar in hypabyssal- and diatreme-facies kimberlites (apart from some instances of greater incidences of angular, broken grains in the latter), it is evident that high level abrasion of olivine macrocrysts is also very limited.

(v) Many olivine macrocrysts display features which can in part be ascribed to magmatic corrosion (Dawson, 1971). Thus many macrocrysts are rounded in form or display irregular curved embayments and protuberances or (less frequently) exhibit marginal, angular re-entrants. The latter usually occur at the margins of cross-cutting cracks. However, corrosion on a scale sufficiently extensive to completely convert abundant large phenocrysts to anhedral macrocrysts can be rejected on several grounds. Firstly, as in (iv) above, an unlikely degree of selectivity is implied whereby larger phenocrysts are entirely rounded while smaller crystals retain idiomorphic morphologies. Secondly, if abundant 'disguised' (corrosively rounded) phenocrysts are present among macrocryst populations many more examples of partly preserved idiomorphic crystals such as that illustrated in Plate 6.1g would be expected. In fact such phenocrysts are extremely rare. Two other features of the macrocryst also suggest that corrosion is not generally severe. Firstly, the shapes of most macrocrysts can be matched with olivines in the texturally varied ultramafic nodules present in kimberlites. Secondly, the macrocrysts are comparable in size with the olivines in these peridotitic inclusions.

(vi) Many idiomorphic olivines in some KIMFIK intrusions contain oriented, rod-shaped, rutile inclusions. Similar inclusions are not present in co-existing macrocrysts. The absence of such inclusions indicates that the macrocryst population does not include rounded phenocrysts; unless the latter constitute a third, discrete, olivine population (i.e. two phenocryst populations and xenocrysts are present).

(vii) If the macrocryst populations include both phenocrysts and xenocrysts bimodal compositional clustering would, at least in some cases, be expected. It would be too much of a coincidence to suppose that compositional modes of phenocrysts and xenocrysts would always be coincident. No bimodal distributions have, however, been reported although the point has not been fully tested because relatively few data are available.

Mitchell (1973a) has argued that the rarity of orthopyroxene relative to olivine among the macrocrysts precludes a xenocrystic origin for many of the olivines, i.e. olivine : orthopyroxene ratios are much higher among macrocrysts than in the supposed source peridotite/pyoxenite mantle rocks. However, Boyd and Clement (1977) and Skinner and Clement (1979) have noted the absence of orthopyroxene as a primary mineral in kimberlites and have suggested that upper mantle-derived orthopyroxene xenocrysts may not be preserved in kimberlites. The lack of orthopyroxene may be related to extensive resorption or, in part perhaps, to alteration which precludes identification of the original mineral.

An important point with respect to Mitchell's argument (Mitchell, 1973a) is that although orthopyroxene abundances are low relative to olivine macrocrysts similar arguments have not been raised in respect of the other common peridotite components (clinopyroxene and garnet). Thus if the relative abundances of olivine, clinopyroxene and garnet macrocrysts reasonably reflect expected upper mantle ratios the argument that orthopyroxene has been destroyed is strengthened. In fact Mitchell's argument can be reversed. If ∿50% of the olivine macrocrysts are assumed to be phenocrysts then the ratios of garnet and clinopyroxene macrocrysts against the remaining olivine macrocrysts (xenocrysts) would probably be too high.

6.2.2 Chemistry of olivine macrocrysts

The forsterite content of olivine macrocrysts from kimberlites in the Kimberley area is shown diagrammatically in Figure 6.1b. These data indicate:

(i) A considerable compositional range (Fo_{94-83}).

(ii) A pronounced peak in the histogram at Fo_{93-92}.

(iii) That 70% of the olivines which have been analysed have a forsterite content of 90% or more.

(iv) That the macrocrysts are generally more magnesian than olivine phenocrysts (cf. Figures 6.1a and 6.1b).

It is also evident from Figure 6.1b that, as in the case of olivine phenocrysts, the macrocrysts within specific kimberlite intrusions display considerable compositional variations. This is not, however, always the case. Thus 12 macrocrysts from a kimberlite sill cut by the Wesselton pipe have compositions between $Fo_{92,71}$ and $Fo_{92,99}$ (Hill, 1977).

It was concluded previously (section 6.2.1) that macrocrysts represent populations of xenocrysts in kimberlites derived from peridotitic and/or dunitic upper mantle rocks. If this conclusion is valid correlation between the chemistry of macrocrysts and the olivines found in ultramafic (mainly peridotitic) xenoliths would be expected. An accurate comparison of these populations requires the analysis of representative suites of olivine macrocrysts and xenolith olivines from the same kimberlite intrusion. No such analyses are available for any of the KIMFIK pipes. Some generalized conclusions can, however, be drawn by comparison of Figures 6.1b and 6.1c.

Figure 6.1b shows the compositions of macrocrysts from several Kimberley localities. Figure 6.1c shows the compositions of olivines in ultramafic nodules from Bultfontein mine (Lawless, 1978). Examination of these diagrams indicates:

(i) That there is generally good correspondence between the compositional ranges exhibited by the two populations.

(ii) That the maxima on the histograms are coincident.

(iii) That a somewhat greater proportion of the macrocrysts have compositions below Fo_{90} than do olivines from nodules.

(iv) That rare macrocrysts have compositions which are less magnesian than any of the nodule olivines analysed.

(v) That there are some distinct differences in compositional distributions of olivine macrocrysts from different sources (Figure 6.1b).

(vi) That there are some differences in composition between olivines derived from different types of nodules, e.g. as a group harzburgites (including garnetiferous varieties) are more magnesian than coarse garnet lherzolites.

In general there appears to be no reason (from the compositional data) to doubt that the vast majority of macrocrysts are indeed xenocrysts. The difference in proportions of grains with $Fo_{>90}$ (cf. Figures 6.1b and 6.1c) could, however, be cited as evidence for the presence of a notable (but minor) percentage of macrocrysts which are not derived from mantle peridotites, pyroxenites or dunites. It should, however, be noted that:

(i) Relatively few data are available and the histograms may not be entirely representative of either olivine population.

(ii) The macrocryst and nodule olivines plotted are from different, albeit nearby, sources. Underground observations have clearly revealed that, even in the same pipe, different kimberlite intrusions have different ultramafic nodule assemblages and, as previously noted (Figure 6.1c), different types of mantle peridotites (and dunites) have differing average olivine compositions. It is thus possible that one, or more, of the macrocryst sources plotted (Fig. 6.1b) has sampled a higher proportion of rocks with relatively low Mg-content olivines than has the Bultfontein pipe.

(iii) A proportion of the 'low' magnesia macrocrysts may be derived from 'polymict' peridotites (Figure 6.1d) although many of the grains analysed from such sources (Lawless, 1978), which have $Fo_{<90}$, are small recrystallized neoblasts and are unlikely to occur as part of macrocryst assemblages. Nevertheless polymict peridotites remain a potential source of some macrocrysts (The rarity of polymict peridotites does, however, suggest that at most only a small proportion of olivine macrocrysts could be derived from such sources).

Snowden (1981) believes that only the recrystallized or partially recrystallized macrocrysts in the Dutoitspan kimberlites which she examined are derived from pre-existing mantle rocks. She considers that most, or all, single grain macrocrysts which have not undergone recrystallization are true phenocrysts. This interpretation is based mainly on the fact that the average composition of the macrocrysts from Dutoitspan which she analysed does not match the average value of Bultfontein nodule olivines obtained from the literature (Boyd and Nixon, 1978). Snowden's interpretation can, however, be criticised on the following grounds:

(i) The interpretation is based on very few data and the average compositions applied may not be valid. Furthermore the average macrocryst value used is calculated from analyses which include several relatively low-Mg subgrains in individual recrystallized macrocrysts and is therefore too low.

(ii) To argue that only recrystallized macrocrysts are xenocrystic is simplistic. Granular (unrecrystallized) nodules are much more abundant in most kimberlites than sheared nodules. It therefore seems unreasonably selective to suppose that the former rocks have not contributed to the olivine macrocryst populations in kimberlites. There is, in fact, abundant evidence that they have done so (e.g. Plate 6.1f).

6.3 PHLOGOPITE MACROCRYSTS

Relatively large (commonly 0,2-10,0mm), anhedral macrocrysts of phlogopite are frequently present in the KIMFIK kimberlites (e.g. Plates 6.2a-h). These macrocrysts rarely make up more than 3-4 modal percent of the rocks but they exhibit a variety of features, even on as small a scale as a single thin section.

6.3.1 The nature of phlogopite macrocrysts

Phlogopite macrocrysts are commonly undeformed but they may be bent, twisted, kink-banded, partially recrystallized or exhibit deformation lamellae, roughly parallel strain shadows or irregular undulose extinction.

In some intrusions the phlogopite macrocrysts are entirely unaltered (e.g. Plate 6.2a and 6.2e-h) but deuteric alteration may be severe (e.g. Plates 6.2b and 6.2d). Such alteration commonly appears to involve serpentinization, carbonatization and chloritization (Wagner, 1914). However, the alteration products are generally extremely fine-grained and intimately associated hence individual alteration products may be difficult to identify by optical methods. Alteration is often accompanied by the development of abundant opaque material, both within the grains or as prominent reaction rims (the 'opacite' of Wagner, 1914). Dawson (1980) stated that the opaque rims (Plates 6.2b-d and 6.3e-f) consist of magnetite (plus hydrophlogopite) but Pasteris (1980a) has noted the occurrence of Cr-spinel within the rims of altered macrocrysts from the De Beers pipe.

Macrocrysts in some KIMFIK kimberlites have been replaced by a later generation of mica. The latter phlogopite occurs as randomly oriented small laths within the confines of the original grains or extends beyond them into the surrounding groundmass (Plate 6.2c). These laths are optically similar to phlogopite which occurs elsewhere in the groundmass of the kimberlites and cross-cutting relationships such as those illustrated in Plate 6.2c indicate that the laths are contemporaneous with the latter. It is possible that, as in the case of olivine phenocrysts and macrocrysts, deuteric phlogopitization follows initial serpentinization of phlogopite macrocrysts.

Some form of zoning is often displayed by the macrocrysts and is indicated by colour variations and differences in the intensity and character of pleochroism. Many macrocrysts have cores which display reverse pleochroism and rims which are normally pleochroic (cf. Plates 6.2e and 6.2f) but (as is shown by the phenocrysts described in Chapter 5) the reverse situation also occurs. Unzoned macrocrysts may exhibit reverse or normal pleochroism.

The nature of zoned macrocrysts in the same kimberlite may vary considerably. Rims may be thick or thin relative to the cores, they may vary in thickness around individual cores or they may be impersistent (Plates 6.2c-h and 6.3a-d). In other macrocrysts zoning appears erratic or random so that a clear distinction between cores and rims is not always possible. Some macrocrysts display multiple zones each of which may be highly irregular (Plates 6.3a-b) while neighbouring macrocrysts may be unzoned. Cores may be more-or-less centrally located and conform in shape to the overall form of the macrocrysts (Plate 6.3c) or cores may be rounded, highly irregular or eccentrically located (Plate 6.3d).

Much of the variation described above can be related to magmatic corrosion which may have occurred evenly around individual macrocrysts or may have taken place in irregular fashion, producing irregular, embayed margins and, in some cases, deep indentations or corrosion pits or highly irregular lath terminations. The latter features are diagnostic of magmatic corrosion rather than mechnical abrasion, an alternate (less favoured) mechanism to account for the anhedral nature of many phlogopite macrocrysts. Corrosion is a particularly convincing way of explaining co-existing features such as zoned rims of variable thickness, impersistent rims, eccentrically located kernels and the occurrence of apparently unzoned macrocrysts (cf. Plates 6.2e-h and 6.3a-d).

In some of the KIMFIK kimberlites zoning of macrocrysts is largely restricted to the occurrence of bleached rims (Dawson, 1980; Snowden, 1981). These rims may reflect the development, by reaction with aqueous fluids, of hydrophlogopite or mica-'serpentine' mixtures (Dawson, 1980). Furthermore, erratic or impersistent zoning may, at least in part, result from phlogopite-liquid reaction on an extremely local scale (as indicated in Figure 5.7f). Elsewhere, particularly where gradational colour changes and/or regular, repetitive growth lines are apparent, compositional growth zoning, as a function of continuously evolving liquids, has probably occurred. Thus Snowden (1981) has reported zoning of Dutoitspan macrocrysts towards low-Fe high-Ti and high-Cr rims. Mitchell (1980) has correlated increasing strength of pleochroism with increasing TiO_2 in micas from the Jos dyke in Canada. A similar trend from other occurrences has been reported by Rimsaite (1971) and Gittens et al. (1975).

In addition to the foregoing types of zoning much zoning of phlogopite macrocrysts in the KIMFIK pipes reflects the development of distinct overgrowths (e.g. Plates 6.3e-g). A distinct and possibly prolonged crystallization hiatus between cores and rims is often indicated by the irregularly corroded (Plates 6.3e-f) or rounded (Plate 6.3g) nature of the cores, by the presence of thick opaque reaction rims around the cores (Plates 6.3e-f), by the presence of mineral inclusions in the rims which are not present in the cores (Plates 6.3e and 6.3g) and by strong colour contrasts between cores and rims suggesting considerable compositional differences. In contrast to the anhedral cores the overgrowths are often optically identical to phlogopite which has crystallized at a relatively late stage in the groundmass and the overgrowths and groundmass micas represent a single episode of phlogopite generation (Plate 6.3e).

6.3.2 The origin of phlogopite macrocrysts

Many phlogopite macrocrysts are bent, kink-banded or exhibit distinct deformation lamellae, parallel strain shadows or irregular undulose extinction. Some deformation features may be imposed by the 'wedging' effects (parallel to cleavages) of alteration and late crystallization of groundmass components. Commonly, however, the deformation features evident are consistent with the view that the macrocrysts have been transported considerable distances and that they originated in the upper mantle or lower crust. Furthermore, since in some cases the deformation features cut across zoned crystals (Plate 6.3b) it is apparent that some zoning is developed prior to transportation. If a deep, or relatively deep, origin is accepted the problem to be faced is (as in the case of olivine) whether phlogopite macrocrysts are xenocrysts, phenocrysts or both.

There is abundant evidence (e.g. Plates 6.2b-d and 6.3c-f) that in some of the KIMFIK intrusions the macrocrysts have been out of equilibrium with the host magma during and/or subsequent to the later stages of intrusion. It is therefore tempting to conclude that extensively corroded, highly altered macrocrysts, with prominent reactions rims (e.g. Plate 6.2d), are xenocrysts derived from mica-bearing mantle peridotites. However, identical micas are found moulded around olivine phenocrysts (Plate 6.3h) hence the above assumption is not justified.

Phlogopite-olivine associations (in some cases together with an opaque oxide) such as that illustrated in Plate 6.3h are, in fact, found in several of the KIMFIK intrusions, together with monomineralic phlogopite aggregates (e.g. Plates 6.3f and 6.4a-c). As indicated in Plates 6.3f, 6.4a and 6.4c the phlogopite grains in these small (usually 2 or 3 grain aggregates) are not always altered and in some cases olivine phenocrysts are wholly enclosed within phlogopite. The small size and limited mineralogical range of these aggregates suggests that they are derived from earlier intrusions of kimberlite which, when sampled, were in various (but incomplete) stages of consolidation. Alternatively these aggregates may be derived from cumulate fractions (of earlier intrusions) which formed during temporary halts in upward intrusion. Clearly both models offer a potential source of phlogopite macrocrysts. Such macrocrysts can best be interpreted as cognate xenocrysts; an appellation equally applicable to similarly derived (cumulate) olivine 'phenocrysts' and, in fact, to batch-mixed olivine phenocrysts (in terms of the possible olivine provenance proposed in section 7.7.2). Cognate xenocrysts may also be in part derived from disaggregation of entirely consolidated, earlier, mica-bearing, kimberlite intrusions. Inclusions from such sources are present in many KIMFIK intrusions (e.g. Plates 6.4d and 6.4e).

The foregoing arguments do not rule out mantle peridotite as the source of some (in specific intrusions possibly many) phlogopite macrocrysts. Evidence that the break-up of upper mantle material have provided numerous olivine xenocrysts was presented in sections 5.2.1 and 5.2.2. Since phlogopite occurs in many of the same peridotitic source rocks a similar origin for some phlogopite macrocrysts is a reasonable conclusion and it is supported by chemical data (e.g. Dawson and Smith, 1975). Phlogopite in peridotites may be 'primary' or 'secondary' (Carswell, 1975; Delaney et al., 1980) and both types, which have distinctive compositions (e.g. Carswell, 1975; Boettcher et al., 1977), may occur as discrete macrocrysts following disaggregation of their host rocks.

A third probable source of macrocrysts are the MARID-suite (and other 'glimmerite') rocks described by Dawson and Smith (1977). MARID nodules (Plate 6.4f) are relatively abundant ultramafic inclusions in some KIMFIK intrusions which also carry considerable macrocryst phlogopite. Such associations are unlikely to be fortuitous and may, in fact, be more common than hitherto supposed. In some intrusions MARID nodule components, notably diopside and phlogopite itself, are severely altered (Plate 6.4g). This alteration may well have facilitated breakdown of the xenoliths to individual xenocrysts or rendered microxenoliths difficult to identify so that their true abundance has been underestimated.

Dawson and Smith (1977) favour an igneous origin for MARID-suite rocks. They suggested that these rocks represent cumulates derived from kimberlite magmas, temporarily held at relatively shallow levels in the upper mantle, in intermediate magma chambers. A major argument against such an origin is the paucity of olivine in MARID rocks (minor amounts have been reported by Gurney, 1979). Irrespective of what paragenetic model for olivine is adopted it is generally accepted that olivine is one of the earliest minerals precipitated by, or incorporated into, kimberlite magmas. Olivine should thus be present in MARID nodules if Dawson and Smith's petrogenetic model holds. In spite of the above objection the geochemical arguments presented by the foregoing authors for a link between kimberlite and MARID rocks are convincing, as are their arguments for an igneous (as opposed to metasomatic) origin. A possible solution to the impassè is to invoke a two-stage origin involving:

(i) Separation of cumulus olivine from a kimberlite melt at considerable depth in the upper mantle (at or near the site of magma genesis).

(ii) Upward movement of residual liquid enriched (relative to bulk kimberlites) in alkalies (particularly K), Al, Ti, Fe and OH and depleted in olivine by crystal fractionation.

(iii) Formation of MARID rocks at higher levels as suggested by Dawson and Smith (op. cit.).

It is also possible that some phlogopite macrocrysts are true phenocrysts that crystallized prior to intrusion of kimberlite into the upper crust, i.e. the 'pre-fluidization' micas of Mitchell (1979b) and Snowden (1981). Some support for this view is provided by the even distribution and abundance (3-4 vol.%) of such grains in some KIMFIK intrusions.

It is clear from the above discussion that phlogopite macrocrysts may originate in several ways. From a provenance point of view the problem lies mainly in evaluating the contribution which each possible source has made to the macrocryst populations in specific kimberlites. In some of these kimberlites, even on a hand specimen or thin section scale, the macrocryst assemblages are accompanied by microxenoliths of mica-bearing peridotite, by MARID-suite fragments, by 2-3 grain olivine-phlogopite aggregates, by inclusions of earlier generation kimberlite (containing coarse phlogopite) and by mica-bearing crustal xenoliths (Plate 6.4h). It is thus possible that co-existing macrocrysts may include representatives from all these sources and Pasteris' conclusion (Pasteris, 1980a), made on geochemical grounds, that macrocrysts of similar appearance in certain De Beers kimberlites reflect more than one period of crystallization, is particularly germane in this respect. Unfortunately it is not possible to distinguish between macrocrysts from different sources on an optical basis. The key appears to lie in detailed mineral chemistry. Although a start has been made (e.g. Dawson and Smith, 1975, 1977; Delaney et al., 1980; Mitchell, 1979a, 1979b, 1980) many more data are required before adequate distinction on chemical grounds is possible.

6.4 ILMENITE MACROCRYSTS

Various paragenetic types of ilmenite have been recognized in kimberlites. Included amongst these are macrocrysts consisting of monomineralic aggregates and anhedral single grains.

The abundance of ilmenite macrocrysts varies from kimberlite to kimberlite ranging from zero to 3-4 vol.% according to Dawson (1980). In rare instances (e.g. the Monastery pipe) ilmenite macrocrysts may locally make up >5 wt.% of the rocks.

Ilmenite macrocrysts are prominent accessory components of the Kimberley pipes but have not been found at Finsch or Koffiefontein (in spite of large scale mining for many years).

The relative abundances of ilmenite in the Kimberley pipes have not been accurately determined but clearly vary from intrusion to intrusion within individual pipes (Table 6.1). Visual examination of bulk heavy mineral concentrates suggests that ilmenite is considerably more common in the Wesselton and De Beers pipes than in the other two currently operated mines, Bultfontein and Dutoitspan.

Ilmenite has not been investigated during the course of this study but, according to the literature, the major characteristics of ilmenite macrocrysts are:

(i) Distinctive compositions highlighted by high Mg content and relatively high Cr and Fe^{3+} content (e.g. Pasteris, 1980b; Dawson, 1980; Mitchell, 1973b, 1977).

(ii) Wide compositional ranges; within kimberlite provinces, between neighbouring kimberlites, between adjacent intrusions in discrete kimberlite bodies and even on the scale of individual hand specimens. Thus Mitchell (1973b) recorded a range of 8 wt.% MgO from grains in a single specimen of Wesselton kimberlite and an overall range of 13 wt.% MgO from the suite of Wesselton samples which he studied.

(iii) The occurrence of chemically distinctive ilmenite at specific pipes. This is reflected by different average values and compositional ranges for both major and trace elements (although few data are available for the latter). For example, Mitchell

(1977) quotes the following MgO wt.% ranges: Sekameng 4,6-8,6%, Kao 7,7-12,5%, Monastery 6,8-12,6%, Frank Smith 8,5-17,8%, Wesselton 6,1-19,0%, Premier 9,4-23,1%, Arturo de Paiva 3,9-9,6%. In some of these examples, e.g. Premier, the maximum MgO values quoted are surprisingly high and it is possible that ilmenite of mixed parageneses are included. For instance some groundmass ilmenite grains at Wesselton contain in excess of 20 wt.% MgO (Shee, personal communication). It is also evident that some megacrysts are extremely rich in MgO (e.g. Pasteris et al., 1979). B.A. Wyatt (personal communication) states that the MgO content of ilmenite macrocrysts very rarely exceeds 16 wt.%. Wyatt confirms, however, that the ilmenites from different pipes often form geochemically distinctive suites and that differences in ilmenite chemistry are apparent between ilmenites from neighbouring intrusions in the same pipe. Wyatt's conclusions are based on microprobe analyses of many thousands of ilmenite macrocrysts from dozens of different occurrences. The occurrence of distinctive ilmenite suites at individual localities has also been noted by Boyd and Nixon (1973), Pasteris et al. (1979), Dawson (1980) and Pasteris (1980b).

6.5 OTHER MACROCRYSTS

Garnets are the most prominent of the other minerals which occur as macrocrysts in the KIMFIK pipes but they seldom approach 1 vol.% of the rocks. They occur as dispersed, anhedral, rounded grains or angular, broken fragments commonly ranging in size from ∿0,1mm to 3-4mm. Rare larger grains may be present.

Most macrocryst garnets are surrounded by rims of kelyphite and some grains have been completely kelyphitized. It has been shown elsewhere (e.g. Dawson, 1980; Garvie, 1981) that kelyphite rims consist of turbid, commonly, microcrystalline, combinations of spinel, phlogopite, amphibole, clinopyroxene, orthopyroxene, chlorite and possible plagioclase. Commonly the extreme fineness of grain-size precludes optical identification of most of the components. Kelyphite rims are often zoned. Such zoning reflects sequentially developed textural and mineralogical variations (Garvie, 1981).

Kelyphite appears to have a complex provenance. Much of this material can probably be related to isochemical, subsolidus reaction between garnet and olivine as a consequence of changes in PT conditions (e.g. Reid and Dawson, 1972; Dawson, 1980). However, it is likely that kelyphite is in part related to mantle metasomatism (Harte and Gurney, 1975) involving the introduction of alkalies and volatiles, from kimberlite precursor or other fluids, into garnetiferous mantle rocks. It is also possible that some kelyphite is a later reaction product of disaggregated garnet xenocrysts and kimberlite magma. Some support for the latter view stems from the fact that kelyphite occurs around garnets of widely variable composition and diverse origins. Thus kelyphite often surrounds large (>>1,0cm) megacryst garnets which cannot have been derived from <u>known</u> mantle rocks and which have been interpreted as high pressure phenocrysts in kimberlites or mantle crystal-mush magmas (e.g. Dawson and Stephens, 1975; Mitchell, 1978, 1979b; Dawson, 1980; Harte and Gurney, 1981).

Irrespective of the actual mode of origin it is abundantly clear that kelyphitization of garnet precedes the onset of fluidized conditions during kimberlite emplacement. This is revealed by the character of broken garnets in tuffisitic kimberlite breccias. Such rocks reflect fludized intrusion (Chapter 10) a process which facilitates garnet breakage. A noticeable feature of these broken grains is the presence of kelyphite on unbroken surfaces while broken surfaces are pristine.

Garnets in the KIMFIK kimberlites vary considerably in colour. Common colours are; mauve, cerise, pink, red, orange-brown and orange.

These colour variations reflect diverse chemical characteristics (Gurney and Switzer, 1973; Hawthorne et al., 1979; Garvie, 1981). Such chemical variability has led to several attempts at geochemical classification of garnets in endeavours to relate distinctive geochemical groups to different sources (e.g. Gurney and Switzer, op. cit.). Among the most successful of these classifications are those based on statistical cluster analysis techniques (Dawson and Stephens, 1975; Danchin and Wyatt, 1979). In these classifications discrete macrocrysts and megacrysts of garnets from kimberlites have been grouped and compared with garnets from a variety of ultramafic mantle and deep crustal xenoliths. It is evident from these investigations that most garnet macrocrysts can be matched with garnets in xenoliths, i.e. a xenocrystic origin for the vast majority of garnet macrocrysts is consistent with the data. The origin of the garnet megacrysts is less clear (see section 6.6) as is that of the low-Ca garnet macrocrysts from Finsch (Gurney and Switzer, op. cit.).

Other macrocrysts in the KIMFIK kimberlites (pyroxene, spinel and zircon) are very rare components of the rocks. In rare cases discrete macrocrysts of clinopyroxene (commonly bright green Cr-diopside) and pale green or yellow-green enstatite are evident in hand specimens and both minerals are found in heavy mineral concentrates, together with more abundant macrocrysts such as ilmenite, garnet and olivine. These pyroxenes are also seen in rare instances in thin section. They are optically identical to more commonly occurring pyroxene grains in peridotitic microxenoliths. Thus most discrete pyroxene macrocrysts are considered to be xenocrysts. This interpretation is consistent with the geochemical comparisons made by Stephens and Dawson (1977). The apparent rarity of pyroxene, particularly orthopyroxene, macrocrysts in the KIMFIK pipes is probably due to reaction and resorption after liberation of xenocrysts into kimberlite magma under disequilibrium conditions (Boyd and Clement, 1977).

Some implications of orthopyroxene resorption on kimberlite magma compositions are noted in section 7.8.2.

It is possible that a minority of the pyroxene macrocrysts originate in other ways, e.g. they may be phenocrysts derived from crystal-mush magmas (e.g. Nixon and Boyd, 1973a). However, such an origin is more likely for pyroxene megacrysts (see section 6.6) which are present as coarse (>1,0cm) crystals and cleavage flakes in heavy mineral concentrates from the KIMFIK pipes but are very rarely seen in situ.

Little is known about spinel and zircon macrocrysts in the KIMFIK pipes beyond the fact that they are present in some heavy mineral concentrates. No detailed studies have been undertaken and their paragenesis is in doubt. Zircon-bearing peridotite nodules have been recovered from tailings dumps in the Kimberley area (Gurney, 1979) hence at least some zircon macrocrysts may be xenocrystic. However, other zircons recovered from the Kimberley pipes are very much larger than any found in xenoliths and they may be more appropriately related to the cryptogenic megacryst suite (section 6.6).

6.6 MEGACRYSTS

As noted in section 6.1 macrocrysts in kimberlites are often accompanied by rare, larger (often >>1,0cm) megacrysts (discrete nodules). These megacrysts commonly occur as single crystals (usually rounded) or as monomineralic aggregates. In addition regular to irregular ilmenite-silicate (ortho- or clinopyroxene/olivine) intergrowths are present at numerous localities (Dawson, 1980; Robey, 1981) and small inclusions of one phase in another larger megacryst host are also found (e.g. Jacob, 1977; Nixon and Boyd, 1973a; Robey, 1981). Megacrysts have not been studied during the course of this thesis and the following general observations have been gleaned from the literature.

Chemically distinctive (Cr-rich and Cr-poor) suites of megacrysts have been identified at a North American locality (the Sloan pipe) by Eggler and McCallum (1974). A similar distinction has been recognized at the Orapa pipe (Shee, 1978). However, only megacrysts corresponding to the Cr-poor group are present at most localities (Dawson, 1980).

Although the same minerals are involved, megacrysts can often be distinguished from macrocrysts according to compositional as well as size criteria and, in some cases, the presence of more than one population of specific megacryst minerals has been determined by compositional differences (e.g. Jakob, 1977). Size alone does not serve to entirely isolate megacrysts from macrocrysts as size overlaps occur.

As indicated in previous sections of this chapter many macrocrysts (notably of olivine, garnet and pyroxene) have compositions which overlap with those of equivalent minerals in peridotite nodules. Accordingly such macrocrysts have been interpreted as peridotite-derived xenocrysts. On the other hand most megacrysts cannot be matched compositionally with the equivalent minerals in mantle-derived ultramafic rocks, although they are similar to minerals in high temperature, deformed peridotites (Robey, 1981). A xenocrystic origin involving the disruption of peridotitic rocks therefore appears unlikely; unless it is proposed that megacrysts are derived from extremely coarse-grained mantle rocks that are not represented among known ultramafic nodule suites.

There is fairly general agreement that megacryst assemblages are of direct magmatic origin and that they relate to some type of differentiated magma series (e.g. Robey, 1981; Harte and Gurney, 1981). Several alternative genetic models which fall within these broad constraints have, however, been put forward. For example it has been suggested that megacrysts are xenocrysts derived from crystal-mush magmas dispersed over a considerable (several tens of kilometres) vertical (and hence pressure) range in the upper mantle (e.g. Nixon and Boyd, 1973a; Boyd and Nixon, 1973, 1975, 1978). It has been suggested that the liquids involved may in fact be kimberlitic (Pasteris et al., 1979). Alternatively megacrysts have been interpreted as the products of temperature controlled, isobaric, igneous fractionation of probable protokimberlite liquids in the upper mantle (Harte and Gurney, op. cit.).

Two other aspects of the megacrysts should be stressed. Firstly, with certain exceptions, e.g. in the Monastery pipe, megacrysts are extremely rare relative to the smaller macrocrysts. Megacrysts have not been recovered from some kimberlites and elsewhere only one or two members of the megacryst suite are represented. Secondly, although megacrysts can often be distinguished from macrocrysts by compositional differences (Robey, 1981) this is not always possible. Ilmenite megacrysts and macrocrysts are a good example (section 7.7.3). In such cases separation can only be made on an arbitrary size basis and implied paragenetic distinctions may not be justified.

CHAPTER 7

SOME PETROGRAPHIC AND PETROLOGICAL ASPECTS OF THE KIMBERLITES

7.1 INTRODUCTION

In this chapter the extensive textural and mineralogical diversity exhibited by intrusions within or associated with the KIMFIK pipes is described and discussed in terms of the textural and mineralogical classifications outlined in Chapter 2. In doing so the many gradational relationships that exist between different varieties of kimberlite are emphasized.

Attention is also drawn to the petrographic features that distinguish diatreme- and hypabyssal-facies kimberlites and various genetic aspects of these two major groups of kimberlites are considered in detail. Much discussion is focussed on the origin of pelletal lapilli in diatreme-facies kimberlites and on the development of segregationary textures in hypabyssal-facies kimberlites and kimberlite breccias. Petrographic and genetic distinctions are drawn between the pelletal lapilli of TKB's and globular segregations present in rocks which may be considered intermediate in character between hypabyssal- and diatreme-facies varieties.

In addition the paragenesis of the primary minerals in the KIMFIK kimberlites is discussed, primarily in relation to pre-emplacement and post-emplacement crystallization periods. Finally various aspects of the bulk compositions of a cross-section of KIMFIK intrusions are discussed and some genetic implications of the analytical data are considered.

7.2 DIATREME-FACIES KIMBERLITES

Diatreme-facies kimberlites are preserved in all the KIMFIK pipes except the De Beers occurrence from which they have been removed by mining operations. The presence of these rocks in the Kimberley pipes influenced geological interpretations of kimberlites for many years and led to the long-held belief in some quarters that all kimberlites are clastic assemblages of rock and mineral fragments of diverse origin (e.g. Kennedy and Nordlie, 1968).

Tuffisitic kimberlite breccias (section 2.3.5) are volumetrically by far the most abundant of all the textural variants of kimberlites in the pipes (Table 7.1). In addition to their abundance at current levels of exposure TKB's are inferred to have occupied (almost exclusively) the substantial volumes of the diatreme zones that have been removed by erosion (section 3.4). This conclusion is supported by the distribution of TKB's in the preserved parts of the pipes and by analogy with less eroded pipes elsewhere (e.g. in Tanzania and Botswana).

The other main subdivision of the diatreme-facies group, tuffisitic kimberlites (section 2.3.5), are volumetrically rare in the KIMFIK pipes and commonly occur only as locally developed zones within TKB's. These local zones of TK usually form narrow (<1,0m), impersistent, vertically or near-vertically oriented, lenses within TKB's. They differ from the latter only in that they contain relatively few (<15 vol.%) plus 4mm rock fragments and the contacts between the TK's and TKB's are generally gradational. Such TK/TKB associations (Plate 7.1c) commonly reflect flowage differentiation (Bhattacharji, 1967) within intrusions during fluidized intrusion (Chapter 10). 'Fine-grained' TK zones often occur at the margins of major TKB intrusions.

TK also occurs as intrusive veins or dykes, often irregular in form and distribution, within TKB's. In most cases such veins probably represent the waning stages of major fluidized intrusions but they may also reflect discrete, small-scale, fluidization events.

7.2.1 Megascopic features of tuffisitic kimberlite breccias

The TKB's in the KIMFIK pipes are soft to moderately hard, brownish-, greenish-, or bluish-grey rocks which have a fragmental (breccia) appearance that is imposed by the presence of abundant country rock clasts in a finer-grained matrix. Commonly numerous anhedral, often rounded, altered (soft, earthy or serpentinous) olivine macrocrysts (most measuring between 0,5 and 5,0mm in diameter) are prominent in hand specimen. Other members of the macrocryst assemblage (see section 2.2.3) are usually noticeable but much less abundant constituents. Scattered autoliths and numerous pelletal lapilli of kimberlite (see later comments and sections 7.2.2 and 7.2.3 for descriptions and interpretations of these bodies) are also commonly evident in the KIMFIK TKB's.

Country rock clasts are volumetrically important constituents of the TKB's in the KIMFIK pipes (Tables 7.2 and 7.3, Plates 7.1a-d and 7.3a-b). The characteristically small size of the country rock inclusions in the TKB's has important genetic implications which are discussed in Chapter 10.

Heterogenous assemblages of country rock clasts are present in the TKB's although xenoliths derived from near-surface strata and from formations which have been removed by post-kimberlite erosion predominate (Plates 7.1a-d and 7.10f). Despite the heterogeneity of clasts the KIMFIK TKB's are characterized, in an overall megascopic sense, by a generally homogeneous appearance. This large scale homogeneity reflects considerable mixing of inclusions and other components. The result is a crude degree of textural and lithological consistency; an 'ordering' effect reflecting fairly even distribution and spacing of megascopic components.

As a result of the mixing referred to above, inclusions derived from the upper mantle (peridotitic or pyroxenitic in character), from local country rocks and from overlying (now eroded) formations occur in juxtaposition within the TKB's. In addition cognate inclusions (autoliths), possibly derived in part from early crystallized parts of the magmas which gave rise to the TKB's or, more commonly, from earlier discrete intrusions of kimberlite, also form part of the inclusion assemblages (Plates 3.3h, 6.4d and 7.6g).

Notwithstanding the general mixing of xenolithic material from widely separated stratigraphic levels the TKB's in the KIMFIK pipes consistently contain a high proportion of xenoliths which have moved downwards in the pipes relative to their original stratigraphic positions. In the Kimberley pipes TKB's containing abundant Karoo shale and dolerite fragments persist downwards, for at least 750m, below the base of the Karoo Sequence in the surrounding wallrocks. Similarly at Koffiefontein the K1 TKB (Fig. 3.38), which also contains abundant Karoo xenoliths, persists well below (>>200m) the Karoo-Basement contact. In the Finsch pipe Karoo inclusions (basalt, dolerite, shale and sandstone) dominate the major intrusion (the F1 kimberlite - see Fig. 3.37) xenolith assemblage for at least 500 vertical metres, although no Karoo rocks are preserved in situ in the vicinity of the pipe.

In the KIMFIK TKB's the majority of crustal xenoliths are angular to subrounded in character (Figs. 7.1-7.6 and Plates 7.1a-d, 7.2a-b and 7.3a-b). Rounded and in rare cases extremely well rounded xenoliths are evident but always occur in subordinate amounts. The angularity of many shale xenoliths is a particularly noticeable feature of the TKB's and highly angular, delicate, shard-like fragments with acutely angled terminations are often evident (see Figs. 7.1-7.6).

The foregoing observations are not entirely in accord with those of Wagner (1914) who, while noting the occurrence of angular shale fragments, stressed the well rounded nature of many inclusions. Wagner considered that prolonged, violent attritional milling of the xenolith assemblages occurred during kimberlite emplacement. This resulted in a high degree of rounding of the harder, more durable inclusions, while concomitant impact between xenoliths, and between xenoliths and the wallrocks, resulted in shale xenoliths being broken down into small, angular, fragments.

In general this study has provided little or no supporting evidence for extensive rounding of inclusions or for the mechanism whereby Wagner visualized that this rounding had occurred. There is, as noted above, a paucity of well rounded inclusions in the TKB's and the xenoliths display few of the surface features which might be expected if they were subjected to prolonged attritional milling. For example no grooved, striated, impact-scarred or markedly polished surfaces were noted on 200 small xenoliths recovered from the D11 TKB (Dutoitspan pipe) by disaggregation of the soft, clayey, matrix in water. Furthermore Wagner's supposition that the angular shale fragments reflect breakage, without concomitant rounding, during prolonged 'cup and ball' attrition, seems untenable. While extensive breakage of shale xenoliths might well occur under such conditions it is unlikely that the resulting, relatively soft, fragments would escape abrasional rounding while at the same time (according to Wagner) harder, more durable and competent, inclusions were perfectly rounded. Although a low degree of sphericity would be expected, because of the fissile character of the shales, considerable rounding would be anticipated if milling was both prolonged and severe.

The purpose of the foregoing arguments are not to negate the possibility of some (almost certainly variable) degree of xenolith abrasion and rounding during TKB emplacement. The intention is rather to stress that such effects are not uniquely extensive within kimberlites nor are the xenoliths in TKB's typically very well rounded. The formation of TKB's is considered to have involved the development of gas-rich fluidized systems (see discussion in sections 10.2 and 10.3) hence some abrasion and fragmentation by mutual attrition and impact has undoubtedly occurred. However, the foregoing comments and the discussion in Chapter 10 indicate that this was generally of limited extent.

At least some of the well rounded 'boulders' referred to by Wagner (1914) were rounded by exfoliation (within the diatremes) rather than by abrasion. This is indicated by the presence of concentric fractures which define exfoliation shells. A perfect example of such exfoliation is illustrated by Williams (1932, Plate 74). It should be noted, in relation to the arguments put forward for high level attritional rounding of xenoliths, that the most perfectly rounded inclusions in kimberlites are upper mantle-derived ultramafic nodules. However, these nodules are equally well rounded irrespective of whether they occur in TKB's or in hypabyssal-facies dyke and sill intrusions.

Numerous early researchers (e.g. Rogers and du Toit, 1904; Wagner, 1914) commented on the unmetamorphosed state of the xenoliths in the upper parts (i.e. diatreme zones) of the major South African kimberlite pipes. This study has confirmed that unmetamorphosed nature of many inclusions in the diatreme-facies kimberlites. Commonly even small, black, soft carbonaceous shale fragments have not been thermally metamorphosed following their incorporation into the TKB's (Plates 7.1a-d). A minority of the inclusions in TKB's do, however, display distinct alteration rims (probably due to incorporation prior to fluidization, see section 7.3.1) and low temperature metasomatic alteration, mainly affecting very small (microscopic) inclusions, may be evident. Metasomatic alteration is, however, much less extensive than in hypabyssal-facies kimberlites and kimberlite breccias (section 7.3.1).

In general megascopic directional structures are rare in the KIMFIK TKB's. Commonly inequidimensional inclusions and macrocrysts are randomly oriented but locally parallel orientation of elongate xenoliths is apparent. Such orientation (usually in an approximately vertical sense) commonly occurs over distances varying from a few centimetres to a few metres. In some instances more irregular swirl and eddy structures can be picked out by the orientation patterns of the 'coarse' components of the TKB's. Such features are of a local, small scale, nature (commonly measured in centimetres).

Apart from the TK zones noted previously (which are devoid of 'coarse' clasts) and an often conspicuous tendency for floating reef inclusions to be concentrated near the margins of the TKB's (e.g. at Finsch, see Fig. 3.37) there are no obvious examples of any systematic distribution zoning of xenoliths in respect of type or size, either vertically or horizontally, within the exposed columns of the KIMFIK TKB's. Such zoning has, however, been reported from other localities (e.g. Dawson, 1960, Lock, 1980). It is therefore possible that more detailed studies of individual intrusions would reveal subtle zonal distribution patterns of xenoliths within the KIMFIK TKB's.

Inclusions of kimberlite (autoliths) within the TKB's range considerably in size and type. In rare instances inclusions measuring >1,0m across have been noted but commonly autoliths do not exceed 10cm in diameter and most are much smaller. Many are microscopic in size (e.g. Plates 7.4h and 7.6g). Characteristically the autoliths in the TKB's are composed of hypabyssal-facies kimberlite. The absence of diatreme-facies inclusions is ascribed partly to the fact that TKB's form late within the evolutionary cycles of kimberlite pipes (see Chapter 10) but mainly to their less competent nature. It is likely that most diatreme-facies inclusions would be disaggregated during incorporation into later intrusions due to their generally soft, altered, partly fragmental and hybrid character (section 7.2.2).

Several mineralogical varieties of autoliths are usually present in the individual KIMFIK TKB's. This variety of autolith type testifies to the disruption and partial incorporation of several (possibly many) hypabyssal-facies intrusions during the emplacement of the KIMFIK TKB's. This aspect of the formation of the pipes is discussed in detail in Chapter 10.

7.2.2 Mineralogical and textural features of KIMFIK tuffisitic kimberlite breccias

Reference was made in Chapter 2 to the various classes of tuffisitic kimberlites breccias which occur in kimberlite pipes. These different classes are, however, by no means equally abundant. In particular, TKB's which are classed as crystallinoclastic varieties (section 2.3.5) are rare although they occur, for example, in the Dokolwayo pipe in Swaziland (Hawthorne et al., 1979) and in the Lace pipe near Kroonstad (Merensky, 1909). Examples from the Dokolwayo pipe are illustrated in Plates 7.3b-c.

As at other localities TKB's in the KIMFIK pipes are commonly extensively altered and many original petrographic features are masked by secondary mineralization (mainly involving the formation of clay minerals). Deuteric alteration may also be extensive and hinder the interpretation of some genetic aspects of the rocks (section 7.2.3).

The TKB's in the KIMFIK pipes are commonly characterized by pelletal textures although, at least locally, segregationary textures (as defined in Chapter 2) are present. Pelletal textures reflect the presence of small (usually <2-3cm), well formed, sharply bounded, commonly more-or-less spherical bodies in the rocks. These bodies are termed pelletal lapilli in this thesis. These lapilli were formed prior to and during fluidization events but were essentially solid prior to the cessation of intrusion (section 7.2.3).

Segregationary-textured TKB's usually contain some identifiable pelletal lapilli but these bodies are extensively augmented by irregular to rudely spherical segregations which are the result of primary crystallization (commonly around nuclei which include pelletal lapilli) and essentially contemporaneous deuteric alteration and replacement of other components, during quiescent post-fluidization consolidation (section 7.2.3). The boundaries of these segregations are commonly much more irregular than those of pelletal lapilli (cf. Plates 7.5a-h and 7.9c-d). Many pelletal-textured TKB's have been modified by extensive post-fluidization (quiescent) crystallization and alteration (section 7.2.3).

The KOF1 (Koffiefontein), E1 (Ebenhaezer - satellite pipe adjacent to Koffiefontein) and F1 (Finsch) intrusions are good examples of TKB's with pelletal textures and they are described below.

The general appearances of the KOF1, E1 and F1 TKB's in thin section are illustrated in Plates 7.2a-d and 7.3a-b. Prominent features in these photomicrographs are numerous angular to subrounded microxenoliths of country rock. Sedimentary inclusions (mainly mudstones and shales) dominate the xenolith assemblages in KOF1 and E1 (Plates 7.2a-d) but dolerite and basalt fragments predominate in F1 (Plate 7.3a-b). Also visible in Plates 7.2a-d and 7.3a-b are many generally oval-shaped or circular (spherical or subspherical), smoothly and sharply bounded, pelletal lapilli. These lapilli constitute between 50 and 60 vol.% of the KOF1, E1 and F1 TKB's (recalculated on a xenolith-free basis, Table 7.3). The lapilli range considerably in size but very few exceed 10mm in diameter and the vast majority are very small, ranging from approximately 3-4mm to about 0,05mm.

As noted above the lapilli are characteristically spherical or spheroidal in form but irregular lapilli are not uncommon. The margins of the latter may be marked by smoothly curved embayments and lobate protrusions and rare irregular lapilli have subangular forms. Broken individuals, with sharp, angular broken surfaces (unmodified by abrasion or corrosion), form noticeable but minor proportions of the lapilli populations in all three TKB's. Variations in lapilli shapes are illustrated in Plates 7.2-7.6.

Most pelletal lapilli in the KOF1, E1 and F1 intrusions are composed of relatively large nuclei or kernels surrounded by mantles (usually relatively thin) of fine-grained to cryptocrystalline kimberlitic material. The kernels are generally more or less centrally located but eccentrically located kernels are not uncommon (Plates 7.2-7.6). Some lapilli are devoid of kernels.

Most kernels consist of entirely altered (serpentinized, chloritized and/or clay mineralized) olivine phenocrysts or xenocrysts (Plates 7.2-7.6) but other kimberlite minerals and a variety of xenoliths and xenocrysts also occur as kernels in lapilli. It is important to note, however, that xenoliths and xenocrysts derived from high stratigraphic levels, although abundant components of the KOF1, E1 and F1 intrusions form only a small proportion of lapilli kernels. Futhermore, the kimberlitic material around such kernels often occurs only as very thin 'coatings' or is impersistent. Many country rock fragments are entirely devoid of such mantles (e.g. Plates 7.2c, 7.5a and 7.5e). These observations have important implications with respect to the origin of pelletal lapilli (see section 7.2.3).

Thin kimberlitic mantles frequently consist only of turbid, brown cryptocrystalline material. More robust envelopes are usually porphyritic in character. The phenocrysts are set in turbid, cryptocrystalline material similar to that which forms phenocryst-free mantles or they occur in slightly coarser-grained material within which several primary kimberlitic groundmass minerals can usually be identified. The variable nature of the mantles has important implications regarding the genesis of pelletal lapilli (see section 7.2.3).

In the KOF1 and E1 TKB's the phenocrysts within the fine-grained or cryptocrystalline lapilli consist mainly of serpentinized olivine euhedra and an altered mineral which displays tabular to lath-like habit (length/breadth ratios are commonly >3:1) and in some instances, relict 'peg structure' (Plates 5.9h, 7.4e and 7.5a-d). This mineral is considered to be deuterically altered melilite (see section 5.11). Rare phlogopite phenocrysts, quite distinct from the altered melilite(?), are also present in KOF1 and E1 lapilli. Possible melilite laths are also present (rarely) in F1 lapilli (e.g. Plate 7.6f). Small (usually <0,1mm), equant, subhedral to euhedral perovskite and opaque oxide microphenocrysts are present in many of the lapilli of all three

TKB's (e.g. Plates 7.8a-g). They are, however, less abundant in F1 lapilli than in KOF1 and E1 lapilli.

A prominent feature of many of the lapilli in the KOF1, E1 and F1 intrusions is some degree of concentric orientation of elongate components. This feature is particularly well displayed by lapilli containing abundant melilite(?) or phlogopite phenocrysts (e.g. Plate 7.5). In some instances perfect concentric orientation of phenocrysts is evident throughout the width of lapilli mantles. In other mantles orientation is well developed towards the outer margins but is poor or absent in the central parts. Lapilli that contain no phenocrysts (apart perhaps from a central kernel) occasionally display vague concentric layering. It is evident from Table 7.3 that a considerable proportion of the lapilli in the three TKB's display some degree of concentric orientation.

As implied above a complete transition from perfect concentric orientation to no orientation can be discerned within the lapilli populations of the KOF1, E1 and F1 intrusions. This trend can be broadly correlated with a gradation from spherical to more irregular shapes and (less clearly) from small to relatively large lapilli. Thus, although some spherical lapilli display no orientation features, much higher proportions of regular spherical lapilli contain oriented phenocrysts than do irregular lapilli. The latter lapilli are frequently porphyritic and relatively large (e.g. Plate 7.6a-d) but the enclosed phenocrysts usually display only slight concentric alignment at lapilli margins or they are randomly oriented.

The irregular shapes of some pelletal lapilli, coupled with an absence of concentric features, suggests that such lapilli may be autoliths in the sense of Rabhkin et al. (1962), i.e. inclusions of earlier generations of kimberlite. However, in view of the gradational relationships noted earlier such an origin is in most cases unlikely. These relationships imply that the formation of irregular lapilli must be linked to the origin of regular lapilli including those with concentric textures and the latter cannot be fragments of earlier generations of kimberlite. Furthermore, in all three TKB's, close genetic associations between irregular and spherical pelletal lapilli are indicated by some common textural features (e.g. grain size) and by identical mineralogical compositions (see section 7.2.3).

Notwithstanding the foregoing comments there is no doubt that some inclusions of earlier generation kimberlite are present in these and other TKB's in the KIMFIK pipes (e.g. Plate 7.6g). Such autoliths are characterized by mineralogical and textural features which differ markedly from those of the much more abundant, accompanying pelletal lapilli. Angular shapes and broken mineral grains at the edges of the autoliths also help to distinguish the latter from pelletal lapilli (see also Clement, 1973).

Rare lapilli in the KOF1, E1 and F1 TKB's form complex bodies apparently reflecting accretion in a liquid or plastic state. Examples are shown in Plate 7.7. Other compound lapilli consist of autolith kernels with pelletal lapilli mantles (e.g. Plates 7.4h and 7.8a).

Elucidation of the groundmass mineralogy of the pelletal lapilli in the E1, KOF1 and F1 intrusions is hampered by the generally fine grain size and by alteration (both deuteric and secondary in character). Nevertheless the main features can be determined in some instances.

Where best preserved and coarsest-grained the groundmass of E1 lapilli are seen to consist mainly of fine-grained to microlitic phlogopite and diopside with lesser serpentine and unidentified clay minerals (e.g. Plate 7.8b). Phlogopite occurs as tiny subhedral or euhedral laths and platelets, often measuring less than 0,01mm in size. In many E1 lapilli the phlogopite is intimately intergrown with very fine-grained, euhedral, prismatic or acicular diopside crystals which are usually less than 0,02mm long and are often considerably shorter. Diopside also occurs within very small (often <0,05mm) serpentine segregations within E1 lapilli. The phlogopite and diopside and much of the limited groundmass serpentine in well preserved, relatively 'coarse-grained' lapilli are considered to be primary minerals.

Diopside and, to a lesser extent, phlogopite may also form a substantial part of the cryptocrystalline groundmasses of many lapilli. Such cryptocrystalline material is particularly abundant in relatively small cored lapilli which have only thin mantles (e.g. Plates 7.5g). The probable occurrence of diopside in such lapilli is indicated in some instances by the presence of scattered slightly larger recognizable grains within the essentially cryptocrystalline, generally optically unidentifiable, groundmass material. The presence of diopside in optically unidentifiable mantles is also indicated by apparent gradations from lapilli with relatively coarse-grained groundmass containing abundant tiny diopside needles, to lapilli composed mainly of cryptocrystalline material. It should, however, be stressed that the groundmasses of many lapilli are so fine-grained that they are almost isotropic and glassy in character. The nature of such groundmass material is entirely beyond optical resolution.

Mineralogically and texturally the pelletal lapilli in the KOF1 TKB closely resemble those of the E1 intrusion. However, in some relatively coarse-grained KOF1 lapilli small (usually 0,02mm or less) serpentine pseudomorphs occur which probably represent altered monticellite grains (Plate 7.6h). Similar grains may have originally been present in some E1 lapilli but have been masked by alteration. Lapilli containing possible monticellite are at least in part pre-fluidization in origin (see section 7.2.3).

Pelletal lapilli in the F1 TKB differ in composition from those in the E1 and KOF1 intrusions. F1 lapilli tend to be phlogopite-rich relative to the lapilli in the other two TKB's (Plates 7.8d-h) but diopside also occurs, either as relatively 'coarse' grains ($\sim$0,02mm) intergrown with groundmass phlogopite and serpentine (Plate 7.8f-h), or as fine-grained to cryptocrystalline material similar to that which occurs in E1 and KOF1 lapilli. As indicated in Table 7.3 the E1, KOF1 and F1 TKB's all contain discrete phenocrysts and xenocrysts which do not form the kernels of (or occur within) pelletal lapilli. The majority of these discrete grains are olivine (e.g. Plates 7.2a-d -7.3a-b). A number of these olivines (and some other grains, e.g. garnet) occur as broken crystals (Plates 7.9a-b).

Xenoliths, xenocrysts, discrete kimberlite minerals, autoliths and pelletal lapilli in the E1, KOF1 and F1 intrusions are set in an interstitial matrix which is often turbid and indeterminate due to extensive alteration, mainly involving the development of unidentified clay minerals. Where relatively well preserved the interstitial matrices contain abundant serpentine much of which is considered to be 'primary' (see discussion in section 7.2.3) although 'ghost' outlines within the serpentine indicate that the latter in part replaces other components. The former presence of other minerals or rock particles is commonly shown by subtle changes in the colour, grain size and habit of the serpentine within the 'ghost' pseudomorphs, relative to the surrounding material. In some thin sections of KOF1, E1 and F1 elongate to acicular, very fine-grained (microlitic) diopside occurs within the serpentine of the interstitial areas (e.g. Plate 7.8c). In many other sections the presence of such diopside has almost certainly been masked by secondary alteration.

As indicated in Chapter 2 and earlier in this section some KIMFIK TKB's, locally at least, have segregationary textures (see examples from Koffiefontein illustrated in Plates 7.9c-d). While they generally contain recognizable pelletal lapilli such TKB's are characterized by the presence of irregular to rudely spherical segregations (Plates 7.9c-d) which crystallized after emplacement (section 7.2.3) and reflect, to varying degrees, patchy crystallization of microlitic diopside (± limited phlogopite in some instances?). These segregations often have irregular, rather vague, boundaries which contrast strongly in appearance with regular-shaped, smoothly-bounded, well-formed pelletal lapilli (cf. Plates 7.9c-d and 7.5a-h). Except where they wholly or in part represent modifications (by deuteric alteration and marginal crystallization, see later discussion) of previously formed pelletal lapilli, the segregations are composed very largely (in some instances apparently exclusively) of diopside which occurs as minute, elongate, prismatic to acicular crystals, rarely exceeding 0,01mm in length, and as tiny microlites (Plates 7.9c-d). It should, however, be noted that the diopside of the segregations is often coarser-grained than the cryptocrystalline-glassy material which forms the groundmass of many of the well formed, smoothly bounded, pelletal lapilli (cf. Plates 7.9c and 7.5f).

The diopside in the post-intrusion segregations frequently occurs in irregular or crudely spherical clusters around pre-existing, larger minerals, rock fragments or pelletal lapilli which it commonly, at least in part, replaces. The diopside of the clusters may be more-or-less radially distributed and individual elongate prisms or acicular crystals project outwards into interstitial serpentine. The latter mineral often forms a continuous base from within diopside-rich segregations to interstitial areas.

In some instances diopside forms many small, highly irregular, often vaguely defined patches without any indication that the diopside has nucleated against or entirely replaced pre-existing solid material (Plates 7.8c and 7.9c-d). The primary nature of these patches is, in fact, indicated by transitions between single, euhedral, very small, acicular crystals or microlites within serpentine, through small patches containing relatively few, relatively widely spaced crystals in serpentine, to densely packed diopside aggregates with a much reduced serpentine base.

Patches and clusters of diopside or diopside-rich aggregates are not always clearly separated and locally merge together to form irregular web-like areas with lobate margins. Serpentine occurs interstitially (Plate 7.9c). In some TKB's segregation between diopside crystals and microlites and the other main post-emplacement mineral (serpentine) is very poorly developed. In such cases the relatively even distribution of diopside within a base composed primarily of serpentine leads to TKB's with uniform textures (Scott and Skinner, 1979; Clement and Skinner, 1979). It should be noted, however, that some degree of diopside-serpentine segregation is usually present. The origin of the segregations described above and the replacement and nucleation relationships which they bear to pelletal lapilli are considered in section 7.2.3.

7.2.3 The origin of pelletal lapilli and post-fluidization segregations in tuffisitic kimberlite breccias

The occurrence of pelletal lapilli in southern African diatreme-facies kimberlites was noted by the author during an examination of kimberlites from diverse geographic areas which was carried out in the early part of this study. These small spherical or spheroidal bodies were termed kimberlite pellets and they were considered to be a specific variety of volcanic lapilli, i.e. solidified magma droplets (Clement, 1973). At the same time Ferguson et al. (1973) described similar but generally larger spherical bodies (ranging in size from "...a few mm to 70mm") in certain kimberlite pipes in Lesotho. Ferguson et al. termed these bodies 'autoliths' implying that "...kimberlite enclosed kimberlite...." and they ascribed these autoliths to precipitation of kimberlite around solid nuclei in a magmatic environment. In recognition of this postulated mode of origin Danchin et al. subsequently (1975) substituted 'nucleated autolith' for 'autolith' and Clement (1975) used the term 'globular segregation' to describe similar structures in kimberlite from the

Finsch pipe (from the F2 intrusion).

When the papers referred to above were published no direct genetic association between 'kimberlite pellets' and 'nucleated autoliths' was proposed although several similarities between the bodies were noted. Subsequently the term 'autolithic' has been used to describe kimberlites containing 'kimberlite pellets' and/or 'nucleated autoliths' and no genetic distinctions have been drawn between these generally spherical or subspherical bodies; they have been referred to generally as 'autoliths' (e.g. Lock, 1980).

Prior to considering the validity of the foregoing conceptual development a comment on terminology is warranted. This concerns the term 'autolith'. The use of 'autolith' in the foregoing context is unfortunate and should be discouraged. In its classical geological sense the term is synonomous with 'cognate inclusion' (Gary et al., 1977) thus implying at least two (albeit genetically associated) igneous events. It was in this sense that 'autolith' was first used in a kimberlite context (Rabhkin, 1962). The term should not be applied to bodies which formed with the material that envelopes them during single intrusive episodes (even though considerable changes in intrusion conditions occurred during the crystallization sequence that produced the spherical bodies and their host material, see later discussion and section 7.4).

A review of section 7.2.2 indicates that any theory which seeks to explain the origin of pelletal lapilli must take the following features of these bodies into account:

(i) The lapilli occur characteristically in diatreme-facies kimberlites within the diatreme zones of kimberlite pipes.

(ii) The lapilli commonly make up substantial proportions of some TKB's, particularly if the kimberlites are considered on a xenolith-free basis (Table 7.3).

(iii) The lapilli vary considerably in size but most of them are small (e.g. Plates 7.2-7.8), very few exceed 1cm and the vast majority measure <2-3mm in diameter (many are microscopic).

(iv) Many lapilli in the KIMFIK TKB's have sharp boundaries and smooth surfaces and display substantial degrees of sphericity. However, the extent to which these characteristics are developed varies from intrusion to intrusion and locally within intrusions.

(v) Many of the lapilli contain kernels which are often large relative to the overall sizes of the lapilli. These kernels are commonly more or less centrally located. Taking the two-dimensional exposures in thin section of the lapilli into acount, it is likely that the actual proportion of lapilli with kernels is higher than is indicated, for example, by examination of Plates 7.2 and 7.3a-b.

(vi) Lapilli without central kernels tend, on average, to be more irregular in form than pellets with kernels.

(vii) The kernels of lapilli consist mainly of minerals which crystallized at an early stage of kimberlite evolution (notably olivine) or they are xenocrysts or xenoliths incorporated at depth. Commonly only a small proportion of locally derived (near-surface) country rock fragments form the kernels of lapilli and generally the mantles around such kernels are thin.

(viii) There is no strong correlation between the shapes of pelletal lapilli and the form of their kernels except when the surrounding mantles are very thin. However, in a broad sense lapilli with elongate kernels tend to be more ellipsoidal than lapilli with equi-dimensional kernels. The latter tend to be spherical.

(ix) Elongate phenocrysts in porphyritic lapilli mantles often display varying degrees of concentric orientation and vague concentric layering is sometimes evident in non-porphyritic, fine-grained or cryptocrystalline mantles. There is commonly a complete gradation between lapilli displaying concentric orientation and lapilli lacking the slightest degree of phenocryst orientation.

(x) Small but noticeable proportions of the lapilli in the KIMFIK TKB's occur as broken fragments of originally spherical or subspherical bodies.

(xi) Examples of compound or accreted lapilli representing the coalescence or partial coalescence of individuals occur but are rare.

(xii) Pelletal lapilli in the KIMFIK pipes are mostly extremely fine-grained but in some instances the grain-size is sufficiently coarse for individual groundmass minerals to be recognised. The groundmass of many lapilli in some TKB's ranges from cryptocrystalline to almost isotropic glassy material. A cryptocrystalline, near-glassy, groundmass is characteristic of (but not restricted to) relatively small lapilli, particularly small cored lapilli with thin mantles. Coarser-grained groundmass material is more common among relatively large lapilli.

(xiii) Where the grain size is sufficiently large and the lapilli are sufficiently well preserved to allow positive identification it is apparent that diopside or phlogopite or, more commonly, varying proportions of these two minerals dominate the groundmass assemblages of the pelletal lapilli in the KIMFIK TKB's. In some instances monticellite has been tentatively identified but the low temperature, late-crystallizing, primary groundmass minerals commonly present in kimberlites, such as serpentine, calcite and apatite, are relatively rare or absent in the lapilli. Some predominantly cryptocrystalline lapilli contain substantial quantities of extremely fine-grained diopside ± phlogopite. However, many lapilli are composed only of near-glassy material which cannot be identified optically.

(xiv) The pelletal lapilli are not vesicular.

It is evident from the foregoing summary and the descriptions given in section 7.2.2 that the abundant, small, sharply defined, generally cored, cryptocrystalline pelletal lapilli in the KIMFIK TKB's (e.g. Plates 7.5a-h) are similar to the 'kimberlite pellets' described by Clement (1973) from the Kao pipe in Lesotho. However, larger 'coarse-grained' lapilli resemble the 'autoliths' or 'nucleated autoliths' described by Ferguson et al. (1973) and Danchin et al. (1975). Bearing in mind the different modes of origin were postulated for 'pellets' and 'autoliths' in the above-mentioned publications it is important to establish:

(i) The mode or modes of origin of the pelletal lapilli in the KIMFIK pipes.

(ii) Whether genetic distinctions can be drawn between pelletal lapilli and globular segregations ('autoliths' or 'nucleated autoliths').

(iii) Whether it is possible to trace an evolutionary trend of development that genetically links pelletal lapilli with globular segregations.

The cryptocrystalline 'kimberlite pellets' in the Kao pipe were interpreted as volcanic lapilli composed of solidified (quenched) magma droplets (Clement, 1973). Variations in the appearance of the 'pellets' were ascribed to the droplets having contained varying amounts of incorporated crystalline material prior to their solidification (Clement, 1973). Dissociation of the magma into droplets was ascribed to violent depressurization following explosive breakthrough to surface of volatile-rich kimberlite magma. Such an origin can clearly be considered for at least some of the small, sharply bounded, commonly cored, spherical or near-spherical lapilli which are abundant in some of the KIMFIK TKB's (e.g. E1). In making this evaluation an important factor is the exact nature of the cryptocrystalline groundmass (mantling) material. At least

(iii) There is commonly little or no relationship between kernel shapes and the overall morphology of the lapilli. The lack of correlation is particularly noticeable in cases where euhedral olivine pseudomorphs form the cores of spherical lapilli (e.g. Plate 7.5f).

(iv) Cryptocrystalline material of identical appearance commonly envelopes many types of kernels with diverse chemical and mineralogical compositions.

(v) Contacts between kernels and mantles are often extremely sharp displaying no replacement or reaction relationships (e.g. Plate 7.5f).

In considering the previously listed alternative interpretations of the nature of the cryptocrystalline material in the pelletal lapilli under discussion it is important to note that essentially identical lapilli (except for compositional differences) have been found in a number of other basic or ultrabasic alkaline rocks, notably in certain alnoitic and olivine melilitite diatremes. The lapilli in many of these occurrences are well preserved and, importantly, many are glassy in character (Singewald and Milton, 1930; Rust, 1937; Cloos, 1941; Lorenz, 1979; Williams and McBirney, 1979). The glassy nature of many of the non-kimberlitic lapilli noted above is consistent with the expectedly rapid cooling of magma droplets during explosion /depressurization/fluidization events such as those referred to previously (Clement, 1973) and all the authors listed above favour some form of a quenched magma droplet origin for the lapilli which they have described.

There are striking similarities between the non-kimberlitic lapilli noted above and a proportion of the pelletal lapilli in KIMFIK TKB's such as E1, KOF1 and F1 and, in fact, these lapilli are, to all intents and purposes, identical. It is therefore concluded that some of the cryptocrystalline lapilli in the KIMFIK pipes are consolidated magma droplets reflecting the physical dissociation (disruption) of kimberlite magma (see discussion which follows). Implicit in the above statement is the fact that, apart from any included kernels or phenocrysts, the droplets must have been essentially liquid at the time of their formation. This liquid element of the droplets is now represented by the cryptocrystalline material. The nature of this cryptocrystalline material cannot be fully elucidated by optical examination. It must, however, be poorly crystalline (quenched) primary material; or deuterically altered, originally poorly crystalline primary material; or devitrified glass; or deuterically or secondarily altered glass or devritrified glass; or varying combinations of these alternatives.

To this point in the discussion care has been taken to indicate that an origin by quenching of magma droplets is not applicable to all pelletal lapilli in the KIMFIK pipes. Reservations arise mainly out of consideration of the relatively coarse-grained groundmass of some lapilli. There is also evidence to indicate that some cryptocrystalline material does not in any way represent quenching of liquid in magma droplets.

The groundmasses of some lapilli are sufficiently coarse-grained to be inconsistent with the extremely rapid cooling which would be a mandatory effect of the postulated explosion/depressurization/fluidization events during which the formation of magma droplets is considered to occur (Clement, 1973; this thesis, Chapter 10). Such 'coarse-grained' lapilli are thus better explained by invoking some degree of crystallization prior to physical disruption of crystallizing magmas. This has been done by Lorenz (1979) in interpreting pelletal lapilli in the Swabian olivine melilitite maar-diatremes.

Insofar as the KIMFIK pipes are concerned a genetic model is proposed which accounts for the general features and variations in character of pelletal lapilli according, inter alia, to the degree to which groundmass crystallization preceded disruption of the magmas and the onset of fluidization conditions (in open diatreme systems). The model also takes post-fluidization crystallization into account and establishes relationships between such crystallization (which commonly results in segregationary textures) and pelletal lapilli. Initially (for the purposes of discussion) two situations are considered. In both of these it is accepted that mobile, highly evolved, gas-rich (due to exsolution of volatiles and, possibly, to the assimilation of meteoric water, see section 10.2) kimberlite magma, carrying high temperature phenocrysts together with xenocrysts and xenoliths, has been intruded to high (near surface) levels. The presence of a discrete gas phase is consistent with the views of Mysen (1975), Elthon and Ridley (1979) and Wyllie (1980).

In the first case the formation of lapilli by dissociation of magmas characterized by very little prior groundmass crystallization is considered. It is proposed that very rapid depressurization of such a magma (by explosive outburst) will result in violent disruption of the magma into droplets - the process being fundamentally akin to spraying (i.e. 'fluidization' involving a liquid-gas system, Reynolds, 1954). The process implies a large volume increase as a consequence of the pressure drop accompanying breakthrough to surface.

Considerable variation in character would be expected among the droplets formed in this manner. At the time of formation they would range from essentially liquid droplets to droplets containing varied amounts of variably sized, previously consolidated (prior to magma disruption), solid material (mineral grains and rock fragments). In many instances the droplets might consist mainly of solid kernels surrounded only by thin films or coatings of liquid. The abundance of the latter droplets would relate directly (amongst other things) to the abundance of phenocrystic and xenogenic material in the magma or incorporated at the time of disruption.

The character of many of the cryptocrystalline lapilli described in section 7.2.2, particularly in the E1, KOF1 and F1 TKB's, is entirely consistent with such lapilli being interpreted as the consolidated equivalent of the 'droplet suite' described above. The generally small size of the lapilli suggests that disruption of the magmas concerned must have been extremely violent, i.e. explosive. Such extreme depressurization is considered to relate to the explosions which produced craters at the top of the KIMFIK and other kimberlite pipes (see Chapter 10).

The generally spherical or near-spherical shapes of many lapilli are ascribed to the combined effects of surface tension, turbulent rotation and rapid solidification of droplets during transport within the fluidized systems. Rapid solidification during explosion-induced fluidization would be expected because of adiabatic cooling of rapidly expanding, exsolved and exsolving gas (Dawson, 1967a, 1971, 1980) and to an influx into the system of cool country rock fragments (see Chapter 10 for further discussion). Rapid solidification is indicated by several features of many lapilli in the E1, KOF1, F1 and other KIMFIK TKB's:

(i) The cryptocrystalline (near-glassy) nature of such lapilli (see previous and subsequent discussions).

(ii) The extreme rarity of indented lapilli which could be ascribed to the mutual impact of semi-consolidated or plastic lapilli or to impact between such lapilli and wall rock fragments, xenocrysts, or phenocrysts.

(iii) The scarcity of twinned (dumbbell-shaped) or more complex (botryoidal) welded aggregates of lapilli.

(iv) The rarity with which high level country rock inclusions form the kernels of lapilli.

(v) The occurrence of broken lapilli which indicate that solidification occurred prior to the cessation of fluidization, notwithstanding the conclusion that fluidization events were short-lived (see arguments presented in Chapter 10). Broken lapilli indicate rapid transitions from gas-liquid-solid to gas-solid fluidized sytems.

(vi) The inter-lapilli matrices of most TKB's consist of microlitic minerals (mainly serpentine ± diopside) which are considered to be derived from quenched vapour condensates and are evidence of rapid cooling of the fluidized systems in toto (Chapter 10). Primary inter-lapilli serpentine and diopside matrices are, however, often extensively altered and replaced by clay minerals (due to weathering).

The common occurrence of more-or-less centrally located kernels in many lapilli and the concentric alignment of elongate phenocrysts in numerous lapilli can also be ascribed (with certain possible exceptions - see below), to surface tension and rotation, as proposed by Rust (1937) and Lorenz (1979) for similar lapilli in alnoitic and olivine melilitite diatremes respectively.

The possible exceptions encompass relatively rare robust lapilli (without kernels or consisting of thick mantles relative to kernel diameters) which display concentric layering of generally fine-grained (cryptocrystalline to glassy) material or several layers of oriented phenocrysts (e.g. Plates 7.5a-d).

Lorenz (1979) has argued that similar concentrically layered lapilli in the Swabian diatremes reflect successive liquid accretions around a core, i.e. that such lapilli are the liquid equivalents of the accretionary lapilli, composed of ash grains, which have been described by Moore and Peck (1962). Lorenz's view that "...contact of several droplets prior to solidification apparently caused accretion and thus concentric layering...." is, however, disputed here and is not considered generally applicable to similar concentrically layered KIMFIK lapilli.

The concentrically layered lapilli illustrated and described by Lorenz (op. cit.) and similar bodies in KIMFIK TKB's consist of fairly even shells of mainly fine-grained (cryptocrystalline to glassy) material which may or may not contain several layers of oriented elongate phenocrysts. The individual layers of the lapilli are essentially homogeneous and continuous and, in terms of Lorenz's views, this implies that each successive layer of liquid entirely and evenly surrounded pre-existing (growing) kernels as liquid accretion and crystallization continued. It also implies, in cases where successive layers are similar in appearance, that successive liquid shells invariably contained little or no previously crystallized material, or that the amount and size of such material in successive liquid accretions was effectively constant (to produce successive layers of identical appearance).

In the author's opinion Lorenz's interpretation of concentrically layered lapilli does not sufficiently take into account the variable nature of the 'droplet assemblages' or their rapid solidification. Since, as noted previously, the droplets will and do contain variable amounts of phenocrystic and xenolithic material and also vary in size it is apparent that they will solidify at different rates. The rates of solidification will almost certainly also be influenced by the position of droplets within differentially cooling and differentially fluidized systems (Chapter 10). Under such conditions regular concentric accretion of successive liquid layers seems unlikely. It is much more likely that contact between droplets would commonly involve irregular accretion of lapilli of varying character in various stages of consolidation. However, even irregular accretions are rare (see section 7.2.2) due, presumably, to very rapid solidification of droplets (or to substantial degrees of crystallization prior to the development of explosively induced, open, fluidized systems, see later discussion).

In view of the foregoing comments it is concluded that concentrically layered lapilli are more likely to reflect successive liquid coatings by repetitive immersions of developing, compound (layered) lapilli in magma. Such coatings do, however, imply repetitive movements of such lapilli from more to less fluidized (more 'magmatic') parts of the intrusive systems. An alternative explanation is that concentric layering, at least in part, reflects processes that predate the disruption of partly crystallized magma, during the formation of TKB's. This possibility is considered during the discussion of 'coarse-grained' lapilli which follows shortly.

The absence of vesicles in the lapilli may be due to extensive degassing of the magma prior to explosively-induced disruption. The relatively low volatile content of the lapilli, expressed by the absence or paucity of late-crystallizing, low temperature, residual minerals such as calcite, apatite and serpentine lends support to this interpretation. The absence of vesicles or amygdales may also be related to very low magma viscosities (Dawson, 1980) which facilitate effective degassing.

It was noted previously that the 'magma droplet theory' does not account for all the pelletal lapilli in the KIMFIK rocks. In terms of the generalized genetic model which is developed in this discussion lapilli which do not represent solidified magma droplets are ascribed to varying degrees of crystallization of relatively high temperature minerals prior to the development of open fluidized systems by explosive depressurization. It is concluded that the latter lapilli will typically display coarser grain-sizes than lapilli which represent quenched magma droplets. It should, however, be stressed that the groundmasses of 'coarse' lapilli are also fine-grained compared with typical hypabyssal-facies kimberlites which may, for example, contain poikilitic phlogopite and relatively large diopside, calcite, apatite or monticellite crystals (cf. Plates 7.6h and 7.8a-h with Plates 5.4a, 5.6a and c, 5.6d and 5.7a). The fineness of grain size of even the 'coarse' lapilli is ascribed to relatively rapid crystallization and two general cases can be considered to account for morphologically and texturally different 'coarse' lapilli:

(i) Crystallization, prior to explosive breakthrough (and attendant 'open system' fluidization), under relatively static conditions, of highly vesiculated magma.

(ii) Crystallization, prior to breakthrough, of highly vesiculated magma within turbulent, probably convective, essentially closed system, complex (vapour-liquid-solid) fluidization cells.

Numerous 'coarse' lapilli in the KIMFIK TKB's have more or less centrally located kernels. Under the conditions of case (i) above this could be ascribed to nucleation of relatively early crystallized groundmass components around phenocrysts, xenocrysts or small xenoliths. If extensive vesiculation (i.e. effectively the development of immiscible liquid (± solid) - vapour systems) had occurred prior to this crystallization, or occurred concomitantly, an 'amoeboid' texture might develop (as immiscible vapour phases segregated away from crystallizing centres). At this stage the textures would probably resemble the segregationary textures present in many hypabyssal-facies kimberlites (sections 7.3.3 and 7.3.4 and Plates 5.6f and 7.12d-h) with the low temperature mineral components largely associated with the vapour fractions. If, however, extensive vesiculation and early crystallization of high temperature minerals was followed by explosive breakthrough and concomitant open system fluidization, the segregated 'clots' or 'patches' of relatively advanced crystallization might evolve into pelletal lapilli displaying a variety of morphological and textural features.

If the 'clots' were partly liquid (or plastic) when explosive disruption and fluidization of the crystallizing, vesiculated, intrusions took place they (the segregations) would probably attain sharper, smoother, boundaries and more spherical shapes due to surface tension and the torsional effects of spinning (rotation) within fluidized columns. Furthermore, for similar reasons, if a liquid (or plastic) component was present in the segregations at the time of explosive disruption some degree of concentric orientation of elongate phenocrysts within the modified segregations might be imposed prior to their final consolidation as discrete lapilli. Such orientation would tend to be most pronounced at the edges of the lapilli assuming initial crystallization took place sequentially outwards from phenocrystal (or other) nuclei. Furthermore torsional effects would be most pronounced at the margins of the lapilli.

In terms of the foregoing arguments relatively irregular 'coarse' lapilli may reflect situations where crystallization (prior to open system fluidization) was relatively far advanced. Consequently subsequent morphological or textural modifications of the embryonic lapilli (clots or segregations) during fluidization would be limited.

The degree to which pre-fluidization crystallization involves nucleation around pre-existing solid components probably varies considerably. Such nucleation will depend on many factors, e.g. the rate of cooling, the number of available nucleation sites and the extent to which out-gassing preceded the crystallization of any major, relatively high temperature, groundmass phase. In instances where early groundmass crystallization did not involve extensive nucleation around phenocrysts (or other solid components) explosive disruption and fluidization would probably result in the formation of a relatively high proportion of lapilli without kernels. It is also likely that such lapilli would tend to be irregular and that they would not display concentric internal structures (provided pre-fluidization crystallization was relatively far advanced so that extensive imposition of concentric texture did not occur during fluidization).

Although (as indicated above) much morphological variation could be displayed by lapilli which reflect considerable early crystallization under relatively static conditions (prior to breakthrough) it is unlikely that such lapilli would be characterized by very high degrees of sphericity, or by particularly smooth, sharp, boundaries or by well developed concentric orientation of phenocrysts. Such features may, however, reflect early crystallization under the conditions outlined in case (ii) above, i.e. substantial degrees of pre-breakthrough crystallization of highly vesiculated magma in turbulent, more-or-less closed, fluidization cells. The manner in which such subsurface systems may develop is discussed in section 7.4 where it is also concluded that crystallization under such conditions can lead to the development of globular segregations which are equated with the 'nucleated autoliths' of Danchin et al. (1975). Many such segregations are practically identical in character to well formed, spherical, 'coarse' pelletal lapilli although, as indicated in section 7.4, they form under essentially hypabyssal conditions.

In considering the total juvenile lapilli suites in the KIMFIK TKB's it is important to realize that the conditions under which globular segregations form (section 7.4) may be developed prior to the explosive outbursts which result in the formation of TKB's. Globular segregations (formed under hypabyssal conditions) may, following subsequent explosive disruption and open system fluidization, be incorporated in TKB's together with pelletal lapilli.

It is evident from the foregoing discussion that in extreme instances juvenile lapilli suites may consist of spherical bodies formed by quenching of magma droplets, or by varying degrees of pre-fluidization crystallization with varying degrees of subsequent modification (pelletal lapilli) or they may, in fact, be composed of globular segregations formed under hypabyssal conditions. Kimberlite lapilli suites may also include true autoliths (i.e. cognate inclusions).

The variations in character displayed by juvenile lapilli in individual KIMFIK TKB's suggest, however, that commonly the lapilli assemblages are 'mixed' in the sense that they include lapilli formed in all of the above ways although pelletal lapilli sensu stricto dominate the assemblages.

The actual proportions of lapilli reflecting different modes of formation are difficult to evaluate because gradational (intermediate) situations would be expected to blur genetically induced differences in the character of the lapilli. Furthermore subsequent alteration (discussed later) may also in many instances have blurred any distinguishing features.

Major differences in the proportions of different juvenile lapilli are likely between individual intrusions because the degree to which crystallization takes place prior to, rather than during, breakthrough will be governed by fundamental (often widely different) compositional characteristics and a host of intrusion parameters (e.g. temperature, degree of vesiculation, extent of assimilation of groundwater, volume of magma, rate of intrusion, rate of cooling, etc.).

Since the above parameters will also vary from place to place in individual intrusions lapilli reflecting different modes of origin (and a host of transitional types) would be expected to form consecutively in specific parts, or simultaneously in different parts, of individual fluidized columns. Furthermore different types of lapilli may be zonally distributed. The latter possibility has not been tested but is theoretically likely.

One feature of most of the KIMFIK TKB's which strongly suggests that the lapilli assemblages actually include some hypabyssal globular segregations (nucleated autoliths) is the presence of some (relatively rare) large spherical bodies (up to 10cm or more in diameter). These bodies display diverse groundmass compositions but characteristically they contain low abundances of late-crystallizing, low temperature, kimberlite minerals. Such large and invariably 'coarse-grained' bodies cannot have been formed during short-lived, open system, fluidization events. Furthermore they are unlikely, because of their high degree of sphericity often coupled with internal concentric structure, to represent pre-breakthrough crystallization, except under the conditions outlined in section 7.4 for the origin of globular segregations.

As indicated in section 7.2.2 diopside not only occurs as a primary mineral in pelletal lapilli but often occurs as a fine-grained to cryptocrystalline deuteric replacement product of some pelletal lapilli. Deuteric diopsidization of some lapilli raises the possibility (suggested by E.M.W. Skinner, personal communication) that all cryptocrystalline material within pelletal lapilli is essentially deuteric diopside and that all pelletal lapilli are fundamentally 'pre-fluidization' in origin, i.e. quenched magma droplets do not form part of pelletal lapilli assemblages. This concept may apply to a considerable degree in the case of some TKB's but is considered untenable in other instances for one or more of the following reasons:

(i) The smooth, sharp boundaries and perfect spherical shapes of many small lapilli in some TKB's are more consistent with the droplet theory.

(ii) The fact that many of the lapilli in some TKB's are identical in character to those of other rock types which occur in similar diatremes and contain identifiable primary glass (see earlier discussion).

(iii) The fact that deuteric processes need not be restricted to the replacement of 'coarse' lapilli but could equally replace the glassy material or devitrified glass of quenched magma droplets.

(iv) The difficulty of explaining why some lapilli are entirely cryptocrystalline (i.e. deuterically altered, diopsidized, 'coarse' lapilli according to the foregoing proposition) while adjacent lapilli have remained 'coarse-grained' and unaltered. Such situations are particularly difficult to explain because deuteric diopsidization (as shown later) is primarily a post-fluidization phenomenon.

(v) The unlikelihood (as indicated previously) of the cryptocrystalline material always being composed essentially of diopside or that it is in fact deuteric in origin.

(vi) The unlikelihood that highly mobile, fluid, gas-rich magmas such as kimberlites do not produce juvenile lapilli while such lapilli are abundant components of a wide variety of other basic and ultrabasic volcanic rocks (e.g. Williams and McBirney, 1979).

The formation of discrete, sharply bounded, smooth-surfaced pelletal lapilli has been considered at length in the foregoing discussion. The remaining part of this section is aimed at interpreting the origin of the more irregular post-fluidization segregations described in section 7.2.2 and assessing the manner in which they relate to, or represent modifications of, earlier-formed pelletal lapilli. Three situations are recognized; the post-fluidization segregations may be entirely primary, or they may have grown around (nucleated against) and partly replaced pre-existing pelletal lapilli (or other components), or they may entirely replace pre-existing pelletal lapilli (or other components).

The occurrence of many small euhedral crystals of diopside, ranging from isolated single grains to relatively large, rather vaguely defined, clusters with serpentine bases and/or the occurrence of outward growing needles fringing partly or extensively altered pelletal lapilli, is evidence that considerable diopside in the inter-pelletal matrices of KIMFIK TKB's has a primary origin. Such diopside (Plates 7.8c and 7.9c-d) also provides convincing evidence that crystallization took place under quiescent (stagnant) conditions. If crystallization occurred prior to, or during, turbulent fluidized intrusion delicate, radially disposed, diopside crystals within discrete segregations or fringing pelletal lapilli would not have been preserved.

In view of the foregoing considerations it is concluded that, although primary diopside segregations may start to form during the rapid cooling which accompanies post-breakthrough fluidization (Chapter 10) most of the crystallization occurs during the final stages of consolidation of TKB's, i.e. post-fluidization. Furthermore it is concluded that this diopside and the generally abundant accompanying serpentine crystallizes from vapour phase condensates (which are probably contaminated by meteoric water, see Chapter 10).

The intimate association between primary diopside in the segregations and deuteric diopside (which replaces pre-existing minerals and pelletal lapilli) indicates that primary crystallization and deuteric alteration were essentially contemporaneous (both processes are mainly post-fluidization phenomena).

The occurrence of relatively abundant post-fluidization diopside in some TKB's and the comparative paucity of similar material in others may in part be controlled by the extent to which individual magmas are dissociated by explosive outburst, by the extent of pre-breakthrough crystallization and by the overall volatile content of the systems. In cases of extremely violent explosion and, consequently, extremely extensive and rapid depressurization, residual, very highly evolved magma fractions may be more or less entirely vaporized. Thus derivation from abundant vapour condensates (Clement, 1979; Scott and Skinner, 1979) would account for abundant microlitic, post-fluidization, diopside and serpentine (reasons for the condensate minerals being restricted largely to diopside and serpentine are considered in Chapter 10). Alternatively, if depressurization was less violent and pre-fluidization crystallization was less advanced, a vapour phase plus relatively large magma droplets might develop. The latter, together with 'pre-fluidization' lapilli, could result in relatively unaltered juvenile (mainly pelletal) lapilli being abundant rather than post-fluidization microlitic segregations.

In most TKB's which contain relatively abundant post-fluidization diopside microlites some degree of segregation of inter-pelletal diopside and serpentine is usually evident. This usually reflects nucleation of diopside around solid particles or clustering of diopside crystals and microlites prior to the consolidation of serpentine (the last mineral to crystallize). However, in some cases the degree of diopside/serpentine segregation is very limited and the TKB's are regarded as having uniform textures (e.g. Scott and Skinner, 1979; Clement and Skinner, 1979). Limited segregation may be related mainly to condensate composition or to rate of quenching.

The abundance of post-fluidization serpentine as an apparently primary post-fluidization mineral may be due to the incorporation of considerable meteoric water into open, TKB-forming, fluidized systems. The abundance of post-fluidization diopside, which crystallizes under similar conditions to the serpentine, is discussed in section 10.3.2.

In summary this investigation of the textures of TKB's has shown (as indicated in Table 2.3) that four textural varieties of TKB's can be recognized:

(i) Pelletal-textured TKB's characterized by the presence of abundant juvenile (mainly pelletal) lapilli that originated prior to and/or during gas streaming (lean-phase, open-system fluidization).

(ii) Segregationary-textured TKB's characterized by the presence of segregations which post-date fluidization and commonly consist mainly of fine-grained to microlitic diopside. The segregations commonly reflect nucleation of diopside around pre-existing solid components including pelletal lapilli.

(iii) Uniformly-textured TKB's which contain few or poorly developed pelletal lapilli and where little or no post-fluidization segregation has occurred.

(iv) Crystallinoclastic TKB's which contain little interstitial, primary, post-fluidization, matrix material (i.e. post-fluidization diopside and/or serpentine).

Relationships between these four textural varieties are entirely gradational and most commonly TKB's have textures reflecting combinations of (i) and (ii) above, i.e. they are pelletal-textured rocks in which post-fluidization segregations have also formed. Post-fluidization diopside has nucleated around (and often partly replaced) pelletal lapilli and other relatively large components. It is likely that gradations between all four textural types may occur (vertically or horizontally) in some individual intrusions but this has not yet been established.

7.3 HYPABYSSAL-FACIES KIMBERLITES

Hypabyssal-facies kimberlites and kimberlite breccias (as defined in Chapter 2, Section 2.3.5) are characteristic of the root zones of kimberlite pipes although they are also present, mainly as minor intrusions or cored-out relicts of major intrusions (Chapter 3) at higher levels. These variants of kimberlite are therefore better exposed in the Kimberley mines, which have been worked to greater actual and relative depths, than have the Finsch and Koffiefontein pipes.

Many of the kimberlites and kimberlite breccias exposed in the mines are extremely fresh and well preserved compared with most surface or near-surface exposures of similar petrological varieties. Thus, in contrast to the latter (and most exposures of diatreme-facies kimberlites), they are amenable to detailed petrographic studies.

7.3.1 Xenoliths in hypabyssal-facies kimberlites

The kimberlites and kimberlite breccias in the KIMFIK pipes generally contain mixed assemblages of xenoliths including exotic (derived from the upper mantle) and locally derived crustal varieties.

Exotic inclusions vary considerably in abundance and type among the intrusions. In some kimberlites (e.g. D10 at Dutoitspan, Plate 6.1f) such inclusions locally make up $\sim$15 vol.% of the rock which is thus, in places, a kimberlite breccia (as defined in Chapter 2) in which the xenolithic components are of upper mantle origin. Abundances of this order are, however, extremely rare and most hypabyssal-facies intrusions in the KIMFIK pipes contain around 1,0 vol.% or less (often much less) incorporated mantle xenoliths.

The nature and relative abundances of different mantle-derived inclusions (which are invariably well rounded) varies considerably in different intrusions. For example the nodules in D10 consist almost entirely of garnet-free harzburgites (Plate 6.1f). In contrast other intrusions (e.g. W3 at Wesselton and F2 at Finsch) contain a relatively high proportion of garnet lherzolites. Yet others (e.g. D5 at Dutoitspan and W3 at Wesselton) contain abundant MARID-suite (Dawson and Smith, 1977) nodules.

Crustal xenolith assemblages in the KIMFIK kimberlites and kimberlite breccias are generally heterogeneous (Plates 7.1e-h) but in most cases the assemblages are more restricted than in the tuffisitic kimberlite breccias. Inclusions in hypabyssal-facies intrusions in the Kimberley pipes consist mainly of Ventersdorp lava and Basement fragments, the latter composed mainly of granitic or gneissic types with subordinate amphibolite and rare talc schist. In most intrusions relatively small amounts of shale (and other sedimentary rocks) and dolerite are also present. However, rare intrusions are devoid of Karoo-derived xenoliths (e.g. DB 2) while others contain inclusions derived from a single source only (e.g. the D1 intrusion at Dutoitspan contains only Basement-derived xenoliths). Although detailed studies have not been carried out it is apparent that both the abundance of xenoliths and the nature of the xenolith assemblages varies from place to place in some intrusions.

For example around the 550m level near the southern boundary of the De Beers pipe, the DB 2 inclusion assemblage consists almost solely of Ventersdorp lava fragments. However in the central part of the intrusion Basement fragments are also abundant and the overall xenolith content is higher.

The crustal xenoliths in the hypabyssal-facies kimberlites have more restricted size ranges than those in diatreme-facies kimberlites. Very large floating reef masses (section 3.6) do not occur in the hypabyssal-facies variants and most of the incorporated xenoliths in these rocks range between 0,5 and 5,0 cm in size (Plates 7.1e-h). Only a small minority of inclusions exceed 10,0cm with very rare inclusions measuring up to ~1,0m in diameter. Microscopic inclusions are rare in comparison with the abundance of microxenoliths in TKB's.

Xenoliths in the hypabyssal-facies kimberlites tend to be highly irregular in form compared with those in diatreme-facies kimberlites (cf. Plates 7.1a-d and 7.1e-h). This irregularity reflects substantial degrees of magmatic corrosion and metasomatic alteration of the inclusions in the hypabyssal-facies intrusions.

The most outstanding differences between xenoliths in hypabyssal-facies and diatreme-facies kimberlites lies in the extent to which the xenoliths have been altered. In contrast to the generally unaltered or only slightly altered (metasomatized) nature of most xenoliths in TKB's, the inclusions in the hypabyssal-facies intrusions have often been extensively metasomatized, frequently in zonal fashion (Plates 7.10d-e). Commonly this metasomatism involves partial to complete replacement of the xenoliths by minerals which also occur as primary groundmass phases in the kimberlites. The minerals involved include monticellite, diopside, calcite, phlogopite, serpentine, perovskite and opaque oxides. The most common metasomatic minerals in the xenoliths are calcite, serpentine and phlogopite.

In addition to the formation of the kimberlite minerals noted above metasomatism in some instances results in hydrous minerals such as zeolites (mainly natrolite) and pectolite forming in and adjacent to xenoliths (see discussion in section 5.12). The extent of metasomatism of the xenoliths is due to the volatile-rich nature of the magmas and the relatively high temperatures and slower cooling of hypabyssal-facies kimberlites relative to TK's or TKB's. Most of the inclusions in the latter are incorporated during short-lived fluidization events and hence are not severely altered (see Chapter 10).

7.3.2 Mineralogical variations in the hypabyssal-facies kimberlites.

In Chapter 2 the mineralogical classification used in this thesis was described and it was noted that kimberlites could be classified in terms of five primary minerals (diopside, monticellite, phlogopite, calcite and serpentine). In addition, in rare instances olivine could probably also be used for classification purposes. The hypabyssal-facies kimberlites in the KIMFIK pipes display extensive mineralogical diversity which is reflected primarily by the presence or absence and relative abundances of all the minerals listed above except groundmass olivine. In addition differences are imposed by variations in the abundances of several minerals which generally occur in accessory amounts. Notable among these minerals are apatite, perovskite and opaque oxides.

The mineralogical diversity of the KIMFIK intrusions is illustrated by the modal analyses given in Tables 7.4 and 7.5, by the detailed descriptions given in section 7.5, by the abbreviated descriptions given in Appendix I and by many photomicrographs in Volume 2 of this thesis.

Wagner (1914) viewed the Kimberley and neighbouring pipes as a province of 'basaltic' kimberlites (see Chapter 2) thereby implying a paucity of mica in the rocks. Based on the results of this study three main comments on the foregoing statement can be made:

(i) The term 'basaltic kimberlite' has misleading implications and should be dropped. As noted by Mitchell (1970), Skinner and Clement (1979) and Dawson (1980) kimberlites are not basaltic in character, either from textural or mineralogical standpoints.

(ii) Phlogopite kimberlites are present among the intrusions in the Kimberley pipes; this kimberlite 'province' is not restricted to the equivalents of Wagner's 'basaltic' kimberlite.

(iii) The mineralogical variability of the KIMFIK pipes is much more extensive than previously supposed (section 7.5, Appendix I and Table 7.4). The intrusions are clearly not restricted to monticellite- and olivine-rich rocks (Wagner, 1929) or to kimberlites with groundmasses of secondary carbonates and serpentine (Wagner, 1914).

Several other features relating to the mineralogical diversity of the KIMFIK hypabyssal-facies intrusions are worth noting and are tabulated below:

(i) Pronounced mineralogical contrasts are evident between the different intrusions in individual pipes as well as in the KIMFIK group as a whole (see sections 7.5 and 7.6).

(ii) Notwithstanding (i) above some of the discrete intrusions in individual pipes are very similar to one another, although field relations indicate that they must have been emplaced at different times. The W3 and W8 intrusions at Wesselton are good examples.

(iii) Similarly, neighbouring pipes may contain intrusions which are petrographically identical in character. The DB2 and D2 intrusions (in the De Beers and Dutoitspan pipes respectively) provide a good example. Although both these intrusions display minor local variations in character they are in general mineralogically and texturally indistinguishable (section 7.5 and Table 7.6).

(iv) In contrast to the DB2/D2 situation most of the intrusions in individual pipes cannot be closely matched with kimberlites in neighbouring pipes although mineralogically similar varieties are evident.

(v) Mineralogical variations within individual intrusions are often evident. These variations occur on microscopic and megascopic scales. The variations mainly reflect the effects of: flowage differentiation; different concentrations of volatiles and hence volatile-rich, late-crystallizing minerals; varying contents of country rocks; different degrees and types of alteration; and varied rates of cooling. Some examples of mineralogical variations within intrusions are given in section 7.5.

(vi) With minor exceptions (some late-stage dykes) all the kimberlites in the Finsch pipe contain abundant matrix (phenocrystal or groundmass) phlogopite and varied but usually substantial amounts of diopside (e.g. Table 7.4). This abundance of phlogopite and diopside is coupled with a paucity of perovskite and opaque oxides (e.g. Plates 5.6a, 7.11a, 7.13c-h and 7.14a-b). In contrast many intrusions in the Kimberley pipes contain very little phlogopite, no diopside and relatively abundant perovskite and opaque oxides. Regional mineralogical distinctions of this type are common (Dawson, 1980) and reflect the fact that the Finsch and Kimberley pipes form parts of different kimberlite provinces.

(vii) Mineralogical characterization of the KIMFIK kimberlites is generally imposed by the groundmass minerals rather than by phenocrystal phases. However, in some instances intratelluric phlogopite phenocrysts are extremely abundant and strongly influence mineralogical classification (e.g. Plate 5.3g).

7.3.3 Textural features of hypabyssal-facies kimberlites.

In hand specimen the hypabyssal-facies intrusions (kimberlites and kimberlite breccias, as defined in Chapter 2) in the KIMFIK pipes are dark grey, greenish, greyish-brown or grey-black (largely depending on the groundmass mineral assemblage and degree of freshness), moderately hard, competent (although some are brittle) rocks. The xenoliths in the kimberlite breccias have, however, frequently been metasomatized (commonly carbonatized or serpentinized) and, because of the resulting pale colours, stand out against the dark backgrounds of the host kimberlites (e.g. Plates 7.1e-h). A mottled overall appearance is thus often apparent.

The most prominent macroscopic textural feature of the hypabyssal-facies kimberlites is normally the porphyritic character imposed by the presence of abundant anhedral macrocrysts (mainly of olivine or its alteration products). Such macrocrysts usually make up between 10 and 30 vol.% of the rocks and usually form 30-60 vol.% of the total olivine content. The macrocrysts are set in aphanitic matrices. Entirely aphanitic kimberlites (see Chapter 2) are rare. They are restricted to minor dykes and, less frequently, sills associated with the pipes, or to local areas of major pipe intrusions where they reflect flowage differentiation (Plate 5.10d). In the latter instances they occur as narrow zones or bands commonly less than 1m in width. In addition to flowage differentiation other crystal fractionation processes may separate out macrocrysts prior to, or during, final intrusion. These processes include gravitative settling and filter pressing.

Transitional relationships exist between aphanitic and macroporphyritic kimberlites. Kimberlite breccias consist of exotic and/or cognate rock fragments set in igneous matrices composed of aphanitic or macroporphyritic kimberlites and, similarly, these breccia matrices may be transitional in character.

As indicated in Chapter 2 the term 'macroporphyritic' is to some extent misleading as most 'macrocrysts' are probably xenocrysts not phenocrysts (see for example the discussion in section 6.2.1) and M.E. McCallum (personal communication) has suggested that the term 'macrocrystic' be substituted. Insofar as the porphyritic character of kimberlites is concerned it should be stressed that both aphanitic kimberlites and the equivalent matrices of macroporphyritic (macrocrystic) kimberlites are porphyritic but the bulk of the phenocrysts (mainly euhedral olivine crystals) measure less than the 0,5mm in size and are not easily visible macroscopically. The term microporphyritic is therefore appropriate. To this extent all the KIMFIK kimberlites have porphyritic textures, one of the most common characteristics of lavas and rapidly cooled hypabyssal rocks (Cox et al., 1979).

Directional structures in the KIMFIK kimberlites and kimberlite breccias are not common but are more frequently evident in these textural variants than in diatreme-facies rocks. Usually directional fabric (Plates 5.10d and 7.10g) has been imposed by magmatic flow (the different types of flow structures evident in the KIMFIK pipes were tabulated in section 3.5.1). Good examples of such structures have been documented by Dawson and Hawthorne (1970), particularly in minor intrusions (dykes and sills) where they tend to be best developed.

One of the structures described in section 3.5.1 is the parallel orientation of elongate components of the rocks. It should be noted that although such orientation (which occurs on megascopic and microscopic scales) is often vertical or subvertical, this is not always the case. For example, in the southern part of the De Beers pipe, on the 595m and 720m levels elongate country rock xenoliths in the DB 2 kimberlite are locally oriented at angles of 0°-30° to the horizontal. These flat angles of orientation are evident towards the margins of the intrusion and strongly suggest some form of convective overturn during the emplacement of this kimberlite. Non-vertical, probably convective, intrusion is also indicated by the low angle of orientation of xenoliths in the DB 1 kimberlite. These xenoliths are oriented parallel to a flat-dipping contact between the DB 1 intrusion and an underlying contact breccia (Fig. 3.23). The breccia has been partly stoped out by the DB 1 intrusion.

As noted previously euhedral (or subhedral) olivine crystals (which are often altered) are usually the dominant microphenocrysts (e.g. Plates 5.1a-b and 6.1a, c and d) in the KIMFIK intrusions. These microphenocrysts commonly measure between 0,05 and 0,5mm (see section 5.2.2). In most intrusions microphenocrystal habits are also displayed by other minerals. These intratelluric crystals include one or more of the following minerals: phlogopite, diopside, perovskite and part of the opaque oxide assemblages (e.g. Plates 5.3h, 5.6b and d, 5.9g). In rare instances phlogopite is the dominant microphenocrystal phase (Plate 5.3g).

Unlike intratelluric phenocrysts, which may display prominent parallel flow orientation, the groundmass minerals (sensibly equivalent to post-intrusion minerals in the KIMFIK pipes, section 7.7) commonly display no directional fabric. Several other characteristic primary groundmass textures can, however, be recognized. These textures reflect specific relationships between particular groundmass minerals or characteristic distributions of certain groundmass minerals.

A granular texture is often exhibited by small ($\sim$0,02mm) subhedral to euhedral monticellite crystals (Plates 5.3a and 5.7a). This granular texture is particularly prominent when the relatively high relief monticellite grains are closely packed and are set in a base of low relief serpentine with or without accompanying calcite. The minerals of the base may be extremely fine-grained and occur interstitially between monticellite grains or they may occur as larger, commonly anhedral, optically continuous patches enclosing abundant, closely packed, monticellite crystals.

A prominent sieve texture is often displayed by relatively coarse-grained, subhedral to euhedral groundmass phlogopite laths due to crowding of the laths by many small inclusions of earlier-crystallizing minerals. Often the most abundant inclusions are monticellite crystals but equant opaque oxides and perovskite grains may also be abundant (Plates 5.4c-g). It should be noted that in some instances the distinction between granular and sieve textures is merely an artefact of the optical prominence of the minerals concerned. Thus in some instances abundant closely packed monticellite grains with high relief are set in spongy calcite or serpentine areas with low relief. The abundant monticellite grains stand out as a granular aggregate while the minerals of the base are optically insignificant. Conversely, in other intrusions phlogopite crystals, which enclose many small monticellite crystals and other grains, are also optically prominent (particularly when zoned) and hence a prominent sieve texture is evident (cf. Plates 5.3a and 5.4d-f).

Phlogopite and, to a lesser extent, calcite also occur as relatively large poikilitic plates enclosing other groundmass and phenocrystal phases (Plates 5.4a-b). There is no quantitative distinction between poikilitic and sieve textures but the host grains of the latter are smaller than those of the former and contain more closely packed inclusions (cf. Plates 5.4a and d).

A mosaic texture is displayed by some calcite-rich kimberlites. Calcite occurs as relatively large (usually 0,1-0,3mm), anhedral to subhedral, irregular to polygonal plates which form an interlocking mosaic. Such calcite mosaics incorporate scattered grains of other minerals and individual calcite grains are partly moulded around relatively large phenocrysts such as olivine.

Some phlogopite-rich kimberlites and kimberlites containing abundant phlogopite and diopside have extremely fine-grained groundmasses (in part cryptocrystalline). In these rocks a felty groundmass texture is often present composed mainly of intimately intergrown and overlapping laths and platelets of phlogopite with acicular and/or prismatic crystals of diopside (when the latter mineral is present). Examples are shown in Plates 5.5a and d and 5.6a. Such groundmasses reflect rapid cooling and crystallization (quenching). Other 'quench' textures include radiating (spherulitic) clusters of acicular apatite crystals (Wyllie et al., 1962) and skeletal crystals of apatite (see Plate 5.9b and Wagner, 1914) and, rarely, opaque oxides.

Intergranular-intersertal textures are often evident and reflect the crystallization of late-forming components in the spaces between earlier-crystallized components. Trachytic textures are often displayed by intrusions which contain abundant phlogopite phenocrysts (Plates 5.3g-h).

The intratelluric components of the matrices and the groundmass components may be uniformly spread through the kimberlites in an orderly, more-or-less even fashion. Commonly the best examples of such uniform textures in the hypabyssal-facies rocks are reflected by those kimberlites which can be interpreted as volatile-poor (degassed) intrusions, because they have precipitated groundmass minerals such as monticellite and/or have crystallized little or no late -crystallizing calcite or serpentine (e.g. Plates 5.1a-c, 5.3a, 5.7a, 5.9g and 7.10g).

In contrast to the uniformly textured kimberlites noted above many hypabyssal-facies intrusions within or associated with the KIMFIK pipes display distinct segregationary textures. Such textures reflect the crystallization of certain minerals (or groups of minerals) in discrete areas of the rocks, often in well defined 'pools' or segregations. These segregations are of considerable interest because they often facilitate interpretation of the detailed crystallization history of the kimberlites and, in some instances, they are indicative of physical changes in the nature of the evolving fluid phases during the consolidation of the rocks (sections 7.3.4 and 7.4).

The degree to which segregationary textures are developed varies considerably, not only between intrusions, but also on a local scale within individual intrusions (even on a hand specimen or thin section basis). Some intrusions only contain small isolated segregations (Plates 7.10h and 7.11a) within matrices which are otherwise uniformly textured. There is a complete gradation from such poorly segregated (or essentially uniform) examples to kimberlites in which segregations are sufficiently abundant to have linked up to form more or less continuous, irregular networks (Plates 5.5f-h and 7.12b-h).

The most spectacular (and common) examples of segregationary textures involve the separation of minerals which crystallized 'early' and 'late' in the paragenetic sequences of the kimberlites (and kimberlite breccias) concerned (section 7.7). Commonly pools or segregations of 'late-crystallizing' minerals occur within host material which incorporates earlier-crystallized components. The segregations are often irregular in character with lobate, cuspate, amoeboid or compound curvilinear boundaries (Plates 5.5f-h, 5.6e-h, 7.12b-h and 7.13a-b), particularly where coalescence of individual segregation bodies has occurred. However, where best developed individual segregations are globular in character and display subspherical or subellipsoidal forms. In certain extreme cases the late-forming components are so abundant that the host/segregation relationships described above are reversed. 'Late' material forms a continuum enclosing areas or patches of 'early' material. Some of the latter rocks are best interpreted as transitional in character between hypabyssal- and diatreme-facies kimberlites (see section 7.4).

Phillips (1973) has drawn attention to the existing complex nomenclature relating to globular structures in igneous rocks. He has pointed out that some of the terms used to describe such bodies (e.g. amygdale, ocellus, orbicule, spherulite, variolite) originally had precise descriptive and/or genetic connotations although they are frequently used somewhat indiscriminately. The globular structures in some of the hypabyssal-facies KIMFIK intrusions are considered to have originated in more than one way (sections 7.3.4 and 7.4) but all reflect some form of segregation of components. Consequently they are referred to in general terms as globular segregations in this thesis.

Calcite is frequently the most abundant and may be the only component of many 'late-crystallizing' segregations (e.g. Plate 7.11b). Primary 'serpentine' is also frequently abundant. This serpentine commonly occurs as massive isotropic 'serpophite' (e.g. Plate 7.9f) or as very small spherulitic aggregates (e.g. Plate 7.11c). As noted in section 5.9 much serpophite may represent amorphous colloidal deposits and some spherulites may be secondary as a consequence of devitrification. Serpentine commonly occurs together with calcite in segregations (e.g. Plates 5.8a-h, 7.9f, 7.10h, 7.11d-h and 7.12b-h) but also forms monomineralic pools. Other common minerals in segregations are apatite and phlogopite (Plates 5.5c, 5.9a, 5.10a, 711a and 7.14d-e) while pectolite and zeolites such as natrolite (Scott-Smith et al., in press; Kruger, 1980) are relatively rare components (Plates 5.6e and 7.4a). Although generally subordinate in amount to calcite and/or serpentine each of the other minerals listed above may be the dominant (or even the only) constituent of individual segregations.

In addition to exhibiting varied mineralogical compositions (often on a local scale) reflecting different combinations and abundances of relatively late-crystallizing minerals, the segregations in the KIMFIK pipes display wide variations in form and size. Contrasting contact features and internal textures are also evident.

In some (relatively rare) instances there is evidence of the formation of distinct segregation boundaries at an early stage of segregation development i.e. prior to the crystallization of the fluids within the segregations and the residual melt in the surrounding host material. It is also apparent that, subsequent to the formation of the original segregation boundaries, some crystallization occurred contemporaneously within and outside the segregations. In the discussions which follow such segregations are termed Type 1 segregations.

In contrast to Type 1 segregations other late-stage segregations consist of minerals which effectively post-date crystallization of the host material. This paragenesis is commonly unequivocably indicated by the presence, within the segregations, of inward-growing crystals which have nucleated against the fringing, sensibly solid, host material. Such segregations are henceforth referred to as Type 2 segregations. Type 2 segregations are common in the KIMFIK kimberlites and good examples are shown in Plates 5.6g-h, 5.8e-f and 7.12b-h.

A third broad group of late-stage segregations (Type 3 segregations) also occurs in some of the KIMFIK intrusions. Unlike Type 1 and Type 2 segregations these segregations do not have well-defined (sharp) boundaries and contemporaneous crystallization (at least for a time) has occurred within segregation and host areas alike. In this respect the Type 3 segregations resemble Type 1 segregations but they differ from the latter in that they are devoid of any features indicative of earlier-formed boundary surfaces and are much more irregular (cf. Plates 7.11a-b and 7.11d-e).

7.3.4 The origin of segregationary textures

It is only during the last decade that attention has been drawn to the occurrence of distinct mineral segregations in hypabyssal-facies kimberlites. Although several references to such structures have been made during this period (Dawson and Hawthorne, 1973; Clarke and Mitchell, 1975; Clement, 1975; Mitchell, 1979; Clement and Skinner, 1979; Kruger, 1978; Pasteris, 1980a; Snowden, 1981; Donaldson and Reid, in press; Clement et al., in press) relatively little comment regarding their origin has been forthcoming.

Based on the examination of a kimberlite dyke in the wallrocks of the De Beers pipe Donaldson and Reid (op. cit.) alluded to three possible ways in which globular segregations may form in hypabyssal-facies kimberlites: they may be due to liquid immiscibility; or they may represent segregations of melt; or they may be infilled gas cavities. If either of the last two possibilities are accepted several questions arise concerning the derivation and physical state of the fluids. Are they residual melts, gas condensates, hydrothermal fluids or gas-melt-hydrothermal fluid combinations? If the segregations are infilled gas cavities were the infilling fluids primary or secondary? It could be argued that none of the foregoing modes of origin are applicable. The segregations, or at least some of them, might, for example, represent nucleation centres for specific minerals or groups of minerals. Alternatively 'segregations' might, at least in part, represent altered, incompletely assimilated inclusions or they might be extensively metasomatized xenoliths or xenocrysts or deuterically altered early phenocrysts. The various possible modes of origin are discussed individually below.

Central nucleation

There are undoubtedly instances in hypabyssal-facies kimberlites where clusters or segregations of minerals within the rocks are best explained by assuming that a particle (usually a mineral grain) has acted as a nucleus around which similar grains or some other mineral has clustered. Examples include 'necklace textures' which involve peripheral clustering of opaque oxide and perovskite crystals around much larger olivine or serpentinized olivine grains (such textures can be regarded as unusual forms of glomeroporphyritic aggregates). Good examples of central nucleation are also provided by clusters of prismatic to acicular diopside crystals, commonly around earlier-formed phenocrysts or around xenocrysts or microxenoliths. Central nucleation is not, however, an acceptable mode of origin for the segregations described in section 7.3.3 and illustrated, for example, by Plates 5.5c, 5.5f-h, 5.6e-h, 5.8a-h, 5.10a, 7.9f, 7.10h, 711a-h, 7.12b-h and 7.13a-b. In the majority of cases such an origin is precluded by the distribution and general textures of the minerals in the segregations and by the fact that crystallization has clearly proceeded inwards (e.g. Plates 7.12e-h), often from almost entirely crystalline margins. Central nucleation implies that crystallization proceeds outwards and it commonly gives rise to spherulitic or radial growth (Phillips, 1973). Thus, although the crystallization of some of the components in certain segregations may have been initiated by central nucleation (as shown by the presence of spherulitic serpentine and radial aggregates of apatite, Plates 7.11c and 5.9a) such nucleation is not responsible for the actual presence of the segregations.

Alteration and metasomatism

There are no valid grounds for supposing that the segregations under discussion represent metasomatized or otherwise altered xenoliths, xenocrysts or phenocrysts. The distribution and shapes of many segregations are entirely inconsistent with such an origin, no relicts of the original xenoliths or minerals are ever observed and the segregations in no way resemble actual examples of xenoliths or individual minerals which have been metasomatized by kimberlite liquids (e.g. Plates 7.1e-h and 7.10d-e).

Liquid immiscibility

Mainly as a result of Bowen's conclusions (Bowen, 1928) liquid immiscibility was for many years (following early enthusiasm) regarded as a rare, unimportant and largely unproven process in igneous petrogenesis. In recent years the pendulum has, however, swung to a considerable extent, as experimental proof of both silicate and carbonate-silicate liquid immiscibility, under geologically plausible conditions, has accumulated. At the same time a considerably body of field evidence favouring (or at least entirely consistent with) liquid immiscibility has been gathered (e.g. see reviews by Ferguson and Currie, 1971; Roedder, 1979 and Philpotts, 1979).

In kimberlites the presence of 'emulsion' or 'amoeboid' textures has been ascribed to carbonate-silicate liquid immiscibility (e.g. Clement, 1975; Mitchell and Fritz, 1975; Mitchell, 1979a) because these textures to a considerable extent reflect the separation of mafic and carbonate minerals into discrete areas of the rocks. Experimental support for this interpretation has been provided by the work of Wyllie and co-experimentalists (e.g. Franz and Wyllie, 1967; Koster van Groos and Wyllie, 1973). However, carbonate-silicate liquid immiscibility is not considered a tenable explanation for the development of the various types of segregations described in section 7.3.3. In each case immiscible carbonate and silicate liquids can be rejected for one or more of the following reasons:

(i) The fact that host and segregation fractions can be equated with 'early' and 'late' phases in the rocks or vice versa in certain extreme cases (section 7.4). In many instances the earliest crystallizing minerals within 'late' segregations have clearly grown inwards after having nucleated against essentially solid 'early' host material. Such situations do not entirely rule out liquid immiscibility since it could be argued that one liquid solidified later than the other, at a much lower temperature. However, separation of 'early' and 'late' phases can be related to a single crystallization sequence, possibly with a crystallization gap between early (high temperature) and late (low temperature) minerals. There appears to be no pressing need to postulate the co-existence of immiscible liquids at any stage.

(ii) There is a virtual absence of any early crystallized matrix minerals, macrocrysts, or xenogenic material (incorporated at an early stage of intrusion) within 'late' segregations. Such components are restricted to the relatively early crystallized parts of the rocks (e.g. Plates 7.11d-h and 7.12a-h). There gaps in the segregation linings, these gaps are not entirely filled by calcite. Frequently the central massive, isotropic serpophite extends into the gaps and in some instances is apparently continuous with serpophite in the groundmass of the host material. The latter also contains abundant small anhedral grains of poikilitic calcite (Plate 7.11g).

At least three modes of origin can be considered for these segregations. They may be amygdales sensu stricto; or they may represent segregations of residual liquids; or they may be segregation vesicles as described previously.

The possibility that these bodies are true amygdales can be ruled out for two reasons: they are not infilled by secondary components and there is evidence of contemporaneous crystallization within and outside the segregations.

Critical to distinguishing between the remaining two possibilities is the time of formation of the fibrous linings in relation to the formation of the calcite and serpophite in the segregations. Two alternatives can be considered. In the first and favoured case the linings are interpreted as having formed prior to the crystallization of the inward-projecting calcites and the consolidation of the serpophite. It is further suggested that the linings developed at the interface between gas vesicles and the surrounding partly crystallized host material. Subsequently, following cooling, condensation of the vapour phase and a consequent fall in gas pressure, it is concluded that the linings were breached locally and infilled by residual melt derived from the host material. Calcite and serpophite crystallized from this residual melt and, as noted previously, in some instances calcite crystals and serpophite cross the gaps in the linings. A possible criticism of this mode of formation is that it requires some serpentine (the fibrous linings) to crystallize before calcite. While no explanation for this is at hand there is abundant evidence from the detailed descriptions given in Chapter 5 that, at least with respect to deuteric alteration processes, the formation of serpentine commonly precedes the formation of several apparently higher temperature minerals (e.g. diopside, phlogopite and monticellite).

The alternative interpretation (that the calcite crystals formed prior to the fibrous rims) is more consistent with the view that the segregations were not gas vesicles but segregations of melt. Such an origin is not, however, favoured because:

(i) The sharp boundaries of the segregations are inconsistent with such an origin (Ferguson and Currie, 1971).

(ii) There appears to be no reason for some residual melt to segregate into sharply bounded, globular bodies devoid of early crystallized minerals while much of the same residual melt crystallizes interstitially or poikilitically outside the segregations.

(iii) It is highly unlikely that fibrous serpentine linings postdate the inward-projecting calcite because under such circumstances the linings would probably have nucleated against and been continuous around the calcites. Instead the discontinuous linings commonly terminate against cross-cutting calcite crystals.

Somewhat more irregular segregations (Plates 7.11h and 7.12a) occur in another dyke near the De Beers pipe. Although serpophite is present in most of the segregations in this dyke they are calcite-rich compared with the examples described above. Furthermore the serpentine linings which are prominent in many of the segregations of the internal dyke are absent in this case. The margins of the bodies may, however, be clearly defined by rows of small opaque crystals and other components of the host material (Plate 7.12a). Excellent examples of individual calcite grains cutting across these margins (which are also interpreted as 'early' vesicles boundaries) are evident (Plate 7.12a).

In the first example described above the retention of more-or-less globular forms may be ascribed to a degree of rigidity imposed by the early-formed fibrous linings and probably also to the establishment of some degree of rigidity in the surrounding host by partial crystallization (the formation of a framework of crystals with interstitial fluids). The much more irregular segregations of the second example probably reflects a situation where vapour condensation in the segregations occurred before an equivalent degree of rigidity had been established. It is possible that in both cases condensation and infilling were contemporaneous and complementary. In the second case such a situation appears to be mandatory to account for the preservation of the original segregation boundaries, even though these boundaries may have been considerably distorted during the later stages of the formation of the segregations.

While the segregation vesicle origin provides a reasonable explanation for Type 1 segregations it cannot readily be applied to the Type 2 segregations. As noted previously the latter are highly irregular bodies commonly characterized by inward-growing crystals which have nucleated against earlier-crystallized material at the segregation margins (Plates 7.12b-h). Reasons for rejecting the segregation vesicle concept in these cases include:

(i) The often extreme irregularity of the bodies.

(ii) The fact that the first minerals to crystallize within the segregations nucleate against essentially solid host material. While there is probably a limited amount of contemporaneous crystallization within and outside the segregations there is no indication that residual melt within host areas was sufficiently abundant or sufficiently mobile to flow into the segregations after their bounding surfaces had been established.

Only two possible explanations for Type 2 segregations appear to be feasible. They may represent 'pockets' of residual melt which had segregated prior to crystallization of the host areas and remained liquid until crystallization of the latter was far advanced. Alternatively they may be the products of condensates of exsolved gas phases, following extensive degassing under conditions which hindered the escape of exsolved volatiles.

If Type 2 segregations are segregations of residual melts it is difficult to understand:

(i) Why in so many cases such extremely efficient segregation has taken place (e.g. Plates 5.6g-h, 5.8c-f, 7.11c and 7.12b-h)? The commonly near-perfect degrees of segregation are particularly difficult to explain because the major minerals in these segregations, calcite and serpentine, occur in considerable amounts in other kimberlites where they are intimately associated with other groundmass minerals and are more-or-less evenly distributed through the rocks. It is also surprising (in terms of the melt segregation hypothesis) to note the almost complete absence of early crystallized (pre-segregation) minerals or xenogenic material in the late-stage segregations. It might be argued that this a function of such pre-segregation components acting as nuclei during the crystallization of the groundmass of the 'early' fractions of the rocks and hence being preferentially incorporated in the latter. It is likely that this would have occurred to some degree but complete or almost complete separation by such nucleation seems unlikely. The increasing relative amounts of residual melts as crystallization proceeded would mitigate against this.

(ii) The sharp boundaries between 'early' and 'late' parts of the rocks are at odds with the idea of melt segregations. Ferguson and Currie (1971) have pointed out that most late-stage segregations have diffuse boundaries unless they have been mobilized and intrude their host. Such a situation applies in the case

(x) Instances of accretion or welding of the segregations in F2 (main segregation zone) although rare, are relatively common (Plate 7.13e) compared with the frequency with which these features are evident among the pelletal lapilli of the TKB's described previously.

(xi) Although conclusive evidence is lacking the F2 intrusion (which recent mapping by P.J. Bartlett has shown is a part dyke, part plug-like complex, Figure 3.37) probably did not reach surface (the Dutoitspan dyke undoubtedly did not).

(xii) Unlike the KIMFIK TKB's the F2 intrusion contains relatively few country rock inclusions and those that are present are relatively large (1-2cm or more). The Dutoitspan dyke is virtually devoid of xenolithic materal.

(xiii) 'Early' crystallized groundmass phases (mainly diopside and phlogopite in F2) are concentrated in the segregations, 'late' phases (mainly calcite and serpentine) occur between segregations. Similarly 'early' minerals in the Dutoitspan dyke are restricted to the segregations.

The author previously ascribed the globular segregations in F2 to carbonate-silicate immiscibility during fluidized intrusion of a liquid-solid system in an open diatreme (Clement, 1975). Such an origin is no longer favoured because of:

(i) The arguments raised against any form of liquid immiscibility in section 7.3.4.

(ii) The fact that calcite has subsequently been found to be less abundant in F2 than previously supposed.

(iii) The fact that the F2 intrusion probably did not reach surface.

In the light of previous deductions concerning the origin of globular segregations composed of late-crystallizing minerals (section

7.3.4) it seems likely that the development of these 'early' (relatively high temperature) segregations may in some way also be associated with the separation and concentration of fugitive components. However, simple separation of volatile phases does not explain the globular nature of these segregations or the occurrence of 'concentric' textures within some of them or the extreme efficiency of separation between 'early' and 'late' components (Plates 7.3c-d and 7.13c-h).

It is suggested that these 'high temperature' globular segregations (like some of the late-stage relatively low temperature, segregations described in section 7.3.4) reflect the development of vapour-liquid-solid systems produced by extensive degassing under essentially closed (i.e. hypabyssal) conditions. However, subsurface fluidization of these three-phase systems, probably in a turbulent, convective, manner is considered to be an additional prerequisite for the formation of these segregations.

It is proposed that rapid boiling of volatiles, possibly aided by temporary pressure drops induced, for example, by hydraulic fracturing during intrusion, could lead to the development of the necessary three-phase fluidization regimes. The liquid fractions of the systems are likely to have been moderately to highly viscous by the time fluidization cells developed due to cooling, some degree of crystallization, extensive degassing and complementary enrichment in the residual liquid fractions of elements such as Si and K.

It is suggested that as a consequence of turbulent fluidization the viscous magma fractions, containing considerable solid material (e.g. phenocrysts, xenocrysts and xenoliths and some groundmass material that may have crystallized around or nucleated against intratelluric or xenogenic components), were disrupted into 'fragments'. These 'fragments' subsequently attained spherical or ellipsoidal shapes and concentric internal textures due to surface tension and the torsional effects of rotation during fluidization. Nucleation around relatively large components (e.g. xenocrysts and xenoliths) during partial pre-fluidization crystallization may also have influenced the shapes of at least some segregations.

Examination of Plates 5.10f, 7.3c-d, 7.13c-g and 7.16c shows that almost all the intratelluric and xenogenic components (mainly olivine phenocrysts and xenocrysts) occur within the globular segregations of the F2 and Dutoitspan dyke intrusions. This situation is consistent with the genetic model. It is likely that most pre-existing solid components would be concentrated in the increasingly viscous liquid fractions as cooling and degassing of the intrusions proceeded.

The late-stage material hosting the globular segregations of the F2 and Dutoitspan dyke intrusions consists mainly of finely spherulitic serpentine (Plate 7.13h) or serpophite together with calcite (Plate 7.13f), which is relatively abundant in the latter intrusion (fairly extensive deuteric carbonatization of the segregations in the Dutoitspan dyke has also occurred). The low temperature host minerals are considered to be derived from the vapour phases of fluidized systems and, as the globular segregations are generally not closely packed, size-sorted or deformed (Plates 5.10f and 7.3c-d), condensation and crystallization of the host material is considered to have proceeded rapidly once convective fluidization waned.

While the bulk of the crystallization of the globular segregations is considered to have been completed prior to consolidation of the interstitial vapour-derived residua it is apparent that crystallization of diopside in F2 continued after turbulence and/or convective movements had ceased. This is indicated by the occurrence of delicate aggregates of extremely fine-grained diopside needles within the interstitial matrix (Plate 7.14b) and by similar needles projecting outwards into serpentine from the edges of some rather vaguely defined segregations. Such projecting needles are unlikely to have been preserved during turbulent motion.

The absence of vugs within the host material is ascribed to deflation of the intrusive systems as condensation and crystallization of the vapour phases proceeded.

The origin of the 'intermediate' zone of the F2 intrusion is uncertain. It may represent an area where relatively large segregations settled or were otherwise segregated (by convective flow?) and where 'squeezing out' of much of the interstitial vapour-derived fraction occurred prior to final crystallization. Alternatively the 'intermediate' zone may represent an area where volatiles were never concentrated to the same extent as in the main area of globular segregation development.

Segregationary textures such as those displayed by F2 and the Dutoitspan dyke are rare in the KIMFIK pipes, no other similar intrusions have been noted. This rarity may imply that kimberlite magmas are seldom suffficiently evolved to be able to produce the necessary abundance of volatiles or, more likely, that conditions that will allow the required degree of retention of volatiles are rarely attained.

As is evident from the foregoing discussion the processes advocated for the formation of the globular segregations in the F2 intrusion and in the Dutoitspan dyke are very similar to those proposed for the formation of some pelletal lapilli (section 7.2.3) in TKB's. The main established or postulated differences are:

(i) The development of globular segregations appears to take place on a relatively small scale; major explosive outburts at surface do not initiate, control or directly influence any aspect of the development of the segregations.

(ii) The fluidized systems are closed or partially closed so that the escape of exsolved volatiles is prevented or inhibited. Consequently calcite is a relatively abundant mineral in the globular segregation zones, mainly as an interstitial mineral.

The relative abundance of calcite can be ascribed to the retention of CO_2 in the more-or-less closed systems.

(iii) Relatively few country rock inclusions are incorporated in the globular segregation zones. This is a function of the essentially closed systems and insulation, from the country rocks, at least in the examples described, by surrounding hypabyssal-facies kimberlite.

(iv) The groundmass grain size of globular segregations and interstitial material may be comparatively 'coarse' relative to that of the equivalent pelletal lapilli and interstitial material in TKB's. Where this pertains it can be ascribed to more rapid cooling of the pelletal lapilli. In the case of the F2 intrusion and the Dutoitspan dyke, cooling of the fluidized zones was probably inhibited by the associated, fringing, hypabyssal material.

(v) Globular segregations may attain much larger sizes (Plate 7.16c) than pelletal lapilli and individual segregations exceeding 10cm in diameter occur in F2. Furthermore, in some cases (e.g. the Dutoitspan dyke) the globular segregations are not only consistently larger but also display a much more restricted size range than the pelletal lapilli in the KIMFIK pipes (Plate 5.10f). These characteristics are ascribed to relatively prolonged cooling, crystallization and convective fluidization under relatively closed conditions.

As noted above segregationary textures typical of those which occur in the F2 intrusion and in the Dutoitspan dyke in many respects resemble the pelletal textures of some diatreme-facies kimberlites and they are also associated with fluidization processes. However, they form entirely in hypabyssal environments and are gradationally related to typical hypabyssal-facies kimberlites. They form therefore a textural class of kimberlites that by gradational variations in evolutionary parameters may logically trend either towards typical hypabyssal- or to diatreme-facies kimberlites. Many of the large 'nucleated autoliths' of Ferguson et al. (1973) and Danchin et al. (1975) are probably

'globular segregations' which may have been incorporated in TKB's with numerous pelletal lapilli. However, because of the gradational evolutionary trends mentioned above transitional relationships between these spherical bodies should be recognized (as noted previously in section 7.2.3).

7.5 DETAILED PETROGRAPHIC DESCRIPTIONS

Detailed petrographic descriptions of all the hypabyssal intrusions in the KIMFIK pipes would make this section inordinately long and would involve considerable repetition. In order to provide an indication of how neighbouring kimberlite intrusions in individual pipes vary, petrographically detailed descriptions of the four main intrusions in the De Beers pipe (DB 2, DB 3, DB 5 and DB 6, see Figures 3.34-3.35) and of four adjacent intrusions in the west part of Dutoitspan (D2, D5, D16 and D18, see Figures 3.31-3.33) are given below. Abbreviated descriptions of a number of the other intrusions are given in Appendix I (Vol. II).

7.5.1 The DB 2 kimberlite

DB 2 is a dark, grey-black, hard, locally brittle, macroporphyritic kimberlite with a dense, aphanitic groundmass. In hand specimen the most conspicuous features are numerous anhedral olivine macrocrysts and scattered country rock xenoliths. Olivine macrocrysts make up 10-20 vol.% of the rock and the xenolith content ranges between 5 and 15 vol.% (generally near 5%). Scattered macrocrysts of phlogopite, garnet, ilmenite and chrome diopside are also evident but are not abundant. The country rock xenoliths consist mainly of subangular Ventersdorp and Basement rock types which have often been partly replaced by calcite.

Most xenoliths are smaller than 10cm across and many have major dimensions of 1-5cm. Occasional ultramafic (mainly peridotitic) nodules are exposed in mine workings. At some levels distinct flow structures occur in DB 2 towards the centre of the intrusion, i.e. near the contact with the later-intruded DB 3 occurrence. These flow structures consist of alternating, flow differentiated, coarse (macrocryst-bearing) and fine (macrocryst-free) bands. In addition near-vertical, parallel flow orientation of elongate components is evident. Near the southeast margin of DB 2, on the 595m level, elongate xenoliths are oriented subhorizontally. This orientation contrasts strongly with the near-vertical alignment in the central part of the intrusion. It is possible that the variation in orientation reflects convective overturn during the intrusion of DB 2.

In thin section two populations of olivine are evident; large (commonly 1-6mm), anhedral, rounded or irregular macrocrysts and generally smaller (usually <0,5mm) euhedral or subhedral phenocrysts. The idiomorphism of the phenocrysts contrasts strongly with the anhedral character of the macrocrysts (Plate 5.1b). The macrocrysts are considered to be xenocrysts derived from the same, or similar, upper mantle rocks that provide the ultramafic nodules found in the kimberlite (section 6.2.1). Many of the xenocrysts display strained extinction, kink-banding and/or substantial degrees of recrystallization (Plate 6.1b). These features are not displayed by the phenocryst population. The latter olivines also differ from the macrocrysts in that they often contain small (usually <0,05mm long), rod-like, oriented rutile inclusions.

The extent to which olivine has been altered varies considerably within the intrusion. Most individual grains are partly serpentinized and many small phenocrysts have been entirely altered. Locally olivine is, however, extremely well preserved (Plate 5.1b).

Alteration of DB 2 olivines is largely limited to serpentinization but some replacement by calcite has occurred (after initial serpentiniof the 'detached diapirs' described by Dawson and Hawthorne (1973) from the Benfontein Sills. It does not, however, apply to the Type 2 segregations described and illustrated in this thesis.

In view of the objections to melt segregations given above the author believes that Type 2 segregations are best interpreted as the products of gas condensates. Degassing, at least in some instances, was probably extended by retrograde boiling after intrusion had ceased. In terms of this hypothesis Type 2 segregations are considered to have crystallized from the condensates of exsolved gaseous fractions. This conclusion presupposes that escape of the exsolved material was prevented or at least inhibited i.e. the systems were partially closed. This hypothesis offers a ready explanation for the extreme degrees of segregation apparent in many instances (e.g. Plates 7.12b-h) since gas-liquid immiscibility will in effect have occurred. The hypothesis also accounts for the sharp boundaries displayed by the segregations (e.g. Plates 7.12b-h and 7.13a-b) as it appears that condensation and subsequent crystallization of the minerals in the segregations was deferred until the co-existing host melt fractions had largely (but not entirely) crystallized. This interpretation is supported by the boundary nucleation (Phillips, 1973) and inward growth of crystals within the segregations. The irregular shapes of the segregations can be ascribed to distortions induced by post-intrusion compaction and shrinkage during the latter stages of crystallization of the host (early) fractions of the rocks and during the cooling, condensation and crystallization of the segregations. It should be stressed that, in order to accomodate the volume changes which accompany gas condensation and crystallization within the segregations, the host material cannot be entirely solid when the segregation minerals form (although host crystallization may be far advanced at this stage).

The condensation hypothesis rests heavily on the conclusion that extensive vesiculation occurs. Such vesiculation is consistent with the volatile-rich nature of kimberlites and with experimental evidence indicating that gas phases can co-exist with kimberlite liquids at relatively low temperatures and pressures. Furthermore, Mysen (1975) has demonstrated that CO_2, which is relatively abundant (Table 7.7) in kimberlites (concentrated as $CaCO_3$ in the segregations of hypabyssal-facies kimberlites displaying segregationary textures), has a low solubility in silicate melts below 15kb. At all temperatures a gas phase is associated with the melts. Elthon and Ridley (1979) have proposed that kimberlite melts are coninuously saturated with CO_2 during their formation and that they must degas large volumes of CO_2.

The condensation hypothesis also depends on the escape of exsolved volatiles being inhibited to varied (often considerable) degrees in order to explain the mineral composition and volumetric abundance of the Type 2 segregations in some kimberlites. The retention of volatiles may be related to the hypabyssal nature of the intrusions and to relatively advanced stages of crystallization having been reached prior to extensive vesiculation. The presence of relatively abundant crystalline material and the probable presence of relatively viscous, volatile-poor residual melt fractions (as a result of cooling, degassing and crystallization) would hinder the escape of exsolved volatiles.

The most likely segregations to represent simple accumulations of residual melts are the Type 3 segregations. These segregations tend to be volumetrically limited and are usually relatively small in comparison with Type 2 segregations. They are commonly irregular in shape and have diffuse boundaries. A common feature is the occurrence of boundary zones composed of one or more of the major groundmass minerals of the surrounding material. However, within the boundary zone the mineral or minerals concerned have grown considerably larger and display much higher degrees of idiomorphism than elsewhere (Plates

7.11a-b). These features of the boundary minerals are ascribed to more extensive or rapid crystal growth or to more prolonged crystallization within or immediately adjacent to concentrations of residual, volatile-rich, melts. The final minerals to crystallize within the segregations (mainly calcite and/or serpentine) commonly crystallize centrally or in an interstitial relationship to the euhedral grains of the boundary zones (Plates 7.11a-b).

7.4 TRANSITIONS BETWEEN HYPABYSSAL AND DIATREME-FACIES KIMBERLITES

It was suggested in section 7.2.3 that crystallization concomitant with extensive vesiculation, prior to post-breakthrough vapour-liquid-solid fluidization, could lead to the formation of a variety of embryonic pelletal lapilli (i.e. partly crystallized segregations within vesiculating magma). These embryonic lapilli would subsequently attain their final morphological and mineralogical character during (and after) fluidization events generated by explosive breakthrough to surface. However, it was also suggested that globular segregations that closely resemble some pelletal lapilli and may, in fact, subsequently be incororated in some juvenile lapilli assemblages, might be formed without ever having been subjected to the open system fluidization events associated with the formation of TK's and TKB's (section 7.2.3 and Chapter 10). Some examples of such segregations, from the F2 intrusion in the Finsch pipe and an internal dyke in the Dutoitspan pipe are shown in Plates 5.10e-f, 7.3c-d and 7.13c-g. These segregations are so similar to many pelletal lapilli (cf. Plates 7.2 and 7.3) that some common genetic factors can be inferred.

In considering the origin of the F2 and Dutoitspan dyke globular segregations several features of the segregations themselves and the rocks as a whole must be taken into account:

(i) Well formed globular segregations such as those illustrated in Plates 7.3c-d and 7.13c-g are a local development within the F2 intrusion. The zone containing these segregations is approximately 5m wide where it is exposed on the 348m level of the mine. Globular segregations in the Dutoitspan dyke (Plate 5.10f) are even more locally developed. They occur in pockets, with gradational boundaries, which measure <1,0m across. These pockets are concentrated along one side of the dyke.

(ii) To the west and east of the globular segregation zone (i.e. towards the margins) the F2 intrusion occurs as hypabyssal-facies kimberlite which contains abundant groundmass phlogopite and diopside. In these areas the kimberlite has a more-or-less uniformly-textured groundmass. The globular segregation 'pockets' of the Dutoitspan dyke are surrounded by uniformly textured hypabyssal-facies kimberlite which is locally aphanitic.

(iii) An 'intermediate' zone occurs between the hypabyssal-facies zone and the west side of the zone containing globular segregations in the F2 intrusion. The intermediate zone contains segregrations which are relatively large and closely packed compared with those shown in Plates 7.3c-d. Many of the segregations within the 'intermediate' zone are several centimetres in diameter and, although some are spherical, many are irregular. This irregularity apparently reflects distortions induced by close packing of the segregations before they were entirely solid. Many of the segregations are in contact with one another in the intermediate zone. This contrasts with the situation in the main globular segregation area where segregations in contact with one another occur but are rare.

(iv) Increasing amounts of primary groundmass calcite and serpentine are evident through the sequence of uniform F2 → intermediate F2 → F2 with loosely packed globular segregations.

(v) Calcite is locally an abundant component of the main globular segregation zone of the F2 intrusion although it is not a major component overall. Furthermore, calcite in the main F2 globular segregation zone has been partly replaced by serpentine (Plate 7.9h) and hence was originally more abundant than at present (Clement, 1975). Calcite is an abundant mineral between the globular segregations in the Dutoitspan dyke and has also in part deuterically replaced the segregations.

(vi) Some globular segregations in F2 contain concentrically oriented phlogopite phenocrysts (Plates 7.13c-e). When present the orientation is poorly to moderately developed and may be restricted to the outer parts of the segregations only. Concentric structures are a common feature of the globular segregations in the Dutoitspan dyke.

(vii) Most of the segregations in F2 have kernels (more-or-less centrally located) which generally consist of altered (serpentinized) olivine grains (Plates 7.3c-d and 7.13c-e). Kernels are present in some of the Dutoitspan dyke segregations but are less abundant than in F2 segregations.

(viii) The segregations in F2 consist mainly of fine-grained, intimately associated,phlogopite and diopside (Plates 7.13c-h). The groundmass is, however, generally coarser than that of most of the pelletal lapilli described in sections 7.2.2 and 7.2.3,which contain the same minerals (cf. Plates 7.14a and 7.8b-h).

(ix) Some of the globular segregations in F2 (main segregation zone) appear to have relatively smooth outer surfaces (Plate 7.13c). Others are less regular with outward-projecting diopside needles (Plates 7.13g and 7.14b). Some extremely irregular diopside aggregates are also evident (Plate 7.14b). The segregations in the Dutoitspan dyke generally have smooth boundaries.

zation) and the marginal zones of serpentinized olivine may contain a little deuteric phlogopite. This mica occurs as very pale brown, faintly pleochroic, laths (maximum length of <<0,1mm) or as tiny sheaf-like aggregates (less common) of similar overall dimensions.

The scattered macrocrysts of phlogopite present rarely exceed 1-2mm in size. They have irregular, corroded margins and have been extensively altered (to the extent that many are barely recognizable). Serpentine is an abundant alteration product and replacement by calcite is also common. Some altered phlogopite macrocrysts contain small, irregular, haphazardly oriented, laths of later (deuteric) phlogopite while many phlogopite macrocrysts are fringed by prominent borders of opaque oxide(s).

The macrocryst assemblage and scattered country rock xenoliths are set in a fine-grained groundmass which, in most thin sections, is dominated by monticellite (Plate 7.15a and Tables 7.4 and 7.5), a common groundmass mineral in kimberlites (Clement et al., 1975; Skinner and Clement, 1979; this thesis, section 5.4). Monticellite occurs as extremely small (∿0,02mm), subhedral, colourless grains showing straight extinction, low birefringence and moderate relief. The most idiomorphic grains resemble olivine crystals in cross section. In many thin sections monticellite is well preserved but in some areas it has been extensively serpentinized and/or carbonatized.

Phlogopite is on average the next most abundant silicate phase in the groundmass but the phlogopite content varies locally from ∿1,0-10 vol.% (Table 7.4). It occurs as colourless or very pale brown, variably sized (<0,01-0,3mm) laths scattered fairly evenly through the groundmass. The larger phlogopite laths often contain inclusions of earlier-crystallized minerals (e.g. monticellite, perovskite, opaque oxides).

The DB 2 kimberlite contains abundant opaque minerals in the groundmass (Plates 5.1b and 7.15a). Opaque minerals in DB 2 and other

De Beers kimberlites have been extensively investigated by Pasteris (1980a). She has recognized groundmass ilmenite and a complex suite of spinels (ranging mainly from chromites to titanomagnetites). Opaque grains in the groundmass of DB2 may exceed 0,1mm but commonly are smaller than 0,05mm. Perovskite is another common accessory groundmass mineral (equant grains up to 0,2mm in size) but DB 2 is characterized by a high ratio of opaque oxides to perovskite (Plate 7.15a) and the abundance of opaque minerals is a distinctive feature of this kimberlite.

Other primary minerals in the groundmass of DB 2 are serpentine, calcite and apatite. Generally DB 2 displays a uniform distribution of groundmass constituents but the latter three relatively rare minerals may be locally concentrated in small (usually <0,5mm) segregations (mainly small Type 2 or Type 3 segregations, section 7.3.4).

Serpentine, calcite and apatite also occur interstitially or as small euhedral prisms (apatite only) elsewhere in the groundmass. Much of the serpentine and calcite in non-segregated areas occurs as deuteric replacement products of phlogopite and monticellite. In terms of the mineralogical classification of kimberlites proposed by Skinner and Clement (1979) DB 2 can be classed as a monticellite kimberlite (Table 7.5). Texturally it is a macroporphyritic kimberlite with a uniform groundmass texture.

7.5.2 The DB 3 kimberlite breccia

In hand specimen DB 3 locally (near its margin) resembles the surrounding DB 2 kimberlite. Generally, however, it is a paler-coloured, grey or grey-green, moderately hard, macroporphyritic rock containing a much greater abundance of country rock fragments ($\sim$20 vol.%). In terms of the textural-genetic classification used in this thesis (Chapter 2) this intrusion is classified as a kimberlite breccia. The xenolith suite differs from DB 2 in that, in addition to Ventersdorp and Basement inclusions, numerous fragments from the Karoo Sequence (mainly shale) are present. The xenoliths display a greater variation of size in DB 3 than DB 2, mostly ranging from a few millimetres to $\sim$25cm in diameter. Most of the xenoliths are subangular and many, particularly the smaller inclusions, have been extensively altered.

In thin section DB 3 displays more extensive petrographic variation than any of the other major intrusions in the De Beers pipe and there is some indication of a rudely concentric, zonal pattern of variation. This variation is manifested in three main ways: by differences in the relative proportions of the groundmass minerals; by grain-size variations, particularly of groundmass phlogopite; and by differences in the nature of olivine alteration (see later comments).

Two populations of olivine and scattered macrocrysts of altered phlogopite, similar in most respects to those occurring in DB 2, are consistently present in thin sections. In some samples alteration of olivine is limited (as in DB 2) to serpentinization. Elsewhere olivine alteration, particularly of smaller phenocrysts, is extremely complex (Plates 5.2c-g) involving the development of several zones of serpentine, extensive carbonatization and phlogopitization and minor chloritization. In rare instances a little monticellite has formed within the marginal rims of altered olivine grains.

Near the contact with the surrounding DB 2 intrusion the DB 3 kimberlite often contains substantial quantities of monticellite. However, in these areas DB 3 generally contains more groundmass phlogopite than DB 2 and the complex alteration of olivine described above is characteristic.

Locally near the margin and more persistently some distance inwards from the periphery the DB 3 kimberlite contains abundant phlogopite and in some slides phlogopite forms $\sim$70 vol.% of the groundmass (Plate 5.4a). In such areas phlogopite occurs as a mosaic of coarse laths and interlocking anhedral to euhedral basal plates (up to 1,0mm or more across). This phlogopite pokilitically encloses abundant inclusions of earlier crystallized minerals including perovskite, opaque oxides, apatite, rare monticellite and small olivine phenocrysts. In the mica-rich areas of DB 3 calcite may also be a prominent constituent but monticellite is rare or, more commonly, absent. Apatite is a constant acessory mineral which occurs as relatively large euhedral grains (up to 0,2mm in length).

In the central part of the DB 3 intrusion calcite is the most abundant groundmass mineral and a segregationary texture is apparent. Earlier crystallized groundmass components (notably heavily altered phlogopite) tend to be concentrated around olivine grains while calcite has crystallized in irregular, interstitial, linked 'pools'. Some serpentine (massive or finely spherulitic) is associated with the calcite in these segregations.

In the central area of DB 3 both the macrocryst and groundmass generations of phlogopite are extensively corroded and altered (serpentinized and carbonatized). Groundmass phlogopite laths rarely exceed 0,25mm in length and are distinctly smaller than in the mica-rich areas described previously. Apatite is an abundant accessory mineral in the central part of the intrusion and euhedral prisms up to 0,3mm long are common.

Small, equant, perovskite grains (0,1mm or smaller) are a noticeable constituent of the DB 3 kimberlite but, in contrast to DB 2, opaque oxides are relatively rare and perovskite is usually more abundant than opaque material.

The distribution of petrographic variants within the DB 3 intrusion has not been well established and requires a much more detailed investigation. It is, however, tentatively concluded that during the final stages of emplacement concentration of volatiles towards the central part of the intrusion was largely responsible for the rudely concentric pattern of variation described.

A feature of the DB 3 intrusion is the presence of a mineral with high birefringence which occurs as fibrous, radiating aggregates. This mineral was identified by Kruger (1978) as cebollite but subsequent XRD and electron microprobe analyses have indicated that it is pectolite (Scott-Smith et al., in press). It is often associated with natrolite (Kruger, op. cit.) and the two minerals frequently occur as metasomatic alteration products of small country rock xenoliths. Spherulitic aggregates of pectolite and natrolite are also present in the DB 3 kimberlite.

Extensive metasomatism of xenoliths is a feature of the DB 3 intrusion. Serpentinization, carbonatization, phlogopitization and minor chloritization have occurred. In addition opaque oxides, perovskite, apatite and monticellite may be present in altered xenoliths. In effect the inclusions have been 'kimberlitized'.

As previously noted, similarly complex deuteric alteration of olivine tends to be most pronounced in areas where monticellite is a relatively abundant groundmass component. In the latter areas it appears that residual kimberlite fluids, instead of producing late-crystallizing, primary groundmass phases, have engendered extensive, complex, deuteric alteration and replacement of previously serpentinized olivine. Conversely, where abundant late-crystallizing, primary groundmass components such as phlogopite, calcite and apatite occur, olivine alteration is largely restricted to serpentinization.

Accurate mineralogical classification of DB 3 requires numerous modal analyses because of the mineralogical variability which the intrusion displays. It is tentatively classed as a calcite-phlogopite kimberlite breccia. The matrix is macroporphyritic and the groundmass texture varies from uniform to segregationary.

7.5.3 The DB 5 kimberlite breccia

In hand specimen DB 5 is greenish-grey to greenish-black in colour and is generally similar in appearance to DB 3. However, it contains far more Karoo shale inclusions which, in underground exposures, may exceed 1,0m in size. Country rock inclusions make up 15-25 vol.% of the intrusion.

Both populations of olivine have been extensively altered, mainly to serpentine, although calcite is also often present. Altered olivine grains may also be fringed by minor, extremely fine-grained, phlogopite. The serpentine of the pseudomorphs is commonly zoned. Frequently the central parts of altered olivines consist of pale yellow serpentine while the margins have been chloritized and are slightly pleochroic in shades of green. As in other kimberlites of the De Beers pipe olivine phenocrysts may contain oriented rutile inclusions.

DB 5 differs considerably from the other De Beers pipe kimberlites described in that it contains numerous phlogopite macrocrysts (Plate 6.2h) which, although commonly corroded with opaque rims, are generally well preserved. The grains occasionally exceed 2mm in size but most are between 0,2 and 1,0mm. This generation of phlogopite is strongly pleochroic from medium orange-brown to straw-yellow. The pleochroic scheme is, however, commonly reversed. Some grains are zoned with normally pleochroic rims.

Another diagnostic feature of the DB 5 intrusion is the common occurrence of garnet in thin sections. In individual thin sections several grains are often apparent, usually with prominent kelyphite rims. This suggests that garnet is more abundant in DB 5 than in the other De Beers pipe kimberlites described here. Garnet is rarely seen in slides cut from samples of the other intrusions.

DB 5 is also unique among the De Beers kimberlites described here for the abundance of small country rock xenoliths present. Many fragments, particularly of Karoo shale, are evident in thin section. They range from $\sim$0,1mm to a centimetre or more in size. Zonal alteration of these inclusions is common (Plate 7.10d).

The dominant groundmass minerals are calcite, phlogopite and serpentine. Phlogopite appears to have been particularly abundant but has been extensively carbonatized, serpentinized and chloritized. This phlogopite is extremely fine-grained, individual laths often measuring <0,05mm. Much groundmass phlogopite occurs as minute, remnant wisps and shreds and the boundaries of individual grains are difficult to discern. Calcite and serpentine occur mainly as deuteric alteration products of phlogopite but some calcite occurs as clear, anhedral to euhedral, grains up to 0,25mm in diameter and is considered to be primary.

Apatite occurs as minute crystals (<<0,1mm in length) but has been extensively carbonatized. Perovskite grains (up to 0,2mm) and opaque minerals (generally smaller than 0,1mm) are scattered fairly evenly through the groundmass but perovksite grains are also concentrated around altered olivine grains.

The petrography of the DB 5 kimberlite varies little with depth. However, in one slide (from the deepest available sample) monticellite is a prominent groundmass mineral. No monticellite has been positively identified from higher levels but it is possible that some highly altered material is present (The monticellite-rich specimen may, however, represent an inclusion in DB 5).

It appears that the groundmass mica is finer-grained at higher levels and that calcite becomes more abundant upwards. Further work is, however, required to confirm these trends.

Mineralogically DB 5 can be classified as a serpentinized and carbonatized phlogopite kimberlite. Texturally it is a kimberlite breccia with a macroporphyritic matrix. The groundmass of the matrix has a uniform texture.

7.5.4 The DB 6 kimberlite

In hand specimen this kimberlite is a dark, hard, grey-black macroporphyritic rock containing scattered xenoliths, mainly of Basement gneiss. It is generally similar in appearance to DB 2.

In most thin sections both macrocrysts and phenocrysts of olivine are highly altered and only the largest macrocrysts have vestiges of fresh olivine preserved as more-or-less central kernels. The main alteration product is serpentine and optically different varieties occur in roughly concentric zones. Calcite is also a prominent mineral within altered olivines. Some serpentinous material, near the margins of olivine pseudomorphs, is green or yellow-green and slightly pleochroic and appears to have been chloritized (cf. DB 5).

Scattered macrocrysts of phlogopite (which in rare instances reach 5mm in diameter) are present in this kimberlite but are much less abundant than in DB 5. They occur as stubby laths and irregular, anhedral or rounded grains. They are commonly severely altered, in the same manner as those in DB 2 and DB 3. Where fresh, this phlogopite often exhibits reverse pleochroism.

Some thin sections contain rare macrocrysts of ilmenite, in some instances rimmed by a little perovskite which results from reaction between the ilmenite and Ca-bearing fluids. No other macrocrysts were noted in the thin sections examined but the other 'normal' kimberlite indicator minerals (garnet and chrome diopside) are evident in hand specimens.

The main groundmass minerals in DB 6 are calcite, serpentine and phlogopite (Plate 5.5h). Apatite is a constant accessory mineral. Other accessory minerals are perovskite and opaque oxides. Much of the serpentine and some of the carbonate clearly replaces extremely fine-grained (<0,01-0,05) phlogopite. In this respect the kimberlite is similar to DB 5.

This kimberlite displays a well developed segregationary texture (Plate 5.5h). Phlogopite and carbonatized/serpentinized phlogopite, together with much of the apatite, opaque oxides and perovskite in in the rock, is concentrated in well defined, commonly lobate or rudely circular, patches. These patches also contain, generally as more-or-less centrally located kernels (Plate 5.5h), much of the altered olivine in the rock. These silicate-rich zones are separated, or partly separated, by interstitial material composed mainly of primary calcite.

The calcite of the segregations occurs as an interlocking mosaic of clear, colourless, anhedral to subhedral, grains up to 0,3mm across. This calcite contrasts strongly with the limited amount present in the silicate-rich areas. The latter calcite is extremely fine-grained (microlitic), turbid in appearance and, as noted, is mainly a deuteric replacement product of mica. Serpentine, massive or finely speckled between crossed nicols and often yellowish in colour, also occurs in the carbonate 'pool' areas (Plate 5.5h). This serpentine is considered to be primary in contrast to the green serpentine (in part chloritized) which replaces groundmass phlogopite.

Apatite is an abundant accessory mineral in DB 6. It occurs as elongate euhedral crystals seldom >0,06mm long and well formed euhedral basal sections are frequently evident. Apatite has in part been carbonatized. Such replacement is prominent where apatite occurs in calcite 'pool' areas. In the latter situation its original presence can often be recognized by 'ghost' euhedral crystals. Perovskite occurs as equant and irregular grains up to 0,2mm across but often <0,05mm in size. It is intimately associated with opaque material which occurs at the edges and/or in the cores of many perovskite grains. Complex twinning is indicated by the presence of numerous re-entrant angles at the grain margins. Opaque oxides range in size from ~0,1mm to dust-sized particles.

The DB 6 intrusion is classified as a phlogopite-calcite kimberlite. It should, however, be noted that many groundmass phlogopite grains have been partly or wholly serpentinized. Texturally DB 6 is a macroporphyritic kimberlite with a segregationary groundmass texture.

7.5.5 The D2 kimberlite

The D2 kimberlite is a hard, brittle, dark grey-black, macroporphyritic rock carrying scattered country rock xenoliths.

The porphyritic aspect results from the presence of numerous anhedral, often rounded olivine macrocrysts. The longest dimension of these grains in some cases exceeds 10mm but the majority are much smaller (<3mm). In some hand specimens much of the olivine is fresh, yellow-green and glassy but in others considerable alteration of the olivine has occurred, particularly of smaller grains.

The scattered country rock xenoliths are commonly highly altered and white to buff in colour. Many have been strongly carbonatized and some have been almost entirely replaced by coarsely crystalline calcite. The xenoliths are extremely variable in size ranging from a few millimetres to more than 25cm. Most measure less than 10cm across. Country rock xenoliths are commonly angular to subangular or have highly irregular corroded shapes. Granitic and gneissic fragments derived from the Precambrian Basement are the most abundant inclusion types. Ventersdorp lava xenoliths are also common and Karoo inclusions (mainly shale) occur but are rare. These shale fragments have prominent reaction rims.

Normal kimberlite indicator minerals are rare components of the D2 kimberlite. Their rarity is emphasized by the fact that no garnet, chrome diopside or ilmenite grains were seen on the surfaces of 5 cut slabs of the kimberlite ranging in size from 30 to 100cm^2. All three minerals have, however, been noted at underground exposures and in some hand specimens.

Rare ultramafic nodules are evident at underground exposures. They include lherzolites and harzburgites (with and without garnet) and scattered MARID-suite nodules.

At some levels zoning of the D2 intrusion is evident. On the 550, 555 and 585m levels a narrow (3-6m) inner zone (adjacent to the central D5 plug which pierces the D2 intrusion) is characterized by prominent near vertical flow structures. These flow structures take the form of alternating coarse and fine-grained, flow differentiated bands. The coarse-grained bands contain abundant, easily visible olivine grains.

These olivine grains are virtually absent in the fine-grained zones which are essentially aphanitic. Within the coarse-grained bands elongate olivine grains may be oriented parallel or sub-parallel to the direction of flow. The widths of individual coarse or fine-grained bands rarely exceed 20cm and they are often much narrower. Similar flow structures are not present elsewhere in the D2 intrusion.

The flow differentiated zone also differs from the remainder of the kimberlite in that it contains virtually no country rock xenoliths. Although such xenoliths are nowhere particularly abundant they do locally approach 10 vol.% of the kimberlite outside the flow differentiated zone.

The contact between the inclusion-poor, flow differentiated inner zone and the inclusion-bearing outer zone is gradational over a short distance (commonly <0,5m). It is conceivable that the two "zones" in fact represent discrete intrusive pulses of similar magma and hence can be regarded as separate intrusions, probably emplaced in rapid sucession. However, kimberlite from the two areas is, on a thin section scale, mineralogically and texturally almost identical and two macroscopically distinctive zones are not always evident in underground exposures.

In thin section the D2 intrusion displays the two populations of olivine typical of kimberlites. One population consists of relatively coarse grains which are typically anhedral. These grains are a prominent feature of the thin sections examined. They range up to 5mm in size and generally measure more than 0,5mm in diameter. Relatively few exceed 3mm but many exceed 1,0mm. Grains smaller than 0,5mm which display morphological or optical features characteristic of this olivine population are rare and some appear to be broken grains.

These olivine anhedra are commonly rounded but may be ragged or highly irregular (Plates 6.1c-d) with prominent embayments and protuberances. The irregularities reflect magmatic corrosion which has, in some instances, imposed angular re-entrants, bounded by straight or slightly curved surfaces, on the grain boundaries.

The majority of the anhedral olivine grains are strained (Plate 6.1d) and in one thin section all clearly anhedral grains >0,5mm in size display undulose extinction. Some of these strained grains show prominent deformation lamellae, in some cases aligned in two directions. Further evidence of the dynamic metamorphism to which this olivine population has been subjected is partial or, in some cases, complete recrystallization (Plates 6.1c-e).

These olivine grains often contain 'trails' of extremely fine-grained 'dusty' opaque material and minute 'bubbles' (Fluid inclusions?) which are commonly oriented in parallel or sub-parallel fashion.

The relatively large anhedral olivines have not (except locally) been severely altered. Generally alteration is restricted to serpentinization (and subsequent partial chloritization) of the margins of the grains. Serpentinization has also occurred along cracks or partings traversing the grains.

Other macrocrysts present in thin sections of D2 kimberlite include rare, anhedral grains of ilmenite, in some instances mantled by narrow, more-or-less continuous rims of perovskite. The maximum size of these ilmenites is 2,5mm. In one slide a single red-brown translucent spinel macrocryst was observed.

The second population of olivine in the D2 kimberlite consists of subhedral to euhedral phenocrysts which display a considerable range in size (Fig. 5.2). Most of these olivine crystals do not, however, measure more than 0,5mm across and at least 50% are smaller than 0,25mm.

Rare phenocrysts exceed 1mm in size and very rarely phenocrysts of between 2 and 3mm occur (Plate 7.14c).

The deformation textures typical of the cryptogenic anhedra are absent in the phenocrysts although rarely 'large' idiomorphic grains may be slightly strained. Very rare large phenocrysts are strongly corroded due to reaction with the magma (Plate 6.1g).

The extent to which olivine phenocrysts have been altered varies considerably from place to place within the intrusion. However, even where alteration is most severe some fresh olivine is preserved. Although, as a function of their smaller size, the phenocrysts are more extensively altered than the macrocrysts even grains as small as 0,2mm may have kernels of fresh olivine. As in the case of the macrocrysts the phenocrysts are predominantly altered to serpentine which has in part been chloritized (bluish or pale yellowish-brown chlorite). Chloritization appears to be most pronounced in areas where prior serpentinization of olivine has been most extensive. Minor marginal phlogopitization of serpentinized olivine is locally evident and in rare instances a little monticellite has replaced the edges of serpentinized olivine grains.

Scattered, irregular, often rounded and corroded basal plates and stubby laths of highly altered phlogopite occur in the D2 kimberlite and make up about 1 vol.% of the rock (Plate 6.3h). The grains range in size between ∿0,1 and 3,0mm but commonly measure between 0,5 and 1,5mm. Alteration of many of the grains is so severe that their original nature can be inferred only by examining a suite of progressively less altered grains. These phlogopite grains have been partly or entirely replaced by serpentine, by a later generation of fine-grained, colourless or near-colourless, randomly oriented, irregular flakes and laths of deuteric phlogopite (optically similar to phlogopite occurring in the groundmass), by minor calcite, by opaque minerals and rarely, along the grain boundaries, by monticellite. A prominent dark reaction rim of opaque minerals may be present (Plate 6.3h). In very rare instances the altered phlogopite grains are moulded around olivine phenocrysts (Plate 6.3h) and they must therefore have crystallized directly from kimberlite magma. These micas can therefore be interpreted as phenocrysts.

The coarser components of the kimberlite (macrocrysts, phenocrysts and country rock xenoliths) are set in a generally fine-grained groundmass consisting of monticellite, serpentine, phlogopite, calcite and apatite, together with scattered, equant, accessory grains of perovskite and opaque oxides. Although these minerals are generally evenly distributed some segregation of the later-crystallizing components is evident in all the thin sections examined. Locally, for example near the east contact of the kimberlite on the 585m level, a segregationary texture is fairly well developed but for the most part an essentially uniform texture is characteristic (Plates 5.3a and 5.7a).

Monticellite is the most abundant groundmass mineral in the D2 kimberlite. It occurs as small, colourless, moderately high relief, fairly closely packed, subhedral to euhedral crystals (Plate 5.3a and 5.7a). Commonly these grains range between 0,01 and 0,05mm in size and many contain minute 'bubbles' which may represent trapped gas or liquid-filled cavities. Monticellite grains are noticeably larger, reaching 0,13mm across, where they occur at the borders of segregations of late crystallizing minerals (Plate 5.7b).

The monticellite grains (together with accessory microphenocrysts of perovskite and opaque oxides) are set in a base which consists mainly of colourless, almost isotropic serpentine (serpophite) together with colourless, low birefringence mica and a little calcite. The latter mineral is generally concentrated in monticellite-free segregations. The low birefringence of the groundmass mica often makes it difficult to distinguish from the serpentine of the base. Consequently groundmass phlogopite migh be more abundant than is reflected by modal analyses

(Table 7.4). These phlogopite grains rarely exceed 0,25mm in diameter and are often smaller than 0,1mm in size.

As indicated above a little calcite occurs locally as a base to the abundant monticellite grains. In such areas calcite (like groundmass phlogopite) is moulded around or poikilitically encloses monticellite crystals (together with scattered opaque oxide and perovskite grains). Limited replacement of monticellite by calcite is evident in these areas.

Most of the calcite in this kimberlite occurs in segregated pools (Plates 5.7b, 5.10a and 7.14d). Calcite in these areas commonly occurs as anhedral grains up to 0,4mm in diameter (much coarser-grained than in the monticellite-rich areas where calcite grains often measure much less than 0,1mm).

Closely associated with calcite in the 'pool' areas are serpentine and apatite. The serpentine occurs as massive almost isotropic material (Plate 7.14c). Apatite in these areas varies from anhedral to perfectly euhedral crystals as much as 0,7mm in length (Plates 7.14d-e). In some instances individual crystals stretch entirely across and sometimes beyond the confines of segregations (Plate 7.14e and Fig. 5.9c). Apatite also occurs as discrete crystals scattered throughout segregations (Fig. 5.9d) and forms radiating groups of crystals growing inwards from the edges of segregations. Apatite is not entirely confined to the segregations but is best developed and attains maximum size in these areas.

A little colourless or very pale brown phlogopite is also present in some of the late-crystallizing segregations. It occurs as subhedral to euhedral laths and basal plates. Traces of pectolite also occur in some of these areas.

Perovskite occurs as honey-yellow, equant, complexly twinned and frequently zoned crystals often intimately associated with opaque spinel. Most perovskite occurs as discrete grains scattered through the groundmass of the rock but is also found rimming serpentinized olivine grains. Perovskite rarely exceeds 0,1mm in size and most grains are considerably smaller. Opaque oxides, including atoll-textured spinels, are generally smaller than 0,05mm and, like perovskite, are scattered fairly evenly through the groundmass of the kimberlite (Plate 5.3a).

Mineralogically the D2 intrusion can be classified as a monticellite kimberlite. Texturally it is a hypabyssal-facies macroporphyritic kimberlite with a uniform groundmass texture (late-stage segregations are relatively rare).

A detailed description of the D2 kimberlite has been included here for two reasons. Firstly, it is a good example of a monticellite kimberlite. Secondly, it is strikingly similar to the DB 2 kimberlite in the De Beers pipe (see section 7.5.1). The two intrusions provide an excellent example of discrete intrusions in neighbouring diatremes which cannot be differentiated megascopically or in thin section.

7.5.6 The D5 kimberlite breccia

The D5 intrusion is a dark grey, moderately hard kimberlite breccia containing abundant pale country rock xenoliths which impose a mottled appearance on the rock. This mottled appearance is enhanced by the kimberlite adjacent to inclusions being altered and lighter in colour than elsewhere.

Olivine macrocrysts are common but for the most part have been extensively altered to dull yellow, earthy material. These macrocrysts are anhedral and occasionally exceed 10mm across although most have maximum dimensions of less than 3mm. Scattered mica grains (rarely >3mm and commonly much smaller) are present and garnet and ilmenite are evident in most hand specimens.

The content of country rock xenoliths varies considerably from place to place in the intrusion ranging from a low of about 15 vol.% to a maximum of around 30 vol.% (visual estimates). The xenoliths are generally small. The vast majority measure <5cm across and many lie in the 1-3cm range. These xenoliths are characteristically altered and highly irregular in shape, often displaying distinctly corroded outlines with prominent embayments (Plate 7.1f-g). The most abundant inclusion types are granitic and gneissic varieties, frequently soft and extensively altered. Ventersdorp lava fragments are present but are less common than Basement rocks. Karoo shale inclusions are present but are relatively rare. They frequently display prominent reaction rims.

On a gross scale the D5 kimberlite appears to be homogeneous throughout the vertical and horizontal extent of the exposure. It does not display any obvious directional or orientation structures, zonal variations in character or flow differentiation effects. Local variations in xenolith content are, however, apparent.

Although diagnostic characteristics are always evident thin sections from different parts of the D5 intrusion display considerable variation, notably with respect to the extent and nature of deuteric alteration of various components. In addition some primary mineralogical differences are evident and limited textural variation occurs.

All the thin sections examined display the two populations of olivine characteristic of 'classic' kimberlites. Anhedral, rounded or irregularly corroded macrocrysts reach 7mm in diameter and commonly exceed 1mm. In contrast the bulk of the subhedral or euhedral phenocrysts in the rock, which may contain tiny rutile inclusions, are smaller than 0,5mm. Some phenocrysts do, however, exceed 1mm in size (up to 2mm) hence there is a size overlap between the macrocryst and phenocryst populations.

Both populations of olivine are extensively altered (mainly by serpentinization). One effect of this alteration is blurring of grain margins producing diffuse boundaries between olivine and surrounding groundmass. This effect is more pronounced in some thin sections than others. Where blurring is extensive the idiomorphism of individual olivine crystals may be destroyed hence it is not always possible to assign specific olivines to macrocryst or phenocryst populations. A further complication in this respect is the presence of some highly irregular, frequently angular, broken olivine grains which may belong to one or both olivine populations.

In most thin sections only the largest macrocrysts (several millimetres across) contain preserved relicts of unaltered olivine. Most olivine, particularly the phenocryst population, has been pseudomorphously replaced (commonly in zonal fashion) by serpentine which has, in some instances, been altered to chlorite. In addition a little phlogopite, in the form of minute laths, may occur as an extremely minor replacement product of serpentinized olivine. Such phlogopite is not evident in all the thin sections examined and when present is confined to the rims of serpentinized olivine grains. In very rare instances serpentinized olivine has been partly replaced by fine-grained diopside (Plate 5.3e). Where olivine macrocrysts are partly preserved typical features of this olivine population such as strained extinction and partial recrystallization are evident.

Scattered, often irregular and corroded, stubby laths and anhedral basal plates of relatively large (up to 3mm) phlogopite phenocrysts occur throughout the D5 kimberlite. Most of these grains are between 0,25 and 1,00mm in size. In some instances these phlogopites are partly moulded around (or contain inclusions of) olivine (Plate.6.4a). Commonly inclusions of opaque oxides are also present (Plate 7.14f-g). These phlogopite grains are therefore clearly primary in origin. It is, however, possible that two populations of phlogopite are present. A small proportion of the grains occur as strained, bent or kink-banded crystals which occur singly or in small aggregates.

The phlogopite phenocrysts exhibit diverse features. Some grains are unzoned and display normal pleochroism throughout. More commonly unzoned grains display reverse pleochroism. Zoned grains are common.

many having thin normally pleochroic rims while the cores of the grain exhibit reverse pleochroism. Elsewhere reversals in the pleochroic scheme and variations in colour occur in patchy fashion within complexly zoned crystals. In some instances individual grains are crowded with inclusions, in others few or no inclusions are present, while in yet others the inclusions are zonally distributed (Fig. 7.14f). In some cases corroded grains are fringed by an opaque oxide indicating reaction with the magma. In other instances distinct growths of phlogopite around such grains are evident.

Much less common than the abundant phlogopite phenocrysts present in the D5 kimberlite are scattered diopside phenocrysts (Plates 5.7c-d). These diopside phenocrysts may reach 0,5mm in size but they are generally considerably smaller. They are commonly somewhat corroded and irregular in shape but do occur (in some instances) as euhedral crystals. Many diopside crystals appear clearly to have been out equilibrium with the magma that post-dated their crystallization. In addition to having corrosion embayments they have been partly melted and recrystallized? (Plate 5.7d). Relatively coarse diopside crystals occasionally occur as aggregates and may be intimately associated with phlogopite grains (Plates 5.7e-f).

Numerous country rock xenoliths are present in the thin sections examined. They are extensively altered, so much so that identification of the original nature of the inclusions is almost invariably precluded. The alteration reflects a process of 'kimberlitization' (section 7.3.1) in which the inclusions are replaced by minerals which occur as primary or deuteric groundmass constituents of the kimberlite. The replacement products include diopside (very common), serpentine, chlorite, phlogopite, minor calcite, pectolite and apatite. In rare instances isolated grains of perovskite and opaque oxides occur within altered xenoliths. Where diopside replaces country rock xenoliths the grains tend to be small; commonly less than 0,1mm ranging down to micrelitic material.

In the majority of thin sections the most abundant groundmass mineral is fine-grained phlogopite (Plate 7.15b). Many of these grains measure less than 0,05mm across and they grade down to barely discernible microlitic wisps and shreds. Only a few of the groundmass phlogopites exceed or even closely approach 0,1mm in size and where alteration is not severe they occur as a fairly dense overlapping mat of anhedral-subhedral grains (felty texture).

Serpentine is also a fairly abundant component of the groundmass and in one of the thin sections examined it is more abundant than phlogopite. The serpentine generally has a finely speckled appearance between crossed nicols, in contrast to the commonly platy appearance of earlier serpentine which pseudomorphously replaces olivine.

Although some of the groundmass serpentine is probably primary most is considered to be deuteric in origin. Groundmass phlogopite is the mineral most heavily affected (Plate 7.15b). There is clear evidence of considerable replacement of phlogopite by serpentine. This is indicated by prominent, often irregular embayments of serpentine in phlogopite laths and plates; by the occurrence of minute, closely associated but entirely discrete shreds of phlogopite (in serpentine) which extinguish simultaneously between crossed nicols and represent relicts of an original single grain; and by 'shadows' in serpentine (areas of slightly different colour, texture and birefringence) which outline the boundaries of discrete pseudomorphs of original phlogopite crystals. Similar 'shadows' also indicate the presence of very small olivine microphenocrysts which were initially serpentinized at a relatively early stage and were subsequently 're-serpentinized'.

As noted previously diopside occurs as scattered phenocrysts in the D5 kimberlite and also formed during the "kimberlitization" of inclusions. In addition diopside occurs as a subhedral to euhedral primary mineral in the groundmass of the kimberlite. The amount of diopside present varies considerably among the thin sections examined and in general it tends to be irregularly distributed, commonly occurring as variably-sized, irregularly-shaped clots or patches (Plates 5.6c and 5.7g-h). Commonly this diopside, which is always fine-grained (commonly <0,1mm ranging down to barely discernable microlites) and may be acicular, is concentrated around highly altered country rock xenoliths which have in part themselves been replaced by diopside.

Several examples of diopside occurring in apparent ocelli or geode-like structures are evident in the thin sections examined. In some cases these bodies may represent replaced xenoliths or xenocrysts (Plate 7.15c and 5.7h) but others may be primary structures.

In rare instances the uniform distribution of diopside and other minerals within patches and the general form of the patches suggests that they represent inclusions of an earlier generation diopside kimberlite. One clearly identifiable inclusion of a diopside-phlogopite kimberlite was noted in a thin section from specimen 173/33/K6/38 (Plate 6.4e). However, most diopside-rich patches clearly do not represent kimberlite inclusions. Such an origin is not consistent with the extreme irregularity of many of the patches nor is it consistent with the mineralogical gradations evident at many of the boundaries of the patches. Furthermore in many cases both diopside-rich patches and 'normal' (phlogopite-rich) groundmass areas partly envelope individual olivine grains or microxenoliths of country rock.

Although volumetrically limited, apatite is a prominent mineral in most thin sections. It commonly occurs as euhedral prismatic crystals, smaller than 0,1mm, scattered throughout the groundmass (e.g. Plate 5.6c). Generally the groundmass of the D5 kimberlite has a uniform texture but locally minor, irregular, segregations of late-crystallizing groundmass minerals have developed. Apatite occurs in some of these segregations and in this situation may occur as subhedral or anhedral grains measuring between 0,1 and 0,2mm across.

A little primary serpentine occurs in the locally developed segregationary pools and calcite also occurs infrequently in this situation. Calcite is in fact a rare mineral in most specimens of D5 kimberlite and was not encountered during point counting of two thin sections (Table 7.4).

In some thin sections pectolite is prominent. It occurs as a constituent of the segregations (Plate 7.4a) and also occurs fringing and within altered country rock xenoliths.

Scattered, equant, brown perovskite grains occur in the D5 kimberlite (Plate 5.6c). They range up to about 0,1mm in size but commonly measure between 0,02 and 0,06mm. Perovskite is intimately associated with opaque material and perovskite grains are often partly or entirely rimmed by the latter. Scattered rare ilmenite macrocrysts, larger than 0,2mm in diameter, are evident in the D5 kimberlite but discrete groundmass opaque minerals are not abundant in this kimberlite. Those present are invariably very small (<0,05mm) and many measure less than 0,01mm. They include atoll-textured spinels (Pasteris, 1980a).

The D5 intrusion is classified as a phlogopite kimberlite. It is a hypabyssal-facies macroporphyritic kimberlite breccia. The groundmass texture is generally uniform but locally a considerable degree of segregation is evident.

7.5.7 The D16 kimberlite breccia

The D16 intrusion is a dull brown kimberlite breccia containing an estimated 20-30 vol.% country rock xenoliths. Shale is the dominant inclusion type, sometimes with well defined reaction rims evident. Subordinate Basement gneiss, Karoo dolerite and Ventersdorp lava inclusions occur. The xenoliths are variable in size, commonly ranging from a few millimetres to about 10cm in diameter. The vast majority of the smaller fragments are shale. Olivine macrocrysts are not prominent although some exceed 10mm in diameter. They are highly altered and dull brown to black in colour. Rare macrocrysts of phlogopite are visible, scattered through the rock. Garnet and ilmenite are present but are rare constituents relative to many other KIMFIK kimberlites. They are seldom seen in hand specimens.

As in most kimberlites two populations of olivine are visible in thin section. Some extremely large (up to 10mm diameter) olivine macrocrysts occur. Several of these macrocrysts exhibit a high degree of rounding but irregularly corroded macrocrysts are also evident. The numerically much more abundant olivine phenocrysts in D16 are mainly subhedral and commonly their margins tend to be blurred as a result of alteration. Alteration of olivine is very severe and fresh olivine is only preserved as the kernels of relatively large macrocrysts (grains having diameters in excess of 3-4mm). Both populations of olivine exhibit wide size variations. Most macrocrysts exceed 1mm ranging up to $\sim$10mm. The majority of the phenocrysts are considerably smaller than 0,5mm but a few exceed 1mm. Rare broken olivine crystals which may relate to the phenocryst and/or macrocryst populations are present.

Olivine alteration mainly involves serpentinization which has been followed by considerable chloritization. Altered olivine grains are commonly pale yellow-green (Plate 5.2b) and slightly pleochroic. The serpentine replacing olivine is replaced to a limited extent by phlogopite (Plate 5.2b). Such phlogopite is concentrated around the rims of individual serpentine/chlorite pseudomorphs. Serpentine in olivine pseudomorphs is also in some instances replaced to a very minor extent by diopside. This diopside is generally restricted to the extreme outer edges of the grains. The alteration products are commonly zonally distributed; diopside occurs at the extreme outer edges of the grains, phlogopite forms an inner rim and serpentine/chlorite occupies the main (central) parts of the grains.

In addition to olivine, scattered phenocrysts of phlogopite and rare phenocrysts of diopside occur. Phlogopite phenocrysts range between 0,1 and 0,4mm and in some instances are bent and strained. Diopside phenocrysts are commonly subhedral, seldom exceed 0,2mm and may display prominent overgrowths. An example of such an overgrowth is shown in Plate 5.6b. The diopside phenocryst illustrated displays several interesting features. The original grain appears to have had its boundaries modified before crystallisation of the overgrowth. Subsequently the overgrowth has been heavily corroded and locally this corrosion has removed the overgrowth entirely so that the original crystal and the overgrowth remnants are pitted and etched by the last corrosion episode.

The dominant groundmass minerals in this kimberlite are phlogopite, diopside and serpentine of which phlogopite is by far the most abundant (Table 7.4). This phlogopite is extremely fine-grained. Individual grains rarely approach 0,1mm in size, are commonly smaller than 0,02mm and many are smaller than 0,01. Thus phlogopite tends to form an overlapping mat of tiny platelets and laths which are often difficult to discern individually (Plate 5.5e).

Groundmass diopside occurs as very small slender prismatic and acicular crystals (Plate 5.2b). Like phlogopite the diopside grains are extremely fine-grained, commonly ranging between 0,005 and 0,01mm. Rare grains approach 0,1mm and groundmass diopside grades down to microlitic material.

The groundmass of the D16 kimberlite breccia displays a fairly well developed segregationary texture reflecting uneven distribution of the major groundmass components (Plate 5.5f). Phlogopite and diopside tend to occur together and to be largely but not entirely separated from serpentine which for the most part occurs in discrete irregular pools (Type 2 segregations). Locally within phlogopite/diopside areas the diopside is strongly clustered in clots or irregular patches. In the D16 kimberlite fine-grained diopside does, in fact, display several modes of occurrence. It occurs as:

(i) A mineral intimately associated with and sometimes clustered within groundmass phlogopite as randomly oriented individual crystals and microlites.

(ii) A mineral replacing (together with phlogopite) serpentinized olivine.

(iii) A mineral replacing country rock xenoliths.

(iv) A mineral fringing or in some instances growing radially inwards at the edges of serpentine pools.

(v) In geodè-like (ocellar, variolitic or vesicular) structures of uncertain origin.

The diopside referred to under (i) above is clearly a primary crystallization product of the magma. This diopside probably represents a second period of diopside crystallization which corresponds to the sharply defined overgrowths present around scattered, larger, early-generation diopside phenocrysts (Plate 5.6b).

The material referred to under points (ii) and (iii) above represents metasomatic and deuteric replacement products of pre-existing rocks and 'early' crystallized kimberlite minerals.

In some instances irregularly shaped clusters of diopside in phlogopite-rich areas surround or partly surround serpentine pools and at the margins of the pools diopside has sometimes grown inwards in radial fashion (Case (iv) above). Neither the boundaries of the entire diopside clusters nor the boundaries of the radial structures are sharp and all the diopside is considered to be primary. These radial structures differ somewhat from others where, although the distribution of diopside and serpentine is similar, the outer limit of the structure is marked by a sharp boundary (similar to that shown in Plate 7.15c). In the latter cases, while a sharp boundary is evident, diopside often occurs on either side of the boundary line. The diopside inside the structure shows distinct radial orientation while that outside and adjacent to the boundary is commonly randomly oriented. It is likely that many of these structures are not primary features but represent serpentinized and diopsidized microxenoliths of country rock (or possibly pre-existing mineral grains) around which primary diopside has crystallized. Frequently it is, however, difficult to reliably assess the origin of these structures. Some may be primary occelli.

The presence of diopside-rimmed serpentine pools which are primary features has some paragenetic implications with respect to the crystallization of diopside and phlogopite. Frequently these pools are surrounded by a groundmass that is composed essentially of phlogopite. This implies that the diopside of the pools (and the associated serpentine) crystallized from residual 'pockets' of liquid after the major groundmass phase - phlogopite. However, outside the pool areas diopside commonly occurs as inclusions in phlogopite or the latter mineral is partly moulded around the former. It is apparent therefore that, relative to phlogopite, diopside crystallized over a long period of time and that diopside crystallization continued after phlogopite crystallization had ceased.

Perovskite and opaque oxide granules are scattered through the matrix of the kimberlite. They range up to about 0,1mm in size but are commonly smaller. They often occur as subhedral to euhedral equant grains. Perovskite is often partly or completely rimmed by opaque spinel and irregular perovskite/opaque oxide intergrowths are also common.

Apatite is in rare instances a prominent mineral in serpentine segregations. Apatite laths are up to 0,1mm long and commonly display typical cross-partings.

Calcite is present in some "kimberlitized" country rock inclusions but is very rare or absent in most thin sections.

The D16 intrusion is classified mineralogically as a phlogopite kimberlite (Table 7.4) and texturally as a hypabyssal-facies macroporphyritic kimberlite breccia which has a segregationary groundmass.

7.5.8 The D18 kimberlite breccia

D18 is a moderately hard, pinkish-brown kimberlite breccia which contains an extremely variable quantity of country rock xenoliths (5-40 vol.%). The groundmass of the rock has a glittering appearance due to abundant fine-grained mica. Numerous lighter-coloured irregular patches of coarser-grained groundmass phlogopite are a prominent feature of the rock.

Most of the country rock xenoliths range between 1 and 4cm in size but xenoliths measuring up to 30cm occur. They include Karoo-derived sandstone, shale and dolerite, Ventersdorp lava and Basement gneiss. Many of the xenoliths are highly altered and they commonly display prominent reaction borders. The xenoliths are subangular or angular and characteristically have irregular shapes.

Olivine macrocrysts range from green and glassy when fresh to dull greenish-yellow and black when highly altered. Ilmenite is the most common of the normal kimberlite indicator minerals but is not abundant. Rare ultramafic nodules are present.

In thin section it is evident that the olivine content of the D18 kimberlite is relatively low compared with the other Dutoitspan intrusions described here. Anhedral, rounded or irregularly corroded, macrocrysts and subhedral to euhedral phenocrysts make up approximately 30 vol.% of the rock (Table 7.4).

The macrocrysts reach 5mm in diameter and are partly or entirely serpentinized and chloritized. The larger macrocrysts usually contain remnants of unaltered material (Plate 5.5a) which commonly display features that are typical of the olivine macrocryst populations in kimberlites (undulose extinction, numerous irregular cracks and partings, partial recrystallization).

The much smaller olivine phenocrysts have been extensively altered. They have been entirely replaced by serpentine which has subsequently been partly or completely altered to darkish green chlorite. The serpentine/chlorite alteration products commonly extend, in ragged fashion, beyond the original grain boundaries into the surrounding groundmass. Consequently the original degree of idiomorphism of many phenocrysts has been reduced. Apart from serpentine and chlorite the only other minerals evident in the olivine pseudomorphs are minute granules of opaque oxide (magnetite) and, in very rare instances, minor amounts of calcite.

The only other macrocrysts evident in thin section are rare anhedral ilmenite grains. The largest grain noted measures 3mm across.

The groundmass of the D18 kimberlite is mineralogically simple. The most abundant minerals are phlogopite and calcite (Plates 5.5a-c). Also present are scattered equant opaque oxide and perovskite grains. Apatite (Plate 5.9d) is a prominent constituent and considerable serpentine is also present (Plate 5.5d). Some of the serpentine is primary but much serpentine occurs as a deuteric replacement product of phlogopite. A little diopside is also present in the groundmass (Plate 5.5d) and rare pectolite occurs as flamboyant radial aggregates between crossed nicols.

The D18 kimberlite has a somewhat unusual segregationary texture. Generally the major groundmass minerals are fairly uniformly spread between olivine crystals which display an orderly distribution (Plate 5.5a). A segregationary texture is, however, produced by the presence of areas which are devoid of olivine (or its pseudomorphs). These areas contain all the major groundmass minerals but tend to be calcite-rich relative to the uniformly-textured main part of the rock (Plates 5.5b and 7.11b). In addition the segregations are relatively coarse-grained compared with the "normal" kimberlite (Plate 5.5c).

The segregations range in size from less than 1mm in diameter to at least 10mm. Many are highly irregular in shape, others are rudely circular or oval. The boundaries of the segregations are invariably diffuse (there is no sharp line of contact between segregations and "normal" groundmass). Some segregations are prominently 'zoned' (Plate 5.5c) with phlogpite concentrated at the margins and calcite in the interiors. In others no preferred orientation or distribution of minerals is evident.

Calcite and phlogopite are by far the most abundant minerals in the segregations. Apatite is concentrated in these areas and pectolite is virtually confined to these areas. Opaques and perovskite are commonly absent.

Where the segregations are large and regular in shape, they form discrete, relatively distinct patches. Elsewhere, however, small irregular patches often link by pinching and swelling and essentially gradational relationships between segregations and 'normal' areas are evident.

Calcite in the segregations commonly occurs as a mosaic of interlocking, interstitial grains with straight (more rarely smoothly curved) boundaries. Individual grains reach 1,25mm in size and may enclose phlogopite in poikilitic fashion. A feature of much of the phlogopite in the segregations is perfect idiomorphism (Plate 5.5c). It occurs as well formed, perfectly euhedral, stubby laths (ℓ x b ratios commonly between 2:1 and 4:1) and hexagonal basal plates. Phlogopite crystals in the segregations reach 0,4mm in size but generally range between 0,1 and 0,2mm. Apatite also displays a high degree of idiomorphism. It occurs as elongate prisms and often displays perfectly euhedral basal sections (Plate 5.5c). Apatite is commonly partly replaced by calcite (Plate 5.9d).

Phlogopite grains in areas of 'normal' groundmass may be characterized by extremely complex zoning and pleochroism but many grains are characterized by dark orange-brown centres and lighter (pale straw-brown) rims. Both cores and rims exhibit reverse pleochroism. The dark cores are often irregular in shape and off-centre with respect to the grain boundaries, i.e. often they bear little geometric relationships to grain boundaries. These phlogopite grains are subhedral to euhedral, rarely exceed 0,2mm in size and commonly measure between 0,05 and 0,15mm. This phlogopite is associated with serpentine, calcite and a little diopside (Plate 5.5d). Serpentine occurs as primary serpophite and is an abundant deuteric mineral replacing phlogopite and, to a limited extent, calcite. In contrast to the calcite of the segregations, calcite grains in the 'normal' areas are very fine-grained, usually between 0,05 and 0,1mm in size. Many of the grains are highly irregular and, in areas where partial serpentine replacement has occurred nearby calcite wisps and shreds (remnants of larger grains) extinguish simultaneously between crossed nicols. Diopside occurs as small prismatic crystals which occasionally reach 0,1mm in length but are generally much smaller; many grains measure less than 0,01mm (Plate 5.5d).

Perovskite is dark brown and often intergrown with opaque material. This opaque material occurs as inclusions within perovskite or as more-or-less continuous rims around perovskite. In other, complex, grains the two minerals are apparently intergrown in random fashion. Equant perovskite grains reach 0,15mm in size but are commonly smaller than 0,08mm.

Texturally this rock is classified as a hypabyssal facies kimberlite breccia. The kimberlite component of this breccia is macroporphyritic. In thin section a segregationary groundmass texture is evident. Mineralogically D2 is classified as a phlogopite kimberlite (Table 7.4).

7.6 MODAL ANALYSES

Modal analyses of 40 thin sections representing 27 discrete kimberlite intrusions in the KIMFIK pipes are given in Table 7.4. Most of these analyses have been recalculated on a xenolith-free basis or xenoliths were not included in the original point counting procedures. A few analyses do, however, show low xenolith contents. In Table 7.5 all 40 analyses have been recalculated to exclude xenoliths (if this had not already been done) and all phenocrystal and xenocrystal olivine (preserved or altered). The main effect of these recalculations is to highlight the relative proportions of the main groundmass minerals in the kimberlites, i.e. the components which in the main are responsible for the mineralogical variations exhibited by kimberlites (as indicated in Chapter 2 and by Skinner and Clement, 1979).

No claim is made that the tabulated modal analyses cover the full range of mineralogical diversity exhibited by the intrusions in the KIMFIK pipes. These analyses do, however, undoubtedly indicate that considerable diversity exists.

For the most part only 500 points per thin section have been counted and in many instances only one slide per intrusion has been counted. Consequently the analyses are best regarded as semi-quantitative representations of modal proportions. In most cases they do, however, provide a reasonable indication of the nature of the kimberlite which they purport to represent. This is borne out by comparison of the results obtained from individual thin sections in cases where more than one slide from an intrusion has been counted. For example, although there is some variation in detail, comparison of analyses DB 1 to DB 4 (Tables 7.4 and 7.5) clearly indicates that the DB 2 intrusion is a monticellite kimberlite and analyses DTP/1 and DTP/2 both indicate that D16 is a phlogopite kimberlite. Similarly 3 of the 4 analyses of kimberlite from the D2 intrusion show that it is a monticellite kimberlite. On the basis of the 4th slide D2 would be classified as a phlogopite-serpentine-monticellite kimberlite but the average of all 4 slides indicates the generally monticellite-rich nature of this kimberlite. Nevertheless, as noted elsewhere in the text (sections 7.3.2 and 7.5) some kimberlites display considerable mineralogical variation (even on a local scale) and for entirely accurate modal representation a large number of thin sections would have to be point counted.

In general the modal analyses quoted in Table 7.4 are considered to be most accurate with respect to the fine-grained groundmass components. The volumetric abundances of scattered relatively large grains, such as olivine phenocrysts and xenocrysts, are considered to have been much less accurately determined. The recalculated analyses of groundmass components given in Table 7.4 probably provide reliable comparisons of the groundmass phases in the analysed rocks.

Examination of the analyses given in Table 7.4 reveals several features of interest:

(i) The ubiquitous presence and general abundance of olivine is apparent. Approximately 75% of the analyses indicate >40 vol.% olivine.

(ii) Phlogopite phenocrysts/macrocrysts are often present in the KIMFIK intrusions but rarely reach major modal proportions.

(iii) Phenocrysts or xenocrysts of diopside are very rarely encountered in thin sections. The same conclusion is applicable to the macrocryst assemblage in general (with the exception of olivine).

(iv) A considerable range of mineralogical varieties of kimberlite are represented (see also sections 7.3.2 and 7.5 and Appendix I, Vol. II).

(v) Phlogopite kimberlites are relatively abundant in the KIMFIK pipes.

(vi) Groundmass diopside is generally a prominent groundmass component of Finsch kimberlite intrusions. It is much less abundant in the Kimberley pipes and in many instances is entirely absent (see also section 7.3.2 and Appendix I).

(vii) Many dyke intrusions are calcite-rich, probably indicating retention of volatiles during their crystallization.

(viii) Apatite is a common accessory mineral.

(ix) The total opaque oxide content may assume considerable modal proportions of some kimberlites (see also Clement and Skinner, 1979).

(x) The petrographic similarity of the DB2 and D2 intrusions noted in section 7.5.5 is borne out by the averaged modal analyses of thin sections from the two intrusions (Table 7.6). D2 appears to have a slightly higher content of volatile components.

7.7 PARAGENETIC CONSIDERATIONS

A detailed assessment of the parageneses of the minerals in the KIMFIK intrusions is handicapped by the hybrid nature of the rocks. Furthermore this problem is compounded by the complex, often repetitive, reaction/breakdown/resorption processes (Pasteris, 1980a) to which many of the minerals have been subjected. Many examples of the results of these processes have been documented in Chapters 5, 6 and 7.

An additional problem in determining the paragenesis of the minerals in the kimberlites is that their formation needs to be considered in relation to at least six potential periods of crystallization:

(i) In the upper mantle.

(ii) During ascent (which is generally considered to have been rapid to allow upward transport of high density xenoliths and to preserve diamond outside its stability field).

(iii) In temporary 'magma chambers' reflecting 'rest periods' during intermittent ascent.

(iv) In a hypabyssal environment following ascent to near-surface levels.

(v) During vapour-solid fluidization following breaching of the surface.

(vi) During near-surface post-fluidization crystallization.

A third major paragenetic problem is posed by the phenocryst/xenocryst question. Firstly, which macrocrysts (and megacrysts) are phenocrysts and which are xenocrysts? Secondly, if xenocrysts are present, when are they incorporated, particularly in relation to the crystallization of early phenocryst phases? Although this thesis is primarily concerned with near-surface processes some discussion of the foregoing problems is presented below.

7.7.1 Olivine crystallization

As indicated in section 6.2 macrocrysts of olivine are interpreted essentially as peridotite-derived xenocrysts and the euhedral olivine crystals are considered to be early-crystallized phenocrysts. The latter interpretation is contrary to the views of some authors. For example, Mitchell (1978), Elthon and Ridley (1979) and Snowden (1981) consider that many or all euhedral olivines post-date 'fluidized' intrusion and crystallize at or near surface as 'groundmass' phases. While the probability of an entirely distinct, extremely fine-grained, post-intrusion, quenched olivine groundmass phase exists (as indicated by Skinner, see section 5.2) a post-intrusion, near-surface, origin for the euhedral phenocrysts is considered untenable. Several reasons for this can be advanced, as discussed below.

The olivine phenocrysts in all the KIMFIK intrusions are similar in character (except with respect to late-stage deuteric alteration) irrespective of whether they occur in major pipe intrusions, in minor dykes (of different relative ages), or in sills or veins and irrespective of the texture or the groundmass mineralogy of the rocks. Irrespective of the size or type of body in which they occur the phenocrysts exhibit similar or identical size ranges (0,05-0,5mm with rare larger and smaller grains) and have invariably crystallized as well formed polyhedra (although ideal morphology may have been somewhat modified by alteration and limited magmatic corrosion). The phenocrysts in different intrusions also display similar compositional ranges (Figure 6.1A). The consistent character of the olivine phenocrysts in otherwise with phenocryst compositions, a case is made to indicate that, in fact the euhedral olivines crystallize in the upper mantle, at an early stage in the evolution of kimberlite magmas.

7.7.2 Major element chemistry of olivine phenocrysts

A detailed assessment of compositional variations within olivine phenocrysts from the KIMFIK pipes (and neighbouring intrusions) is handicapped by a paucity of data.

Analyses of 14 euhedral olivines from the DB 2 kimberlite in the De Beers pipe have indicated a wide range of core compositions ($Fo_{93,1-87,2}$; data from Boyd and Clement, 1977, Pasteris, 1980a and S.R. Shee, personal communication). Similarly Snowden (1981) recorded a range of $Fo_{91,1-87,4}$ from the cores of 22 phenocrysts from the D2 kimberlite in the Dutoitspan pipe and Dawson and Hawthorne (1973) noted a range of $Fo_{93,1-87,9}$ from only 8 analyses of phenocryst cores from the Benfontein sill complex.

Similar ranges of core compositions have been noted elsewhere. For example, Nixon and Boyd (1973b) recorded a range of Fo_{95-88} for euhedral olivines from the Liqhobong satellite pipe in Lesotho (16 analyses). Mitchell (1973) also recorded a considerable compositional range ($Fo_{93-88,5}$) from 10 phenocrysts in the Ison Creek occurrence (U.S.A.) and from 44 olivines ($Fo_{93,5-88}$) in the Elwin Bay kimberlite in Canada (Mitchell, 1978).

More extreme compositional ranges have been found by Emeleus and Andrews (1975) from minor kimberlite intrusions in Greenland. They recorded a range of Fo_{92-76} and state specifically that both idiomorphic (phenocrysts) and xenomorphic (xenocrysts) grains may be Fe or Mg-rich. Similarly Mitchell (1980) noted a compositional range of $Fo_{93,5-86}$ (70 analyses) in the Jos dyke in Canada and he stated that the bulk of this variation could be related to compositional differences among phenocrysts. In contrast to the wide ranges in core compositions quoted above the olivines in some kimberlite intrusions display more restricted ranges. Thus Mitchell (1973a) recorded a range of $Fo_{91,5-88,5}$ from 24 analyses of olivine phenocrysts from the Wesselton pipe (probably the W3 kimberlite). Elthon and Ridley (1979) obtained core compositions ranging from approximately $Fo_{90,5}$ to about $Fo_{87,5}$ from olivines in the Premier pipe (probably, in fact, the Wesselton pipe; see Scott and Skinner, 1979). Euhedral olivines in Greenland kimberlites examined by Scott (1981) showed virtually no compositional variation with forsterite contents centred around Fo_{85} and Hill (1977) found a very limited range ($Fo_{86,9-85,5}$) in the core compositions of eight phenocrysts from a narrow kimberlite sill associated with (and cut by) the Wesselton pipe.

It is evident from the foregoing review that in a number of instances the cores of olivine phenocrysts in specific kimberlite intrusions display marked chemical inhomogeneity. Furthermore in several of the examples quoted the number of analyses are limited hence the true compositional ranges are likely to be greater than those recorded. It should also be noted that the variable core compositions recorded are not due to the derivation of the analysed grains from markedly different parts of individual intrusions. For example, Boyd and Clement (1977) recorded a range of 5,9 mol.% Fo from 10 analyses in a single thin section.

While the core compositions of olivine phenocrysts often vary substantially in specific kimberlites compositional zoning of individual phenocrysts is generally limited. For example, Boyd and Clement (op. cit.) noted generally homogeneous compositions across phenocrysts from the DB 2 kimberlite with minor major element zoning (equivalent of 0,5 mol.% Fo or less) restricted to thin rims (Fig. 7.7). Similarly in only one case out of 16 core-rim comparisons from the D2 kimberlite at Dutoitspan does the forsterite content differ by more than 1,5% and in most cases the differences are less than 0,5% (Snowden, 1981). Data

Also there does not appear to be any suitable mechanism for producing almost unzoned phenocrysts of widely different composition during rapid intrusion to surface. Rapid intrusion would promote rapid crystallization and the latter would facilitate zoning (Cox et al., 1979). In addition the habit of the olivines and the lack of relationships between phenocryst sizes and compositions are inconsistent with rapid crystallization during intrusion.

The early crystallization of euhedral phenocrysts has an important temporal implication with respect to the anhedral xenocrysts (macrocrysts) of olivine in kimberlites. The xenocryst assemblages display wider compositional ranges than the euhedral phenocrysts with which they co-exist but on average are generally more magnesian (e.g. Fig. 6.1B). Like the phenocrysts the co-existing xenocrysts also display relatively little zoning and such zoning as occurs is again generally restricted to very thin rims (e.g. Boyd and Clement, 1977; Snowden, 1981). If the xenocrysts were already present in liquids which later crystallized less magnesian phenocrysts within the upper mantle, then much greater degrees of re-equilibration of xenocrysts towards phenocryst compositions would be expected. Alternatively overgrowths (with phenocryst compositions) might occur around xenocryst cores. Since neither extensive re-equilibration nor overgrowths have been recorded an appropriate inference is that euhedral phenocrysts crystallized prior to the incorporation of xenocrysts. At the time the xenocrysts were incorporated, the liquid(s) could not have been crystallizing or extensively resorbing olivine and such processes could not have occurred to any great extent subsequently.

It could alternatively be argued that most olivine xenocrysts were incorporated 'early' in xenoliths and were not liberated into kimberlite magmas until the foregoing conditions pertained.

crysts and which are xenocrysts? Secondly, if xenocrysts are present, when are they incorporated, particularly in relation to the crystallization of early phenocryst phases? Although this thesis is primarily concerned with near-surface processes some discussion of the foregoing problems is presented below.

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Rapid post-intrusion olivine crystallization is also precluded by the morphology of the phenocrysts. Donaldson (1976) considers that well formed polyhedra are indicative of relatively slow cooling - a condition which, as noted above, is unlikely to have been satisfied after kimberlite intrusion. In this respect it is interesting to compare the morphology of olivine phenocrysts in the KIMFIK kimberlites with those in Precambrian dykes, which have kimberlite affinities (Clement et al., 1979) and are located near the Wesselton and De Beers pipes. In these dykes much (possibly all) olivine undoubtedly crystallized after intrusion and considerable textural variation is evident across the intrusions. Corresponding changes in olivine morphology occur. Skeletal (hopper) crystals (Plate 7.14h) occur near the margins of the dykes while crystals displaying greater morphological perfection occur in the central areas. Such variations are common in olivine-rich basic and ultrabasic minor intrusions (e.g. Drever and Johnston, 1957; Cox et al., 1979) but are not present in any of the KIMFIK intrusions.

In view of the foregoing discussion there is little doubt that olivine phenocrysts crystallize prior to final intrusion and in some kimberlites this is unequivocably indicated by parallel orientation (due to flow) of these crystals. In the following section, which deals with phenocryst compositions, a case is made to indicate that, in fact the euhedral olivines crystallize in the upper mantle, at an early stage in the evolution of kimberlite magmas.

7.7.2 Major element chemistry of olivine phenocrysts

A detailed assessment of compositional variations within olivine phenocrysts from the KIMFIK pipes (and neighbouring intrusions) is handicapped by a paucity of data.

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Several of the authors mentioned above stress that compositional zoning is commonly restricted to very thin rims. The most detailed study in this respect is that by Boyd and Clement (1977). Several authors (e.g. Boyd and Clement, 1977; Emeleus and Andrews, 1975; Snowden, 1980; Dawson and Hawthorne, 1973) also stress that the generally restricted compositional zoning recorded may be extremely complex, i.e. normal, reverse or oscillatory. The net effect of such zoning is for the compositions of the extreme edges of phenocrysts in individual kimberlites to converge upon a common compositional band (e.g. Boyd and Clement, 1977; Snowden, 1981; Emeleus and Andrews, 1975).

In terms of the foregoing discussion the most notable features of the major element compositions of olivine phenocrysts are:

(i) The commonly considerable compositional variation exhibited by the cores of individual crystals in discrete intrusions (Fig. 6.1A).

(ii) The limited nature of major element zoning within such phenocrysts.

(iii) A pronounced tendency for the thin, often complexly zoned, rims of phenocrysts to trend outwards to a common composition.

These features have several implications with regard to the origin of olivine phenocrysts in kimberlites. With respect to core compositions two features are particularly relevant. Firstly, there appears to be no relationship between grain size and composition and, secondly, in the majority of examples quoted the compositional ranges exhibited by phenocryst cores are not matched by concomitant degrees of zoning within the phenocrysts themselves. Thus, while core compositions between phenocrysts in a single intrusion may vary by 5-10 mol.% Fo , zoning in individual phenocrysts is often limited to 1 wt.% MgO or less. Such features are difficult to reconcile with any model for phenocryst generation from a single magma body.

If temperature, pressure and/or liquid composition of a single magma source changed sufficiently during a continuous crystallization event, to produce ranges in core compositions of 5 or more mol.% Fo, major zoning of the phenocrysts would be expected. The absence of zoning cannot be ascribed to equilibrium crystallization as core compositions are so variable. Zoning should be particularly evident in relatively large (and hence presumably early-nucleated) phenocrysts. Relatively small (later-crystallizing) phenocrysts should show reasonable compositional correspondence (Fe-enriched assuming a normal crystallization trend) and should correspond with the outer zones of larger crystals. In fact a general correspondence between compositional variation and grain size would be expected. However, as noted previously, neither major zoning nor composition/grain size relationships are evident among the olivine phenocryst populations that have been studied.

Recourse to discontinuous crystallization from a single magma body to explain phenocryst compositional variations is also unsatisfactory. If discontinuous olivine precipitation occurred and early phenocrysts did not re-equilibrate with the later, more evolved liquid, many phenocrysts would be expected to have distinct, compositionally different overgrowths around earlier formed cores - a common feature of other minerals in the kimberlites. Furthermore, in view of the wide range of core compositions recorded, several generations of overgrowths might be expected and the number of overgrowths should relate to phenocryst size. None of the studies discussed previously provided any evidence for the occurrence of such overgrowths.

To obtain the wide compositional variations recorded, unaccompanied by concomitant zoning, it could be argued that crystallization took place, from a fractionating liquid, in a non-convective magma chamber. The absence (or paucity) of zoning could then be ascribed to removal of successive compositionally different phenocrysts as cumulates. Genetic models of this generalized type are not favoured, however, because vertical compositional and temperature gradients are likely to develop in non-convective magma chambers. In such circumstances fortuitously delicate balances between crystal settling rates, liquid compositions, and cooling rates would be required to allow the accumulation of compositionally varied but unzoned phenocrysts.

A possible way of explaining the unusual chemical features of the olivine phenocrysts is to adopt some form of incremental batch melting model (Cox et al., 1979). A batch mixing model has been proposed by Boyd and Clement (1977). These authors envisaged extensive mixing (during intrusion) of phenocrysts derived from batches of magma with substantially different bulk compositions.

In terms of an incremental batch melting model successive partial melt extracts could separately produce, under equilibrium conditions, unzoned phenocrysts with different compositions. Subsequent mixing of crystals (and liquids) could account for the compositionally variable but relatively unzoned phenocrysts found in many kimberlites. Such a model is, however, subject to one stringent constraint; olivine crystallization must be essentially complete in all melt batches prior to mixing. If olivine remained for long periods above or close to the liquidus following the mixing of compositionally variable phenocrysts then considerable degrees of reaction, corrosion, major and minor element zoning or overgrowth development would be expected.

Pre-mixing crystallization strongly implies that olivine phenocrysts must have formed at very early stages in the evolution of discrete kimberlite intrusions, i.e. within the upper mantle.

As noted by Cox et al. (op. cit.) the later liquids resulting from incremental melting will be severely depleted in incompatible elements. With respect to the batch melting model advocated above severe depletion of such elements in individual batches might be avoided by assuming:

(i) That the overall degree of melting is limited (cf. the incipient melting hypothesis of Dawson, 1980).

(ii) A dynamic melting, open system model in which a proportion of successive melts is mechanically trapped within the source material and mixes with the next-formed liquid (Langmuir et al., 1977).

If the deductions arising from the foregoing admittedly speculative arguments are valid, two main conclusions can be drawn:

(i) That euhedral olivine phenocrysts crystallize in the upper mantle at an early stage in the evolution of kimberlite magmas.

(ii) That these phenocrysts crystallize from discrete magma batches and are subsequently mixed.

The first of these conclusions is supported by the fact that none of the other potential crystallization sites is suitable. Post-intrusion crystallization has already been rejected. Crystallization of compositionally different phenocrysts, in successively higher magma chambers, during intermittent uprise, is conceivable. However, there is no explanation of why early phenocrysts are not partially or completely made over to new compositions during successive crystallization periods.

Also there does not appear to be any suitable mechanism for producing almost unzoned phenocrysts of widely different composition during rapid intrusion to surface. Rapid intrusion would promote rapid crystallization and the latter would facilitate zoning (Cox et al., 1979). In addition the habit of the olivines and the lack of relationships between phenocryst sizes and compositions are inconsistent with rapid crystallization during intrusion.

The early crystallization of euhedral phenocrysts has an important temporal implication with respect to the anhedral xenocrysts (macrocrysts) of olivine in kimberlites. The xenocryst assemblages display wider compositional ranges than the euhedral phenocrysts with which they co-exist but on average are generally more magnesian (e.g. Fig. 6.1B). Like the phenocrysts the co-existing xenocrysts also display relatively little zoning and such zoning as occurs is again generally restricted to very thin rims (e.g. Boyd and Clement, 1977; Snowden, 1981). If the xenocrysts were already present in liquids which later crystallized less magnesian phenocrysts within the upper mantle, then much greater degrees of re-equilibration of xenocrysts towards phenocryst compositions would be expected. Alternatively overgrowths (with phenocryst compositions) might occur around xenocryst cores. Since neither extensive re-equilibration nor overgrowths have been recorded an appropriate inference is that euhedral phenocrysts crystallized prior to the incorporation of xenocrysts. At the time the xenocrysts were incorporated, the liquid(s) could not have been crystallizing or extensively resorbing olivine and such processes could not have occurred to any great extent subsequently.

It could alternatively be argued that most olivine xenocrysts were incorporated 'early' in xenoliths and were not liberated into kimberlite magmas until the foregoing conditions pertained.

A possible criticism of the phenocryst mixing theory is that in some cases mixing of a relatively large number of phenocryst batches would be required to account for the spread in Mg/Fe ratios of phenocryst cores. Such mixing does not, however, appear unrealistic when the complex nature of many kimberlite bodies is considered. Kimberlite pipes, dykes and sills are often composed of successive magma pulses which differ substantially in bulk composition. Furthermore there is no doubt that in many instances these pulses were intruded in rapid succession; certainly close enough in time for earlier injections to have undergone little or no post-intrusion crystallization before later intrusions were emplaced (see Chapter 10). Classic examples of the latter situation, involving very limited amounts of magma, have been described by Dawson and Hawthorne (1973) and Donaldson and Reid (in press). If almost contemporaneous magma pulses have been injected in rapid succession along common intrusion paths from mantle depths it is not unreasonable to suppose that in some instances actual mixing of pulses has occurred prior to final emplacement; probably at various depths along the intrusion paths (see Chapter 10).

The possible incorporation of xenocrysts subsequent to the crystallization of phenocrysts is a somewhat radical suggestion. It is, however, surprising, if the reverse situation pertains, that the generally more magnesian macrocrysts do not show either a greater degree of re-equilibration towards phenocryst compositions or the occurrence of less magnesian overgrowths (corresponding to phenocryst compositions). The absence of such features is even more surprising when the morphological perfection of the phenocrysts is considered. As indicated previously (section 7.7.1) olivine phenocrysts commonly occur as well formed, polyhedral crystals which are indicative of relatively slow crystallization (Donaldson, 1976). Such crystallization conditions should favour extensive reaction between the xenocrysts (if they were already in the magma) and the melt or the formation of prominent 'phenocryst' overgrowths.

As noted previously the re-equilibration problem can be overcome by assuming delayed liberation of xenocrysts from incorporated xenoliths but is doubtful if this argument can be applied to very large proportions of the olivine xenocrysts in kimberlite.

It is difficult to envisage a situation in which, following phenocryst generation and xenocryst incorporation, kimberlite melts are sufficiently far from forsteritic olivine liquidi to prevent reaction or resorption; particularly when the varying compositions of both phenocrysts and xenocrysts are considered. In fact the thin zoned rims of both phenocrysts and xenocrysts noted by Boyd and Clement (1977) and others testify to attempts by both olivine populations to re-equilibrate towards common compositions. It is likely, however, that extensive degrees of re-equilibration following mixing are prevented by relatively rapid intrusion to near-surface levels. This implies that olivine xenocrysts are incorporated at, or just before, the time when relatively rapid intrusion to surface is initated.

7.7.3 Other macrocrysts

Garnet, clinopyroxene and orthopyroxene are associated, in various combinations, with olivine in the peridotites and pyroxenites from which the olivine xenocrysts are derived. It follows therefore that these minerals must also be xenocrysts (as indicated in section 6.5) and they would be incorporated into the kimberlites at the same time as the olivine xenocrysts.

The origin of ilmenite macrocrysts remains enigmatic despite fairly extensive investigations and considerable debate in the last decade. The extreme points of view are represented by:

(i) Those who consider such ilmenites to be early phenocrysts, precipitated at depth (within the upper mantle) from kimberlite melts (e.g. Mitchell, 1973b, 1977; Elthon and Ridley, 1979).

(ii) Those who consider such ilmenites to be xenocrysts which were incorporated into kimberlite melts at upper mantle depths (e.g. Boyd and Nixon, 1973, 1975; Haggerty, 1975; Haggerty et al., 1979; Pasteris, 1980a).

Resolution of this problem is beyond the scope of this thesis. It should however, be noted that if ilmenite macrocrysts are to be interpreted as phenocrysts some aspects of the compositional variation which they display require explanation. These aspects pertain mainly to the extent and nature of the compositional variations evident between ilmenite grains in individual kimberlite intrusions (Section 6.4).

The extent to which the wide compositional ranges recorded may be due to growth or reaction zoning of ilmenite has not always been indicated in the literature. There are certainly some instances where extensive marginal zoning has occurred (e.g. Haggerty et al., 1979; Pasteris, 1980a). There is, however, no doubt that in other cases wide compositional ranges between grains from discrete intrusions are accompanied by little or no zoning (Mitchell, 1973b; Dawson, 1980) or zoning is restricted to very thin rims (Wyatt, personal communication). In such cases the situation mirrors that displayed by olivine phenocrysts and macrocrysts (section 7.7.2); large compositional variations are exhibited by neighbouring crystals which display little or no indication of re-equilibration with evolving liquids. Furthermore there are no indications that early formed crystals (or xenocrysts incorporated at an early stage into the liquids) have been protected by armouring in the form of overgrowths. Mitchell (1973b) accounted for this anomalous situation by postulating extensive fractional crystallization under equilibrium conditions. He envisaged the build-up of chemically braded ilmenite cumulates as progressively less magnesian, unzoned, ilmenite phenocrysts (macrocrysts) were removed from kimberlite liquids. Later (1977), Mitchell concluded that the distribution of trace elements in such ilmenites did not support the fore-going concept of differentiation-related series of crystals. He therefore suggested that the compositional differences, evident in ilmenites from individual kimberlites, reflected changes in some other parameter such as temperature, oxygen fugacity or silica activity.

The rationale behind Mitchell's (1977) views is an assumption that the change in Fe/Mg ratios of the ilmenite is not accompanied by significant changes in the Fe/Mg ratio of evolving kimberlite liquids. Essential to this interpretation is a further assumption that the amount of ilmenite crystallizing is small in relation to the total amount of magma available (so that ilmenite crystallization will not markedly affect the bulk composition of the liquid). There is, however, fairly general agreement that kimberlite melts are volumetrically limited. It is also known that modal amounts of ilmenite macrocrysts approaching 3-4% occur relatively frequently (Dawson, 1980) and in rare cases exceed 5% (E.M. Skinner, personal communication). Under such circumstances it is doubtful whether the crystallization of significant (albeit accessory) amounts of ilmenites (varying in composition by 5-15 wt.% MgO in some cases) would be unaccompanied by changes in Fe/Mg ratios of the bulk compositions of the evolving liquids.

Alternatively a phenocrystic origin for ilmenite macrocrysts might be postulated if it supposed that they crystallized in discrete, compositionally distinctive, batches of kimberlite magma and were subsequently mixed, prior to final intrusion (see section 7.7.2). Such a possibility has been alluded to in principle by Pasteris et al. (1979). However, an alternative hypothesis is to suppose that the ilmenites are xenocrysts derived from heterogeneous, pre-existing crystal mush-magma assemblages sampled by kimberlites, i.e. they are discrete nodules in the sense of Nixon and Boyd (1973a), Boyd and Nixon (1975) and Pasteris et al. (1979).

There are at present no chemical criteria for separating discrete macrocryst and megacryst (discrete nodule) ilmenite assemblages. An entirely megacrystic origin (in the sense of the above-mentioned authors) is, however, unlikely. In addition to possible phenocrysts, ilmenite-bearing peridotitic and MARID-suite rocks are also potential sources of at least part of the coarse ilmenite assemblages in kimberlites.

In view of the unresolved genetic problems of ilmenite macrocrysts their temporal relationships with olivine phenocrysts and known xenocrysts (peridotitic and/or eclogitic) cannot be evaluated with confidence. If, however, many ilmenite macrocrysts are batch-crystallized phenocrysts it is possible that they, like olivine phenocrysts, crystallized prior to the incorporation of ultramafic, nodule-derived, xenocrysts.

The 'ages' of all the megacrysts (discrete nodules) found in kimberlites (in relation to early crystallized minerals such as olivine) are difficult to assess because of doubts as to their origin. As noted in section 6.6 there is currently considerable debate as to whether megacryst-producing liquids are 'kimberlitic' or whether megacrysts crystallized from unrelated melts. In a sense arguments as to whether the liquids are 'kimberlite', 'proto-kimberlite', 'kimberlitic' or 'kimberlite-like' are pedantic. The product of any limited partial melt of four-, five- or six-phase (Ol+Cpx+Opx+Gt±Ph±Carb.) mantle peridotite is likely to be 'kimberlite-like'. The important aspect is whether the liquids which produce the megacrysts also produce, or are closely associated with, the host kimberlites. If this is the case most authors would agree that megacrysts crystallize before early primary minerals such as euhedral olivine (see discussion in Robey, 1981). The problem of variable compositions without attendant zoning is, however, again evident; more so, in fact, because of the generally Fe-rich nature of many megacrysts (e.g. olivine) in relation to phenocrysts and/or macrocrysts of the same minerals (e.g. Gurney et al., 1979b). It would appear therefore that megacrysts do not crystallize from the liquid(s) which produce their host or associated kimberlites.

7.7.4 The remaining matrix minerals

The crystallization history of the primary minerals in the KIMFIK pipes (i.e. those minerals that constitute the 'matrix' as defined in Chapter 2) can for the most part be interpreted in terms of two crystallization periods. In essence 'pre-intrusion' and 'post-intrusion' minerals can be recognised. It should, however, be noted that the implied intrusive events are those that involve the relatively rapid transport of kimberlite melt (plus some crystalline material) to near-surface levels. The subdivision does not relate to any relatively slow movements of melts (±crystals) within the upper mantle which might be linked with the actual genesis of kimberlite melts and/or the initial stages of their emplacement and crystallization. Similar groupings or 'early' and 'late' minerals have been recognized by other authors although there is a lack of agreement as to whether an actual and significant hiatus in crystallization separates the two groups of minerals (Clarke and Mitchell, 1975; Mitchell, 1978; Elthon and Ridley, 1979; Snowden, 1980). Clarke and Mitchell (op. cit.) introduced the terms 'pre-fluidization' and 'post-fluidization' to describe the two groups of minerals. These authors have, however, commonly used these terms in relation to hypabyssal-facies kimberlites which have not been fluidized in the classic kimberlite sense of gas-solid fluidization (see discussion in section 10.2.3).

Postulation of a period during emplacement when little or no significant crystallization occurs is largely prompted by the rapidity with which intrusion from the upper mantle to near surface levels must take place, i.e. fast enough to prevent complete resorption of diamonds and to prevent sedimentation of ultrabasic nodules. These constraints do not imply that intrusion from mantle depths took place in a matter of a few hours in the form of accelerating fluidized systems which attained speeds of ∿400 metres/second by the time the surface was reached (e.g. McGetchin et al., 1973). Evidence from the KIMFIK pipes indicates that at least some intrusions were emplaced relatively slowly in near-surface zones (Chapter 10).

The validity of the proposed crystallization hiatus is heavily dependant on the conclusion that the major phenocryst phases in kimberlites (olivine and, in some cases, phlogopite) crystallize prior to 'rapid' intrusion. If the latter hypothesis is correct there is little chance of any modally abundant mineral having crystallized during rapid intrusion as olivine (±phlogopite) phenocrysts plus various macrocrysts (xenocrysts) plus the major post-emplacement groundmass minerals (some combination of diopside, monticellite, phlogopite, calcite or serpentine) commonly make up 85-95 vol.% of the KIMFIK kimberlites (Table 7.4).

The remaining (accessory) minerals commonly consist of apatite, opaque (and other) oxides and probably minor sulphides. Of these apatite is clearly a post-intrusion mineral which often occurs in late-stage segregations. Elucidating the paragenesis of the opaque grains has not been a goal of this thesis but it is evident from the limited work carried out and from the results of other researchers that a considerable part (if not all) of the opaque mineral assemblages either pre-date or post-date rapid intrusion.

Aspects of the genesis of ilmenite were noted previously. It was concluded that if ilmenite macrocrysts were phenocrysts they are likely to pre-date rapid intrusion. This view is consistent with those of several other authors who have concluded that anhedral picroilmenite grains and a variety of Cr-Mg-Al spinels crystallized in the mantle (e.g. Mitchell, 1973b, 1977, 1978; Snowden, 1980; S.R. Shee, personal communication). Rutile commonly occurs as inclusions in olivine phenocrysts (section 5.2.4) and hence is also interpreted as a pre-intrusion mineral.

Insofar as the remaining opaque minerals are concerned there is a consensus of opinion that many, either in the form of reaction products, overgrowths or discrete primary crystals, post-date emplacement. The minerals concerned include perovskite, magnetite, Ti-Mg chromite, Ti-Mg-Al chromite, and sulphides (Mitchell, 1978; Elthon and Ridley, 1979; Pasteris, 1980a; Snowden, 1980). S.R. Shee (personal communication) also considers groundmass ilmenite to be a post-intrusion mineral. Some small ilmenite crystals occur, however, as inclusions in olivine phenocrysts (Shee, in press).

It is evident from the above discussion that many of the generally minor constituents of the KIMFIK pipes can also be assigned to pre-intrusion and post-intrusion crystallization periods. However, a minor amount of primary crystallization during relatively short upper mantle to surface ascent periods cannot be ruled out. This applies particularly to certain oxide minerals. Thus Mitchell (1978) has indicated that some Ti-Mg chromites crystallize during emplacement and Pasteris (1980) has also advocated the crystallization of some spinels and ilmenite during intrusion to high crustal levels.

In addition to limited primary crystallization during intrusion (notably of opaque phases) it is likely that modification of some opaque and silicate phases of breakdown/reaction/resorption processes occurred. For example, Elthon and Ridley (1979) proposed that reaction of picro-ilmenite with kimberlite magma during intrusion resulted in the formation of perovskite and magnetite. They considered that this process conveniently separated 'early' and 'late' periods of crystallization. In so far as silicates are concerned it is likely that reaction zoning producing compositionally different thin rims and early serpentinization of olivine occurred during the later stages of intrusion. In fact 'early' serpentinization of olivine (and, in some instances, phlogopite) often appears mandatory to allow for the subsequent complex deuteric replacement of altered olivines by other minerals as described in section 5.2.3. The concept of early serpentinization is consistent with the views of Pasteris (1980) and Donaldson and Reid (in press). In view of the foregoing comments it should, however, be stressed that many of the complex reaction/replacement/resorption processes that are characteristic features of the mineralogy of kimberlites, have been clearly recognized as post-intrusion phenomena. This is evident from many of the descriptive and interpretative comments made in this thesis (Chapters 5, 6, 7 and 10) and has been recognized by other authors (e.g. Dawson, 1980; Pasteris, 1980a; McMahon et al., 1979).

Considerable mineralogical diversity is displayed by the intrusions within the KIMFIK pipes (sections 7.2.3, 7.5 and Appendix I). This variability probably reflects intrinsic or primary differences in the magmas or magma batches concerned and variations in mode of emplacement (section 10.2). Consequently, although a generalized crystallization sequence can be established it may not necessarily be applicable in detail in every case. This generalized crystallization sequence is shown in Figure 7.8 in relation to the various processes which are considered to have operated during the evolution of the KIMFIK pipes (these processes are discussed in Chapter 10). It is important to note that the depths and temperatures indicated in Figure 7.8 are approximations. They are 'ballpark' figures based loosely on data from Wyllie (1966), Franz and Wyllie (1967), Wyllie and Huang (1976), Clarke and Mitchell (1975), Mitchell and Clark (1976), Mitchell (1978) and Pasteris 1980a. Somewhat higher near-surface temperatures have been postulated in respect of the unusual Benfontein Sill Complex near Kimberley (McMahon et al., 1979).

The primary crystallization sequence of the common post-intrusion minerals in the KIMFIK intrusions is usually readily apparent from fabric studies and deuteric alteration and replacement relationships. Such intergrain and replacement relationships have been described and illustrated elsewhere in this thesis (Chapters 5, 6 and 7). However, although the relative ages of the onset of crystallization of different minerals can be fairly readily established there is no doubt that the crystallization of many of the post-intrusion (and pre-intrusion) minerals overlaps and the extent of these overlaps has not been well defined.

In general (as shown in Figure 7.8) the initiation of post-intrusion crystallization in hypabyssal-facies KIMFIK intrusions is considered to take place in the following sequence: perovskite/spinels/ilmenite, diopside, monticellite, apatite, phlogopite, calcite and serpentine. Perovskite and some opaque oxides are considered to be the earliest post-intrusion minerals to form because they occur as euhedral microphenocrysts (or complex aggregates), they form 'necklace' textures around olivine phenocrysts and are commonly included within other relatively early-crystallizing post-intrusion minerals, notably phlogopite (Plates 5.4a-g). Diopside commonly occurs as inclusions within phlogopite as does monticellite (Plates 5-4c. 7.15d-e). Apatite sometimes occurs as inclusions in phlogopite (Plate 6.3g) but is also common in volatile-rich segregations (Plates 5.10a, 7.14d-e and Fig. 5.9d) in which circumstances it may post-date phlogopite crystallization. Both apatite and phlogopite are often deuterically replaced by calcite and/or serpentine which appears to be contemporaneous and physically continuous with primary material. Similarly late-stage serpentine often replaces calcite (Plates 5.6g-h and 5.8a-f).

The generalized sequence outlined above does not hold with respect to diatreme-facies kimberlites. In the latter cases crystallization sequences have been affected by loss of volatiles and, probably, by the incorporation of contaminants during fluidization. The main effect appears to be a delay in the crystallization of diopside which may crystallize mainly after the cessation of fluidization and after minerals such as monticellite and phlogopite have formed (section 7.2.3).

It appears also (based on replacement relationships) that sequences of deuteric replacement occur which broadly follow (in attenuated forms) the post-intrusion primary crystallization sequence outlined above, except that an initial stage of serpentinization (or, possibly, steatization in the case of some highly micaceous kimberlites) is a prerequisite, at least in respect of deuteric alteration of olivine (section

5.2.3). Deuteric replacement by one or more of the post-intrusion groundmass minerals commonly appears to be more-or-less contemporaneous with primary crystallization of the same minerals. This is generally indicated by physical continuity between the deuteric product and the primary groundmass phase (e.g. Plates 5.2c-d). In some instances deuteric minerals are larger than the same minerals in the groundmass or specific minerals occur only as deuteric alteration products of earlier minerals. Such situations suggest that deuteric alteration was initiated prior to groundmass crystallization because physico-chemical conditions favoured deuteric replacement rather than primary crystallization and that subsequent changes in conditions (P,T and magma compositions) inhibited or precluded groundmass crystallization.

Thus far in the discussion it is apparent that a reasonable case can be made in favour of all, or almost all, kimberlite crystallization taking place either in the upper mantle (prior to rapid intrusion) or near the surface after intrusion. Substantial degrees of crystallization during ascent are not favoured. However, the possibility of intermittent ascent with crystallization during 'rest periods' in temporary magma chambers has not yet been considered. Perhaps the best argument in favour of such crystallization is the occurrence of scattered (rare), coarse-grained, usually simple two or three grain aggregates of olivine and phlogopite (±opaque minerals) or, less frequently, of diopside and phlogopite (Plates 5.7e, 6.3f and h and 6.4a-b).

These aggregates are not xenoliths derived from mantle rocks as the olivine, opaque grains and diopside are commonly euhedral. The phlogopite is anhedral to subhedral and is often partly moulded around or poikilitically encloses olivine, diopside or opaque euhedra.

These aggregates may originate in two ways. They may be autoliths derived from earlier kimberlites which consolidated at considerable depth (the coarse grain size suggests slow cooling). Alternatively their simple mineralogy suggests that they may represent disrupted cumulates entrained in later intrusive pulses.

There is no hard evidence to indicate where the disrupted kimberlites or cumulates crystallized. Some form of relatively deep, but intermediate level magma chamber is a possibility. However, in view of the previous discussion of olivine compositions (section 7.7.2) it is suggested that these aggregates may represent batch melt fractions which partly or wholly crystallized in the upper mantle, i.e. before they were incorporated into subsequent kimberlite magmas and transported rapidly to surface. Possibly of significance in terms of this interpretation is the generally small size of the aggregates. Mechanical (±chemical) disruption may have been facilitated by the extreme transport distance and the major changes in physical conditions that would occur.

7.8 GEOCHEMISTRY

In this section 53 new whole rock analyses of kimberlites from the KIMFIK pipes are presented. In addition to major element results trace element data for 43 of the analyses are included (Table 7.7).

Major and trace element compositions were determined using X-ray fluorescence and gas chromatography techniques. All analytical data were determined by the Kimberlite Research Unit of the Department of Geochemistry of the University of Cape Town under the direction of Dr. J.J. Gurney. The XRF operating conditions and data reduction processes applicable when the analyses were carried out are given by Willis et al. (1971, 1972) and Gurney and Ebrahim (1973).

It has been pointed out by Dawson (1980) that because of the hybrid and complex nature of kimberlites, whole rock analyses will commonly reflect considerable differences from the probable bulk compositions that pertain prior to the final stages of emplacement. To a degree this conclusion is applicable to all igneous rocks but the problem is relatively severe in the case of kimberlites because these rocks:

(i) May have been extensively contaminated by crustal xenoliths.

(ii) May have interacted with abundant groundwater.

(iii) Have often lost abundant volatiles and associated mobile elements.

(iv) Are prone to extensive weathering with concomitant leaching of original elements and/or the addition of others.

All four of the above factors are most applicable in respect of diatreme-facies kimberlites and Dawson (op. cit.) stated that these effects can be minimized by analysing either hypabyssal-facies dykes and sills or the fine-grained selvedge of 'nucleated autoliths', i.e. pelletal lapilli or globular segregations (sections 7.2 and 7.4).

In so far as pelletal lapilli and globular segregations are concerned there is an implication that the fine-grained mantles of these bodies can be equated (for analytical purposes) with chilled margins. However, rather than representing situations where representative portions of original magmas are preserved by rapid chilling the fine-grained mantles of pelletal lapilli and globular segregations represent devolatilized magma fractions. They are not representative of original magmas because they are segregated from the often abundant volatile components and mobile elements associated with the latter.

Dawson (op. cit.) is critical of "simplistic" attempts to purify analytical material by breaking out visible xenoliths because considerable finely fragmented (on microscopic or submicroscopic scales) material is often present. Analyses of 7 KIMFIK TKB's (after attempts to remove xenolithic material) have verified this conclusion in respect of diatreme-facies kimberlites (section 7.8.1) but the technique can often be successfully used to obtain meaningful analyses of many hypabyssal-facies kimberlite breccias because:

(i) Far fewer xenoliths are usually present.

(ii) The xenoliths display smaller size ranges. Microscopic fragments are often rare and may be absent.

7.8.1 Assessment of the degree of crustal xenolith contamination

Since crustal xenolith contamination can strongly influence bulk compositions attempts have been made to derive empirically a measure of the degree to which contamination has occurred. Ilupin and Lutz (1971) considered that Si/Mg and Mg/Fe ratios could be used for this purpose and Fesq et al. (1975) stated that a Si/Mg ratio in excess of 1,2 would indicate excessive contamination.

In this thesis a slightly more complex contamination index is used to evaluate the extent of crustal contamination and, to a considerable degree, the extent to which weathering has occurred. This index is calculated as follows:

$$C.I. = \frac{SiO_2 + Al_2O_3 + Na_2O}{MgO + 2K_2O}$$

Application of this formula to analyses of hypabyssal-facies kimberlites which detailed petrographic examination reveals to be optically uncontaminated (and unweathered) results in contamination indices of around unity. Given the mineralogical and concomitant geochemical diversity of kimberlites some variation around this figure is, however, to be expected. The highest C.I.'s of apparently 'clean' kimberlites are obtained from phlogopite and/or diopside-rich and volatile-poor varieties. In such cases the contamination indices may reach 1,5 although they are usually less than 1,3.

The xenoliths which may most profoundly affect the bulk compositions of the KIMFIK intrusions are Karoo shales, dolerites and basalts, Ventersdorp andesites and quartz porphyries and a variety of acid basement rocks. These contaminants are all much richer in silica, alumina and soda than kimberlites and commonly contain much less magnesia. Consequently they have high 'contamination indices'. This point can be tested by considering random analyses taken from the literature. Thus an 'average' continental tholeiite (Manson, 1967) has a C.I. of 9, an 'average' andesite (Irvine and Baragar, 1971) has a C.I. of 14 and a biotite-gneiss from the VanRhynsdorp area (Kroner, 1968) has a C.I.

of 7. It is evident from these figures that the incorporation of relatively limited amounts of such contaminants will have pronounced effects on the C.I.'s of kimberlites.

High C.I.'s will generally also be obtained if weathered kimberlites are analysed. Commonly weathering involves enrichment of SiO_2, Al_2O_3 or Na_2O either directly or relatively by leaching of elements such as Mg and K.

Contamination indices for the 53 new whole rock analyses presented in this thesis are given in Table 7.7. Among these, analyses 12, 14, 22, 27, 31, 35 and 40 are of TKB's and have high C.I.'s (close to or in excess of 2). Of the remaining analyses only 6 (Numbers 7, 13, 34, 44, 49 and 50) have C.I.'s in excess of 1,30. Analysis 7 is of a highly micaceous kimberlite (D1) in the Dutoitspan pipe and contamination, if present, is considered to be very limited. Similarly Analysis 34 is of a kimberlite rich in phlogopite, diopside and primary serpentine (the F2 intrusion in the Finsch pipe) and contamination if present is again considered to be slight. Analyses 13 and 44 are of kimberlite breccias and in these two cases the relatively high C.I.'s are probably due to minor contamination despite attempts to remove crustal xenoliths. The levels of contamination are not, however, high and in the case of Analysis 13 this conclusion is supported by trace element concentration levels which are generally within the ranges displayed by the analyses of kimberlites with lower (1-1,3) C.I.'s. It should be noted that in several other instances the removal of xenoliths has resulted in apparently uncontaminated samples being analysed (e.g. Analyses 1, 2, 3, 25 and 26). The two remaining analyses with relatively high C.I.'s (Nos. 49 and 50) represent samples taken from very thin (<1,0m) kimberlite sills within 40m of surface. These rocks are considered to have been somewhat weathered.

In the light of the sensitivity of the contamination index to relatively acid contaminants it is interesting to compare various 'average' kimberlites (see Table 7.8). Column 1 (Table 7.8) shows the average of 14 kimberlite precursor and internal dykes associated with the Kimberley pipes (individual analyses are given in Table 7.7). Column 2 shows the average composition of 5 precursor dykes. In Column 3 the average composition of 16 analyses of kimberlite and kimberlite breccias from the Kimberley pipes is given and a combined average for the 14 Kimberley dykes and 16 pipe intrusions is given in Column 4. Careful selection, cleaning and petrographic examination of all these samples was carried out to ensure the 'cleanest' possible material for analysis and attention is drawn to the low C.I.'s calculated for Columns 1-4(Table 7.8).

In contrast to the above, the average composition quoted by Gurney and Ebrahim (1973) for 80 analyses of South African kimberlites has a C.I. of 2,05 (Column 5). This high figure suggests that many of the kimberlites analysed were badly contaminated and/or weathered. In view of the relatively high contamination index similar conclusions are drawn in respect of Gurney and Ebrahim's 'average Lesotho kimberlite' (Column 6). Dawson's 'average basaltic kimberlite' (Dawson, 1980) may also be based on some 'impure' analyses as the contamination index is slightly high (Column 7) and is, unexpectedly, higher than that obtained from this author's 'average micaceous kimberlite' (Column 8). In contrast the averages quoted by Ilupin and Lutz (1971) and Scott (1979) give C.I.'s closer to unity (Columns 9 and 16 respectively). Columns 10-15 are examples of the generally high contamination indices obtained from badly contaminated diatreme-facies kimberlites (TKB's in these examples).

7.8.2 Major element chemistry

The geochemical character of the KIMFIK kimberlites and the compositional variations which these rocks display are shown in Tables 7.7, 7.8 and 7.9 and in Figure 7.9. Several important features are evident:

(i) The highly ultrabasic character of the kimberlites which have an average SiO_2 value of ~30 wt.%. Compared with an 'average' ultrabasic rock (Vinogradov, 1962; quoted by Dawson, 1980) these kimberlites are enriched in TiO_2, Al_2O_3, CaO, K_2O, P_2O_5, CO_2 and H_2O^+. They generally contain less SiO_2, FeO^T*, MgO and Na_2O.

(ii) Wide ranges in the abundances of major oxides are evident within the KIMFIK kimberlites as a whole, within individual groups, e.g. the Kimberley occurrences, within specific types of geographically associated intrusions, e.g. Kimberley dykes and between intrusions which are located within a single pipe, e.g. Dutoitspan.

(iii) The extremely high volatile content of the kimberlites (CO_2 and H_2O^+ contents in excess of 10 wt.% occur). On average the volatile contents of kimberlite sills and dykes are higher than those of major pipe intrusion because the latter have been more extensively degassed during emplacement (Chapter 10).

(iv) Relatively few analyses of Finsch intrusions have been carried out but it is apparent that the kimberlites within this pipe generally differ in respect of chemical composition from the Kimberley group. The former are relatively enriched in SiO_2, Al_2O_3 and K_2O and are depleted in TiO_2, CaO and volatiles. However, highly evolved volatile-rich minor intrusions which geochemically resemble similar bodies in the Kimberley pipes also occur. Analysis 5 (Table 7.7) of an internal dyke in the Finsch pipe is an example.

(v) The generally high K_2O/Na_2O ratios of the kimberlites are a noticeable feature.

Figure 7.10 is a ternary discrimination diagram used by Dawson (1967b) to distinguish between kimberlites and other ultrabasic/ultramafic rocks. It is evident from this diagram that:

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*FeO^T= Total iron expressed as FeO.

(i) There is little overlap between the field defined by 'clean' KIMFIK analyses and other potassic magnesian rocks.

(ii) There is a considerable degree of separation between Kimberley pipe and dyke intrusions, due primarily to the higher volatile contents of the latter. Inclusion of Wesselton sill analyses (Analyses 50-53, Table 7.7) with the dykes would enhance the separation.

(iii) There is a fair degree of coincidence between the field of kimberlites given by Scott (1977) and that defined by the KIMFIK analyses. The latter field does, however, extend somewhat further towards the volatile end of the ternary.

Generally good correlations are apparent, at least on a qualitative basis, between the abundances of major oxides and the mineralogy of the rocks, particularly in relation to groundmass mineralogy. For example:

(i) High SiO_2 is associated with abundant groundmass diopside, cf. Analyses 34 (Table 7.7) and FINSCH/2 (Table 7.4) of the F2 intrusion. Lower but still relatively high SiO_2 frequently reflects the presence of abundant phlogopite, e.g. the D1 and D5 intrusions in the Dutoitspan pipe are both phlogopite-rich and correspondingly contain considerable SiO_2 (Analyses 7 and 44 respectively, Table 7.7).

(ii) Depression of the SiO_2 content below $\sim$30 vol.% is generally coincident with increased volatiles and CaO. Mineralogically this situation is reflected by the presence of abundant calcite (and serpentine in some instances), cf. Analyses 5, 28 and 36 (Table 7.7) with Analyses DB/11, WESS/4 and FINSCH/7 (Table 7.4), respectively.

(iii) TiO_2 is held primarily in ilmenite and perovskite. A good example is provided by a dyke in the Finsch pipe. This dyke has a high content of fine-grained groumdmass opaque oxides (much of which is probably ilmenite) and perovskite (Analysis FINSCH/7, Table 7.4). The high TiO_2 content coincides with high CO_2 and H_2O^+ levels, an association which indicates the relatively evolved nature of the magma (probably a volumetrically limited residuum emplaced after major episodes of pipe formation and infilling). A similar kimberlite at Bultfontein contains considerable TiO_2 (Analysis 41, Table 7.7) but less than the previous example. The difference is probably due to less ilmenite being present in the groundmass opaque oxide suite. Many of the KIMFIK intrusions contain phlogopite which can incorporate considerable titanium. However, the amount of TiO_2 held in phlogopite appears to be relatively limited as no association between high TiO_2 and high phlogopite levels is apparent. For example, the mica-rich kimberlites in the Finsch pipe all contain less than 1,0 wt.% TiO_2 (Analyses 32-34 and 37-38, Table 7.7).

(iv) Iron does not display major variations within the KIMFIK intrusions but higher values are generally coincident with low SiO_2 and abundant volatiles. Analysis 36 (Table 7.7) is an example of a SiO_2-deficient, FeO^T and calcite-rich kimberlite which contains abundant groundmass opaque oxides. A modal analysis of this rock is given in Table 7.4 (Analysis FINSCH/7).

(v) High MgO appears primarily to reflect abundant olivine, e.g. the D2 kimberlite (Analyses DTP/3, 4, 5 and 15, Table 7.4 and Analyses 9, 10 and 45, Table 7.7). A second example is the DB 2 intrusion (Analyses DB/1-4, Table 7.4 and Analysis 2, Table 7.7).

(vi) There is a very strong positive correlation between CaO and calcite and between CO_2 and L.O.I. and calcite (cf. Analyses 28, Table 7.7 and WESS/4, Table 7.4). The foregoing example represents a dyke in the Wesselton pipe and a similar example occurs at Bultfontein (cf. Analyses 41, Table 7.7 and BULT/1,

Table 7.4).

(vii) Moderately high CaO levels (without accompanying high CO_2 and L.O.I. values) are displayed by monticellite-rich kimberlites (e.g. the DB 2 kimberlite, Analysis 2, Table 7.7 and Analyses DB/1-4, Table 7.4). None of the KIMFIK intrusions that were analysed contain sufficient groundmass diopside to have produced significantly high bulk CaO values. The most diopside-rich kimberlites occur in the Finsch pipe but diopside-bearing intrusions such as F2 in fact contain relatively little CaO. Evidently an enhancement of CaO due to the occurrence of diopside is offset by the paucity of calcite in the rocks. In addition to the minerals mentioned above some CaO in the KIMFIK kimberlites is held in perovskite and apatite. These minerals are important and consistent accessory components of the intrusions but are rarely present in sufficient abundance to markedly affect bulk CaO levels.

(viii) P_2O_5 is concentrated in apatite in the KIMFIK kimberlites but correlations between high apatite and high P_2O_5 are poor in some instances, e.g. in the DB 3 kimberlite (cf. Analyses DB/5, Table 7.4 and 1, Table 7.7). Analysis 1 indicates that DB 3 has only slightly more P_2O_5 than DB 2 (Analysis 2, Table 7.7) but modal analysis indicate much more apatite in DB 3 than DB 2 (cf. Analyses DB 1-5, Table 7.4). It is possible that very fine-grained apatite was not recognized during modal analysis of DB 2.

(ix) None of the primary minerals in the KIMFIK intrusions is a major host of Na hence soda levels are low (except where crustal contaminants are present). Phlogopite is the dominant host mineral for K_2O, e.g. note the relatively high K_2O contents of the phlogopite-rich D1, F2 and D5 intrusions (Analyses 7, 34 and 44 respectively in Table 7.7).

7.8.3 Trace element chemistry

The concentrations of various trace elements in KIMFIK kimberlites are presented in Tables 7.7 and 7.8 and are indicated in the form of histograms in Fig. 7.11. In general the analytical results reinforce Dawson's conclusion (Dawson, 1962, 1967b and 1980) that the trace elements in kimberlites can be divided into two major groups, i.e. those which occur in amounts characteristic of ultramafic rocks (i.e. the compatible trace elements) and those that are enriched relative to other ultramafic rocks (i.e. the incompatible trace elements). Comparison of the averages quoted in Column 4 of Table 7.8 (analyses of KIMFIK pipe and dyke intrusions) with the trace element content of an average ultramafic rock quoted by Dawson (1980) indicates that:

(i) The Kimberley kimberlites contain considerably higher concentrations of Nb (X130), Zr (X20), Y (X6), Rb (X55), Sr (X52) and Ba (X46). The figures in brackets indicate the average enrichment factors.

(ii) Zn, Cu, Ni, Co, Cr and V occur in the kimberlites in concentrations which are comparable with the average ultramafic rock quoted by Dawson (op. cit.).

7.8.4 Genetic considerations

Simple Harker-type variation diagrams using MgO wt.% as a differentiation index are presented in Fig. 7.12. These plots are restricted to analyses of kimberlites from the Kimberley group of intrusions which, because of geographic proximity and similar ages of intrusion (Allsopp and Barret, 1975) are most likely to be consanquineous. It is evident from the variation diagrams that with increasing MgO the kimberlites show:

(i) A poorly defined increase in SiO_2.

(ii) Reasonably well defined decreases in Al_2O_3, TiO_2, CaO, CO_2 and L.O.I.

(iii) Poorly defined decreases in K_2O and Na_2O.

(iv) Scatter for P_2O_5, FeO and Fe_2O_3.

Scott (1977) noted that Harker-type variation diagrams are rarely used for petrogenetic interpretations of kimberlites and related rocks, mainly because of the high content of carbonate, volatiles and xenoliths in these rocks. Much of the scatter in the variation diagrams of Figure 7.12 may thus be related to high and variable carbonate (plus associated volatiles) levels or to xenolithic contamination. However, since only analyses with low contamination indices have been plotted the contamination effects are likely to be limited.

Scott (op. cit.) attempted to overcome the carbonate/volatiles problem by removing H_2O^+ and CO_2, the latter as $Ca(Mg)CO_3$, and recalculating her analyses to 100%. This approach was justified in part on the assumption that kimberlites develop immiscible carbonatitic (volatile-rich) and silicate fractions which evolve separately thereafter. However, as indicated in section 7.3.4 serious arguments against carbonate-silicate immiscibility can be raised.

Various authors (e.g. Scott, op. cit.; Dawson, 1980) have noted that the amount of carbonate and commonly associated minerals (mainly primary serpentine and apatite) in kimberlites is strongly influenced by the degree to which volatile components have been retained during emplacement. Clearly random losses of volatiles during emplacement may strongly influence the abundance of other components. Such losses are likely to be most pronounced with respect to major pipe intrusions since these intrusions are likely to be relatively open systems during emplacement compared with kimberlite dykes and sills. Accordingly some reduction in the scatter of the variation diagrams would be expected if the data for sills and dykes only are plotted. This has been done in Figure 7.13 and it is apparent that improved trends are obtained for most major oxides. Figure 7.13 shows that with increasing MgO there are:

(i) Well defined decreases in TiO_2, Al_2O_3, CaO, K_2O, CO_2 and L.O.I.

(ii) A well defined increase in SiO_2 and a poorly defined increase in FeO.

(iii) A poorly defined decrease in Fe_2O_3.

(iv) Doubtful, very vague, decreases in P_2O_5 and Na_2O. In the case of P_2O_5 considerable scatter remains and the generally low levels of Na_2O make it difficult to decide whether any real trend occurs.

Trace element variation diagrams are presented in Figures 7.14 (Kimberley dyke and pipe intrusions) and 7.15 (dykes only). No data for the Wesselton sills are available. In general the results are similar but better trends are defined in the plots of dyke concentration because, as in the case of major elements, the scatter of data points is reduced. It is apparent from Figure 7.15 that:

(i) There are strong positive correlations between MgO and Ni and Co.

(ii) There are moderately to very well defined negative correlations between MgO and Zn, Cu, Zr, Ba, Sr, V, Rb and Nb.

(iii) There is a poorly defined negative correlation between Cr and MgO.

The generally coherent trends revealed by the variation diagrams suggest a close genetic association between the Kimberley intrusions. Such an association would be entirely consistent with the geographic and temporal associations of these kimberlites and some possible genetic relationships are discussed below. It is emphasized that the possible relationships are considered in general terms only and that much of the argument put forward is speculative.

Various authors (e.g. Scott, 1977; Danchin et al., 1975 and Ferguson et al., 1975) have attributed the bulk composition variations of kimberlite suites to olivine fractionation from parental magmas. For at least two reasons this proposal also merits consideration in respect of the Kimberley intrusions:

(i) Olivine is the major mineral which is universally present in the kimberlites.

(ii) The major element and trace element variation trends are consistent with the view that olivine is the dominant controlling mineral during crystal-liquid fractionation.

Although inconsistent with the batch mixing hypothesis proposed in section 7.7.2 the simplest explanation of the trends displayed by the variation diagrams is that they reflect continuous magma evolution, either by single stage fractional crystallization or progressive partial melting (in which olivine plays the major role). Some scatter in the data points and variation in individual oxide trends (Figs. 7.12-7.13) could be ascribed to contamination and/or alteration and to the inherent difficulties of handling the bulk compositional data of such complex (carbonate/volatile-rich) rocks. Such simple genetic models are, however, entirely negated by the fact that the implied liquid line of descent requires a continuous trend of plotted composition points in relation to time. The examples in figure 7.16 show that such trends do not occur. It is evident that:

(i) The different pipes were not infilled in a sequence which can be correlated with single stage crystal fractionation or progressive partial melting (Figs. 7.12-7.15).

(ii) Discrete intrusions within individual pipes do not show simple correspondence between degree of chemical evolution and time (Fig. 7.16).

Since no single crystal fractionation or partial melting model can account for the observed distributions of compositional data the reasonable coherency of the geochemical trends must be explained in some other way. The most obvious solution is to suppose that repetitive, closely reproducible, fractionation processes of some type occurred. In attempting to postulate such a solution it is important to note:

(i) That in addition to crystallizing primary olivine (as phenocrysts) the kimberlites contain abundant upper mantle-derived (peridotitic) olivine xenocrysts (section 6.2).

(ii) That, contrary to expectations (assuming a garnet lherzolite/harzburgite upper mantle), the kimberlites contain very few or no orthopyroxene xenocrysts (section 6.2.1).

(iii) That although some scatter in trends is evident inspection of the major element variation diagrams suggest evolutionary control by an extract/additive mix which approximates 80 wt.% olivine and 20 wt.% orthopyroxene (e.g Fig. 7.13). This 80:20 olivine/orthopyroxene mix is based on the assumed 'average' upper mantle olivine and orthopyroxene compositions that have been plotted in Figures 7.12-7.15.

(iv) That within individual pipes reversals in trends related to time occur (Fig. 7.16), i.e. there is an apparent implication that in some instances (within individual pipes) discrete intrusions are related by crystal fractionation while other intrusions (in the same pipe) may be related by increasing degrees of partial melting.

If, as the variation diagrams suggest (and ignoring for the moment the possibility of batch-mixing and early phenocryst crystallization), magma evolution is controlled primarily by some form of olivine plus minor orthopyroxene fractionation the possibility is raised of control merely by the addition or subtraction of xenocrysts. The presence of variable (generally abundant) quantities of olivine xenocrysts *in* kimberlites has already been referred to (section 6.2.1) and the rarity of orthopyroxene xenocrysts has been ascribed to extensive resorption (section 6.2.1). Since orthopyroxene xenocrysts are extremely rare in the KIMFIK intrusions it can be argued that resorption is initiated at an early stage of magma evolution (probably in the upper mantle) and is virtually complete at the time of final emplacement.

A major problem with respect to simple xenocryst control of bulk composition variation is the nature of the indicated control line. As noted above this control line suggests the extraction or addition of an $\sim$80 wt.% olivine and 20 wt.% orthopyroxene mix. If, in fact, magma composition variations relate solely to simple xenocryst additions or subtractions an additive/extract mix of approximately 65 wt.% olivine and 30 wt.% orthopyroxene (and 5 wt.% other xenocrysts, e.g. garnet, clinopyroxene) would be expected; the approximate proportions of the two minerals in 'average' upper mantle 'peridotite'. Simple xenocryst control of bulk compositions therefore appears to be extremely unlikely.

The observed 80:20 ol-opx ratio implies enrichment of olivine relative to orthopyroxene and in this respect early crystallized olivine phenocrysts may be important (see discussion below). Importantly the 80:20 mix also implies that magma evolution cannot be ascribed to crystal fractionation involving subtraction of olivine while orthopyroxene is resorbed. Such a process would shift the control line towards the orthopyroxene side of the 65:30 (ol:opx) mantle ratio. For this reason it is concluded that the magmatic evolution of the Kimberley intrusions reflects an additive (enrichment) process rather than one involving mineral subtraction (i.e. crystal settling mainly involving olivine).

It might be possible to devise a model whereby the necessary olivine enrichment could be related to increasing amounts of olivine xenocrysts only. However, olivine phenocrysts are probably also involved as they are considered to have crystallized at an early stage of kimberlite genesis (within the upper mantle, see section 7.7.1).

Thus far in the discussion the possibility of batch-mixing of crystals and liquids (as suggested in section 7.7.2) has been ignored but in attempting to suggest a possible mechanism for the enrichment of olivine (xenocrysts and/or phenocrysts) some previous suggestions and comments in this respect need to be taken into account. In particular attention must be paid to the occurrence, in discrete intrusions, of essentially unzoned olivine phenocrysts and xenocrysts which display considerable compositional ranges (sections 6.2.2 and 7.7.2). Note must also be taken of the kimberlite-free, root-zone contact breccias associated with most of the Kimberley pipes (section 3.3.1). These contact breccias indicate that at least some intrusions have been extensively degassed; to the extent that in some cases volatiles have separated entirely from the non-volatile fractions of the original liquids (Chapter 10). In other instances it follows that while complete separation did not occur considerable vertical differentiation developed, leading to the formation of volatile-rich tops and early crystal-enriched bases of intrusions.

Varying degrees of olivine enrichment could be ascribed to devolatilization. Such degassing will, however, proportionately increase both olivine and orthopyroxene (although the latter may be resorbed) if only mantle-derived xenocrysts are present. Under such conditions the 'average mantle' control line should still apply (65:30 ol:opx), provided that resorbed orthopyroxene plus any remnant crystals are retained in the 'involatile' liquid plus crystals fraction. If the latter situation pertains, enrichment in olivine by devolatilization can only occur if, in addition to mantle xenocrysts, olivine phenocrysts are present. Evidence favouring the early precipitation of olivine phenocrysts was put forward in section 7.7.2.

The presence of mixed populations of olivine phenocrysts (section 7.7.2) suggests that simple enrichment in olivine (phenocrysts plus xenocrysts) by devolatilization (as proposed above) is unlikely and that olivine enrichment must reflect more complex situations. It is therefore suggested that the bulk composition variations displayed by Kimberley group intrusions may be due to:

(i) Mixing of crystal/magma batches which have probably been subjected to varying degrees of olivine and orthopyroxene enrichment by the addition of 'average' upper mantle before and/or after mixing.

(ii) Further olivine enrichment (involving xenocrysts and phenocrysts) by degassing which might also occur before and/or after mixing.

The foregoing genetic model requires the olivine phenocrysts to have crystallized before batch-mixing.

The lack of correlation between bulk composition and time shown by the variation diagrams (e.g. Fig. 7.16) is easily explained in terms of the foregoing generalized model. Since the bulk compositions of successive intrusions will be governed primarily by the amount of incorporated 'mantle' and the degree of olivine enrichment (by degassing and mixing) a constant 'one-way' trend in chemical evolution would not be anticipated. In fact such a uni-directional trend would be highly unlikely as, assuming similar starting liquids, it would imply constant or exponential increases (or decreases) in the amount of incorporated 'mantle' and/or constant or exponential increases (or decreases) in the degree of olivine enrichment by degassing in batch-mixed intrusions.

A fair degree of scatter of data points in the variation diagrams would also be consistent with the foregoing, admittedly highly speculative, model. Since magma mixing and olivine enrichment could be 'random' in respect of the number, nature, volumetric proportions and degree of evolution of the mixed magma batches, scatter would be expected. In fact the apparent $\sim$80:20 ol:opx control line may not be real. It may merely be an artefact of the constraints imposed by 'mantle' and 'olivine only' control lines, i.e. it may be a 'pseudo-control line' reflecting varying combinations of 'mantle' and olivine phenocrysts. The data points would be expected to be constrained within the 65:30 'mantle' control line and a pure olivine control line.

Doubts regarding the validity of the 80:20 ol:opx control line are reinforced by the fact that some of the major oxide (e.g. TiO_2 and K_2O) and trace element (e.g. Ni and Nb) trends are 'off line'. Such deviations may reflect mixing situations (see previous discussion) or fractionation processes involving minerals such as phlogopite and ilmenite in addition to olivine and orthopyroxene. Alternatively these deviations may not be real but may relate to the inherent difficulties of handling the bulk compositional data of such complex rocks as kimberlites.

CHAPTER 8

DIAMONDS IN THE KIMFIK PIPES

8.1 INTRODUCTION

Production from the KIMFIK pipes in 1980 (the latest year for which figures were available at the time of writing) totalled in excess of 4,5 million carats (Table 8.1). Consequently a dissertation dealing with the geology of the pipes would be incomplete without reference to the diamonds which they contain. Diamonds have not, however, been examined during the course of this study and the discussion which follows is restricted to some aspects of the distribution of diamonds in the kimberlites and to brief comments concerning the nature of diamonds from different sources. More detailed comments on the nature of the diamonds in the KIMFIK pipes can be found in publications by Wagner (1914), Williams (1932), Harris et al. (1975, 1979), Robinson (1980) and Clement et al. (in press).

8.2 DISTRIBUTION OF DIAMONDS IN THE PIPES

Some features of the distribution of diamonds in the pipes are illustrated by the examples from individual pipes which are given below. The De Beers pipe is a particularly instructive example (see also Clement et al., in press).

The De Beers pipe was discovered in 1871 (Williams, 1948) and exploited until 1908 when operations ceased due to a recession in the diamond industry. The mine was subsequently re-opened in 1963 and is still operating. Production figures for the two periods are given in Table 8.2. It should be noted, however, that subsequent to the re-opening of the mine in 1963 ore from this source has been mixed with that from other operating Kimberley mines (Bultfontein, Dutoitspan and Wesselton) and treated in a central 'pool' recovery plant. Diamond yields and grades quoted in Tables 8.1 and 8.2 are therefore reconciliations based on the results of:

(i) run of mine bulk samples (treated in a separate bulk test plant).

(ii) bulk samples of specific mining blocks (treated in the same bulk test plant).

(iii) development samples taken to prove additional ore reserves in the pipes (treated in a separate development test plant).

Although the quoted figures are unlikely to be grossly in error it should be stressed that they do not reflect physically measured diamond yields but are portions of total production allocated to individual mines.

Similarly, during part of the initial stages of mining of the De Beers pipe, ore from this occurrence and the Kimberley mine was blended before treatment. The overall production figures quoted in Table 8.2 for production up to 1908 have been taken from an internal unpublished De Beers report by R.A. McCallum (1969) and the original sources have not been verified. The high yield reported is supported by De Beers Annual Reports for some of these years. All the other figures quoted in Tables 8.1 and 8.2 are taken from more recent Annual Reports.

The average grade at any particular level in the De Beers pipe is strongly dependant (amongst other factors) on the relative areas occupied by each of the five most extensive kimberlites, i.e. the DB1, DB2, DB3, DB5 and DB6 intrusions (cf. the geological plans shown in Figure 3.34). These kimberlites differ markedly with respect to diamond content as is indicated by the sampling results (averages for each kimberlite are given in Table 8.3). A sharp contrast in grade between neighbouring kimberlites is particularly well illustrated by the individual sample results obtained from the DB2 and DB3 kimberlites on the 720m level in the southeastern part of the pipe (Fig. 8.1).

A notable feature of Table 8.2 is the very high average yield obtained during the early years of mining of the De Beers pipe; considerably more than double the grade obtained during any subsequent year of production. Part of the difference can be ascribed to 'high-grading' since the relatively low grade (DB5) northwestern part of the mine was left largely unmined up to 1908 whereas much of more recent production has come from this area.

'High-grading' does not, however, entirely explain the high early (up to 1908) yields. A contributing factor is that (as indicated in Figures 3.34 and 3.35) the area of the DB 3 kimberlite (by far the richest of the intrusions) increases rapidly, relative to the other intrusion, as the surface is approached. Near surface DB3 occupies the whole of the central and southeastern part of the pipe and it must have been effectively the only intrusion mined during pre-1908 exploitation of the ore body.

Based on the grades quoted in Table 8.3 it is apparent that the combined effects of 'high-grading' and the major increase in the area of DB3 upwards, relative to the other intrusions, also fail to account for the high production grades of the early mining period. These results are further explained by the fact that the grade of the DB3 intrusion decreases with depth. Thus in 1890, shortly after underground operations had started, G.F. Williams (the then General Manager) stated that the average yield of the mine (effectively the DB3 kimberlite) was equivalent to between 165 and 183 carats/100 tons (Annual Report, De Beers Consolidated Mines Ltd.). As indicated in Table 8.3 the recoverable grade of DB3 on the 595m level is 61 carats/100 tons while on the 720m level a grade of 39 carats/100 tons was obtained from development sampling. Based on sampling results and production records L. Kleinjan (personal communication) has calculated that the decrease in grade is essentially linear at a rate of 18 carats/100 tons/100 vertical metres.

Wagner (1914) suggested that decreases in grade with depth in the Kimberley and De Beers pipes could be correlated with a transition from kimberlite tuff to kimberlite (i.e. from diatreme-facies to hypabyssal-facies kimberlite). He suggested that diamonds might be concentrated in 'kimberlite tuffs' by the "...elimination, probably in the form of volcanic dust, of the less resistant constituents of kimberlite...".

In the light of Wagner's theory it is tempting to conclude that the increase in grade upwards within the DB3 kimberlite can be equated with a transition from hypabyssal- to diatreme-facies kimberlite. The nature of the DB 3 kimberlite near surface is not known as this part of the pipe was mined out to almost 600m before the mine was re-opened in the early 1960's. However, a transition such as that postulated is consistent with the author's views on pipe formation and infilling (Chapter 10) and such a transition (within the DB5 intrusion) appears definitely to have occurred in the northwestern part of the pipe (section 10.3.1). Wagner's theory is, however, largely negated by evidence from other intrusions including DB5. The latter intrusion displays no noticeable vertical grade variation across the inferred transition zone between TKB and kimberlite breccia. Sampling of this intrusion on the 560m level (where the kimberlite is hypabyssal in character) produced a grade of 15 carats/100 tons while numerous bulk samples and earlier development sampling at higher levels indicated grades of 15-18 carats/100 tons. An 1890 result of 16,7 carats/100 tons was quoted in the De Beers Annual Report for that year. The last-mentioned figure must represent a surface or near-surface grade as underground operations had only just begun in 1890. The kimberlite (as indicated by Wagner, 1914) occurred as a typical TKB close to surface.

Apart from the effects of dilution by 'floating reef' TKB's generally have more consistent grades (on a mining scale) than hypabyssal-facies kimberlites, over considerable vertical distances. For example, the grade of the D11 TKB at Dutoitspan is consistently low, 5-8 carats/

100 tons, over a vertical distance of 870 metres. Similarly the grade obtained from the F1 intrusion at Finsch on the 348 metre underground sampling level matches (allowing for dilution effects and for discrepancies between production and test plant efficiencies) the grades recorded in the open pit between surface and 200 metres. The KOF1 intrusion is a further example of a TKB with a reasonably consistent grade (~10 carats/100 tons between 100 and 500m in depth). Similarly, relatively consistent grades, over substantial vertical intervals, have been recorded in TKB's in the Premier and Letseng pipes.

The most likely explanation for the relative consistency of grade in TKB's is that considerable mixing of components occurred during fluidization (Chapter 10). Reynolds (1954) has pointed out that fluidization can provide ideal conditions for the mixing of solid particles.

The consistency of grades within TKB's is further evidence against Wagner's 'winnowing out' theory. Such winnowing should not only result in an overall enrichment in the diamond content of TKB's as a whole but should result in upward enrichment <u>within</u> TKB's. The latter situation, an expected reflection of winnowing being increasingly effective upwards, is not shown by the examples noted above.

Two other arguments can be raised against the 'winnowing out' theory. In the first instance there is no evidence that TKB's characteristically contain greater abundances of other heavy components (e.g. ilmenite macrocrysts or ultramafic nodules) than do hypabyssal-facies kimberlites. Such situations would be the expected norm if extensive winnowing out of small, light, components had occurred. Secondly, a fundamental aspect of the TKB's in the KIMFIK pipes is that they have incorporated as much as 50 vol.% country rock material (section 10.3.8). Since the relative abundances of the kimberlitic components may differ slightly but are not markedly atypical of the relative abundances of the same components in hypabyssal-facies kimberlites, a dilution rather than enhancement of diamond content might be expected in TKB's.

The emphasis in the foregoing comments on the relative consistency of the diamond content of TKB's is not meant to imply that more or less constant grades never occur in hypabyssal-facies intrusions; only that the frequency with which the latter situation occurs is lower. Reference to the constant grade of the lower (hypabyssal-facies) part of the DB5 intrusion has already been made and similarly the DB2 intrusion in the De Beers pipe has a consistent grade (∿3-5 carats/100 tons) wherever it has been sampled. There is too little information available to draw any firm conclusions regarding grade trends within other De Beers pipe intrusions.

Variations in grade between different intrusions is a common feature of the KIMFIK pipes although the differences are not always as spectacular as the contrast between the DB3 and DB2 intrusions in the De Beers pipe. Examples from the Dutoitspan and Wesselton occurrences are illustrated in Figures 8.2 and 8.3. The D5/D2 relationships in the Dutoitspan pipe (Figure 8.2) mirror the DB3/DB2 situation in the De Beers pipe. The grade of D2 is consistently low whereas the diamond content of D5 is relatively high and increases upwards. DB3 and D5 are, however, the only intrusions where this linear or more-or-less linear grade trend has been reliably established.

The grade figures quoted in the foregoing discussion are averages for pipes or for specific intrusions at particular levels. On a smaller scale the diamond content of most KIMFIK intrusions varies considerably. This variation is illustrated in Figure 8.1 where the results of individual samples (each weighing ∿20 tons) from the DB3 and DB2 intrusions are shown. Individual samples within DB3 have grades varying between 2 and 134 carats/100 tons. On even smaller scales grades are still more variable as the recovery of even one moderately large stone in a sample will markedly effect the calculated yield. Variable grades probably reflect heterogeneous diamond distributions in kimberlites but, because of the rarity of diamond, the exact natures of these distributions have not been quantitatively established.

Comments relating to grade trends and variations such as those made in the foregoing discussion, should, for several reasons (see Wagner, 1914), be treated with caution. Although the conclusions drawn are considered to be broadly valid two factors in particular can adversely affect the interpretations. Firstly, recovered grades do not represent the pristine diamond content of the kimberlites being mined or sampled. Recovery grades are an artefact of recovery plant design parameters which have changed from time to time due, for example, to developments in engineering design, to the adoption of more sophisticated recovery techniques or to market demands. Secondly, variations in the physical characteristics of different intrusions (e.g. hardness, density or crushing parameters such as particle or fragment shape) will strongly influence the efficiency of recovery operations.

8.3 POSSIBLE EXPLANATIONS FOR DIFFERENT GRADES AND DISTRIBUTION PATTERNS

It could be argued that the different diamond contents of neighbouring intrusions in the pipes, or at different places within individual intrusions, reflect (at least in part) varying degrees of resorption of diamonds; either within the pipes after emplacement or at sub-pipe levels during intrusion. Such resorption explanations for variations in diamond content have, however, been refuted by Robinson (1980 and in Clement et al., in press) who has examined diamond parcels from the DB1, DB2 and DB3 intrusions at various levels in the De Beers pipe. Not only do the diamonds from different levels in the DB3 intrusion (which shows systematic vertical grade variations, section 8.2) show similar levels of resorption but resorption has similarly affected the diamonds from different levels in all three intrusions. More recently Robinson (personal communication) has found no material differences in resorption characteristics between the diamonds from a number of different intrusions in the Dutoitspan pipe, although the diamond content of these kimberlites varies substantially.

It should be noted that the De Beers and Dutoitspan diamonds examined do display considerable degrees of resorption but Robinson (1980) has concluded that this resorption was essentially complete prior to the diamonds being emplaced within the pipes. Support for this view is provided by the fact that horizontal variations in the extent of resorption of diamond within the pipes are never apparent. If extensive very late (post-emplacement) resorption had occurred horizontal variations in the degree of resorption of diamond, related to the configuration of the pipes, would be expected. Such variations would reflect the spatially related differences in physico-chemical conditions which must pertain after emplacement (e.g. differing cooling rates related to the proximity of pipe contacts or to the configuration of the pipes). Horizontal variations in the extent of resorption have not been noted during valuation exercises involving diamonds derived from sampling operations. During these exercises the diamonds are examined in relation to their actual sample positions within the intrusion (as indicated, for example, in Figure 8.1).

Robinson (1978, 1980) has argued convincingly that most, if not all, diamonds in kimberlite are xenocrysts although others (e.g. Fesq et al., 1975; Harte et al., 1979, 1980) favour a phenocrystal origin. If a xenocrystic origin is accepted then grade variations between individual kimberlites in the KIMFIK pipes might relate essentially to differential sampling of diamondiferous mantle sources. Different intrusions might sample different mantle sources with different diamond contents or successive intrusions might sample the same sources less and less effectively; perhaps because upward movement became progressively easier due to more favourable intrusion conditions being established. Alternatively sampling might become progressively less effective as the source materials originally closest to repetitively utilized intrusion paths became depleted in diamonds. Robinson (1980) has suggested that certain differences in the proportions of diamond of different colours indicate a slight difference in derivation from the mantle between diamonds from the DB1 and DB2 kimberlites, on the one hand, and the DB3 intrusion, on the other hand.

It is interesting to speculate (in terms of a xenocrystic origin) on why some intrusions, such as DB3 and D5 display systematic decreases in grade with depth while, in other intrusions, zones with different grades appear to be randomly distributed in relation to the morphology of the intrusions.

The fact that the grade variations in DB3 and D5 are systematic suggests that these intrusions represent single pulses of magma. Batch mixing situations, after relatively rapid intrusion to surface has been initiated, are unlikely to produce systematic grade changes such as those displayed by DB3 and D5. In terms of this initial premise the grade trends of the two intrusions could be explained by either of two possible sampling models:

(i) The first model assumes that prior to, or during, the early stages of intrusion the magma extends over a considerable vertical distance. The upper part of the 'reservoir' lies within a relatively highly diamondiferous source region, or enters such a region. Thus the upper part of the magma 'column' incorporates relatively abundant diamonds while fewer diamonds are incorporated lower down in the column. When the lower grade 'tail' of the intrusion passes through the 'high grade' source zone it incorporates few additional diamonds because this zone has already been impoverished in respect of diamond content. Alternatively (as noted previously) intrusion of the tail through the 'high grade' source zone may be relatively easy (due to the intrusion along a previously used passageway) thus inhibiting the processes responsible for diamond incorporation (partial melting, mantle stoping, shearing?).

(ii) Alternatively it is not necessary to assume a source area that is stratified with respect to diamond content. The systematic grade variations of DB3 and D5 may relate only to temporal variations in sampling effectiveness as noted in (i) above.

Considered in the foregoing simplistic terms one factor mitigating against source area stratification, at least on a small or local scale, is that repetitive, similar (i.e. decrease with depth) grade distribution patterns would not be the only systematic pattern developed. In at least some instances reverse situations (i.e. grade increases with depth) might be expected. No regular increases of grade with depth in pipes or dykes are known to the author and where local increases are apparent they can be related to obvious controls such as decreases in the amount of xenolithic dilution.

It has been suggested (e.g. section 7.7.2) that some intrusion in the KIMFIK pipes reflect mixing of more than one batch of magma. It is concluded (Chapter 10) that such mixing might occur at various depths along the emplacement paths. In terms of this hypothesis it follows that such mixing may not always be complete and, in fact, complete homogenization may be a rare feature of mixed batches because mixing is inhibited by comparatively rapid rates of intrusion and because of the large amount of solid material present. Incomplete mixing of magma batches with different diamond contents offers a reasonable explanation of irregular and apparently randomly distributed grade zones within major intrusions in the KIMFIK pipes. However, if such incomplete mixing has occurred discrete areas of the intrusions would be expected to display additional petrographic contrasts; not just differences in diamond content. A possible example of such a mixed intrusion is the W3 kimberlite in the Wesselton pipe.

The W3 kimberlite occupies a substantial part of the irregular root zone of the pipe and has been sampled for diamonds on several mining or exploratory levels. These sampling operations have revealed considerable contrasts in grade across the body (Fig. 8.4). On the 930m level a low grade area in the south central part of W3 is undoubtedly due to dilution by included country rock blocks (Fig. 8.4) but elsewhere substantial grade differences cannot be related to varying degrees of dilution.

The W3 intrusion also displays substantial petrographic differences from place to place. Macroscopically the two most noticeable differences are pronounced variations in the content of phlogopite macrocrysts and peridotite and MARID-suite nodules. In some areas phlogopite macrocrysts are so abundant that as many as 10 individual grains may be evident in a single thin section. In other areas the kimberlite appears to be devoid of phlogopite macrocrysts or only rare, highly altered individuals are present. Similarly in some areas mantle-derived nodules are abundant whereas in other parts of the intrusion they are very rare in underground exposures. Heavy mineral analyses have also revealed gross differences from place to place in the content of other macrocrysts such as garnet and picroilmenite. Petrographic variation within W3 is not restricted to the coarse-grained components or to xenogenic material. The groundmass of the rock also varies considerably due to differences in the primary mineral assemblages and to contrasting styles of deuteric alteration (Appendix I). Sharp contacts between the petrographic variants are, however, never evident.

Petrographic variants such as those displayed by the W3 kimberlite need not necessarily reflect incomplete mixing of magma batches. They could also result from variations in intrusion parameters (e.g. velocity, viscosity and temperature gradients across the intrusion) and/or chemical variations due, for example, to local concentration or impoverishment of volatiles. However, in general such variations should cause zonally distributed features in the intrusion. There is no obvious evidence of zonally distributed petrographic variations in W3 that can be related to the form of the intrusion. However, the petrographic variants within W3 (as in most other KIMFIK intrusions) have not been sufficiently accurately mapped to allow close comparison with diamond variations to be made. Thus, while the incomplete batch mixing hypothesis is attractive, it must remain speculative until more detailed investigations have been carried out.

Thus far in the discussion grade variations within and between intrusions have been discussed in terms of a xenocrystic origin for the diamonds. In the previously suggested mixing model the origin of the diamonds (xenocrystic or phenocrystic) is irrelevant with respect to the explanations offered for different kinds of diamond distributions. This is not so, however, with respect to systematic grade decreases with depth such as those displayed by the DB3 and D5 intrusions. If, as seems likely, diamond growth takes place, under relatively static conditions, in the upper mantle at a depth of at least 154km (Robinson, 1980), it is difficult to understand how the distribution patterns in DB3 and D5 could occur if the diamonds are phenocrysts. On the contrary a gravitative tendency for the reverse situation to develop during and subsequent to intrusion is more likely.

During crystallization under static conditions diamonds would be expected to form cumulates at the bottom of the magma chamber (although gravity settling may not be particularly efficient because of the small sizes of many diamonds). Subsequent intrusion should result either in initial barren or low grade kimberlites (tapped from the upper parts of the chambers) or, if the intrusion was of sufficient magnitude to more-or-less empty the chamber, the resulting kimberlite would normally display increasing grade downwards. It could be argued that turbulence and/or convection during intrusion to the surface may have homogenized diamond grade within the ascending system. There does not appear, however, to be any reasonable mechanism for concentrating phenocrystal diamonds at the head of the intrusion and tailing the concentration off downwards in systematic fashion.

The suggestion made in the foregoing discussion is that the intrusion of low grade, more evolved magma would take place ahead of less evolved, more diamondiferous kimberlite. This situation is atypical of the common intrusive sequence of basic (less evolved) to acid (more evolved) differentiates in igneous complexes. By analogy with such complexes low grade kimberlites should post-date richer intrusions. However, the proposed sequence is supported to some extent by the fact that the precursor dykes and sills associated with the KIMFIK pipes are amongst the most evolved of all the intrusions. The Wesselton sills are a particularly good example (Table 7.7, Analyses 50-53).

In view of the foregoing comments there are probably only two ways in which the diamonds in DB3 and D5 might be interpreted as phenocrysts. One is to suppose that diamonds, crystallizing from a liquid in the upper mantle, were incorporated in another intrusive phase rising from below. Alternatively it could be argued that a rising, relatively diamond-rich magma pulse, was overtaken by, and incorporated in the forefront of, a more rapidly ascending, relatively diamond-poor, magma pulse. Neither alternative is favoured because in both cases delicate mixing balances would be required to produce systematic vertical variations in diamond content. It could also be argued that the systematic depth-related grade variations of DB3 and D5 reflect incomplete gravity settling after intrusion. However, such processes should produce systematic vertical variations in diamond sizes. Such variations do not exist in DB3 or D5.

The distribution of diamonds in DB3 and D5 is, in view of the foregoing arguments, considered to be more consistent with a xenocrystic rather than phenocrystic origin and, if this interpretation is correct, it is concluded that the diamonds in other kimberlite intrusions in the pipes are also mainly xenocrysts. This conclusion is based on Robinson's (1980) observations that the diamonds in the DB1, DB2 and DB3 intrusions are essentially similar. Furthermore Harris (in

Clement et al., in press) has concluded that the diamonds from the Bultfontein, Dutoitspan, Wesselton and De Beers pipes "...exhibit close similarities with respect to those characteristics which can reasonably be associated with diamond formation in the Earth's upper mantle." Accordingly the bulk of the diamonds in the Kimberley group of pipes can be interpreted as xenocrysts.

Harris' conclusion, which is based on the examination of thousands of diamonds from the four pipes, is at odds with the views of earlier authors. Both Wagner (1914) and Williams (1932) refer to representative parcels of diamonds from the different pipes being easily recognizable but neither author presented any quantitative data to substantiate their statements. It does, however, appear that certain specific groups of diamonds (e.g. relatively abundant, large yellow octahedra at Dutoitspan) can be related to specific pipes. This does not, however, negate a xenocrystic origin since it might be ascribed to neighbouring kimberlites in part sampling slightly different mantle sources. Viewed as three geographic groups the Kimberley, Finsch and Koffiefontein diamonds are clearly derived from different populations (Harris et al., 1975, 1979; Robinson, 1980; Clement et al., in press).

Although the diamond content and distribution in the KIMFIK kimberlites may broadly reflect mantle sampling processes it is likely that modifications to diamond content and distribution probably do or can occur during and after intrusion. For example Robinson (1980) has suggested that resorption and high temperature etching probably occurs at depths shallower than 100km, i.e. during intrusion. Mechanical processes such as flow differention (section 3.5.1) within intrusions must influence diamond distributions and the possibilities of batch mixing during intrusion have already been referred to.

CHAPTER 9

THE ORIGIN OF KIMBERLITE PIPES: A REVIEW OF PREVIOUS THEORIES

Much of the work carried out during the course of this study has been aimed at determining the mode of origin of the KIMFIK pipes. This is an aspect of kimberlite geology which has received intermittent attention since the late 19th Century. In more recent times interest in the mutually related problems of pipe formation and kimberlite emplacement has been boosted primarily by the discovery of the Siberian occurrences (in the mid-1950's) and by J.B. Dawson's work on pipes in Lesotho (Dawson, 1960, 1962). The result of this enhanced interest has been a multiplicity of hypotheses; exemplified by such diverse views as pipe formation by explosive volcanism (e.g. Trofimov, 1971) and by diapiric intrusion (Mikheyenko, 1972). The complexity and often conflicting conclusions of published theories justifies a detailed critical review.

PART I: DIATREMES AND BRECCIA PIPES IN GENERAL

In Part I of this review an attempt is made to collate and classify existing general theories relating to the formation of a broad cross-section of diatremes and breccia pipes. This review is primarily concerned with processes occurring within the crust; particularly within the near-surface (0-5± km depth) environment.

In view of the recognition given to fludization as a major process in the formation of kimberlite pipes (e.g. Dawson, 1960, 1962, 1967a, 1971, 1980) considerable attention is also given in Part I to reviewing fluidization concepts and the geological features that can be produced by fluidized intrusion.

9.1 GENERAL THEORIES OF ORIGIN

Bryner (1961) defined breccia pipes as "...crudely cylindrical (often downward tapering) inclined or vertical structural units composed wholly or partly of angular or rounded rock fragments, with or without a matrix." This definition is non-genetic and sufficiently broad to include pipe-like bodies or diatremes with distinct volcanic associations (including kimberlite pipes) and other occurrences of diverse origins unrelated to volcanism. Various theories have been put forward to account for specific breccia pipes. Most occurrences have been explained by one of seven major processes or by modifications or combinations of these processes. These seven processes are briefly reviewed below.

9.1.1 Explosion

Explosive activity leading to the formation of breccia pipes is commonly associated with vulcanism or hypabyssal igneous intrusion. Such pipes are often apically situated with respect to underlying intrusions (Bryner, 1961).

The formation of such pipes is ascribed to the sudden and violent release of fluids (frequently gases) which have accumulated under high pressures. The vents or diatremes produced may be partly or wholly filled by fragments of juvenile igneous parentage, by country rock fragments or by varying proportions of both. Bryant (1968) noted that explosion vents tend to be vertical and circular, a form imposed by the violent expansion of the fluids during explosion and the concomitant requirement of immediate pressure release.

Explosive activity is not restricted to the formation of vents. The formation of breccia pipes in a hypabyssal environment has been ascribed to repeated, brecciating, subsurface explosions, prior to, or entirely without, later breaching of the surface (Bowes and Wright, 1961; Wright and Bowes, 1968). Such subsurface explosive activity is thought to have played a significant part in the early stages of formation of the KIMFIK pipes (section 10.2.1).

Subsequent to explosive vent formation brecciation and modifications of the shapes and sizes of breccia pipes may result from counter-explosions as a result of pressure release. This was proposed by Rust (1937) to explain 'authigenic breccias' in lamprophyric pipes in Missouri. Rust considered that the counter-explosions could be directly ascribed to pre-breakthrough build-up of gas pressure in the country rocks. Later breaching of the surface and the accompanying pressure drop in the vent led to violent inward decompression (implosion) of volatiles which had been temporarily held in the adjacent wallrocks.

A juvenile origin for the source of the fluids necessary for explosive acitivity is generally advocated or is implied by the commonly cited volcanic and/or intrusive igneous associations of explosion pipes (e.g. Gates, 1959). The origin of the fluids, generally gases, has, however, often been a contentious issue. It has also been proposed (e.g. Singewald and Milton, 1930) that fluids of magmatic origin are in some instances augmented by additional volatiles, derived from assimilation of country rocks, during magmatic intrusion at pre-explosion depths. Numerous authors (e.g. Lorenz, 1971a, 1971b, 1973, 1975, 1979; McBirney, 1959, 1963; Wolfe, 1980) consider that the explosions which cause the formation of many pipe-like bodies result from rising magma intersecting groundwater (phreatomagmatic explosions). Wolfe (1980) has pointed out that a great number of closely spaced explosions are involved in the formation of some maars and that the centres of the explosions migrate downwards with time.

9.1.2 Tectonic movements

Breccias are often formed as a result of folding, faulting, or shearing (Bryner, 1961; Bryant, 1968) and columnar breccia masses may be formed at the intersections of faults or shear zones or at bends in faults (Johnston and Lowell, 1961). Such breccias can generally be recognised by their close association with other tectonic features. According to Johnston and Lowell (op. cit.) tectonically produced breccia columns tend to be irregular in form and often change markedly in character on passing from one wallrock type to another. Bryant (1968) noted that breccias of this type are found at the site of disruption or have been dragged distances which correspond to the amount of fault or fold movement. The fragments comprising tectonic breccias are usually angular.

9.1.3 Igneous intrusion

The formation of breccia pipes or dykes can result from intruding magma disrupting and/or entraining sufficient solid material to form a breccia. Several examples of intrusion-breccias and intrusive intrusion-breccias (terminology of Wright and Bowes, 1963) have been described (e.g. Lovering, 1949; Wright and Bowes, op. cit.).

Bryner (1961) noted that the fragments in these breccias may be derived by magmatic stoping, by abrasion of the wallrocks if the magma is viscous, or by the engulfment of previously formed breccias (in the latter case actual fragmentation is usually due to some other process; commonly explosion).

Bryant (1968) pointed out that breccias formed by the push or drag of intruding magma are generally restricted to igneous contacts. The fragments may consist of broken and sheared wallrock or may be derived from chilled marginal zones of the intrusions.

In addition to direct brecciation and incorporation of fragments by rising magma columns, intruding magma may cause brecciation of overlying rocks by distension without actual magma penetration. Closely allied to this mechanism is the formation of breccia columns by hydraulic ramming of hydrothermal or volatile solutions into overlying rocks. This mechanism was proposed by Kents (1964).

9.1.4 Corrosion and solution

Breccias formed in situ, which may be pipe-like in character, can result from the removal of material by the movement of corrosive solutions along fractures or joints (Kuhn, 1938; Johston and Lowell, 1961). The fragments in these breccias may be angular but they are often well rounded. This rounding is due to enhanced corrosion at the corners and edges of the original joint or fracture-bounded blocks. Corrosion of this nature has often been invoked as a prerequisite for the formation of 'collapse breccias' (e.g. Locke, 1926).

9.1.5 Meteoric impact

Circular structures sometimes pipe-shaped but usually with high width to depth ratios and similar in form to many volcanic craters, have been ascribed to the impact of meteors. These structures commonly contain abundant brecciated material and may resemble in other ways some breccia pipes of terrestrial origin. However, further description or discussion is unwarranted as these structures are clearly irrelevant to the formation of kimberlite pipes.

9.1.6 Subsidence

Numerous pipe-like breccia bodies have been ascribed wholly or in part to the collapse of previously solid rock. Obviously this theory can only be applied when evidence is available which indicates that the breccia masses have moved downwards relative to the original stratigraphic position of the rocks which now comprise the breccia fragments. This downward movement has often been associated with peripheral ring faults or fractures (Francis, 1962, 1971; Whyte, 1964, 1968; Lorenz, 1967, 1968, 1971b, 1973, 1975; Watson, 1967 ; Hearn, 1968; Eggler and McCallum, 1974; McCallum and Mabarak, 1976).

Several explanations have been advanced for the formation of sufficient space to allow subsidence and to cater for the expansion that accompanies brecciation. Proposals include; mineralization stoping (Locke, 1926), magmatic stoping (Daly, 1933), slackening of intrusive pressure (Bryner, 1961), complex up-and-down pumping actions of igneous intrusions (Bryant, 1968) or associated hydrothermal solutions (Kents, 1964), evacuation of magma chambers by subterranean magma withdrawal or by evacuation through related nearby eruptions (Perry, 1961; Frechen, 1962; Francis, 1971), groundwater solution of limestone (Bryner, 1961) and melting induced by compressed gases in advance of an intruding magma column (Bryner, 1961). In addition Norton and Cathles (1973) have suggested that collapse may take place into a void created by disruption of a vapour bubble (produced by exsolved volatiles) apically situated at the head of a rising pluton.

9.1.7 Fluidization

Recognition of fluidization as a major geological process is a comparatively recent development although Cloos (1941) proposed an essentially similar mechanism, involving gas-tuff streaming, to account for the formation of certain olivine-melilitite tuff pipes in the Swabian district of Germany (Figure 9.1).

Later, in a classic paper, Reynolds (1954) reviewed Cloos' work and that of several other authors who had proposed intrusive gas-solid emplacement systems (Fairbairn and Robson, 1942; Walton and O'Sullivan, 1950; Patterson, 1952; King, 1953).

Reynolds recognized the similarity between these geological systems and the industrial process of fluidization and postulated several other instances of fluidized intrusion. Her examples encompassed a variety of rock types and forms of intrusive bodies. She also recognized several examples of extrusive fluidized systems.

Since 1954 a number of authors have advocated fluidized systems (mainly gas-solid systems) to explain a variety of non-kimberlitic intrusive bodies many of which are pipe-like in form (e.g. Shoemaker and Moore, 1956; Brindley, 1957; Francis, 1959, 1960, 1961, 1962; Whitten, 1959; French and Pitcher, 1959; Barrington and Kerr, 1961; Bowes and Wright, 1961, 1968; Coe, 1966; Bryant, 1968; Dunham, 1968; Lorenz, 1975, 1979; McCallum et al., 1976; Anderson, 1979). Experimental support (in the form of physical modelling) for the formation and infilling of diatremes by fluidization processes has come from Woolsey et al., 1975; McCallum, 1976; and McCallum et al., 1976.

Most of the general theories of diatreme and breccia pipe formation discussed above have been applied, either directly or in some combined form, to the origin of kimberlite pipes. These applications are discussed in Part II of this review.

9.2 THE NATURE OF FLUIDIZATION

The industrial process of fluidization is "...the operation by which fine solids are transformed into a fluid-like state through contact with a gas or liquid" (Kunii and Levenspiel, 1969). Most geological examples of fluidization have, however, been ascribed to gas-solid rather than liquid-solid fluidization (section 9.3).

In most industrial applications gas is transmitted through a bed of fine-grained solids to provoke or enhance gas-solid chemical reactions. However, in a geological content the physical changes imposed on the particles, during the passage of gas, are equally (or more) important.

Two fundamental parameters influence the behaviour of particles in gas-solid systems. These are the rate of gas flow and the size of the particles. The effects of variations in these (and other parameters) are noted below (data from Reynolds, 1954; Lewis, 1968; Kunii and Levenspiel, 1969).

At slow flow rates gas will pass through a particle bed without agitating the particles, i.e. the gas merely percolates through the void spaces between stationary particles. In industrial fluidization terminology this is a 'fixed bed'.

At more rapid gas flow rates the bed will expand and the packing density of the particles will decrease, i.e. the particles move apart and some may vibrate and move about in restricted zones of the bed. This is termed an 'expanded bed'.

Further increases in gas velocity will result in the free suspension of solid particles in the bed. At this stage the frictional forces between particles and through-flowing gas counterbalance the weight of the particles and the bed is considered to be fluidized. It is termed an 'incipiently fluidized bed' or 'a bed at minimum fluidization'. The gas velocity at which minimum fludization occurs is termed the minimum fluidization velocity (U_{fm}). Unless a particle bed is very shallow the top will fluidize before the bottom. The gas will be at a higher pressure at the bottom and will expand as it passes through the bed so that the velocity is highest at the top.

The voidage within incipiently fluidized beds is generally of the order of 30-40 vol.% and U_{fm} is proportional to the density of the particles and to the square of the particle diameter (assuming particles of a standard size).

Gas velocities in excess of U_{fm} will causes further separation of the particles in a fluidized bed. However, unless the particles are smaller than 0,05mm, the degree of separation will not exceed $^1/_{16}$ of the particle diameter before the bed becomes unstable. At this stage agitation becomes more violent, bubbling and channeling of gas becomes evident and the movement of particles becomes more vigorous. The bed does not, however, expand, much beyond its volume at U_{fm}. An 'aggregative' or 'bubbling' fluidized bed is formed.

Flow rates which produce bubbling fluidized beds have potentially important geological applications as such beds provide ideal conditions for mixing and attritioning of solid particles (section 9.3) and for chemical reactions between gases and solids.

Under bubbling fluidized bed conditions the phenomenon of 'slugging' may occur if the fluidized bed is deep in relation to its cross-sectional area. 'Slugging' results from gas bubbles growing and coalescing as they rise. Eventually the coalescence of bubbles may lead to gas extending across the bed and carrying plugs of solids up before it, i.e. slugging results in transportation of solids at gas velocities which normally would only cause mixing under 'bubbling fluidized bed' conditions.

Up to the 'bubbling fluidized bed' stage both gas- and liquid-fluidized systems are considered to be 'dense-phase fluidized beds'. Dense-phase gas-fluidized beds have the appearance of boiling liquids and in many ways they behave like liquids. They have definite, roughly horizontal upper surfaces and relatively light objects will float on these surfaces. Also, the beds will flow like liquids and the difference in pressure between any two points in a bed is roughly approximated by the static head of bed between the points. The effective density of the beds is the expanded volume divided by the weight of the particles and is analagous to the densities of liquids.

At sufficiently high gas flow velocities (beyond those necessary for bubbling fluidized beds) the terminal velocity of the solid particles in a fluidized bed will be exceeded. As a result entrainment of particles occurs and the upper surface of the fluidized bed disappears. A 'disperse-, dilute- or lean-phase fluidized bed' is developed and pneumatic transport of solids occurs.

The foregoing properties of fluidized beds are those that are reflected under conditions which, for industrial purposes, are made as ideal as possible. Commonly the particles are small and of uniform size and shape (often spherical); conduits are smooth-walled and regularly shaped; gas flow is regular and the operations are small-scale in relation to many natural geological situations. Changes in these parameters can have pronounced effects on the behaviour of fluidized systems (Reynolds, 1954; Lewis, 1968; Kunii and Levenspiel, 1969; McCallum et al., 1976; Wolfe, 1980). For example more-or-less open channels may be produced in a fluidized bed containing irregularly shaped particles which tend to stick together, i.e. channels occur between clustered particles (Lewis, 1968). Fine particles remain fluidized under wide variations in gas flow rate even if the particles have a wide size distribution. In contrast beds of large, uniformly sized particles often fluidize poorly with attending bumping, spouting and slugging. The addition of limited quantities of fine material to such systems often dramatically improves the quality of fluidization by acting as a lubricant (Kunii and Levenspiel, 1969).

The behaviour of very large particles in fluidized systems has not been experimentally verified. According to Lewis (1968) gas flow patterns begin to change if the particles are greater than 2-3cm in size. Extrapolations of correlations developed for small particles will therefore not necessarily be valid. However, the ability of a fluid to transport coarse material of higher density has been clearly demonstrated by dredging and commodity transport pipe lines (Bryant, 1968).

In industry relatively large particles (pea-sized and upwards) have fluidized in part, under 'spouted bed' conditions. Spouted beds involve a central jet of upward-moving, 'lean-phase' fluidized material surrounded by a slow downward-moving bed through which gas percolates upwards. The spouted bed process has in effect been invoked to explain the origin and the emplacement of numerous volcanic structures including kimberlite pipes (e.g. Lorenz, 1975; McCallum, 1976; McCallum et al., 1976).

The foregoing discussion and review has been confined mainly to gas-solid fluidized systems; the systems commonly invoked to explain the emplacement of kimberlites in diatremes and, to varying degrees, the formation of the diatremes themselves (section 9.10). The latter inference is based on the ability of through-flowing gas-tuff streams to erode and enlarge their passage-ways.

As noted previously fluidization can also be produced by the flow of liquids through particle beds. Kunii and Levenspiel (1969) note that in liquid-solid systems increases in flow rates above U_{fm} usually result in smooth, progressive expansion of fluidized beds.

Gross flow instabilities are dampened and remain small and large scale 'bubbling' does not normally occur. The particles merely separate further and further apart with increasing flow rate. Reynolds (1954) pointed out that elutriation will occur if the particles fluidized by a liquid vary in size.

The process of spraying is akin to fluidization (Reynolds, op. cit.). It is the gas-liquid equivalent of gas-solid and liquid-solid fluidized systems.

As pointed out by Wolfe (1980) the term 'fluidization' is often very loosely used in geological literature. This is certainly true in relation to kimberlites. Some references to 'fluidized intrusion' of kimberlite imply an unusual or unique intrusive process when, in fact, 'lean-phase' liquid-solid fluidized intrusion has occurred (e.g. Mitchell, 1978, 1979a). The latter process is common in geological situations and has operated, to a greater or lesser degree, in all cases where relatively abundant xenoliths, xenocrysts, autoliths or phenocrysts have been transported upwards by a magma of lower density than the transported solids. Much more specific gas-solid systems were invoked in the original fluidization theory applied to kimberlites (Dawson, 1960, 1962).

Wolfe (op. cit.) draws a distinction between high velocity gas streams transporting solids and fluidization processes. However, the former process is entirely consistent with 'lean-phase' gas-solid fluidized intrusion, a process which has been invoked to explain several classic geological examples of fluidization (e.g. Reynolds, 1954; Coe, 1966).

9.3 GEOLOGICAL FEATURES PRODUCED BY FLUIDIZATION

A fluidized origin has been proposed for a variety of geological bodies. These include regular pipe-shaped bodies and diatremes, extremely irregular pipes, dykes, veins and sills, bodies involving combinations of these forms and extremely complex bodies which defy morphological classification. However, although the shape of the body is not definitive of an origin by fluidization, it would be expected, according to theoretical concepts and the experimentally proven behaviour of fluidized systems (section 9.2), that such processes would impose some distinctive features on geological bodies.

Many factors will influence the number, nature and extent of development of features resulting from fluidization. These factors include: the nature of the rocks traversed by the intrusion, the types of structures present in the country rocks, the nature, rate and duration of fluidized flow, the degree of fluidization attained, the nature and abundance of solid material in the fluidization cell or column, the mobility, density, and temperature of the intrusion and the extent to which other genetic processes (unrelated to fluidization) may have influenced or modified the character of the deposit.

Outside the kimberlite field many examples of fluidized intrusion have been described (references are given in section 9.1.7). For the most part a fluidized origin has been invoked in these cases because of the presence of rocks which have the physical characteristics of tuffs or tuff-breccias but are clearly intrusive, i.e. they are tuffisites or tuffisite-breccias. In several of these instances the evidence for fluidization, specifically gas-solid fluidization, is virtually indisputable. Features displayed by these intrusions are briefly described below for later comparison with those evident in kimberlite pipes (Part II and Chapter 10). This summary is based on information given in the references noted in section 9.1.7.

(i) The occurrence of breccias made up of two distinct components, i.e. relatively large breccia fragments set in a finely fragmental tuffisitic matrix. Commonly the matrices of these tuffisite-breccias are composed of a heterogeneous assemblage of particles derived from the rocks traversed by the intrusion. Juvenile lapilli may be present and the matrices may also incorporate some primary material derived from the gas phase (or phases) present during fluidization.

(ii) The presence of huge country rock masses (which may measure many tens of metres across) which are located within tuffisite-breccias at or below their original stratigraphic levels (Figure 9.1). In some instances the original orientation of these masses (shown by internal structures such as bedding) has been maintained despite subsidence of several hundred metres.

(iii) The occurrence of distinctive distribution patterns which indicate that the positions of the larger fragments (breccia fractions) within the tuffisitic breccias are not entirely random or chaotic. Patterns range from simple 'ordering' (Cloos, 1941), through pronounced horizontal and/or vertical zoning (in relation to fragment size and/or rock type), to extensive homogenenization indicative of thorough mixing of fragments. One of the patterns most diagnostic of fluidization in vertical or near-vertical bodies, is the maintenance of the original stratigraphic succession by breccia fragments, which have not been transported to any extent upwards or downwards, following incorporation in intruding tuffisite. This phenomenon has been ascribed to the mass of the fragments counter-balancing the positive buoyancy imposed by upward flowing gas-tuff streams. In such cases the incorporated breccia fragments are homogeneous with respect to rock type (within specific vertical zones of the bodies) but the matrix (tuffisitic) components are a heterogeneous mixture of transported particles. This situation is probably rare because of the delicate balances of upward and downward-acting forces required. A more common distribution feature is therefore a mixing of locally derived breccia fragments with others that have been transported variable and often considerable distances. The degree of mixing and extent of transportation are dependant, amongst other things, on: the rate and continuity of flow of the transporting medium; its density; and the size and density of the fragments which comprise the breccia fraction of the tuffisite-breccia which is eventually formed.

(iv) Pronounced control on the shape of the intrusions by structures in the country rocks such as faults, fractures, bedding planes, foliations and joints is a common feature of fluidized intrusions.

(v) The presence of contact features which are indicative of non-dilational intrusion. These include: abraded contacts, indicated by slickensided polished, grooved or striated surfaces; partially detached (by erosive gas-solid streaming), often irregular, protuberances or wallrock appendages projecting into tuffisite (Fig. 9.2); and highly irregular or serrated contacts delineated by wallrock structures (Fig. 9.2).

(vi) Intimate penetration of tuffisite as apophyses of major tuffisite and tuffisite-breccia bodies into cracks, crevices, joints and other structural breaks in the country rocks (Fig. 9.3). This type of tuffisite intrusion frequently leads to the formation of complex, reticulate networks of tuffisite veins or the formation of intrusion breccias in contact zones.

(vii) The presence of associated dykes, veins and, less frequently, sills of tuffisite emplaced before or after the main phase of intrusion.

(viii) Internal features indicative of tuffisite intrusion. These include: flowage differentiation within tuffisite; orientation of elongate breccia fragments parallel to flow directions in tuffisite-breccias; similar orientation of elongate matrix components; flow lines exhibited by tuffisite around large fragments; swirl and eddy structures indicative of turbulent tuffisite intrusion; variable flow directions which can be related to structural control imposed by the country rocks and flow structures indicating circulatory movements within confined fluidization cells. Another common feature is the transgression, veining and fragmentation of breccia fragments and large inclusion masses (floating reef) by tuffisite. Some of the above features are illustrated in Figures 9.2a, 9.3, 9.4c and 9.5.

(ix) The presence of highly abraded, rounded, grooved, faceted, polished or striated components reflecting attrition within gas-tuff streams. Inclusions which display one or more of these features (particularly a high degree of rounding) but have not been transported far from their sites of origin, have frequently been cited as evidence of fluidization. The inclusions have been rotated and abraded by intruding tuffisite while remaining (because of their mass) essentially in situ.

(x) The occurrence of erosively-produced fretted surfaces (sand blasted) on some fragments and mineral grains (Fig. 9.4a-b).

(xi) The absence of thermal metamorphic effects on incorporated breccia fragments or on the wallrocks is a feature of fluidized intrusions and is generally taken to indicate low intrusion temperatures.

(xii) The presence of open cavities in tuffisites or tuffisite-breccias, or the presence of cavities partly or wholly filled by secondary minerals.

PART II THEORIES ON THE ORIGIN OF KIMBERLITE PIPES

In Part I the aims of the review were:

(i) To describe briefly various processes which have been invoked for the genesis of a variety of pipe-like structures, particularly those containing breccias of one form or another.

(ii) To assess the nature and effects of fluidization - particularly gas-solid fluidized intrusion in geological situations.

In Part II of this review a representative cross-section of published theories on the genesis of kimberlite pipe are reviewed and evaluated. Critical comment is, however, restricted to points that are considered to be of major significance. More detailed comment is made, where relevant, in Chapter 10 where the origin of the KIMFIK pipes is discussed in detail.

9.4 EXPLOSIVE VOLCANISM: THE CLASSICAL VIEW

The classical interpretation of the origin of kimberlite pipes stems from early studies of the major South African pipes. These led to the pipes being interpreted as volcanic vents formed by volcanic outbursts from depths of $\sim$2km or more. Subsequently an origin involving some form of explosive activity has been fairly generally accepted. Opinions differ, however, with regard to the exact role played by volcanic explosions in the formation of kimberlite pipes and with respect to the depths from which explosive outbursts occurred.

The volcanic affinities of kimberlite were first recognized by Carvill Lewis in 1887 and, shortly after the turn of the Century, several authors presented papers which, directly or by implication, invoked explosive volcanism for the formation of the pipes (e.g. Harger, 1905; du Toit, 1906; Merensky, 1909). The most intensive early investigations were, however, those of Wagner whose results were published in 1914.

Wagner (op. cit.) concluded that the major South African pipes represented "...deeply eroded, funnel-shaped volcanic rocks of the Maar type.... formed by the violent explosive liberation at the earth's surface of highly compressed vapours and gases of magmatic origin." This conclusion was based on his interpretation of: the shape of the pipes; the nature of the pipe contacts; the nature of the infilling material (kimberlite tuff, kimberlite injection breccia and, mainly at deeper levels, porphyritic kimberlite); the presence of slumped blocks or brecciated masses of country rock within the pipes; and the absence of thermal effects on inclusions in the upper parts of the pressure necessary for explosion to build up) in near-surface (2-6km deep) magma chambers.

The intermediate hearth concept is exemplified by Trofimov's 1971 paper. It is assumed that initially alkaline-ultrabasic magma (proto-kimberlite) is intruded, in pulsatory fashion, along deep-seated fractures, from upper mantle depths. Magma ascent is halted when an impermeable cap of sedimentary or volcanic rocks (which post-date the fractures) is reached. Thus subsurface magma chambers are considered to form at depths as shallow as 2km. Trofimov equates the evolution of true kimberlite liquids to continuing high pressures and temperatures within these magma chambers; the processes involved include partial assimilation of country rock and magmatic differentiation, aided by the repeated influx of new magma from depth. Trofimov considers that diamond crystallizes at this stage (and that its crystallization is related to temporary, very high, over-pressures; see also Trofimov, 1980).

According to the above author kimberlite pipes reflect explosive destruction of the magma chambers; the result of pressure build-up, due to exsolution of volatiles, during eventual cooling and crystallization of magma contained in the chambers. He envisages that initially increased volatile pressure causes upward warping of the cap rocks which eventually fracture. The subsequent escape of volatiles, along apically situated fractures, is interpreted, at least in part, as an explosive process which brecciates the surrounding rocks and forms the pipes. Explosively ejected pyroclastic material either falls back in the vents or forms tuff rings around them.

The classic explosive-boring hypothesis has been strongly criticised by Dawson (1960, 1962, 1967a and 1971). His contention that this theory is untenable is supported and amplified by the results of this study.

The complex configurations of kimberlite pipes (prior to erosion) involving morphologically distinct crater, diatreme and root zones are inconsistent with paroxysmal outbursts from depths of 2km or more, particularly as the morphological zones are not related to changes in country rock stratigraphy (section 3.2.2).

The irregularity of the root zones (sections 3.2.2 and 3.2.3) precludes an origin by explosive-boring from root zone depths although it could be argued that the irregularity of these zones reflects counter-explosions in the pipes, following initial explosive outburst (cf. Rust, 1937). Features which negate the latter possibility include:

(i) The occurrence of subsurface, pipe-like, often domed, extensions to the root zones of some pipes and the occurrence of blind satellite pipes (sections 3.2.3 and 3.2.4).

(ii) The lack of direct correlations between root zone irregularities and variations in country rock stratigraphy (section 3.2.2).

(iii) Indications that magmatic stoping has played a significant role in the formation of root zones. Evidence for this is the occurrence of minor sills, dykes and veins of kimberlite (hypabyssal-facies) intruded into the country rocks (along joints and fractures) at the margins of pipes. It is also shown by the occurrence of intrusion breccias at root zone contacts (section 3.3).

Although explosive outburst from root zone depths is untenable it could be argued that explosive breaching of the surface took place from just above the root zones, i.e. from or near the base of the vertically extensive diatreme zones. Kostrovitsky and Vladimirov (1971) and Kostrovitsky (1976) have raised a major objection to this postulate by pointing out that kimberlite pipes are far too deep, relative to their radii at surface, to be ascribed to explosive boring. These authors have noted that the ratios of the radii of the orifices of other volcanic structures, relative to their depths, commonly equal or exceed 0,7. This figure is based on theoretical calculations and numerous empirical data derived from natural volcanic explosion craters and must be large relative to the kimberlite pipes which are considered to be apically located above them. No such chambers have been intersected in pipes which have been mined to considerable depths; almost certainly in excess of 2km below the original (pre-erosion) surface (e.g. Kimberley Mine). In addition, since kimberlites vary extensively in respect of absolute ages (Hawthorne et al., 1975; Davis, 1977), it would be expected that some areas of kimberlite injection have subsequently undergone sufficient post-emplacement erosion for intermediate hearths to be revealed at surface. No such exposures are known. Additional evidence against the occurrence of major magma chambers within a few kilometres of surface is the absence of any geophysical (or other remote sensing) evidence for such subsurface bodies.

In one sense the possibility of near-surface magma chambers being present and instigating higher level kimberlite pipe formation, cannot be entirely ruled out. Near-surface ponding of kimberlite magma in sills certainly occurs (Hawthorne, 1968; Dawson and Hawthorne, 1973) and it is conceivable that pipes rooted in underlying sills might occur. This situation is analagous to the occurrence of Stormberg volcanoes rooted in major basaltic sills reported by Gevers (1928). However, kimberlite pipe/sill relationships of this type have never been reported and it is noticeable that in most areas where pipes and sills occur together that the sills have been intruded at levels well above the base of the pipes (Hawthorne, 1975). Furthermore, known kimberlite sills are of limited extent and are unlikely to have given rise to major pipes, unless they have acted only as linking passageways between the pipes and underlying feeders.

9.5 PIPE FORMATION BY INTRUSION ALONG FISSURES.

In a detailed two-volume treatise Williams (1932) presented a wealth of data relating to many aspects of the major South African pipes, particularly the Kimberley group. His views on the genesis of these pipes have not, however, received wide acceptance.

As a prerequisite for the formation of pipes Williams envisaged disruption of the crust by "...certain agencies" (tectonic movements). This disruption was not considered in any way related to kimberlite magmatism which it pre-dated. Williams inferred that these crustal movements produced a widespread network of fissures and that these fissures were partly or wholly occupied by rock fragments derived from the fissure walls. Kimberlite pipes were ascribed to the later intrusion of kimberlite magma at the junctions of intersecting fissures.

Williams considered that kimberlite pipes reflected slow intrusion along the intersecting fissure zones and that, during intrusion, fragmented country rock, originally present in the fissures, was either incorporated in the magma or removed ahead of the rising column. He concluded that the shapes of the pipes were extensively modified during prolonged upward movement of xenolith-bearing magma. Such modification was considered to reflect an erosive process during which the original irregularities of the intersecting fissure zones were worn away and the pipes were enlarged.

Williams did not believe that large, complex pipes, such as the KIMFIK occurrences, reflected several stages of intrusion and eruption. He considered that major kimberlite pipes were formed by a single continuous, prolonged, intrusive event. Variations in the appearance of the kimberlite and of the diamond content from place to place within individual pipes were ascribed to the mechanical mixing of cooled semi-plastic material and still-rising, more fluid magma.

Both downward and upward movements of incorporated country rock fragments was envisaged during the mixing process. Williams contended that some large inclusions sunk prior to consolidation of the kimberlite magma in the pipes while other fragments were transported upwards by intruding magma. He suggested that the rounded nature of some inclusions was caused by mutual abrasion within intruding magma. Williams recorded

(vi) Williams' emplacement model fails entirely to account for the textural variation exhibited by kimberlites within the KIMFIK pipes (Chapters 2 and 7).

(vii) The clear cut evidence of multiple intrusions in the KIMFIK pipes (section 3.5) clearly negates Williams' postulate of a single, drawn-out intrusive episode.

9.6 KIMBERLITE PIPES AS SUBSIDENCE STRUCTURES

Perry (1961) suggested that kimberlite pipes developed as a consequence of subsurface subsidence. His views stemmed mainly from an examination of the Premier pipe and were strongly influenced by the great down-rafted mass of Waterberg System rocks (mainly quartzite) which is present in this pipe (Frick, 1970; Hawthorne, 1975).

Perry considered that redistribution of magma, during the early stages of formation of the Premier pipe, took place through neighbouring vents or fissure outlets. The subsurface withdrawal of magma resulted in the excavation of a cupola into which the sides and roof of the embryonic pipe collapsed. Perry considered that the column of brecciated rock so formed continued to develop upwards following persistent magma withdrawal. The end result was the development of a pipe-like, subsurface breccia column with an apex near surface.

Perry postulated that this apex was subsequently fractured in response to locally developed stresses. The fractures are considered to have reached the surface and provided passageways for the relief of vapour pressure within the apex. The outward flow of gas, restricted by friction, would have provided, according to Perry, a medium for upward and downward directed mixing, agitation and comminution of the previously formed breccia fragments.

The enormous block of partly brecciated Waterberg floating reef within the pipe was considered to reflect mass collapse, along peripheral fractures, into a void which existed temporarily under the near-surface arch.

the displacement of large wedge-shaped blocks along upthrust faults with "...great amplitudes of displacement." At the same time downward movement of some incorporated fragments is considered to occur because the rising magma circulates in a closed system.

Clearly neither of these diapiric theories has any real application to the formation of kimberlite pipes. This is reflected by the lack of doming, folding, arching, faulting, thrusting and shearing of the immediate country rocks. Some of these features would be mandatory effects of diapiric intrusive processes. Many other arguments can be raised against the formation of kimberlite pipes by diapiric intrusion but further discussion is not warranted.

9.8 PIPE FORMATION BY REPETITIVE SUBSURFACE EXPLOSIONS.

In an attempt to overcome the shortcomings of the classical explosive boring hypothesis Kostrovitsky and Vladimirov (1971) proposed that kimberlite pipes are formed by successive, upward-migrating, subsurface explosions. They concluded that these explosions culminated in crater-forming eruptions from depths of 150m or less.

These authors suggested that initial explosive activity takes place at depths of 5km or less. This results in the development of a collapsed 'crush zone' (breccia zone) and a series of concentric vertical cracks in the rocks above the explosion centre. Thus a new chamber is provided for a subsequent explosion at a higher level. Progressive repetition of this process is envisaged until ultimately breaching of the surface occurs.

Kostrovitsky and Vladimirov ascribed the subsurface explosions to exsolution and expansion of magmatic gases as the confining pressure on uprising kimberlite magma decreased. They suggested that the explosions could occur in two ways: by supersaturation of the melt by adiabatically expanding gas or by the formation of chemically reactive gas mixtures.

attained) and the increased erosive capabilities of gas-tuff streams in the near-surface environment.

It is concluded that all theories that seek to explain the genesis of kimberlite pipe by emplacement models that incorporate points listed under (i)-(iii) above are untenable. Under such 'reaming out' conditions the complex configurations of kimberlite pipes (with discrete vertical zones) would not be formed. In particular the occurrence of many of the complex root zone features described in Chapter 3 is entirely inconsistent with such genetic models. These root zone features include: blind appendages of major pipes; apparently discrete, blind satellite pipes; sill-like protuberances from pipes into the wallrocks; local subhorizontal pipe contacts; hypabyssal-facies kimberlite intrusion breccias at pipe contacts; and the occurrence of contact breccias characterized by local derivation of the component fragments and little or no dislocation of the breccia blocks. The abundances of discrete intrusions within pipes are also inconsistent with high velocity 'reaming out' emplacement models involving deep-seated fluidized intrusion.

Dawson's views (Dawson, 1960, 1962, 1967a, 1971, 1980) were influenced by Cloos' earlier interpretation of the origin of olivine melilitite pipes in the Swabian district of Germany (Cloos, 1941). Thus Dawson envisages a major role for fluidization in enlarging and shaping kimberlite pipes by erosive gas-tuff streaming. In order for this to occur prolonged high velocity gas streaming is mandatory. Taking into consideration the features developed by gas-solid fluidization in other (non-kimberlite) situations (section 9.3) and evaluating them against features displayed by the KIMFIK pipes does not, however, support this conclusion.

If prolonged erosion by gas-solid fluidized systems is responsible for significant degrees of enlargement and shaping of the KIMFIK (or other) kimberlite pipes then extensive penetration of tuffisitic veins that the development of subsurface convective cells required the initial formation of a void by uparching of the country rocks. Such uparching of the wallrocks of kimberlite pipes is extremely rare and known examples are due to expansion) following secondary alteration and weathering of the kimberlite (Wagner, 1914).

(ii) The absence of tuffisitic veining adjacent to the pipes is inconsistent with extensive subsurface fluidization stoping (cf. comments on Dawson's views, page 345).

(iii) McCallum indicates that, following subsurface stoping, breakthrough to surface would be "...accompanied by subsidence of surface layers into the fluidized conduit and the development of pronounced inward dips of wallrock sediments." Inward dips of the wallrocks around kimberlite pipes are extremely rare and have not, to the author's knowledge, been recorded around any southern African kimberlite pipe, including those where the presence of epiclastic kimberlite indicates that they have not been substantially eroded.

(iv) Like all other published theories which ascribe kimberlite pipes to fluidized intrusion (usually involving gas-solid systems) McCallum's emplacement model does not explain some of the root zone features of the KIMFIK pipes (e.g. the contact breccias).

9.11 GENERAL COMMENTS

(i) It is evident from Part II of this review that hypotheses explaining the origin of kimberlite pipes are many and varied but that none is satisfactory.

(ii) The complicated nature of the KIMFIK pipes (described in earlier Chapters of this thesis) indicates that they have undergone complex evolutionary histories and that a variety of processes have played a part in their development. The complexity of kimberlite pipes probably accounts, in part, for the diversity of published views on their mode of origin.

(iii) The importance of the role which has been ascribed to gas-solid fluidization in the formation of kimberlite pipes can be questioned.

(iv) A model for the origin of the KIMFIK (and other) kimberlite pipes is proposed in Chapter 10. This model incorporates some of the concepts noted in this review but involves a more detailed assessment of the processes which are responsible for the complex nature of the pipes.

ring fault results in the formation of a maar-type crater at surface.

Lorenz indicates that the saucer-shaped subsidence structure formed at this stage may be completely destroyed if subsequent subsidence and eruptive activity continue for a sufficiently long period. Eventually all the material within the ring fault may circulate, via a central eruption channel in which upward movement occurs, while slow downward movement occurs in the marginal zone of the developing pipe. In effect a process of tuffisitization develops as a consequence of circulatory movements of material induced by the fluidization which occurs in the central eruption channel.

The intersection of groundwater by rising kimberlite magma would, under appropriate conditions (sufficient groundwater at sufficiently shallow depths), almost certainly trigger surface-breaching phreatomagmatic explosions.

It is also conceivable that, as suggested by Lorenz, the initial explosion channel would be enlarged by repetitive explosions (assuming water recharge of the system occurred) and spalling and subsidence along ring faults (although there is little or no evidence for the latter). Thus the development of saucer-shaped structures as envisaged by Lorenz is a reasonable postulate and de Stadelhoven (1960) has described some pipes in Zaire which have extremely flat marginal dips (as low as 25°). These dips suggest some form of subsidence from well outside the limits of the primary explosion craters. Alternatively, however, such flat dips may reflect infilling of topographic depressions adjacent to vents by pyroclastic deposits (the nature of the material in the deposits described by de Stadelhoven is not known).

Lorenz's further conclusions that the overall morphology of kimberlite pipes reflects additional, prolonged subsidence along ring faults, coupled with zonally distributed tuffisitization and circulatory movements of material, is not supported by this study of the KIMFIK pipes.

The diatreme zones of these pipes do contain tuffisitic material (Chapter 2) but display no indications of being bounded by ring faults. As noted in sections 3.3.1 and 3.3.2 they tend to be bounded by pre-existing structures (mainly joints) in the wallrocks.

There is no evidence in any of the KIMFIK pipes of the well developed, central eruption channels within which, according to Lorenz's model, fluidized intrusion and tuffisitization occurred while contemporaneous subsidence occured in the peripheral areas. These channels are an essential element of Lorenz's views.

It can be argued (as Lorenz has done) that prolonged tuffisitization of circulating material would, by homogenization, have destroyed all traces of the original eruption channels. It is, however, extremely unlikely that this would have occurred in every case. Thus, while there are numerous examples of earlier kimberlites cut and cored out by later intrusions (from which they commonly differ substantially both texturally and mineralogically) no central eruption channel of the type suggested by Lorenz is known.

Lorenz has pointed to the occurrence of down-rafted floating reef masses near the margins of pipes as evidence favouring his theory. Such masses cannot, however, be unambiguously interpreted as evidence for systematic ring faulting over vertical distances of up to 2km± or for his proposed circulatory systems (see section 10.2.3).

Another problem with respect to Lorenz's views on the origin of kimberlite pipes is that they, like most published theories, offer no explanation for the root zone complexities described in Chapter 3. Also, they neglect the complex internal structure of large pipes such as the KIMFIK occurrences. In addition they do no adequately explain the various textural types of kimberlite present in many pipes (Chapters 2 and 7).

9.10 FLUIDIZATION MODELS FOR THE ORIGIN OF KIMBERLITE PIPES

In an unpublished thesis (1960) and subsequently in a series of publications (1962, 1967a, 1971, 1972, 1980) J.B. Dawson proposed that, following initial explosive breakthrough to surface, kimberlite pipes were enlarged, shaped and infilled by gas-solid fluidized intrusion. In 1980 Dawson summarized his views as follows:

"A composite model of kimberlite emplacement would be that calcite-rich kimberlite magma, containing megacrysts and xenoliths is intruded rapidly up a system of deep seated fractures....; speed of ascent must have been sufficiently rapid to preclude sedimentation of high-density peridotite and eclogite xenoliths and also to prevent total resorption of diamonds. Then at preferred points where access to the surface is easiest....and possibly where groundwater percolation is deepest, kimberlites break through to surface from depths of 2-3km. Adiabatic expansion of exsolving CO_2 from the magma and heat absorption by groundwater would cause rapid cooling....and rapid upward transport of deep seated material.... The initial explosion vent would then be enlarged by fluidized fragmental material drilling upwards and taking advantage of the country rock jointing; major marginal expansion could be caused by major wall-rock fragments (the 'floating reef') slumping into the vent; and wall-rock spalling in the initial explosive stage is a possibility. Also some type of pseudo-stoping, perhaps due to hydraulic fracturing, must be invoked to account for 'blind' diatremes, such as one of the lobes of the Wesselton Mine. In some cases later surges emplace later distinctive tuff columns or tuffisite dykes, whilst cavities or consolidation fractures in the vent kimberlite may be infilled by quietly upwelling kimberlite magma to form plugs or dykes of hypabyssal-facies kimberlite and kimberlite breccia."

Following publication of Dawson's hypothesis, which was based on an investigation of pipes in Lesotho, few authors have seriously questioned his conclusion that gas-solid fluidization played a major role in the formation of kimberlite pipes. However, contrary to Dawson's views, many subsequent investigators have concluded that fluidization is not restricted to the near-surface environment. Some authors consider that fluidization is initiated deep within the crust (Davidson, 1967;

Mitchell and Crocket, 1971). Others favour the initiation of some form of fluidized intrusion from the upper mantle (Kennedy and Nordlie, 1968; Harris and Middlemost, 1969; Brookins, 1970; McGetchin and Ullrich, 1973; McGetchin et al., 1973; Ellis and Wyllie, 1979; Wyllie, 1978, 1980; Bailey, 1980).

The genetic models proposed by these authors differ in many ways, particularly with respect to: the nature and state of the fluids involved; the actual depths at which fluidized systems evolve; and whether intrusion is a continuous event or a two-step process in which fluidization is preceded by some other intrusive mechanism. Notwithstanding these (and other) differences the theories advocating fluidization from great depth (deep crust or upper mantle) have several features in common:

(i) All postulate (or imply) that in the near-surface zone gas-solid fluidized systems are involved in kimberlite emplacement and pipe formation. Gas, supercritical fluids, or liquids (kimberlite magma) are held to be the transporting agents at great depth.

(ii) All postulate that intrusion is very rapid following the onset of fluidized intrusion and that the rate of intrusion increases as the surface is approached. McGetchin and co-workers (1973) have attempted to predict, by means of computer-controlled hydrodynamic modelling, actual intrusion rates. They have suggested intrusion rates of $\sim$400m/sec. near surface.

(iii) All postulate (or imply) that at depth the intrusions are confined to relatively narrow channels but that these pathways expand in the low pressure, near-surface zone to form outward-flaring pipes. In all cases a smooth transition from dyke-like to pipe-shaped bodies is a necessary response to the intrusive mechanism proposed. The transition in form is considered to reflect violent breaching of the surface (as terminal intrusion velocities are attained) and the increased erosive capabilities of gas-tuff streams in the near-surface environment.

It is concluded that all theories that seek to explain the genesis of kimberlite pipe by emplacement models that incorporate points listed under (i)-(iii) above are untenable. Under such 'reaming out' conditions the complex configurations of kimberlite pipes (with discrete vertical zones) would not be formed. In particular the occurrence of many of the complex root zone features described in Chapter 3 is entirely inconsistent with such genetic models. These root zone features include: blind appendages of major pipes; apparently discrete, blind satellite pipes; sill-like protuberances from pipes into the wallrocks; local subhorizontal pipe contacts; hypabyssal-facies kimberlite intrusion breccias at pipe contacts; and the occurrence of contact breccias characterized by local derivation of the component fragments and little or no dislocation of the breccia blocks. The abundances of discrete intrusions within pipes are also inconsistent with high velocity 'reaming out' emplacement models involving deep-seated fluidized intrusion.

Dawson's views (Dawson, 1960, 1962, 1967a, 1971, 1980) were influenced by Cloos' earlier interpretation of the origin of olivine melilitite pipes in the Swabian district of Germany (Cloos, 1941). Thus Dawson envisages a major role for fluidization in enlarging and shaping kimberlite pipes by erosive gas-tuff streaming. In order for this to occur prolonged high velocity gas streaming is mandatory. Taking into consideration the features developed by gas-solid fluidization in other (non-kimberlite) situations (section 9.3) and evaluating them against features displayed by the KIMFIK pipes does not, however, support this conclusion.

If prolonged erosion by gas-solid fluidized systems is responsible for significant degrees of enlargement and shaping of the KIMFIK (or other) kimberlite pipes then extensive penetration of tuffisitic veins along joints and fractures in the country rocks would be expected. Similarly, extensive veining of the floating reef bodies within the pipes and of earlier kimberlite intrusions by later tuffisitic kimberlite, would be expected. Enlargement and shaping of kimberlite pipes by gas-solid fluidization requires intimate penetration of tuffisitic material along country rock discontinuities for the detachment, incorporation and eventual tuffisitization of wallrock blocks. Such features were recorded by Cloos (1941) in the Swabian diatremes and excellent examples from Northern Ireland have been described and illustrated by Coe (1966). Similar features are very rare in, or adjacent to, kimberlite pipes. Only two examples of tuffisitic kimberlite breccia intruded along wallrock structures have been noted adjacent to any of the KIMFIK pipes. In both cases single dyke-like offshoots from the pipes, along pre-existing joint planes are evident. The extreme rarity of such features strongly suggests that, although gas-solid fluidized intrusion has occurred within the KIMFIK pipes (Chapter 10), it has not played as fundamental a role in pipe formation as Dawson has proposed.

Features of the KIMFIK pipes which suggest that fluidization was not in fact prolonged are the lack of rounding of many inclusions in the tuffisitic kimberlite breccia (section 7.2.1) and the quench textures frequently exhibited by the interstitial matrix material in these rocks (Clement and Skinner, 1979; Smith and Skinner, 1979; this thesis, Chapters 7 and 10).

The fluidization theories reviewed so far all suggest the fluidization evolved after breaching of the surface, or coincidently with the rapid propagation of passage-ways to surface. McCallum (1976) adopted a different point of view. He suggested that the early stages of pipe formation involved the development of subsurface, convective fluidization cells which stoped out an embryonic pipe prior to eventual breakthrough and 'conventional' fluidized streaming. McCallum envisaged the following stages of pipe formation:

(i) Slow intrusion of kimberlite crystal mush-magma along deep-seated fractures from mantle source areas.

(ii) Stoping and incorporation of upper mantle wallrocks.

(iii) Temporary arrest of intrusion in the lower crust, allowing accumulation of volatiles at the head of the magma column, accompanied by carbonatization and fenitization of local wallrocks and inclusions.

(iv) Renewed intrusion following increased gas pressure and/or the intrusion of additional magma from below.

(v) The onset of fluidization (at upper crustal levels) due to decreasing load pressure at progressively shallower levels.

(vi) Upward migration and stoping of embryonic pipes by fluidization cells.

(vii) Breakthrough to surface accompanied by gas-tuff streaming, subsidence of wallrock blocks and downwarping of vent margins.

McCallum's views are of particular interest because he is one of few authors to suggest that pre-breakthrough, subsurface processes play a major role in the formation of kimberlite pipes. This postulate is strongly supported by the results of this investigation (section 10.2.1). The convective fluidization cells which McCallum has advocated could conceivably be invoked to explain some of the morphological complexities of the root zones of the KIMFIK pipes and the more regular diatreme zones can be equated with post-breakthrough gas-tuff streaming as he suggests (see section 10.2.3). Some evidence for convective subsurface fluidization, at least on a local scale, was presented in section 7.4. In relation to the KIMFIK pipes McCallum's emplacement model can, however, be criticised for the following reasons:

(i) The small scale physical modelling of diatreme formation on which his hypothesis is based (Woolsey et al., 1975) indicated that the development of subsurface convective cells required the initial formation of a void by uparching of the country rocks. Such uparching of the wallrocks of kimberlite pipes is extremely rare and known examples are due to expansion) following secondary alteration and weathering of the kimberlite (Wagner, 1914).

(ii) The absence of tuffisitic veining adjacent to the pipes is inconsistent with extensive subsurface fluidization stoping (cf. comments on Dawson's views, page 345).

(iii) McCallum indicates that, following subsurface stoping, breakthrough to surface would be "...accompanied by subsidence of surface layers into the fluidized conduit and the development of pronounced inward dips of wallrock sediments." Inward dips of the wallrocks around kimberlite pipes are extremely rare and have not, to the author's knowledge, been recorded around any southern African kimberlite pipe, including those where the presence of epiclastic kimberlite indicates that they have not been substantially eroded.

(iv) Like all other published theories which ascribe kimberlite pipes to fluidized intrusion (usually involving gas-solid systems) McCallum's emplacement model does not explain some of the root zone features of the KIMFIK pipes (e.g. the contact breccias).

9.11 GENERAL COMMENTS

(i) It is evident from Part II of this review that hypotheses explaining the origin of kimberlite pipes are many and varied but that none is satisfactory.

(ii) The complicated nature of the KIMFIK pipes (described in earlier Chapters of this thesis) indicates that they have undergone complex evolutionary histories and that a variety of processes have played a part in their development. The complexity of kimberlite pipes probably accounts, in part, for the diversity of published views on their mode of origin.

(iii) The importance of the role which has been ascribed to gas-solid fluidization in the formation of kimberlite pipes can be questioned.

(iv) A model for the origin of the KIMFIK (and other) kimberlite pipes is proposed in Chapter 10. This model incorporates some of the concepts noted in this review but involves a more detailed assessment of the processes which are responsible for the complex nature of the pipes.

CHAPTER 10

THE ORIGIN AND INFILLING OF THE KIMFIK PIPES

10.1 SYNTHESIZED MODEL OF THE PIPES

Due to erosion and generally limited exposures there are no examples of individual kimberlite pipes where the overall morphological and internal characteristics have been accurately determined. Nevertheless, by synthesis of information from a large number of occurrences and particularly by examining open pit and underground exposures resulting from mining operations, it is apparent that the KIMFIK pipes were characterised by three distinctive depth zones (Chapter 3).

The three depth zones differ markedly from each other both in respect of configurations and the nature of the material contained within them. The differences are so pronounced (Chapters 3 and 7) that the formation of the pipes cannot be ascribed to any single mechanism of formation and emplacement. The KIMFIK (and other) pipes must reflect complex evolutionary histories and different processes must have operated at different times and/or different levels. An assessment of the origin of the pipes thus requires elucidation of the nature of these processes, establishment of the sequence and relative depths at which they operated and interpretation of their possible interrelationships. The major features of the three depth zones are summarized below. These features form the basis for many of the genetic interpretations which follow in subsequent sections.

10.1.1 Crater zones

The main features of crater zones are:

(i) They are shallow features usually not exceeding $\sim$300m in depth (section 3.4.2).

(ii) They have relatively flat marginal dips, commonly 50°-75° (section 3.4.2) but occasionally much flatter (section 9.9).

(iii) Country rock contact zones are often brecciated (section 3.4.2).

(iv) They have regular shapes in horizontal cross-section; usually circular or multiple-circular if the centres of explosive outburst have migrated at intervals during the formation of complex pipes.

(v) They contain stratified or unstratified epiclastic kimberlite and, in some instances, basal or interbedded tuffs and tuff-breccias (section 3.4.2; Hawthorne, 1975).

(vi) The axes of crater zones are vertical.

10.1.2 Diatreme zones

The main feature of diatreme zones are:

(i) They have regular shapes although they tend to become slightly more irregular at depth. Cross-sectional areas decrease systematically with depth (sections 3.2.2 and 3.2.3).

(ii) Marginal dips are steep (75°-85°) and contacts are megascopically smooth (sections 3.2.2, 3.2.3 and 3.3.2).

(iii) The zones frequently extend over considerable vertical distances, often 1-2km (section 3.2.2).

(iv) They have vertical axes (section 3.2.3).

(v) Joint-bounded contacts are evident indicating some structural control by discontinuities in the country rock on the form of diatreme zones (section 3.3.2). Note, however, that the influence of joints is much less pronounced than in root zones (sections 3.3.1 and 10.1.3).

(vi) The country rock contacts of diatreme zones may be grooved, polished or striated. Grooves or striations are not always vertically oriented (section 3.3.2).

(vii) Marginal zones of brecciated country rock occur within diatreme zones but are rare compared with their frequency of occurrence in root zones (section 3.3.2).

(viii) The wallrocks of diatreme zones have not been thermally metamorphosed.

(ix) Apart from rare brecciation the wallrocks of diatreme zones have not been structurally deformed (folded, faulted, arched or fractured) during intrusion (section 3.3.2).

(x) Characteristically diatreme zones are occupied by diatreme-facies kimberlites, mainly tuffisitic kimberlite breccias (sections 2.3.5 and 7.2).

(xi) Usually only one or few major intrusions are present in diatreme zones although small, marginal remnants of earlier intrusions may be preserved (e.g. Figures 3.30 and 3.35).

(xii) Diatreme zones contain subsided floating reef masses which tend to be peripherally located (section 3.6) and are often extensively brecciated.

(xiii) The vertical distribution of floating reef bodies may crudely reflect their original stratigraphic sequence (section 3.6.3).

(xiv) Floating reef bodies may be cut by tuffisitic veins (Plate 7.16e) but such veining is limited.

(xv) The TKB's in diatreme zones commonly contain pelletal lapilli, autoliths and abundant, commonly angular - subangular, unmetamorphosed, variably sized but mainly small, country rock xenoliths (sections 7.2.1 and 7.2.2).

(xv) The primary components of the TKB's in diatreme zones have commonly been altered by low temperature deuteric, metasomatic and weathering processes (sections 7.2.2 and 7.2.3).

(xvi) Calcite is an extremely rare primary mineral in the TKB's and TK's of diatreme zones.

(xvii) Microlitic serpentine and diopside are commonly abundant components of the interstitial matrix of TKB's and TK's (sections 7.2.2 and 7.2.3).

(xviii) The primary, xenolithic and xenocrystic components of the TKB's in diatreme zones are normally well mixed so that the rocks commonly have a fairly homogeneous appearance (on a gross scale) over considerable vertical and horizontal distances (section 7.2.1).

(xix) The diamond content of TKB's is relatively constant compared with the variations which are commonly displayed by hypabyssal-facies kimberlites (section 8.2).

(xx) Directional structures in the TKB's of diatreme zones are rare and tend to be only locally developed (section 7.2.1).

10.1.3 Root zones

The main features of root zones are:

(i) Pronounced irregularity, both vertically and horizontally and on large and small scales (sections 3.2.1, 3.2.2, 3.2.3, 3.3.1).

(ii) Considerable evidence of structural control of root zone morphology, mainly by planar discontinuities such as joints and fractures in the country rocks (section 3.3.1).

(iii) The occurrence of extensive contact breccias preserved particularly under overhangs of country rocks. The breccias are composed of local country rock fragments which are commonly angular. Many breccias are devoid of kimberlite and any movement of the breccia masses as a whole is downwards (section 3.3.1). In most instances little or no movement is evident.

(iv) The axes of root zones or parts of root zones may be inclined (section 3.2.3).

(v) The root zones often split into more-or-less discrete columns or channels.

(vi) Blind (subsurface) extensions of root zones and apparently discrete subsurface domes may occur (sections 3.2.3 and 3.2.4). Cappings of contact breccia may occur above subsurface domes (section 3.2.4).

(vii) The outer and inner contacts of contact breccias may be sharp or gradational (section 3.3.1).

(viii) Intrusion breccias and stockworks occur locally at root zone contacts (section 3.3.1).

(ix) Within the root zones minor dyke-like, sill-like and vein-like apophyses of kimberlite locally project outwards into country rock along joints and fractures (section 3.3.1).

(x) Stoped blocks of country rock are locally concentrated near root zone contacts (section 3.3.1).

(xi) In general relatively many discrete intrusions occur in root zones compared with diatreme zones (e.g. Figures 3.30 and 3.35).

(xii) Characteristically root zones are occupied by hypabyssal-facies kimberlites, i.e. kimberlite sensu stricto and kimberlite breccias (sections 2.3.5 and 7.3).

(xiii) Thermal metamorphism of the wallrocks may be evident at the contacts of pipe root zones, dykes and sills (e.g. Dawson and Hawthorne, 1970).

(xiv) Extensive metasomatism ('kimberlitization') of country rock xenoliths is a common feature of kimberlites and kimberlite breccias (section 7.3.1).

(xv) Crustal xenolith assemblages in kimberlite breccias tend to be more restricted than in TKB's and the size ranges of xenoliths in the former rocks are more restricted (section 7.3.1).

(xvi) The hypabyssal-facies kimberlites in root zones display wide ranges of mineralogical and geochemical composition and igneous textures (Chapter 7).

(xvii) The distribution of diamonds in the hypabyssal-facies kimberlites or root zones is commonly less consistent than in diatreme-facies kimberlites. Likewise the distribution of many other components (e.g. xenoliths, xenocrysts, phenocrysts) is generally more variable in hypabyssal-facies intrusions than in diatreme-facies kimberlites (Chapters 3, 7 and 8).

(xviii) The hypabyssal-facies intrusions in the root zones do not contain floating reef masses (section 3.6.3).

(xix) Deuteric alteration within hypabyssal-facies kimberlites is commonly extensive and complex (Chapters 5 and 7).

(xx) Contacts between intrusions within the root zones of the pipes may be sharp or gradational (section 3.5.1).

10.2 GENETIC CONSIDERATIONS

As noted in Chapter 9 the contrasting nature of crater, diatreme and root zones of kimberlite pipes are not adequately explained by published theories of pipe formation. Most theories are based on the characteristics of one or other of these morphologically distinctive zones (generally the diatreme zones) and fail to take the overall character of the pipes into account.

To account for many of the geological complexities described previously in this thesis (and summarized to a considerable extent in section 10.1) emphasis is placed here on the importance of a variety of interrelated subsurface processes. These processes result in the initial formation of subsurface embryonic pipes with the geological characteristics of the root zones.

10.2.1 Embryonic pipes: the development of extended root zones

Geological features of the root zones of the KIMFIK pipes indicate that a number of different processes are responsible for their formation. These processes include hydraulic fracturing and wedging, brecciation by intrusion of magma along wallrock discontinuities, magmatic stoping, intermittent explosive and/or implosive brecciation, spalling and slumping and, possibly, rock bursting (Gates, 1959) from temporary free faces. Evidence for these processes and the sequence in which they operated is discussed below.

The distribution of the contact breccias in the root zones of the KIMFIK pipes (Figures 3.29-3.30, 3.34 and 3.35) and their age relationships with other intrusions (section 3.5.3) indicate that they formed at early stages of individual intrusive events within the pipes. The early stage of formation of the breccias is indicated in particular by their occurrence as preserved remnants under country rock overhangs and by truncation and coring out of the bodies by subsequent intrusions (sections 3.3.1, 3.5.2 and 3.5.3).

The occurrence of the breccias under overhangs and the presence of similar breccias as caps at the top of subsurface domes (e.g. Wesselton and Monastery pipes, section 3.2.4) strongly suggests that these breccias formed prior to breaching of the surface. The local derivation of fragments within the breccias and the limited extent of transport undergone by the breccias, either as a whole, or in terms of individual breccia fragments relative to one another, are features that are consistent with pre-breakthrough brecciation.

The intimate spatial relationships between the contact breccias and the pipes proper clearly indicate that the breccias must be associated with kimberlite volcanism. However, many of the breccias are entirely devoid of kimberlite (section 3.3.1). This clearly implies that the formation of at least some contact breccias must be associated with discrete vapour phases.

An association between vapour phases and the development of contact breccias in the root zones of the pipes prior to breaching of the surface implies also that some (early) kimberlite intrusions must have differentiated, to produce precursor gas phases, prior to the zone of pipe generation being reached. Wyllie (1980) has suggested that initial kimberlite intrusions might generate separate vapour phases by crystallization at depths of $\sim$90km. Alternatively vapour phases may not occur as separate entities but might occur as a gas cap at the head of an intruding, differentiated, probably lenticular, magma body (Clement, 1975). Irrespective of which situation pertains the problem is to deduce how subsurface brecciation might be initiated by, or result from, the ascent of a precursor vapour phase. Several possibilities can be considered.

In the first instance hydraulic ramming (Kents, 1964) of a precursor gas phase into the wallrocks might occur. Penetration would take place mainly along pre-existing discontinuities (e.g. joints, fractures, faults) and some wedging, breaking and dislocation would be anticipated. The direct pressures of vapour injection could not, however, account for several features of the KIMFIK contact breccias in particular:

(i) The extensive degrees of brecciation that are frequently apparent. In some instances considerable volumes of close-packed, fine breccia occur in which the vast majority of fragments measure only a few centimetres across (section 3.3.1).

(ii) The presence of considerable void space (perhaps 5-10 vol.% in some instances) which implies an equivalent volume increase (section 3.3.1).

(iii) The variable degrees of brecciation that characterize some contact breccias (section 3.3.1) such that pockets of extreme brecciation occur within less brecciated areas or even within relatively competent large blocks (section 3.3.1).

Features such as those noted above can be most readily explained if some form of pulsatory or intermittent advance of precursor vapour is postulated or, possibly, if interaction with meteoric water occurred. Intermittent brecciation by a precursor gas phase has been advocated by Gates (1959) to account for the Shoshone breccia pipes in Nevada and a similar mechanism has been proposed by Tweto (1951) to account for brecciation associated with sills in Colorado.

Insofar as the KIMFIK pipes are concerned intermittent advance of vapour phases presupposes that the country rocks (at root zone levels) are inhomogeneous with respect to physical properties so that alternating zones of lesser or greater permeability exist. For explanatory purposes (see below) varying lithologies or a situation of alternating zones characterized by different intensities of jointing, fracturing, bedding, or cleavage and different intrinsic rock strengths can be envisaged.

In terms of the environments postulated above consider a vapour phase migrating upwards under pressure, through a relatively permeable (perhaps highly fractured) zone. During upward advance the pressure of the vapour might wedge open existing fractures and perhaps initiate new ones (Anderson, 1979). Disruption and breaking of country rock blocks might well occur to some extent. Eventually, however, the rising vapour would reach a relatively impermeable zone (a structural or lithological trap zone). Such a trap might cause a halt in vapour ascent thereby allowing vapour pressure to increase. Subsequently, once sufficient pressure had built up, the impermeable or relatively impermeable zone could be explosively ruptured (brecciated) as previously contained vapour decompressed violently into the next relatively permeable zone. At a higher level the process could be repeated, possibly several times, assuming that repetitive geological environments were intersected.

The sequence of events postulated above is essentially that proposed by Wright and Bowes (1968) to account for subsurface explosion breccias. In terms of this genetic model (and if it is assumed that the vapour phase is underlain by progressively less differentiated magma) it would be expected that:

(i) There would be a correlation between the stratigraphic succession (or areas of relatively greater and lesser structural dislocation within the country rocks) and zones of most extensive brecciation, i.e. structural or lithological controls on the sites of brecciation would be apparent.

(ii) Explosive brecciation might be augmented by spalling, wedging, attrition and, possibly, rock bursting due to explosive decompression.

(iii) Much of the explosively produced breccia would be incorporated and transported upwards by gas and/or magma (the latter would be expected to surge up behind precursor vapour as a consequence of decompression). Remnants of breccia zones might be left essentially in situ along the margins of the developing pipes, particularly under country rock overhangs (as described in section 3.3.1).

(iv) Brecciation could occur intermittently over relatively long vertical columns because, after each stage of upward-migrating explosive brecciation, the precursor vapour phase would largely be retained at the head of the rising column. Furthermore this vapour could be supplemented by additional exsolved volatiles as magma rose behind the gas cap. Additional vapour might also be made available for higher level brecciation by the incorporation of meteoric water into the intruding systems. In the case of the Kimberley pipes such water is likely to have been contained mainly in fractures as the country rocks at root zone levels are relatively impermeable. It is, in fact, possible that local intersections of groundwater by rising vapour and/or magma are responsible, at least in part, for subsurface explosive activity. In terms of this theory there is no need for recourse to local structural or lithological traps to explain intermittent subsurface explosive activity.

(v) Kimberlite magma rising behind precursor vapour phases, in addition to incorporating breccia fragments, might also penetrate remnant breccia zones to varying degrees. The decreasing amounts of interstitial kimberlite (outwards towards the country rock) in some breccias are consistent with such penetration as are the outwardly decreasing degrees to which breccia fragments have been dislocated and transported (section 3.3.1).

(vi). The intensity of brecciation (degree of fragmentation) might vary according to the nature of the rocks at specific sites of brecciation (e.g. the extent of pre-existing fracturing or jointing would influence the extent of brecciation), the amount of vapour available, the local abundance of groundwater, and the intensity of decompression involved in breaking through trap zones if they are present.

It is apparent that the processes described above could account for many of the features of the root zones of the Kimberley pipes. In particular the strong structural controls on root zone morphology, the general irregularity of the root zones, the evidence for magmatic stoping and the presence of explosion and intrusion breccias (terminology of Wright and Bowes, 1963) are consistent with the foregoing genetic model. Also consistent with this model is the preservation of contact breccias under overhangs since it can be assumed that the central parts of successively formed breccia 'helmets' were incorporated and removed by the intermittently upward-moving columns. The occurence of 'blind' domes with breccia caps can also be explained in terms of this model. Upward migration of the vapour-magma column and concomitant explosive brecciation could, for example, be terminated by non-violent dissipation of the vapour phase along relatively open fractures, after earlier explosive brecciation at lower levels.

Although the processes described above provide explanations for several prominent root zone features several genetic problems remain unresolved:

(i) There is a space problem. All the processes described previously (e.g. wedging, spalling, magmatic intrusion, explosive brecciation) result in volume increases. In terms of the genetic model proposed above the increasing volume requirements of root zone formation would expectedly be taken up by distortions (e.g. arching or folding) of the surrounding country rocks. However, as noted in Chapter 3 such distortions have not been found although features such as minor up-arching may exist or may have occurred above the present land surface.

(ii) Although there are considerable stratigraphic variations in the country rock of the root zones of the Kimberley pipes (section 1.5 and Fig. 3.1) there is no close association between contact breccias and stratigraphic changes. It should be noted, however, that detailed mapping of country rock structures has not been undertaken hence there may be some correlation between contact breccias and different structural domains.

(iii) Where movement of remnant, kimberlite-free contact breccias has occurred it has taken place in a downward direction (section 3.3.1).

The problems listed above can best be explained if a degree of pulsatory intrusion is accepted, i.e. temporary magma withdrawal occurred at times during the formation of the root zones. Modification of the pipe-forming model in this way allows for the formation of some subsurface breccias by implosion rather than explosion. Thus situations are envisaged where, after hydraulic ramming of a precursor vapour phase into the country rocks (mainly along structural breaks), temporary withdrawal of the underlying magma occurred (cf. Kents, 1964). Such withdrawal would cause a pressure drop at the head of the intrusive system and, if sufficiently pronounced and rapid, explosive, inward decompression (implosion) of the vapour (previously rammed under pressure into the country rocks) would occur. Such implosion could cause major brecciation, probably augmented by subsequent failures (spalling, bursting) at temporary free faces. Temporary magma withdrawal could be caused by a concentration of intrusive activity in other parts of the same pipe or, perhaps, in neighbouring pipes or associated dykes and sills.

The subsurface implosion theory offers a ready explanation for several of the features displayed by contact features in the root zones of the Kimberley pipes. For example, varying intensities of brecciation, particularly on very small scales (section 3.3.1 and Fig. 3.23), could be related to variable degrees of original vapour penetration. Areas of country rock which were highly jointed or fractured or otherwise relatively porous would contain relatively large volumes of gas and, on decompression (implosion) would be relatively extensively brecciated. Areas where less vapour would penetrate would be broken up to lesser degrees. Such variations might occur on small or large scales.

Downward slumping of some remnant breccia masses is more consistent with temporary magma withdrawal and implosive depressurization than with intermittent, partly explosive, gas-magma advance. In addition, the lack of direct correlation between stratigraphy and zones of brecciation is more consistent with the implosion hypothesis unless, as suggested previously, breccia zones are directly related to the distribution of groundwater or fracture zones. Furthermore the implosion hypothesis overcomes the space problem to a considerable degree (although in a sense the problem is only removed one step further).

An anticipated result of the implosion hypothesis is that temporary arches would form at the top of intermittently advancing embryonic pipes and several vertical sections through the Kimberley pipes reveal the probable positions of such structures. In fact the country rock overhangs (with underlying, protected, breccia remnants) are interpreted as the remains of such arches (Figures 3.20-3.21, 3.30 and 3.35).

A feature of root zone development is that intrusion within these zones cannot have been particularly rapid. The irregularity of the zones (including their locally inclined character), the nature of the contacts and the character of the hypabyssal-facies kimberlites which they contain are entirely inconsistent with proposals such as that of McGetchin et al. (1973) who have predicted intrusion velocities (within the zone of pipe generation) approaching 400m/sec. In fact intrusion of magma in the root zones must have been intermittent and relatively slow. Furthermore in some (probably most) cases it has not been laminar but turbulent or convective as indicated in section 7.3.3. It is also evident from the nature of the kimberlites (hypabyssal-facies) and the extensive metasomatism of xenoliths that cooling and crystallization has been relatively slow (in comparison with diatreme-facies kimberlites, see section 7.3.1 and later discussion).

It is concluded that the root zones of the Kimberley pipes are formed by the complex, subsurface sequences of processes (including subsurface explosive and implosive activity) described in the foregoing discussion. It is suggested further that the embryonic pipes so formed must continue to develop upwards to very near surface, if the next stages of pipe formation are to proceed. It is likely that the upward extension of the embryonic pipes will be accompanied by increasing degrees of marginal and overhead brecciation; a function of increasing amounts of vapour being available for explosion and/or implosion (due to increasing exsolution of volatiles from underlying magma and/or increasing amounts of meteoric water being available for incorporation in the system at higher levels).

For ease of discussion the processes responsible for root zone (embryonic pipe) formation have been discussed in terms of a single (albeit in some instances discontinuous) intrusive event. Clearly this need not necessarily be the case. Thus an initial event might be responsible for an embryonic pipe reaching a certain level of development before activity ceased (perhaps due to escape of volatiles and crystallization of the magma). An irregular 'blind' pipe may thus be formed. In other instances subsequent vapour-magma systems, following essentially the same path, could result in further development of an embryonic pipe until a position close to surface is reached and other processes come into play.

The genesis of embryonic pipes as outlined above reflects a complex, often repetitive, sequence of events. An alternative, simpler, interpretation is to assume that the irregularity of root zones and the contact breccias in particular, represent post-breakthrough (rather than subsurface) processes. For instance it could be argued that decompression along the intrusion passageway, following explosive breakthrough at surface, would cause counter-explosions at depth. A venturi effect is implied. Volatiles rammed into the country rocks, prior to breaching of the surface, would implode back into the vent when explosive outburst caused a rapid pressure drop and contact breccias might be formed in this manner (the 'authigenic' breccias of Rust, 1937).

It is likely that authigenic brecciation occurred to some extent (particularly at relatively high levels) during the formation of the KIMFIK pipes (section 10.2.2) but, for several reasons, the contact breccias in the preserved parts of the root zones are not considered to have formed in this manner:

(i) The extent and distribution of authigenic breccias are very closely controlled by the stratigraphy and structure of the country rocks (Rust, 1937). Such control is not evident in the root zones of the Kimberley pipes.

(ii) Marginal or capping breccias occur above 'blind' domes or extensions of major pipes (e.g. at the Wesselton and Monastery pipes).

(iii) There is little doubt that explosive outburst took place from shallow levels (a few hundred metres at most) and there is no evidence of vents open to surface from depths of 2-3km (the pre-erosion depths of the root zones). Arguments against such vents were presented in sections 9.4 and 9.8.

(iv) If, in fact, open passageways were maintained, even temporarily, at depths of 2-3km they are likely to have been tabular in nature and of extremely limited extent, i.e. relatively narrow fissures. The extent and form of many of the contact breccias are entirely inconsistent with implosive activity into passageways of this nature.

(v) The arch configurations evident within the root zones are more consistent with the previously advocated subsurface processes.

(vi) Subsequent magmatic stoping and the formation of intrusion breccias at root zone contacts would not have occurred if a depressurized channelway to surface had already been established.

10.2.2 Explosion outburst: the formation of craters

It is contended that if the embryonic pipes continued to develop upwards a position would be reached where the pressure of the vapour cap exceeded the lithostatic load and explosive breakthrough to surface would occur. It is likely, based on the profiles of known crater zones (e.g. the Mwadui pipe in Tanzania, Edwards and Howkins, 1966), that breakthrough takes place from shallow levels (300-400m below surface at a maximum). It is also likely that in some (perhaps many) cases the explosions reflect the intersection of groundwater and hence are phreatomagmatic in character (as suggested by Lorenz, 1975, 1979).

All the main features of the crater zones are entirely consistent with explosion craters. These features include the relatively flat slopes, the brecciated contacts, the tendency towards circular shape and the presence of epiclastic kimberlite in sedimentary basins. This latter material is indicative of the former presence of tuff/agglomerate cones around the craters (Hawthorne, 1975). The presence of tuffs carrying accretionary lapilli (in the sense of Moore and Peck, 1962) indicates that back-fall of explosively ejected material occurred.

10.2.3 Diatreme zones: modified embryonic pipes

In order to assess the formation of diatreme zones it is worthwhile reviewing the probable nature of developing pipes immediately after explosive breakthrough. In terms of the model proposed here the pipes would at this stage occur (over most of their lengths) as rudely vertical (local inclinations may be present) bodies with highly irregular contacts which dipped outwards locally (breached arch areas). Extensive contact breccias would be preserved and authigenic brecciation and spalling and slumping after breakthrough would probably have augmented subsurface fragmentation of the wallrocks in the zone immediately below the explosion craters topping the embryonic pipes.

The development of diatreme zones is considered to reflect post-breakthrough modification of the basal parts of the crater zones and a considerable part of the upper sections of the underlying irregular pipes. Diatreme formation is ascribed to the development of vapour-liquid-solid fluidized systems (which changed rapidly to vapour-solid systems) from volatile-rich but already partly degassed magmas, as a consequence of explosive decompression when outburst occurred. As a consequence of vaporization and rapid adiabatic expansion such fluidized systems are considered to have evolved backwards down the embryonic pipes. During downward evolution the diatreme zones would be formed primarily by the incorporation of previously fragmented material surrounding the embryonic pipes into the fluidized (gas streaming) systems. Rare remnant contact breccias (e.g. Figure 3.25) are evidence of the former presence of brecciated country rock at diatreme levels but most high level contact breccias are considered to have been incorporated by fluidization into the pipes. Slumping and plucking of joint-bounded country rock blocks into the fluidized system would also occur but diatreme enlargement involving extensive tuffisitization caused directly by the intrusion of fluidized systems is not envisaged.

Floating reef bodies represent down-slumped masses of embryonic pipe side-wall material, either brecciated by previous root zone processes or by authigenic brecciation (in the sense of Rust, 1937) or in the form of relatively solid masses. Slumping is particularly likely to have occurred where overhanging sidewall situations had developed by the time of breakthrough. The marginal location of many floating reef bodies is consistent with slumping from and down the walls of the pipes. Disruption of many previously brecciated floating reef masses during downward slumping will have added many small xenoliths to the rapidly evolving TKB's (see discussion below). Furthermore mixing of xenoliths within TKB's will have been facilitated by derivation of many of the clasts at various levels during descent and break-up of brecciated floating reef. Grooved, striated or polished pipe contacts may reflect: the frictional effects of downward-slumping floating reef masses, or the effects of attrition by particles carried in vapour-solid fluidized systems, or deflation as intrusive activity waned, or secondary movements due to volume changes (mainly expansion) after post-emplacement alteration.

It is important to note that the formation of diatremes is not ascribed primarily to tuffisitization resulting from the erosive capabilities of particle-charged gas streams. The origin of the Swabian olivine melilitite diatremes by such a process was advocated by Cloos (1941) and later application of this process to kimberlite pipes by Dawson (1960, 1962, 1971, 1980) has received considerable support. Arguments against significant degrees of tuffisitization of this nature in the KIMFIK pipes have been put forward in section 9.10.

The TKB's in the KIMFIK pipes provide abundant evidence of fluidization processes and the textures developed within these rocks have been discussed at length in sections 7.2.1-7.2.3. It has previously been noted, however, that fluidization was probably short-lived (section

7.2.3) and several features of the KIMFIK TKB's support this view:

(i) The abundant country rock fragments in the TKB's are commonly angular to subangular whereas prolonged fluidization would result in considerable attritional rounding of xenoliths (section 7.2.1).

(ii) Although a considerable degree of mixing of xenoliths has occurred a crude vertical zoning of xenolithic material (particularly of floating reef) may be evident in TKB's. This indicates that fluidization was not sufficiently prolonged for complete homogenization with respect to xenolith distribution (section 7.2.1).

(iii) Tuffisitic veins penetrating or cutting the country rocks or floating reef bodies are rare (section 9.10).

(iv) The frequent occurrence of microlitic diopside as a post-fluidization mineral derived from vapour condensates (section 7.2.3) implies relatively high post-fluidization temperatures (almost certainly $\sim$500°C or more). Bearing in mind that magmatic temperatures would probably have cooled to around 800°C or less prior to the onset of fluidization (section 7.7 and Fig. 7.8), that adiabatic cooling would accompany fluidization, and that fluidized systems would also be cooled by an influx of relatively cool country rock fragments it is unlikely that fluidization could have been prolonged. Prolonged fluidization implies the attainment of temperatures that would be too low for diopside to crystallize.

It is apparent that, while mixing of material derived from considerable depths and from near-surface has occurred in the KIMFIK TKB's the dominant sense of movement is downwards. Thus in the Kimberley pipes TKB's with xenolith assemblages dominated by Karoo rocks occur adjacent to Ventersdorp or Basement wallrocks. Such situations indicate that, during the waning stages of fluidization, the condensing vapour phases act essentially as a buffering agents, allowing relatively gentle subsidence of solid material. Deflation of the systems probably occurred during condensation and crystallization which must have taken place rapidly to prevent close packing of solid material. The textures of the interstitial material in TKB's are consistent with quenching (sections 7.2.2 and 7.2.3).

10.3 GENERAL COMMENTS RELATING TO THE ORIGIN OF THE PIPES

10.3.1 Textural gradations

The downward development of vapour-solid fluidized systems should result in transitions between such systems and more 'magmatic-looking' material in the KIMFIK pipes. A good example is provided by the DB 5 intrusion in the De Beers pipe. At current mining levels this intrusion occurs as a volatile-rich, hypabyssal-facies kimberlite breccia with a pronounced segregationary texture (Plate 5.5h). Near surface Wagner (1914) has described DB 5 as an 'agglomeratic kimberlite tuff' and a photograph in his book (Plate 25/4) clearly shows that at this level DB 5 occurred as a TKB.

Another example is provided by the W5 intrusion at Wesselton. At depths of more than 800m rapid local transitions are evident between zones which are essentially hypabyssal-facies in character and others which are tuffisitic.

10.3.2 Mineralogical changes

As Dawson (1980) has pointed out the mineralogy of kimberlites will not only reflect intrinsic magma compositions but will depend on the mode of emplacement and the partial pressures of volatiles. Bearing this in mind it is evident that mineralogical variations might be expected to accompany the textural changes that reflect transitions from tuffisitic to hypabyssal-facies characteristics in individual intrusions. Such mineralogical changes are often shown, for example, by the presence of monticellite in the 'magmatic' parts of such intrusions or in pelletal lapilli that crystallized, at least in part, prior to fluidization (section 7.2.3). No examples of monticellite having crystallized during or after fluidization are known, probably because the volatile-rich nature of the systems renders monticellite unstable (cf. Janse, 1971).

The paucity of calcite in TKB's (section 7.2.1) is another noticeable mineralogical feature related directly to emplacement conditions. It is proposed (see also Clement, 1979) that the rarity of calcite reflects CO_2 and, possibly, CO loss during open system, vapour-solid, fluidization. Ca in quenched vapour condensates thus goes mainly into diopside microlites. The abundance of matrix serpentine in TKB's probably in part reflects the assimilation of considerable meteoric water into the open fluidized systems (Sheppard and Dawson, 1975).

10.3.3 Resorption, deuteric alteration and zoning of phenocrysts and macrorysts

A feature of phenocrysts and macrocrysts in some hypabyssal-facies KIMFIK intrusions is complex zoning, deuteric alteration and in some instances considerable resorption. Much zoning is complex and is the result of reactions between crystals and residual liquids (e.g. Boyd and Clement, 1977; Dawson, 1980; Pasteris, 1980a; Snowden, 1981). Deuteric alteration is particularly complex in respect of the olivine grains in certain kimberlites (section 5.2.3) which may show several serpentine zones and/or be replaced by several other minerals. The contacts between alteration zones are often sharp.

The complexity of these zonal reaction or replacement effects and the often sharp, rather than gradational, boundaries between compositionally or mineralogically different zones is difficult to explain if intrusion is viewed as a continuous process involving continuous modification of crystallization or reaction conditions. These features could more easily be explained in terms of the discontinuous processes envisaged during the formation of embryonic subsurface pipes. If crystallization or deuteric alteration was in progress at the time, abrupt changes in pressure and corresponding temperature changes related to intermittent, advancing, stages of pipe formation, would influence the nature of the crystallization and/or reaction products. Some resorption, particularly of phlogopite (sections 5.3.2, 5.3.3 and 6.3), might also be related to changing conditions during embryonic pipe formation. Overgrowths around certain minerals might be similarly explained. Pasteris' (1980a) conclusion that alteration of olivine is generally much less complex (not necessarily less severe) in dykes associated with the De Beers pipe than in the pipe intrusions themselves lends some support to the foregoing broadly speculative comments.

10.3.4 Pipes without root zones

In an early part of this thesis (section 3.2.2) attention was drawn to the fact that of all the major Kimberley pipes only Bultfontein is devoid of a typical root zone. Since the root zones are interpreted as the deeper remnants of embryonic irregular pipes a simple explanation is that vapour-solid fluidization (during diatreme formation) removed all vestiges of the embryonic Bultfontein pipe, i.e. a diatreme zone stretches from the feeder channel (section 4.3) upwards. Similar root zone features have been destroyed in parts of other KIMFIK pipes.

10.3.5 Multiple intrusions

Indisputable evidence that several, in some instances many, episodes of intrusion have occurred within the KIMFIK pipes has been presented in earlier chapters. In order to fully explain the origin of the pipes these discrete intrusive events must be taken into account. Broadly speaking multiple intrusions can be considered in terms of two emplacement situations. In the first instance successive intrusions may follow slightly different but adjacent pathways (at least at or near the zone of pipe generation). Alternatively later intrusions may be emplaced along previous intrusion paths.

Intrusion along different pathways leads to the formation of two or more root zone lobes (which may or may not be linked by dyke-like connections) and to coalescing diatreme or crater zones (Figures 3.26-3.27, 3.29 and 3.34). Coalescence of discrete intrusions implies partial coring out of an earlier kimberlite by a subsequent intrusion (section 3.5.3). Where identical intrusion paths have been utilized later intrusions may form plug-like bodies within earlier kimberlites (e.g. Figures 3.31-3.32 and 3.34) or may have entirely removed earlier intrusions at least within the exposed zones of the pipes. The latter conclusion is supported by the common occurrence of autoliths which cannot be matched with in situ intrusions (section 3.7).

Multiple intrusions imply repetitive intrusive/eruptive cycles of pipe formation and kimberlite emplacement but intrusion parameters are likely to have varied widely in response to many factors, e.g. bulk composition, degree of retention of volatile components, volume of magma, temperature, pressure, nature of the country rock, structural features, etc. Consequently it is unlikely that corresponding processes (e.g. explosive or implosive brecciation, magmatic stoping, vapour-solid fluidization, magmatic intrusion) were operative for similar lengths of time, with similar intensities, during successive emplacement episodes, at the same levels in the pipes. Variations in the foregoing respects could lead to diatreme and root zone characteristics being developed at the same level but in different parts of individual pipes and diatreme- and hypabyssal-facies kimberlites could occur side by side in complex pipes. Both the above situations are common features of the Kimberley pipes.

Variations in the character of different contact breccias and the age relationships exhibited between certain breccias and some kimberlite intrusions (e.g. Wesselton, Figure 3.30) indicate that subsurface brecciation was associated with more than one intrusive cycle in some pipes. There is, however, rarely any indication of contact brecciation where plug-like intrusions occur within and have cored out early kimberlite. A number of reasons can be advanced for the absence of such 'internal' contact breccias.

(i) Successive intrusions followed one another closely and the earlier intrusion was not consolidated, or was only partly consolidated, when the later intrusion was emplaced. Under such conditions the earlier intrusion could have been cored out relatively easily, mixed (incorporated) in the head of the succeeding intrusion, and carried upwards.

(ii) In several instances later plugs cut TKB's (e.g. Dutoitspan, Figures 3.31-3.32) which are relatively soft, inhomogeneous and partly clastic in character. These features probably facilitated TKB breakdown by wedging, slumping and plucking by rapidly intruding, open systems which were vapourizing at higher levels to produce complex fluidized systems. Alternatively coring out may have mainly been achieved by vapour-liquid-solid fluidized systems which developed sufficiently far backwards to establish a plug-like conduit. Subsequently, as fluidization waned rapidly (see comments in sections 7.2.3 and 10.2.3), hypabyssal-facies material from lower down in the differentiated column may have surged up contemporaneously to occupy the lower part of the conduit ('internal diatreme'). In some instances coring out, without substantional earlier subsurface brecciation, may have been facilitated by being initiated along contacts between intrusions or between kimberlite and country rock. Rapid intrusion (streaming) might be facilitated by relatively easier penetration and access to surface along such contacts.

(iii) Subsurface brecciation may have occurred but remnant breccias could have been removed by subsequent vapour-liquid - solid fluidized systems developing sufficiently far downwards to remove the breccias (as proposed to explain the configuration of the Bultfontein pipe, Section 10.3.4)). As in case (ii) above the lower part of the relatively smooth-walled conduits produced by multiphase fluidization might be filled by upwelling magmatic material.

As noted earlier variations in intrusion parameters would markedly influence the nature of individual intrusive/eruptive cycles. Mention has also already been made of incomplete cycles. Thus, for example, blind pipes may reflect non-violent dissipation of precursor gas through favourable channelways (thereby precluding explosive breakthrough). Many other variations are possible. For example, rapid quenching after explosive breakthrough might preclude or severely inhibit the post-breakthrough development of diatreme zones. Thus crater zones directly above irregular embryonic (root zone) pipes may occur in some instances although to the author's knowledge such situations have not yet been documented.

10.3.6 Duration of pipe formation

As noted in section 3.5.1 some intrusions in the KIMFIK pipes are separated by sharp contacts. Such contacts may imply that substantial intervals of time have elapsed between intrusions. In some pipes there is evidence that the time intervals between intrusions and associated volcanic activity must have been measured in at least hundreds and possibly thousands of years. This is indicated by the presence in primary kimberlite of down-rafted blocks of lithified epiclastic kimberlite (section 3.4.2 and Plate 7.16d). The presence of these sedimentary inclusions implies:

(i) Volcanic activity resulting in the development of a tuff ring around a vent.

(ii) Erosion of the ejected deposits and deposition of kimberlitic material in a sedimentary basin within the crater zone of the pipe.

(iii) Lithification of the sediments.

(iv) Disruption of the sedimentary basin by renewed intrusive and volcanic activity and down-rafting and incorporation of disrupted blocks within diatreme-facies kimberlite emplaced by the later volcanic activity.

The above sequence of events clearly indicates that substantial periods of time elapsed between some intrusions in kimberlite pipes but no data indicating quantitative periods of pipe formation are available. No current radiometric dating procedures appear to be sufficiently sensitive to distinguish discrete ages for individual intrusions.

10.3.7 The implications of precursor dykes

So far in this dissertation no comments have been made on what factors govern the precise location of kimberlite pipes. A common comment in the literature is that pipes are located at the intersections of fractures or fracture zones (Williams, 1932; Dawson, 1980), i.e. pipes form at points where upward access by the intruding systems is easiest. It can, however, be argued that easy access along fractures should result in the emplacement of dykes rather than pipes. Consideration of the role played by precursor dykes helps to explain this conundrum.

It was shown in sections 4.2 and 4.6 that the formation of many (all?) pipes is preceded by the intrusion of precursor dykes (and, in some cases, sills). Furthermore, although the dykes may form part of a regional network of hypabyssal intrusions they are most abundant and often reach higher levels of intrusion (some dykes are 'blind') in the immediate vicinity of pipes (cf. Dawson, 1960, 1962; this thesis, Chapter 4).

The concentration of precursor intrusions in the vicinity of pipes strongly suggests that the sites of pipes were areas of relatively abundant jointing or fracturing, i.e. areas which were relatively more favourable for kimberlite intrusion. It is, however, suggested that the formation of the pipes is not related to relatively 'easy' kimberlite emplacement but to the fact that, after following and establishing passageways in the near-surface environment, the precursor dykes and sills effectively 'sealed' the area. Subsequent pipe generation might be the direct result of such sealing processes. Sealing might, for example, be responsible for subsurface pressure build-up, for providing more time for the development of precursor vapour phases to accumulate and, consequently, for the initiation of subsurface embryonic pipe-forming processes.

It should be stressed that the control on kimberlite localization by fractures as suggested above is a near-surface phenomenon. It is not suggested that the dislocations concerned are deep-seated 'fundamental fractures' in the sense of some other authors (e.g. Dawson, 1971). The fractures referred to are considered to have absolutely no bearing on the actual generation of kimberlite in the upper mantle nor do they play a role in the intrusion of kimberlite through the mantle and lower crust. It seems much more likely that at depth kimberlite intrusions propagate their own passageways, probably in accordance with fracture mechanics theory as discussed by Anderson (1979). It also appears unlikely that dyke-like feeder channels, stretching over vertical distances of 100-200km, would remain open during and after intrusion. It is more probable that the channelways would close up behind individual intrusions. Anderson (op. cit.) has indicated the circumstances under which closing-up (downward pinching) might occur (such closing up might also in part explain how ultramafic nodules could be emplaced at relatively slow intrusion rates, i.e. without recourse to the speeds of ascent advocated by McGetchin et al., 1973).

10.3.8 The abundance of crustal xenoliths in TKB's

It is evident from Table 7.3 that small crustal xenoliths (<4mm to microscopic particles) and xenocrysts usually make up between 10 and 30 vol.% of TKB's. It is also apparent (Plates 7.1a-d and 7.10f) that a further 10-30 vol.% of the rocks consist of relatively large inclusions. In total, therefore, $\sim$30 to 50 vol.% of TKB's usually consist of incorporated country rock. This implies that in most cases greater than 50% of the material which originally occupied the diatreme zones of the KIMFIK pipes must have been ejected (minus any floating reef masses).

If, as postulated in section 10.2.2 explosive cratering took place from shallow levels (and there is abundant field and theoretical evidence for this, sections 3.4.2 and 9.4), the ejection of most of this material cannot be ascribed to initial explosive outburst. It must therefore be related to an initial upsurge and ejection of material immediately after breakthrough, i.e. during the early stages of fluidization (gas streaming involving pneumatic transport of solids). Under such circumstances it would be expected that decreasing proportions of country rock would be ejected as distance from the point of explosive break-out increased. The rarity of inclusions derived from the upper part of the Karoo Sequence (despite the evidence for a considerable degree of mixing of fragments from different stratigraphic horizons) in the preserved parts of the Kimberley pipes may in part be related to such differential losses of material.

The foregoing discussion provides further argument in favour of the conclusion that 'lean-phase' gas-solid fluidization (section 9.2) is a short-lived (although repetitive) process during the formation and infilling of kimberlite pipes (sections 7.2.3 and 10.2.3). Other considerations apart it is difficult to see how the necessary gas velocities would be maintained for substantial periods of time once major connections to the surface (explosion craters) had been established. Prolonged fluidization under such circumstances would require extremely rapid regeneration of extensive gas supplies, presumably by extensive unmixing of vapour from less and less differentiated magma immediately below the diatreme zones or, perhaps in part, by continuous and immediate groundwater recharge. Given the small volumes of kimberlite intrusions the first possibility appears remote and has been severely criticised by Lorenz (1975, 1979). The second possibility appears to be equally remote. Firstly there appears to be an insurmountable physical problem. How does sufficient water enter the system sufficiently rapidly to be vaporized and maintain fluidized conditions in major diatremes? Secondly, the nature of the post-fluidization interstitial groundmass in TKB's (mainly diopside and serpentine) is not consistent with a fluidization medium composed almost entirely of steam, i.e. the degree to which juvenile vapour phases are contaminated by groundwater must be limited.

In view of the previous comments it is concluded, therefore, that, following breakthrough to surface, initially high gas velocities would wane rapidly. Initial transport and/or ejection of fragments with attendant mixing (as a function of size, shape, density, time of incorporation and position of fragments in the system) would give way initially to buoyant mixing and then to buffering of generally downward-sinking fragments, prior to final quenching.

10.3.9 The distribution of intrusions in the pipes

Reference was made in section 3.5.2 to the tendency for greater numbers of volumetrically significant intrusions to occur in the root zones rather than in the diatreme zones of the Kimberley pipes. Commonly the situation reflected is that the last TKB(s) to have been emplaced have largely cored out previous intrusions within the diatreme zones (e.g. Figs. 3.30 and 3.35). Probable reasons for the relative ease with which earlier intrusions in the diatreme zones have often been largely or completed cored out include:

(i) Much of the earlier material probably consisted of TKB's which, as noted in section 10.3.5, are relatively soft, inhomogeneous, fragmental rocks which would be more susceptible to coring out by fluidized systems than hypabyssal-facies kimberlites.

(ii) Considerable authigenic brecciation may have been imposed on neighbouring, earlier intrusions by subsequent explosive outbursts. Such brecciation would facilitate incorporation and/or ejection of material by post-explosion fluidized phases. Because of their nature most inclusions of TKB's would probably break down rapidly in fluidized systems while inclusions of hypabyssal-facies material would be more resilient. The relative abundance of the latter autoliths and extreme paucity of the former in TKB's support this conclusion.

(iii) The explosive outbursts associated with the formation of later TKB's may take place from deeper levels than initial crater-forming explosions. This might be due, for example, to the less competent nature of the material within the partly formed pipes. Explosions from deeper levels would facilitate more extensive authigenic brecciation.

10.3.10 A note on internal dykes

A feature of many internal dykes is their limited extent and lenticular nature (section 4.4). Clearly small volumes of magma are involved in the emplacement of such dykes and it seems unlikely that such quantities of kimberlite melt would always be derived directly from the upper mantle as discrete intrusions. Alternative possibilities are that they represent volatile-rich precursors of more substantial intrusions or that they are derived from relatively shallow intermediate magma chambers, as suggested by Donaldson and Reid (in press). Such magma chambers may be ephemeral involving, perhaps, only the temporary halt of a more substantial intrusion from which a volatile-rich fraction separated (section 7.8.4). In fact separation may be triggered by the development of a volatile-rich top fraction and renewed intrusion of the residuum might be triggered by a later pulse of melt from below.

More substantial internal dykes in the KIMFIK pipes may have been emplaced directly from the mantle and some are probably the feeders of intrusions which expanded to form larger plug-like or pipe-like bodies above present levels of exposure.

CHAPTER 11

SUMMARY OF THE MAIN OBSERVATIONS AND CONCLUSIONS

In this final chapter the main observations recorded and conclusions reached during this study are summarized (chapter by chapter) in point form.

Chapter 1

This chapter is introductory in character and inter alia outlines the scope and objectives of the study and the geological setting of the KIMFIK kimberlite pipes.

Chapter 2

In this chapter the petrological status and classification of kimberlites in general are discussed. It is concluded that:

(i) Many existing definitions of kimberlites are inaccurate, incomplete, too restrictive or misleading and that kimberlites are best defined in terms of the petrological definition proposed by Clement et al. (1977). This definition, while emphasizing the characteristic (diagnostic) petrological and geochemical features of kimberlites, is sufficiently broad to encompass the extensive textural, mineralogical and chemical variations which kimberlites display.

(ii) The description and interpretation of kimberlites is facilitated if the components of these rocks are considered in terms of macrocryst (anhedral, cryptogenic, relatively large mineral grains, many of which are probably xenocrysts) and matrix (phenocrysts plus groundmass) minerals. Megacrysts (discrete nodules often >1 - 2cm in diameter) are sparsely present in many kimberlites.

(iii) The genetic implications of the extensive variability of kimberlite justifies the development of textural and mineralogical classifications within which different types of kimberlite can be grouped and related and against which other rock types can be compared.

(iv) For the purposes of this thesis the most useful mineralogical classification is that proposed by Skinner and Clement (1979). In this classification kimberlites are subdivided (and named) according to the relative abundances of five primary matrix minerals - diopside, monticellite, phlogopite, calcite and serpentine . Some other minerals may be sufficiently abundant to qualify, for classification purposes, as characterizing accessories. The advantages of this classification are that it is quantitative, it is based on primary mineral composition and it accomodates the full range of mineralogical variation exhibited by kimberlites.

(v) With minor modifications the textural classification devised by Clement and Skinner (1979) can be applied to the KIMFIK intrusions. In the modified classification several textural classes are recognised within each of three fundamental facies of kimberlite (hypabyssal-, diatreme- and crater-facies kimberlites). The different textural classes are primarily related to different modes of emplacement and particularly to whether explosive outburst with concomitant complex (vapour-liquid-solid) fluidization has, or has not, occurred.

Chapter 3

In this chapter the major megascopic features of the KIMFIK pipes are described. These observations highlight the geological complexity of the pipes and are particularly relevant to the proposed mechanisms of pipe formation advanced in Chapter 10. The main observations and conclusions are that:

(i) Prior to erosion three successive morphologically distinctive depth zones (respectively crater, diatreme and root zones) were present within the vertical extent of the KIMFIK pipes. Due to erosion only the root zones and lower parts of the diatreme zones are preserved (note, however, that the root zones of the Koffiefontein and Finsch pipes have not been exposed by mining operations and note also the comments in sections 3.2.2, 3.2.3 and 10.3.4 regarding the morphology of the Bultfontein pipe).

(ii) The root zones of the pipes are relatively irregular in form reflecting pronounced (often local) variations in contact dips and strikes from place to place. This irregularity is highlighted by the following features. The root zones (or parts thereof) may be locally inclined, they commonly split into discrete columns or channels and 'blind' root zone appendages may occur. Remnant contact breccias composed of locally derived country rock fragments (with or without interstitial kimberlite) occur and are commonly preserved under overhanging, in situ, country rock. The breccia fragments are dominantly small (<5cm), angular to subangular and have not been transported substantial distances within or by intrusive kimberlite systems. In many instances breccia blocks have remained in situ after fragmentation. Intrusion breccias and kimberlite/country rock stockworks occur locally at root zone contacts and minor dykelets and vein-like apophyses of kimberlite locally penetrate the wallrocks of the root zones. Stoped blocks of country rock are in places concentrated near root zone contacts. In general the contact features of the root zones indicate considerable control of morphological features by pre-existing planar discontinuities (mainly joints) in the country rocks.

(iii) Relatively many discrete intrusions occur in the root zones (compared with the diatreme zones) of the KIMFIK pipes. Characteristically these intrusions consist of hypabyssal-facies kimberlite. They vary widely in respect of texture, mineralogical composition and xenolith content and commonly exhibit flow textures, at least locally. The distribution of these intrusions and contact breccias in the root zones (e.g. Figs. 3.29 - 3.37) indicate that the latter were formed at relatively early stages in the evolution of the pipes.

(iv) Preserved root zones may extend vertically for distances in excess of 0,5km.

(v) The diatreme zones of the KIMFIK pipes are much more regularly shaped than the root zones and are more extensive, originally having extended vertically for estimated distances of 1 to 2km. The contrasting character of the diatreme zones (compared with the root zone) is illustrated by the following observations. The diatreme zones have vertical axes and megascopically smooth, steep (75°-85°), inward-dipping margins. Joint-bounded contacts are evident and the contacts may be grooved, polished or striated. Contact breccias are rare and no other structural deformation of the wallrocks (folding, faulting, arching or fracturing), imposed during kimberlite emplacement, is evident. Usually only one or few major intrusions are present, commonly in the form of diatreme-facies tuffisitic kimberlite breccias. In many instances these TKB's contain down-rafted floating reef masses which tend to be peripherally located within the diatreme zones and are commonly extensively brecciated (in many cases the rocks appear to have been shattered). In most instances brecciated floating reef masses are devoid of kimberlite but the latter may form an interstitial matrix, at least marginally. Tuffisitic kimberlite veins may locally cut floating reef masses. The contacts of brecciated floating reef masses may be gradational, reflecting increased assimilation of breccia fragments outwards as xenoliths in the host kimberlite. Different floating reef masses (within a single pipe) may, despite considerable subsidence (in some instances > 1000m), crudely reflect the original stratigraphic sequence of the wallrocks from which they are derived. Smaller xenolithic, xenocrystic and primary components of the KIMFIK TKB's are reasonably well mixed hence, on a gross scale, the rocks are fairly homogeneous in character over considerable (tens to hundreds of metres) vertical and horizontal distances. The abundant xenoliths in the TKB's are commonly angular to subangular; abrasional (or other) rounding effects are limited.

(vi) Unambiguous evidence for the former presence of craters at the top of some KIMFIK pipes is the presence of down-rafted, lithified, epiclastic kimberlite blocks in some intrusions. By analogy with less eroded pipes elsewhere it is concluded that the crater zones of the KIMFIK pipes were shallow features (a few hundred metres deep at most), that they had relatively flat marginal dips (50° - 75° or, possibly, even less), that the adjacent country rock was extensively brecciated, that the craters had regular (often circular) cross-sectional shapes and that they contained stratified and/or unstratified epiclastic (and pyroclastic) kimberlite. The epiclastic kimberlite must have been derived from the erosion of a surrounding tuff ring.

(vii) The original surface areas of the KIMFIK pipes (crater surfaces) are likely to have exceeded 60 ha and in one case (Finsch) may have exceeded 100 ha. The largest of the pipes (Finsch) measures only 17,9 ha at the present erosion level.

(viii) The KIMFIK pipes are in an overall sense much more complex geologically than has been indicated by previous investigators (e.g. Wagner, 1914; Williams, 1932). Over 20 discrete mineralogically and texturally distinctive intrusions are present in one of the pipes. Furthermore, the common occurrence of kimberlite autoliths, which cannot be matched with in situ exposed intrusions indicates that many intrusions have not been mapped (they are not preserved in the accessible parts of the pipes).

Chapter 4

Chapter 4 is concerned with kimberlite dykes and sills associated with the KIMFIK pipes. It is concluded that:

(i) Four main groups of dykes are associated with the pipes. These groups can be differentiated according to their relative ages of emplacement (in relation to the formation of the pipes) and/or according to their field relationships. The four groups are termed precursor dykes, contemporaneous dykes, internal dykes and cross-cutting dykes.

(ii) Precursor dykes are abundant in the vicinity of some of the Kimberley pipes. They occur wholly within the country rocks and were emplaced prior to the formation of the pipes, in steeply dipping (near-vertical), pre-existing joint planes. These dykes rarely exceed 0,5 to 2,0m in width and have strike lengths which vary from a few tens of metres to more than 0,5km. Precursor dykes occur singly, in parallel groups, as anastomosing individuals or they form reticulate networks of kimberlite along intersecting joint sets. Some precursor dykes are off-set in en echelon fashion. A number of these dykes do not reach the present surface and others pinch out downwards as well as upwards and laterally i.e., they are lenticular in character. The levels to which blind dykes have penetrated and the frequency of precursor dyke intrusion fall off with increasing distance from individual pipes.

(iii) Precursor dykes are composed entirely of hypabyssal-facies kimberlite and the dykes in the vicinity of individual pipes display considerable compositional and textural variation. This variation in character strongly suggests that more than one period of precursor dyke intrusion occurred and, in fact, some individual dykes are multiple intrusions.

(iv) The contemporaneous dykes represent off-shoots into the country rock of major pipe intrusions and hence were emplaced during the formation of the associated pipes. In rare instances direct connections between pipes and contemporaneous dykes are not evident (at current levels of exposure) although field relations indicate that such dykes must have been intruded between the onset and cessation of pipe formation (such dykes may be difficult to recognize and hence may be more abundant than currently supposed). Dyke-like off-shoots of pipe intrusions may have strike lengths in excess of 50m and may be several metres wide where they leave the pipes proper. In other instances they are insignificant features. Since they commonly occur as off-shoots of discrete pipe intrusions more than one episode of contemporaneous dyke intrusion has taken place.

(v) Contemporaneous dykes which are off-shoots of major intrusions are commonly hypabyssal in character but rare examples of diatreme-facies (TKB) dykes occur.

(vi) Contemporaneous dykes which are not obviously connected to pipe intrusions are generally similar to precursor dykes in character. They are recognized only because they are emplaced along rare country rock structures which were recognizably imposed by pipe-forming processes.

(vii) Internal dykes consist of hypabyssal-facies kimberlite and have been intruded transgressively within major pipe intrusions or along the contacts between major intrusions or along pipe-wallrock contacts. Internal dykes do not cut across pipe boundaries. These dykes are commonly of limited extent, display pronounced lateral lenticularity and little vertical continuity. Strike lengths of more than a few tens of metres are rare and few internal dykes have been traced vertically for more than 100 to 200m. The internal dykes in individual pipes vary substantially in mineralogical composition and several periods of intrusion can be deduced from their field relationships.

(viii) A major complex of kimberlite sills (40m below the present surface) extending over an area of at least 70 ha is cut by the Wesselton pipe. At least two ages of intrusion (associated with the injection of precursor dykes) have been recognised (Hill, 1977). The emplacement of the sills is directly related to the occurrence of dolerite sills in the Karoo Sequence country rocks (as suggested for other sills in the Kimberley area by Hawthorne, 1968). The dolerite formed a barrier against upward penetration of kimberlite magma, thereby imposing lateral spreading. Individual kimberlite sills range from a few centimetres to 5m in thickness. Other sills closely connected to the KIMFIK pipes are restricted to minor occurrences in the vicinity of the De Beers pipe.

Chapter 5

In this chapter the mineralogy of the matrices of kimberlites is described and reviewed with particular reference to examples from the KIMFIK pipes.

(i) Theoretically as many as four populations of olivine may be present in some kimberlite intrusions but characteristically the KIMFIK intrusions contain two olivine populations; relatively large (0,5 - 5,0mm), anhedral macrocrysts (Chapter 6) and smaller (commonly $<$0,5mm) euhedral phenocrysts.

(ii) Olivine crystals (macrocrysts and phenocrysts) in the KIMFIK intrusions have commonly been deuterically altered. The main deuteric alteration products are serpentine, calcite, phlogopite, magnetite, monticellite, diopside and, probably, some chlorite. Complex, commonly zonal, alteration is evident in many instances and may involve replacement by a variety of different minerals or repeated episodes of replacement by a single mineral (commonly serpentine). With rare possible exceptions alteration always involves an initial stage of serpentinization. The subsequent development of other minerals within olivine pseudomorphs involves reactions between previously formed serpentine and residual kimberlite liquids.

(iii) The development of brucite, bowlingite, goethite, iddingsite, quartz, baryte, pyrite and/or a variety of clay minerals within olivine pseudomorphs is ascribed to secondary (rather than deuteric) processes related to weathering, or to the introduction of metasomatizing fluids, or to extensive contamination of residual kimberlite liquids by meteoric water.

(iv) Phlogopite is often an abundant mineral in the matrices of KIMFIK kimberlites and occurs as subhedral to euhedral phenocrysts and microphenocrysts, as a primary groundmass mineral of diverse habit and variable grain size, in cryptocrystalline residual groundmass 'pools' (segregations), as overgrowths around phenocrysts and xenocrysts of phlogopite, as a deuteric alteration product of earlier-formed minerals and as a metasomatic mineral replacing xenocrysts and xenoliths. Two populations of primary phlogopite (phenocryst and groundmass phases) are common in individual KIMFIK intrusions. Phlogopite crystals are commonly zoned and are usually pleochroic. Unzoned crystals may exhibit normal or reverse pleochroism. Zoned crystals may display normal and/or reversed pleochroism. 'Coarse' groundmass plates of phlogopite commonly poikilitically enclose other primary (earlier-crystallized) minerals.

(v) Monticellite is an abundant groundmass mineral in some KIMFIK intrusions. It occurs as very small (< 0,01 - 0,05mm) primary, colourless, subhedral to euhedral crystals which commonly form closely packed, granular, groundmass aggregates. In other intrusions monticellite crystals occur as prominent oikocrysts in sieve-textured phlogopite laths. Typically minute 'bubbles' in the monticellite probably represent gas or liquid-filled cavities. Monticellite crystals commonly display patchy, poorly defined, zones of contrasting colour and relief, probably due to incipient alteration. Monticellite also occurs as a deuteric replacement product of serpentinized olivine and as a metasomatic mineral in altered xenoliths. Monticellite in the KIMFIK kimberlite is often extensively carbonatized or serpentinized.

(vi) Primary diopside is an abundant component of most intrusions in the Finsch pipe and occurs, generally to a lesser degree, in all the KIMFIK pipes. It occurs as phenocrysts and/or as a fine-grained groundmass mineral; in the latter situation commonly as slender prismatic or acicular crystals. Diopside is also a characteristic interstitial (post-fluidization) mineral in diatreme-facies kimberlites. In the latter intrusions it occurs as extremely fine-grained, acicular crystals (< 0,01mm long) or as microlitic material. Groundmass diopside may occur in distinct segregations, particularly in diatreme-facies kimberlites. Diopside also occurs as a deuteric mineral replacing earlier-crystallized minerals, notably serpentinized olivine. Metasomatic diopsidization of country rock xenoliths is also apparent in some KIMFIK intrusions.

(vii) Calcite is the dominant carbonate mineral in the KIMFIK intrusions and is a volumetrically important component of many of the hypabyssal-facies intrusions. It is a very rare component of the TKB's. Primary calcite commonly occurs as: anhedral to euhedral crystals in late-stage groundmass segregations; as anhedral, poikilitic groundmass minerals which rarely exceed 0,3mm in diameter; as minute, anhedral, interstitial, groundmass grains; or, rarely, as prismatic laths. A number of earlier crystallizing minerals are susceptible to carbonatization hence calcite also occurs abundantly as a deuteric mineral replacing, for example, diopside, monticellite, phlogopite, apatite and serpentine. Like the other groundmass minerals in the KIMFIK kimberlites calcite also occurs as a metasomatic mineral in altered xenoliths. Except when severe weathering has occurred alteration of calcite is generally limited to serpentinization.

(viii) Serpentine is an abundant component of most KIMFIK intrusions and occurs in a variety of guises. It is a common deuteric alteration product of most earlier-formed silicates, calcite and apatite and occurs as a primary mineral in many instances. Primary serpentine commonly occurs in irregular to globular segregations within which it is often associated with calcite. In some of these segregations serpentine occurs as structureless, virtually isotropic serpophite but in others it occurs as an extremely fine-grained (microlitic) felty or spherulitic material. Microlitic or massive serpentine also occurs interstitially between other components of the kimberlites. It is possible that much of the apparently primary, microlitic or massive serpentine that occurs in segregations or interstitially is, in fact, a serpentinous colloidal mesostasis or a mixture of serpentine and other hydrous magnesian silicates rather than serpentine sensu stricto.

(ix) In general the most abundant and consistently present accessory minerals in the KIMFIK kimberlites are apatite, perovskite and opaque oxides (mainly groundmass ilmenite and a variety of spinels). The spinels are often prominently zoned.

(x) Probable pseudomorphs of melilite are present in some KIMFIK intrusions but no unaltered melilite which can be positively identified has been found.

Chapter 6

This chapter is concerned with certain aspects of the nature and origin of macrocrysts and megacrysts. It is shown that:

(i) On the basis of size, shape, deformation features and chemical composition the vast majority of olivine macrocrysts are xenocrysts, derived from the disruption of upper mantle-derived peridotitic and pyroxenitic rocks. This conclusion is supported by the occurrence of ultramafic nodules in various stages of disruption.

(ii) Phlogopite macrocrysts vary considerably with respect to size, shape, colour, nature and extent of pleochroism, nature and extent of deformation features, type and extent of zoning, nature and degree of alteration and degree to which corrosion has occurred. Such variations may occur among the macrocrysts in individual intrusions indicating the presence of assemblages of mixed parentage. Phlogopite macrocrysts may originate in one or more of the following ways; they may (like olivine macrocrysts) be upper mantle-derived xenocrysts, they may be cognate xenocrysts (derived from previously consolidated kimberlites or from disrupted cumulates), they may be derived from MARID-suite nodules or other 'glimmerites' (Dawson and Smith, 1977), or they may be intratelluric phenocrysts corroded (rounded) during subsequent magma evolution and intrusion. Detailed geochemical investigations are required to try and distinguish between these provenances.

(iii) Ilmenite macrocrysts are common in the Kimberley pipes (forming up to approximately 1 vol. %) but are not present in the Finsch and Koffiefontein intrusions. These ilmenites vary widely in composition and form distinctive geochemical populations on a pipe to pipe basis and among intrusions in individual pipes (B A Wyatt, personal communication). The question of whether these ilmenite macrocrysts are phenocrysts or xenocrysts remains to be resolved and the possibility that the assemblages may include both phenocrysts and xenocrysts cannot be ignored.

(iv) Garnet macrocrysts are prominent accessory minerals in all the KIMFIK pipes. They occur as dispersed, anhedral, rounded grains or angular, broken grains rarely larger than 5mm in diameter, Complete or partial kelyphite rims are commonly present. Kelyphite may be due to isochemical, subsolidus reactions between garnet and olivine (e.g. Reid and Dawson, 1972), to mantle metasomatism involving the introduction of alkalies and volatiles from percursor kimberlite or other fluids into garnetiferous upper mantle rocks, or possibly, to reaction between garnet xenocrysts and kimberlite magma (the vast majority and probably all garnets in the KIMFIK kimberlites are considered to be xenocrysts).

(v) Other macrocrysts in the KIMFIK pipes (in the most common estimated order of abundance) are spinel, clinopyroxene, orthopyroxene and zircon. The presence of zircon in all the pipes is likely but has not been proven. The origin of spinel and zircon macrocrysts is in considerable doubt (they may be xenocrysts and/or phenocrysts). Clinopyroxene and orthopyroxene macrocrysts are probably xenocrysts. The paucity of orthopyroxene macrocrysts (probable xenocrysts) relative to other minerals of upper mantle peridotite suites is ascribed to extensive resorption by kimberlite liquids.

(vi) Megacrysts in the KIMFIK pipes have not been studied as part of this thesis investigation. They occur, however, as rare single crystals (commonly > 1 - 2cm in diameter) or monomineralic (in some instances bimineralic) aggregates in at least some of the KIMFIK intrusions. The most common megacrysts are estimated to be olivine, ilmenite, clinopyroxene, orthopyroxene, garnet and zircon but in many instances individual megacrysts are so rare that they are only found in bulk concentrates obtained from mining operations. As is the case with respect to macrocrysts the full suite of megacrysts is not always present in individual intrusions. Some megacrysts can be distinguished from macrocrysts (of the same mineral) by differences in chemical composition and more than one compositionally distinctive suite of specific megacrysts may be present in some kimberlites (e.g. Eggler and McCallum, 1976). Compositional distinctions between megacrysts and macrocrysts are not always evident, e.g. ilmenite macrocrysts and megacrysts cannot be distinguished according to their chemical compositions (Wyatt, personal communication). Controversy exists regarding the origin of megacrysts and hinges on the question of whether megacrysts are 'cognate phenocrysts' or xenocrysts.

Chapter 7

This chapter includes detailed petrographic descriptions of kimberlites in the KIMFIK pipes. These descriptions highlight the petrographic diversity of the rocks. Also discussed in this chapter are the origin of different textural varieties of kimberlite, the paragenesis of the primary minerals, the nature and sequence of deuteric alteration, the bulk compositions of a cross-section of KIMFIK kimberlites and certain genetic considerations arising therefrom. The main observations and conclusions reached are summarized below:

Diatreme-facies kimberlites

(i) Pelletal-textured TKB's are volumetrically the most abundant of all textural variants of kimberlite in the KIMFIK pipes and occur in the diatreme zones of the pipes.

(ii) The TKB's in the KIMFIK pipes consist of abundant country rock xenoliths (and xenocrysts), altered olivine macrocrysts (xenocrysts) and pelletal lapilli together with scattered autoliths, other members of the macrocryst suite and ultramafic nodules, all set in aphanitic, commonly altered, interstitial cementing material.

(iii) Country rock xenoliths in the TKB's are generally small (commonly from $\sim$2 - 3cm down to microscopic particles), little altered, commonly angular to subangular, heterogeneous, fairly well mixed assemblages. Most of these xenoliths have moved downwards in the pipes relative to their original stratigraphic positions.

(iv) Floating reef masses typically occur within the TKB's of the KIMFIK pipes.

(v) A crude, very poorly developed 'stratification' of inclusions and the large floating reef masses may be evident in some TKB's but generally there is little or no evidence of zonal distributions of xenoliths according to size or type.

(vi) Autoliths in the TKB's commonly reflect more than one kimberlitic source but are almost invariably hypabyssal-facies in character.

(vii) Pelletal lapilli in the TKB's are characteristically spherical or spheroidal and commonly measure between 0,005 and 5mm in diameter (many, in fact, conform to ash particles in size). Few pelletal lapilli exceed 10mm in diameter. Most pelletal lapilli consist of relatively large kernels of rock fragments or, more commonly, mineral grains (mostly altered olivine) surrounded by thin mantles of fine-grained to cryptocrystalline kimberlite material (mainly diopside, phlogopite and secondary clay minerals with lesser serpentine). Monticellite and melilite(?) are abundant components of a minority of pelletal lapilli. The mantles surrounding the kernels may be incomplete. Many mantles are porphyritic and contain a variety of phenocrysts the most abundant of which are altered olivines. The kernels of the lapilli are commonly more-or-less centrally located but eccentrically located kernels occur and some lapilli have no kernels. The phenocrysts in pelletal lapilli may be concentrically oriented and phenocryst -free lapilli may have mantles displaying vague concentric layering. Gradations exist between spherical pelletal lapilli displaying well developed concentric structures and irregular lapilli devoid of any concentric features. Accretion of pelletal lapilli is rarely evident, the lapilli are not closely packed and size-sorting of lapilli is not apparent.

(viii) The interstitial material which cements xenoliths, xenocrysts, discrete 'coarse' kimberlite minerals, autoliths and pelletal lapilli in the KIMFIK TKB's commonly consists of serpentine and clay minerals with extremely fine-grained to microlitic, acicular diopside. The latter mineral is often extensively altered and was probably more abundant in most of the TKB's than is indicated by optical examination. Interstitial diopside and serpentine are interpreted as primary minerals although their compositions may reflect modification of residual liquids (gas condensates) by meteoric water and other contaminants.

(ix) Acicular (microlitic) diopside commonly forms segregations within the interstitial matrices of the TKB's and is often clustered around (and replaces to different degrees) pelletal lapilli. Delicate acicular diopside crystals in these clusters may project radially outwards into surrounding serpentine.

(x) It is concluded that the pelletal lapilli in the KIMFIK TKB's consist of variable proportions of lapilli that represent magma droplets quenched in fluidized systems (following explosive disruption of magma) and others that reflect variable degrees of crystallization prior to explosive outburst and the concomitant development of vapour-liquid-solid fluidised systems.

Hypabyssal-facies kimberlites

(i) Hypabyssal-facies kimberlites and kimberlite breccias occur characteristically in the root zones of the pipes but are also commonly present as minor intrusions (mainly dykes) cutting TKB's or, less commonly, as cored-out remnants of major intrusions within the diatreme zones.

(ii) Country rock xenolith assemblages in the hypabyssal-facies intrusions tend to be volumetrically less abundant, less heterogeneous and less variable in size than the xenoliths in the TKB's. The xenoliths in the hypabyssal-facies intrusions are commonly extensively metasomatized ('kimberlitized') and/or magmatically corroded (in contrast to the xenoliths in the TKB's). These features of the xenoliths reflect the volatile-rich nature and slower cooling of the magmas which crystallize as hypabyssal-facies kimberlites.

(iii) Xenoliths in the hypabyssal-facies kimberlites may be orientated by magmatic flow, in some instances with attitudes which suggest that convective overturn occurred during intrusion.

(iv) Some hypabyssal-facies intrusions are devoid of inclusions derived from the Karoo Sequence. This situation suggests that some intrusions were 'blind' and failed to penetrate the Karoo stratigraphy.

(v) The hypabyssal-facies intrusions in the KIMFIK pipes display extensive mineralogical diversity reflected mainly by the presence or absence and relative abundances of groundmass diopside, monticellite, phlogopite, calcite, serpentine, apatite, opaque oxides and perovskite, by variations in xenolith types and abundances and by variations in textural features. Mineralogical variations are evident within individual intrusions, within individual pipes, among different pipes and on a regional basis (e.g. between the Finsch and Kimberley pipes).

(vi) The most prominent macroscopic textural feature of the hypabyssal-facies kimberlites is their macroporphyritic (macrocrystic) texture. The matrices of these rocks are microporphyritic. Xenoliths, xenocrysts and phenocrysts may display parallel or subparallel orientation imposed by magmatic flow (e.g. trachytic texture). Groundmass minerals do not exhibit such orientation textures because the major groundmass minerals crystallize after intrusion. The groundmasses of hypabyssal kimberlites may, however, be characterized by granular, intersertal, intergranular, poikilitic, mosaic, quench or segregationary textures.

(vii) Segregations composed of late-crystallizing, relatively low temperature, minerals are common in many hypabyssal-facies intrusions in the KIMFIK pipes. The segregations are not ascribed, as has previously been suggested (e.g. Mitchell and Fritz, 1975; Clement, 1975; Mitchell,1979), to carbonate-silicate liquid immiscibility. Some of these segregations are gas cavities infilled by residual melt (i.e. segregation vesicles), others are direct accumulations of residual melt and a third group are derived directly from exsolved vapour phases (by condensation and crystallization).

(viii) Globular segregations which closely resemble the pelletal lapilli present in KIMFIK TKB's are present in some of the hypabyssal-facies kimberlites. They consist of segregations of relatively high temperature minerals set in a matrix (host) of low temperature minerals derived from exsolved volatiles. The formation of globular segregations is related to the development of vapour-liquid-solid fluidized systems, under more-or-less closed (hypabyssal) conditions. The fluidized cells are considered to have been turbulently, convective systems within which disrupted magma 'fragments' developed into globular segregations.

(ix) Petrographic descriptions and modal analyses of hypabyssal-facies kimberlites in the KIMFIK pipes indicate that olivine (as phenocrysts and macrocrysts) is a ubiquitous and abundant mineral, that phlogopite macrocrysts/phenocrysts are often present but rarely in major modal amounts, that other macrocrysts are rare, that diopside phenocrysts are rare, that diopside, monticellite, phlogopite, calcite or serpentine may be the most abundant groundmass mineral in the rocks, that diopside-phlogopite kimberlites predominate in the Finsch pipe while a greater variety of mineralogical types occur in the Kimberley pipes, that many kimberlite dykes are calcite-rich, and that the apatite and opaque oxide contents of some KIMFIK intrusions form significant proportions of the rocks.

Paragenetic considerations

(i) The paragenesis of the primary minerals in the KIMFIK pipes involves two essentially discrete crystallization periods; pre-intrusion and post-intrusion crystallization periods are recognised. The primary minerals either crystallize in the upper mantle or in near-surface environments.

(ii) Olivine (phenocrysts) is the major mineral to crystallize in the upper mantle and may crystallize in discrete magma batches that are subsequently mixed. Other pre-intrusion minerals are phlogopite, ilmenite (as inclusions in olivine phenocrysts and, possibly, as phenocrysts), rutile (as inclusions in olivine phenocrysts), some spinel grains and, rarely, diopside phenocrysts.

(iii) Olivine phenocrysts probably crystallize before olivine xenocrysts are incorporated in kimberlite magmas. The latter event probably occurred just before or, more likely, when relatively rapid intrusion of magma to surface began. Other xenocrysts derived from upper mantle ultramafic rocks were probably incorporated more-or-less contemporaneously.

(iv) If ilmenite macrocrysts are, in fact, phenocrysts they probably also crystallized from discrete magma batches and were emplaced shortly after mixing or were mixed during intrusion (in this way, as in the case of olivine phenocrysts, reaction zoning of compositionally different grains would be inhibited).

(v) The major groundmass minerals (diopside, monticellite, phlogopite, calcite and serpentine) are post-intrusion minerals, together with a number of accessory groundmass minerals (e.g. apatite, perovskite, some spinels, ilmenite and sulphides).

(vi) A general crystallization sequence for hypabyssal-facies kimberlites (which ignores substantial crystallization overlaps and thus in effect refers to the onset of crystallization of each mineral) is given below:

(a) Pre-intrusion: ilmenite, Cr-Mg-Al spinels, perovskite, rutile; olivine; diopside; phlogopite.

(b) Post-intrusion: ilmenite, Ti-Mg chromite, Ti-Mg-Al chromite, magnetite, perovskite; diopside; monticellite; apatite; phlogopite; calcite and serpentine.

(vii) The foregoing paragenetic sequence is not applicable to diatreme-facies kimberlites. The crystallization sequence in such rocks is affected by loss of volatiles and, probably, by the incorporation of contaminants during fluidization. The main effect is a delay in the crystallization of groundmass diopside which crystallizes mainly after minerals such as monticellite and phlogopite.

(viii) Deuteric replacement of other components by the post-intrusion groundmass minerals in the hypabyssal-facies kimberlites generally occurs more-or-less contemporaneously with primary crystallization of the mineral or minerals concerned (if, in fact, primary crystallization occurs at all).

Geochemistry

(i) 53 New whole rock analyses (incorporating trace element data for 43 analyses) are presented and indicate wide ranges in the abundances of major oxides and most minor elements. Extensive geochemical variation is not only evident among the pipes but also within individual pipes and among specific types of geographically associated intrusions such as, for example, precursor dykes in the Kimberley area.

(ii) Diatreme-facies kimberlites are contaminated by foreign material to an extent that precludes the determination of meaningful bulk compostitions. Reliable analyses can, however, be obtained from hypabyssal-facies kimberlites and kimberlite breccias (after removal of xenoliths), although the effects of the hybrid, contaminated nature of kimberlite per sê cannot be entirely overcome.

(iii) Application of a 'contamination index' ($\frac{SiO_2 + Al_2O_3 + Na_2O}{MgO + 2K_2O}$) to published kimberlite analyses indicates that many of the analysed rocks must have been badly contaminated or highly weathered kimberlites.

iv) Trace element concentrations in general confirm earlier views (Dawson, 1962, 1967b) that two groups of trace elements can be recognized. One group contains trace elements in amounts that are characteristic of ultramafic rocks. The other group contains trace elements that are significantly enriched relative to their abundances in other ultramafic rocks.

(v) Harker-type variation diagrams using MgO wt. % as a differentiation index are presented for analyses of kimberlites within the Kimberley group. With respect to major elements these diagrams indicate negative correlations between MgO and TiO_2 Al_2O_3, Cao, K_2O, CO_2, L.O.I. and Fe_2O_3. With respect to trace elements negative correlations are apparent between MgO and Zn, Cu, Zr, Ba, Sr, V, Rb, Nb and Cr. Positive correlations exist between MgO and SiO_2 and MgO and FeO. Strong positive correlations are evident between MgO and Ni and MgO and Co.

(vi) It is suggested that the bulk composition variation trends displayed by the KIMFIK pipes may be due to the mixing of magma batches which had probably been subjected to varying degrees of olivine and orthopyroxene enrichment, by the addition of 'average' upper mantle (before and/or after mixing). Further olivine enrichment (involving olivine phenocrysts) probably took place as a result of batch-mixing and degassing. The latter processes may also have occurred before and/or after mixing.

Chapter 8

In this chapter some aspects of the distribution of diamonds in the pipes are illustrated by references to the De Beers, Wesselton and Dutoitspan occurrences. It is shown that:

(i) Discrete intrusions in individual pipes may have markedly different diamond contents.

(ii) Diamond grades within individual intrusions may be relatively constant from place to place, may vary considerably in apparently random fashion or may decrease regularly with increasing depth.

(iii) TKB's tend to have more regular diamond grades than hypabyssal-facies kimberlites and kimberlite breccias. The more consistent grades of TKB's are ascribed to mixing of the constituents, including diamonds, during fluidization.

(iv) Grade variations within and among the KIMFIK intrusions are not related to resorption of diamond before, during or after intrusion.

(v) If the diamonds in the KIMFIK pipes are upper mantle - derived xenocrysts then grade differences between intrusions could be related to differential sampling of diamondiferous upper mantle sources. Different intrusions may sample different mantle sources with different diamond contents or successive intrusions may sample the same source less and less effectively. Sampling may become less effective due to intrusion becoming progressively easier (along pre-established passageways). Alternatively sampling effectiveness may be maintained but the sampled area may become more and more depleted in diamond. Such sampling situations may also explain instances of linear decreases in grade with depth in individual intrusions. In other instances diamond distributions probably reflect varied degrees of mixing during intrusion. Such mixing may involve more than one magma batch or may relate to mixing, due to turbulence or convection, within individual batches.

Chapter 9

In this chapter theories relating to the origin of kimberlite pipes and other diatreme and breccia pipes are reviewed. In addition, because of the postulated importance of fluidization as a kimberlite pipe-forming process, the theoretical concepts and geological products of this process are examined in detail. Based on this review and observations made during the course of this study it is concluded that none of the published theories of the origin of kimberlite pipes can be satisfactorily applied to the KIMFIK pipes. The diversity of published views probably stems from the complex nature of many pipes, coupled with the fact that most authors have had insufficient access to well exposed pipes to allow a sufficiently broad overview to be developed. One of the main conclusions of the review is that the importance of the role hitherto ascribed to fluidization in the genesis of pipes can be seriously questioned.

Chapter 10

In chapter 10 information obtained during the course of this investigation of the KIMFIK pipes is utilized to develop a general theory of kimberlite pipe formation. This theory takes into account the geological complexities (in respect of morphology, contact features, petrography and internal structures) of the pipes and recognizes that a variety of processes must have operated during the complex evolutionary histories which these bodies have undergone. The main elements of this theory are summarized below.

To account for root zone features emphasis is placed on the importance of a variety of interrelated subsurface processes which result in the initial formation of embryonic pipes. The embryonic pipes develop (migrate) upwards and culminate in explosive breaching of the surface (which may, at least in part, be phreatomagmatic in character). The subsurface processes involved include hydraulic fracturing and wedging, magmatic stoping, brecciation by intrusion of magma along discontinuities in the wallrocks, intermittent explosive and/or implosive brecciation, spalling, slumping, and possibly, rock bursting from temporary free faces. Some of the subsurface processes noted above require that differentiation of rising magma occurred to produce precursor gas phases at the head of at least some intrusions, before the zone of pipe generation was reached. Embryonic pipes are considered to have developed upwards to within a few hundred metres of the land surface.

It is suggested that the subsequent development of kimberlite pipes is activated by explosive breaching of the surface and the consequent formation of explosion craters. The vertical profiles of preserved crater zones indicate that breakthrough takes place from shallow depths (commonly <300 - 400m).

Explosive breakthrough is likely to have been accompanied by authigenic brecciation at levels below the base of the crater zones. Such brecciation would enlarge and intensify the zones of broken and brecciated wallrock which resulted from the prior formation of embryonic pipes by root zone (pre-breakthrough) processes.

The development of diatreme zones is ascribed to post-breakthrough modification of the basal part of the crater zones and considerable parts of the extended root zones, i.e. the embryonic pipe columns. The rapid depressurization accompanying explosive breakthrough would result in an upsurge of precursor gas phases and partly degassed magma and the nature of the TKB's in the diatreme zones of pipes indicates the development of vapour-liquid-solid fluidized systems at this stage. As a consequence of vaporization and rapid adiabatic expansion these systems are likely to have evolved backwards down the embryonic pipes. During this process diatreme zones would be formed, mainly by the incorporation into the fluidized systems of previously brecciated material which formed the sidewalls of the embryonic pipes. Relatively smooth-walled, regular-shaped conduits would result. In rare instances all remnants of the embryonic pipes (i.e. root zones) might be removed, as is suggested in the case of Bultfontein.

The fluidization events are considered to have been short-lived. This is indicated by several features of diatreme zones and the TKB's within them. These features include the angularity of many country rock inclusions in TKB's, the lack of abrasion marks on the surfaces of xenoliths and the presence of a relatively high temperature, primary, mineral such as diopside as a microlitic quench product, derived from vapour condensates, in the matrices of TKB's.

Following the cessation of fluidization, concomitant deflation and interstitial consolidation, the final stage of the formation of kimberlite pipes is the deposition of epiclastic kimberlite (derived from the erosion of tuff rings) in the crater zones.

Most major kimberlite pipes reflect multiple intrusive events hence the cycle of formation summarized above may have been repeated several (in some instances many) times. However, intrusion parameters are likely to have varied considerably in respect of each cycle of events hence the relative importance and duration of each of the processes involved is likely to have varied and individual cycles may not always have continued to completion. Thus non-violent dissipation of precursor gas through favourable channelways, might result in crystallization prior to breakthrough and the formation of fluidized systems. Furthermore variations in intrusion parameters may result in diatreme and root zone characteristics being developed at the same level in different parts of a pipe.

NOTE ADDED IN PROOF

It has been suggested (M E McCallum, personal communication) that objections may be levied at the use of the term 'pelletal lapilli' to describe the ovoid bodies in TKB's which have also been termed 'magma pellets', 'autoliths' or 'nucleated autoliths' (Section 7.2.2 and 7.2.3). It is likely that such objections would primarily be based on the sizes of the bodies and assumptions that 'lapilli' must be pyroclastic bodies, i.e. volcanic ejectamenta.

In terming these bodies 'pelletal lapilli' the following considerations were taken into account:

(i) There is little point in introducing new petrological terms if precedents can be followed. 45 Years ago Rust (1937) described identical spherical bodies from lamprophyric diatremes in Missouri. He used the terms 'pellets' or 'lapilli' (commonly with the adjective 'explosive') to describe these bodies and suggested processes of formation not too dissimilar from those advocated in this thesis. His terms have been combined to 'pelletal lapilli' which aptly describes the bodies.

(ii) There is little justification for placing much emphasis on the size of pelletal lapilli. According to the AGI Glossary of Geology (Gary et. al., 1977) 'lapilli' is a non-genetic term applied to (pyroclastic) rock fragments which have been variously defined as measuring between 1 and 64mm in diameter. Some definitions place the lower cut-off at 4mm, others at 2mm. Such limits may be convenient for screening analyses but have little genetic importance. Pelletal lapilli very tremendously in size (from <0,1 - >10mm) hence could be termed mixtures of pelletal lapilli and pelletal ash grains. However, many pelletal lapilli lie within the defined size ranges of volcanic lapilli and it is these relatively large bodies which, on hand specimen or larger scales, commonly impose a noticeable characteristic on the TKB's in which they occur.

(iii) Statements that all 'lapilli' must by definition, be pyroclastic ejectamenta do not appear to be based on a full appreciation of the term. Rust (op. cit.) clearly indicated that the lapilli in Missouri diatremes formed within the diatremes and he implied that many never reached the surface. A similar use of the term 'lapilli' is evident in modern texts. For example, Williams and McBirney (1979) refer to identical bodies in the Montana kimberlite pipes described by Hearn (1968) as 'spheroidal and ovoid lapilli'. The former authors state further that ....' the lapilli formed underground'. It is true that most definitions of 'lapilli' state or imply that all such bodies are ejectamenta and as a general case this is correct. However glossary-type definitions are general cases catering for general situations, pedantic application of such definitions should be avoided.

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