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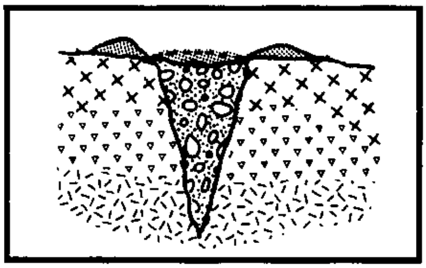

Kimberlites and Lamproites: Primary Sources of Diamond

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INTRODUCTION

A variety of mantle-derived igneous rocks comprise the primary sources of diamond, with the principal hosts being kimberlite and lamproite. Primary diamonds or graphite pseudomorphs after diamond are also known to occur in some lamprophyres (Jaques, Kerr *et al.*, 1989), alkali basalts and alpine type peridotites (Kaminskii, 1984). Significant quantities of diamond have not yet been found in these rocks.

Secondary diamond deposits are formed from these primary source rocks by weathering and transportation. These deposits are commonly very rich in high-quality diamonds. Examples include those of the Ural Mountains (USSR), the marine deposits of Namibia and the alluvial deposits of West Africa, Brazil and Venezuela. The identity of the rocks which were parental to these types of deposits is not always evident from the mineralogy of the detrital phases present. The nature and origins of secondary diamond deposits are not discussed further in this work.

Currently, diamonds are extracted from both kimberlites and lamproites and most exploration activity for diamond is directed toward the discovery of further exploitable deposits in these rocks. For these purposes, it is important to be able to determine rapidly the correct identity of a potentially diamondiferous rock, as exploration and assessment techniques for kimberlites and lamproites are different. Determining the correct identity of such rocks in some instances is not trivial or easy, as many rocks belonging to different petrological clans are petrographically similar.

Identification of a rock as kimberlite or lamproite does not guarantee that that it will contain economic amounts of diamond. There are two reasons for this observation. Firstly, it is now accepted that diamond is a *xenocryst* in both rock types. Secondly, magmatic processes may act to resorb and eliminate any entrained diamonds. Thus, a given barren intrusion may never have contained diamonds, due to failure to incorporate xenocrysts or any originally present may have been completely resorbed during emplacement and cooling of the magma. *It is particularly important to realize that kimberlites and lamproites are merely vehicles which transport diamond from the upper mantle to the crust.*

Detailed discussion and description of current hypotheses regarding the origin of diamond are beyond the scope of this work. However, some understanding of diamond genesis is essential to appreciate the distribution of diamonds in the primary source rocks.

DIAMOND GENESIS

It has long been known that primary diamonds are not identical in composition and/or morphology. For example, the presence or absence of nitrogen has led to the recognition of two major groups of diamonds, termed type I and II respectively. Diamonds also exhibit an extremely wide range (+5 to −35‰ $\delta^{13}C$) in their carbon isotopic composition (Harris, 1987).

Morphological differences, *i.e.*, octahedral *versus* hexahedral habits, may result from formation under different PTX conditions or in different environments, *i.e.*, solid state porphyroblastic growth *versus* precipitation from a magma.

Diamonds exhibit a wide range in size. Those which are smaller than 1mm in their maximum dimension are referred to as microdiamonds, and those larger than 1mm are termed macrodiamonds. Rare megadiamonds (>100 CM)[1], such as the Cullinan diamond (3106 CM), are also found. Most studies of diamond have been undertaken upon macrodiamond populations. It should be understood that hypotheses deduced for the origin of this group may have no relevance to micro- or megadiamonds.

The compositional and morphological differences noted above are so profound that it must be conceded that several diamond-forming processes must exist. The discussion below is concerned primarily with type I macrodiamonds.

For many years, it was believed that diamond was a phenocryst in kimberlite. However, hypotheses of diamond genesis were revolutionized with the discovery that diamonds are older than their host rocks. Richardson *et al.* (1984) determined that the Sm-Nd model age of inclusions in diamonds in the 90-100 Ma Finsch and Kimberley kimberlites (South Africa) was 3300 Ma. These results confirmed earlier, less definitive U-Pb studies by Kramers (1979) and demonstrated the antiquity of diamonds and presumably their mantle sources. Subsequent studies of inclusions in diamonds from the Premier (South Africa) kimberlite and Argyle (Australia) lamproite also gave Sm-Nd ages in excess of the age of these intrusions (Richardson, 1986). These data imply that diamonds cannot be phenocrysts in kimberlite or lamproite, and that their origins must lie within rocks sampled by these magmas during their ascent toward the crust.

Additional evidence regarding the origin of diamonds has stemmed from studies of upper mantle-derived xenoliths and mineral inclusions in macrodiamonds.

During the past two decades, detailed studies of upper mantle-derived xenoliths found in kimberlites and basaltic rocks have enabled petrologists to develop models of the petrological structure of cratonic regions. Reviews of this work may be found in Nixon (1987) or Harte and Hawkesworth (1989). Figure 1 illustrates a hypothetical cross-section of an Archean craton and adjacent Proterozoic mobile belt. The principal feature of this model relevant to diamond genesis is the

[1] CM = metric carat; 1 CM = 0.2 g

presence beneath the craton of a keel of rigid lithospheric mantle. The boundary between this keel and the underlying asthenospheric mantle acts as a major discontinuity, which separates mechanically and chemically disparate regions of the mantle. The boundary acts as a focal point for diverse reactions involving ascending magmas or fluids, and as a potential site for the underplating of subducted material. The boundary may lie at depths of 200-250 km (Nixon and Davies, 1987) or as deep as 400-600 km (Jordan, 1978). The lithospheric mantle is depleted in basaltic components and is believed to consist of spinel and garnet lherzolite, harzburgite and dunite. Scattered throughout this laterally and vertically heterogeneous assemblage are eclogitic rocks, which may represent either basaltic magmas crystallized at high pressure or remnants of ancient subducted oceanic basaltic rocks. The asthenospheric mantle is believed to be relatively homogeneous and to consist of convecting mantle material. This material has the potential to generate mid-oceanic ridge type basalts and the rocks are considered to be "fertile" in contrast to the "barren" lithospheric mantle. Partial melting of rising plumes or diapirs may give rise to voluminous melts which may pool at the lithosphere-asthenosphere boundary or erupt as continental flood basalts. Asthenosphere-derived melts may interact with the lithospheric mantle during their transit through this material.

Diamonds commonly contain small inclusions of silicates, oxides and sulphides. These inclusions are interpreted to be samples of the material which co-existed with the diamond during its growth. Reviews of this topic are given by Meyer (1987) and Gurney (1989). These studies have demonstrated the existence of two principal groups of inclusions, termed the peridotitic and the eclogitic suites. The inclusions of the peridotite suite consist of Cr-rich garnet, Cr-diopside, forsteritic olivine and enstatite. Although similar to the constituents of lherzolite and harzburgite which form the lithospheric upper mantle, they are distinctly richer in Cr. The Cr-rich pyrope garnets in diamonds are depleted in CaO and exhibit solid solution towards knorringite ($Mg_3Cr_2Si_3O_{12}$) rather than uvarovite ($Ca_3Cr_2Si_3O_{12}$). This compositional peculiarity is shared only with garnets found in rare examples of highly depleted, diamond-bearing garnet harzburgite (Pokhilenko *et al.*, 1977; Nixon *et al.*, 1987). On the basis of this evidence, it is now believed that *some* diamonds are derived by the disaggregation of such source rocks. Thus, the recognition of purple subcalcic chrome pyrope xenocrysts, known colloquially as G10 garnets, in kimberlites is commonly regarded as a key indicator for the presence of diamond (see below).

The inclusions of the eclogite suite consist principally of orange pyrope-almandine, omphacitic pyroxene, kyanite and coesite. The assemblage is similar to that which characterizes the eclogite xenoliths found in many kimberlites. Many of these xenoliths are diamond bearing and their disaggregation thus provides a realistic source for the eclogitic suite of diamonds.

The inclusion data suggest that diamond xenocrysts may originate from at least two sources, *i.e.*, garnet harzburgite or eclogite. Studies of southern African diamond deposits have demonstrated that there is no correlation between the xenolith suite found in a kimberlite and the diamond inclusion suite. A kimberlite rich in eclogite xenoliths may contain diamonds with predominately peridotitic suite inclusions and *vice versa*.

Studies of diamond-free and diamond-bearing lherzolites, harzburgite and eclogite

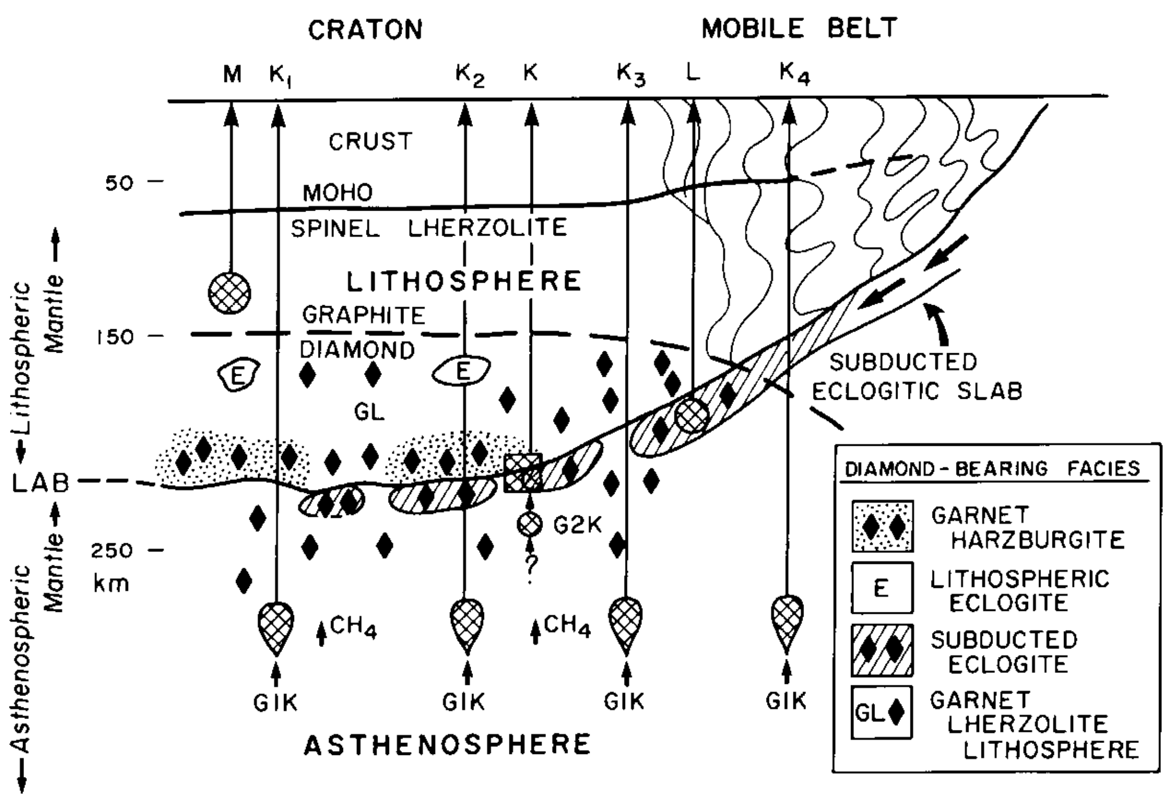

Figure 1 *Hypothetical cross-section of an Archean craton and adjacent cratonized mobile belt, showing the location of the lithosphere-asthenosphere boundary (LAB) relative to the stability fields of diamond and graphite. The diagram illustrates why different group 1 kimberlites (G1K) differ with respect to sources of xenocrystal diamond. K_1 may contain lithospheric and asthenospheric garnet lherzolite diamonds together with garnet harzburgite-derived diamonds. K_2 contains diamonds from the aforementioned sources plus diamonds derived from lithospheric eclogites and subducted eclogites,* i.e., *five distinct sources. K_3 contains only lithospheric and asthenospheric garnet lherzolite diamonds. K_4 does not pass through any diamond-bearing regions and is barren of diamonds. Group 2 kimberlites (G2K) are shown originating at the LAB and contain diamonds derived from garnet harzburgites and subducted eclogitic sources. An asthenospheric component may be involved in their genesis. Lamproite (L) contains diamonds derived from subducted eclogite and lithospheric garnet lherzolite sources. Melilititic (M) magmas are shown to be derived from depths within the graphite stability field and hence they are barren of diamond.*

xenoliths found in kimberlites show that the minerals present have equilibrated at pressures [50-60 kbar (150-250 km)] and temperatures (900-1400°C) characteristic of the upper mantle. These PT conditions are within the stability field of diamond as defined by the diamond-graphite univariant curve (Figure 2). It is assumed from these data that diamond growth must also have occurred in the upper mantle at similar, or greater, pressures and temperatures. However, it does not follow from this conclusion that diamonds formed in lherzolites and eclogites were necessarily produced by identical processes.

Current models of the process of diamond formation differ in particular with respect to the source of the carbon. One group of hypotheses suggests that the carbon is juvenile. Deposition of the carbon as diamond occurs as methane or other hydrocarbons are oxidized during ascent through the upper mantle (Taylor and Green, 1989) or at the lithosphere-asthenosphere boundary (Haggerty, 1986). This hypothesis is favoured for the generation of diamonds containing the peridotitic suite of inclusions.

A second group of hypotheses suggests that the carbon is introduced into the mantle by subduction processes (Schulze, 1986; Kesson and Ringwood, 1989). The carbon is thus not juvenile and may even be ultimately of biogenic origin (Milledge *et al.*, 1983). Proponents of subduction hypotheses commonly cite the wide range in carbon isotopic compositions observed in diamonds in support of this process. A subduction origin for diamonds containing the eclogitic suite of inclusions seems highly probable. It is important to note that diamonds derived from an eclogite source will not be associated with subcalcic chrome pyrope. Recently, Kesson and Ringwood (1989) have presented a model which attempts to link all varieties of diamond to subduction processes.

Both groups of hypotheses have in common the concept that diamond-bearing rocks originate at depths greater than 150 km, and primarily at or just above the continental lithosphere-asthenosphere boundary (Figure 1). Diamond formation is ultimately related to the long-term development of continental cratons (Boyd and Gurney, 1986). It is not known whether diamond-forming processes operated only in the Archean and Proterozoic or are still operative today.

Diamond preservation for billions of years requires that the mantle be held at low oxygen fugacities (Haggerty, 1986; Taylor and Green, 1989). Under such conditions diamond is "indefinitely" stable. However, passage of oxidized fluids rich in CO_2 and H_2O through diamond-bearing horizons would result in oxidation of the diamond to CO_2 or its conversion to graphite.

In summary, current hypotheses of diamond formation postulate that the roots of continental cratons contain diamond-bearing horizons. The vertical and lateral extent, diamond content and uniformity of these zones are unknown. Given the capriciousness of geological processes, a uniform distribution of diamond is highly unlikely. Disruption and disaggregation of such diamond-bearing zones by the passage of magmas ascending from greater depths will result in the incorporation of diamonds as xenocrysts in the magma. The type and amount of xenocrystal diamonds cannot be predicted.

The subsequent fate of xenocrystal diamonds entrained in the magma is dependent upon its oxygen fugacity and rate of ascent toward the crust. Slow transport in highly oxidized magma may result in the complete resorption of all diamond originally present. Studies of their morphology demonstrate that diamonds in most kimberlites and lamproites appear to have undergone varying degrees of resorption during transport (Harris, 1987).

Bearing the above in mind, it appears that the formation of a primary diamond deposit depends upon: (1) Development of an ancient diamond-bearing horizon at depths greater than 150 km in the continental upper mantle; (2) Passage of the transporting magma through diamond-bearing zones in the mantle. During transit, diamond xenocrysts derived from the disaggregation of mantle material are incorporated into the magma; and (3) Preservation of the xenocrystal diamonds in the magma during ascent.

It is not surprising that the diamond tenor of kimberlites and lamproites is highly variable, given the potentially wide and unpredictable variation in these parameters. Thus, the search for diamond deposits consists of locating and identifying rocks which have crystallized from magmas that have transported (and preserved) diamond from great depth in the upper mantle. Clearly, magmas which are derived from depths above the diamond-bearing zones in the mantle will be barren of diamonds (Figures 1 and 2). Unfortunately, some of these magmas, *e.g.*, melilitites and ultramafic lamprophyres, have petrographic similarities to kimberlites. *Correct petrological identification of such rocks will prevent wasted exploration efforts.*

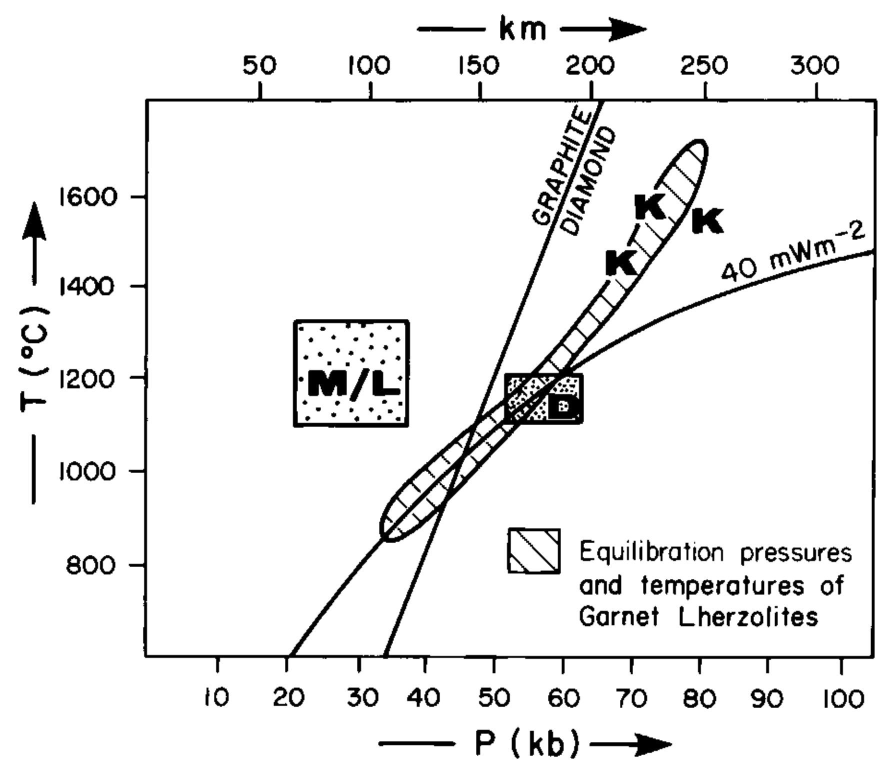

Figure 2 *Equilibration pressures (depths) and temperatures of diamond-free and diamond-bearing (field D) garnet lherzolite xenoliths found in kimberlites. Temperatures are calculated from the clinopyroxene-orthopyroxene solvus. Pressures are estimated from the Al_2O_3 content of orthopyroxene in equilibrium with garnet. Maximum depths recorded in the xenolith assemblage indicate the minimum depths of kimberlite magma generation. Kimberlites (K) are derived from within the diamond stability field. Melilitites (M) and ultrabasic lamprophyres (L) originate at much shallower depths within the graphite stability field. A representative continental shield geotherm is also illustrated. Some xenoliths have equilibration parameters which lie along this geotherm. These are considered to be mantle material which has not been affected by kimberlite-xenolith thermal interactions. Xenoliths which have higher equilibration parameters may record kimberlite-xenolith thermal interactions and/or are restites derived from the partially melted asthenospheric material which was the source of the kimberlite magma.*

It should be especially noted that neither kimberlites nor lamproites are members of the lamprophyre clan, as they are derived from petrologically distinct magma types (Mitchell and Bergman, 1991). Inclusion of these rocks within this clan, as suggested by Rock (1989), therefore, serves no petrological purpose, and leads only to further confusion as to their character.

KIMBERLITES

Kimberlites remain the principal source of primary diamond despite the discovery of high-grade deposits in lamproites. Recent mineralogical and Nd-Sr isotopic studies have shown that two varieties of kimberlite exist:

- Group 1 or olivine-rich monticellite serpentine calcite kimberlites.
- Group 2 or micaceous kimberlites.

Groups 1 and 2 correspond to the original "basaltic" and "micaceous or lamprophyric" kimberlites of Wagner (1914).

Smith (1983) showed that most group 1 and 2 kimberlites are derived from sources which are depleted or enriched, respectively, in light rare earths and Rb relative to the bulk earth reference composition (Figure 3). This division is profoundly significant, in that it demonstrates that group 1 and 2 kimberlites may be derived from asthenospheric and lithospheric sources, respectively, and thus cannot be genetically related.

Group 1 Kimberlites.

Group 1 kimberlites are complex hybrid rocks consisting of minerals that may be derived from (1) the fragmentation of upper mantle xenoliths (including diamond), (2) the megacryst or discrete nodule suite, and (3) the primary phenocryst and groundmass minerals. The contribution to the overall mineralogy from each source varies widely and significantly influences the petrographic character of the rocks. Consequently, group 1 kimberlites comprise a petrological clan of rocks that exhibit wide differences in appearance and mineralogy as a consequence of the above variation, coupled with differentiation and the diverse styles of emplacement of the magma.

Figure 4 illustrates an idealized kimberlite magmatic system, showing the relationships between effusive rocks, diatremes and hypabyssal rocks. Currently, three textural-genetic groups of kimberlite are recognized, each being associated with a particular style of magmatic activity in such a system. These are: (1) crater facies, (2) diatreme facies and (3) hypabyssal facies. Rocks belonging to each facies differ in their petrography and primary mineralogy, but may contain similar xenocrystal and megacrystal mineral assemblages.

Definitions of kimberlites in the literature are based primarily on the character of hypabyssal kimberlites, as these are the most amenable to petrographic study. Satisfactory unambiguous definition of kimberlite is clouded by the uncertainty regarding the nature of the megacryst suite. [N.B. The terms megacryst and macrocryst are used in a non-genetic sense to refer to minerals whose relationship to their host rock is unknown.] Some petrologists consider that megacrysts are actually xenocrysts, *e.g.*, Skinner and Clement (1979), Clement *et al.* (1984), while others, *e.g.*, Mitchell (1986), regard them as cognate minerals of high pressure origin that need not necessarily be phenocrysts in the current hosts.

In this work, following Mitchell (1989), group 1 kimberlites are defined as a clan of volatile-rich (dominantly CO_2) potassic ultrabasic rocks. Commonly, they exhibit a distinctive inequigranular texture, resulting from the presence of macrocrysts (and in some instances megacrysts) in a fine-grained matrix. The megacryst/macrocryst assemblage consists of rounded anhedral crystals of magnesian ilmenite, Cr-poor titanian pyrope, olivine, Cr-poor clinopyroxene, phlogopite, enstatite and Ti-poor chromite. Olivine is the dominant member of the the macrocryst assemblage. The matrix minerals include second generation euhedral primary olivine and/or phlogopite, together with perovskite, spinel (titaniferous magnesian aluminous chromite, titanian chromite, members of the magnesian ulvöspinel-ulvöspinel-

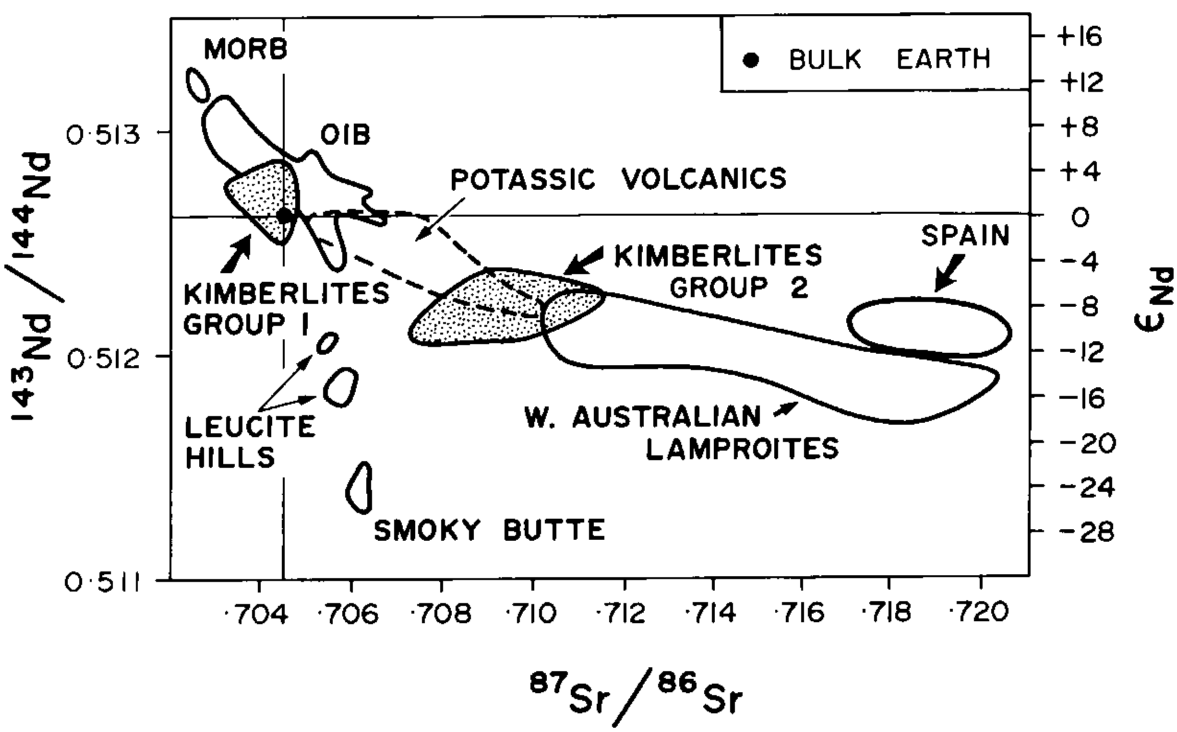

Figure 3 *Nd versus Sr isotopic compositions of kimberlites and lamproites relative to those of the bulk earth reference composition, oceanic island basalts (**OIB**), mid-oceanic ridge basalts (**MORB**) and potassic volcanic rocks. Rocks with isotopic compositions close to those of bulk earth and within the upper left quadrant of the diagram are conventionally interpreted as being derived from asthenospheric sources. Rocks with isotopic compositions which plot in the lower right quadrant are believed to be derived from ancient enriched lithospheric sources. Trends toward the bulk earth composition may reflect mixing with asthenospheric components. Note that Leucite Hills and Smoky Butte lamproites appear to be derived from sources which were depleted in Rb relative those of the West Australian and Spanish lamproites. See Mitchell (1989) for data sources.*

magnetite series), monticellite, apatite, calcite and primary late-stage serpophitic polygonal serpentine (commonly Fe-rich). Some kimberlites contain late-stage poikilitic eastonitic phlogopite or Ti-poor tetraferriphlogopite. Nickeliferous sulphides and rutile are common accessory minerals. The replacement of early formed olivine, phlogopite, monticellite and apatite by deuteric serpentine (lizardite) and calcite is common. Evolved members of the clan may be devoid of, or poor in, macrocrysts, and composed essentially of calcite, serpentine and magnetite together with minor phlogopite, apatite, rutile and perovskite.

Depending upon the presence or absence of macrocrysts, it is possible to recognize:
(1) Macrocrystal group 1 kimberlites, *i.e.*, those containing (>5 vol.%) rounded to anhedral crystals 5-10 mm in diameter.
(2) Aphanitic group 1 kimberlites, in which macrocrysts and megacrysts are absent or present in small quantities (<5 vol.%).

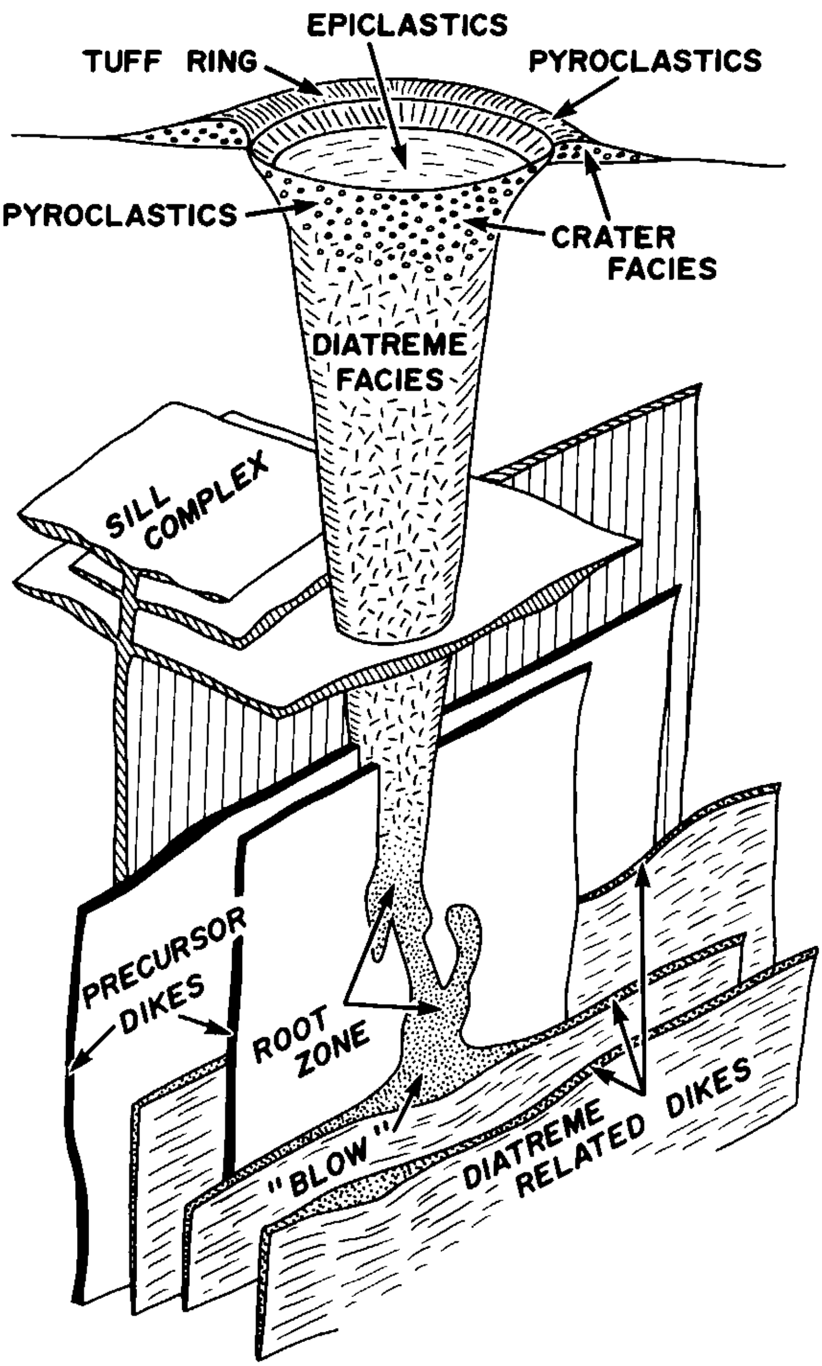

Figure 4 *Model of an idealized kimberlite magmatic system (not to scale) illustrating the relationships between crater, diatreme and hypabyssal facies rocks. The diatreme root zone is composed primarily of hypabyssal rocks. (After Mitchell (1986), reproduced by permission of Plenum Press).*

Key mineralogical features essential for the recognition of group 1 kimberlites include the presence of:
(1) The megacryst suite, especially Mg-ilmenite (3-23 wt.% MgO, but typically >8 wt.% MgO), Cr-poor (0-3 wt.% Cr_2O_3) titanian pyrope and subcalcic diopside.
(2) Ti-poor (<2 wt.% TiO_2) phlogopites that range in composition from phlogopite (phenocrysts or macrocrysts) to eastonitic phlogopites or Ti-poor tetraferriphlogopite in the groundmass.
(3) Spinels that range in composition from titanian magnesian aluminous chromites to magnesian ulvöspinel-magnetite solid solutions.

Characteristically absent from group 1 kimberlites are aluminous diopside and augite, andraditic garnets, feldspars, amphiboles, leucite, nepheline and melilite. Group 1 kimberlites do not contain primary diopside. When diopside is present, it is invariably a secondary mineral which crystallizes as a result of contamination of the magma by silica-rich country rocks. Mantle-derived *xenocrysts* of magnesian aluminous chromite, Cr-diopside and Cr-pyrope are common.

All kimberlites are classified on the basis of their groundmass mineralogy (Skinner and Clement, 1979). The method assumes that it is not always possible to distinguish between the ubiquitous xenocrystal and phenocrystal olivine. Prefixes are used to indicate the dominant groundmass minerals, *i.e.*, monticellite serpentine kimberlite, perovskite calcite kimberlite, or phlogopite kimberlite. [N.B. In this work, a phlogopite kimberlite, that is a group 1 kimberlite modally enriched in phlogopite, is not synonymous with a group 2 micaceous kimberlite. Standard petrographic terms may be used to describe group 1 hypabyssal kimberlites.]

Diatreme facies kimberlites differ from hypabyssal kimberlites in that they are volcaniclastic breccias. They consist of clasts of country rocks, fragments of hypabyssal kimberlite (autoliths) and rounded pelletal lapilli set in a matrix of microcrystalline serpentine and diopside. Pelletal lapilli consist of an olivine macrocryst nucleus surrounded by hypabyssal-like material. Commonly lath-shaped minerals, now altered to carbonate, are arranged tangentially around the nucleus. Formation of the groundmass diopside is considered to be a consequence of contamination. Monticellite is typically absent, although pseudomorphs may indicate its former presence. Diatreme facies kimberlites are commonly referred to as tuffisitic kimberlites. However, use of this term implies knowledge of the processes that have led to their formation, and, as the origin of kimberlite diatremes is still in dispute, the non-genetic term volcaniclastic kimberlite may be preferable.

Crater facies group 1 kimberlites are volumetrically insignificant in most kimberlite provinces due to extensive erosion. They include epiclastic deposits and rocks which may represent tuffs. Kimberlite lavas have not yet been recognized. Crater facies rocks represent significant sources of diamond, as demonstrated by the important deposit found at Orapa (Botswana), and where preserved should be regarded as prime exploration targets. Unfortunately, studies of crater facies kimberlites have not yet been published and they remain one of the least understood aspects of kimberlite petrology.

Diamond deposits in group 1 kimberlites. Figure 4 shows that the diatreme and its root zone are typically major components of kimberlite magmatic systems. The majority of diamonds extracted from group 1 kimberlitic sources come from these regions. Our knowledge of the structure of diatremes and their root zones has been derived primarily from mined South African kimberlites as described by Clement (1982) and Clement and Reid (1989).

Diatremes are vertical or steeply inclined cone-shaped bodies consisting primarily of tuffisitic kimberlite breccia (or volcaniclastic kimberlite breccia). The typically constant marginal dip (75-85°) and downward tapering of the diatreme results in the cross-sectional area decreasing regularly with depth. Approximately circular or elliptical outcrop plans are characteristic of the diatreme zone. Diatremes grade with increasing depth into a root zone, which consists of hypabyssal kimberlites. Cross-sections of root zones (Figure 5) are highly irregular, in marked contrast to those of the diatreme proper. The axial lengths of diatremes are estimated to range from 300 m to 2000 m. Discussion of the formation of diatreme systems is beyond the scope of this work. Reviews of the fluidization and hydromagmatic hypotheses of diatreme genesis can be found in Clement (1982), Clement and Reid (1989) and Mitchell (1986).

Diatremes and root zones are typically filled by several distinct varieties of volcaniclastic or hypabyssal kimberlite, respectively (Figure 5). Each of these intrusions differ in diamond content, as well as in the amount of megacrystal and xenolithic material present. Figure 6 indicates how diamond grade is correlated with kimberlite type within the root zone of the Dutoitspan (South Africa) mine. *The grade variations illustrated in Figure 6 are substantial and demonstrate the need for accurate petrographic description and mapping of the distribution of kimberlites found in a given intrusion. Failure to do this will result in unpredictable grades and/or the mining of low-grade ore.*

Significant intra-kimberlite variations in diamond content also are common. Figure 7 illustrates such variations within the W3 root zone kimberlite unit at the Wesselton (South Africa) mine. Variation in diamond content results in part from dilution of the ore with country rock clasts, *e.g.*, the low-grade,

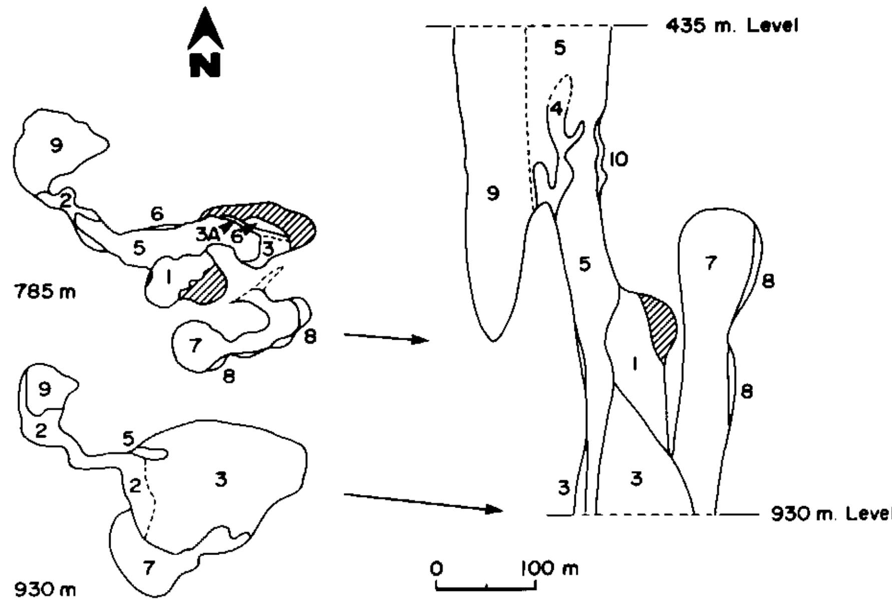

Figure 5 *Vertical and plan sections of the root zone of the Wesselton diatreme, South Africa. Numbers signify petrographically distinct types of kimberlite. Diagonal shading indicates kimberlite breccias. Note the presence of a "blind diatreme" (units 7 and 8) which failed to extend to the surface. (After Clement (1982)).*

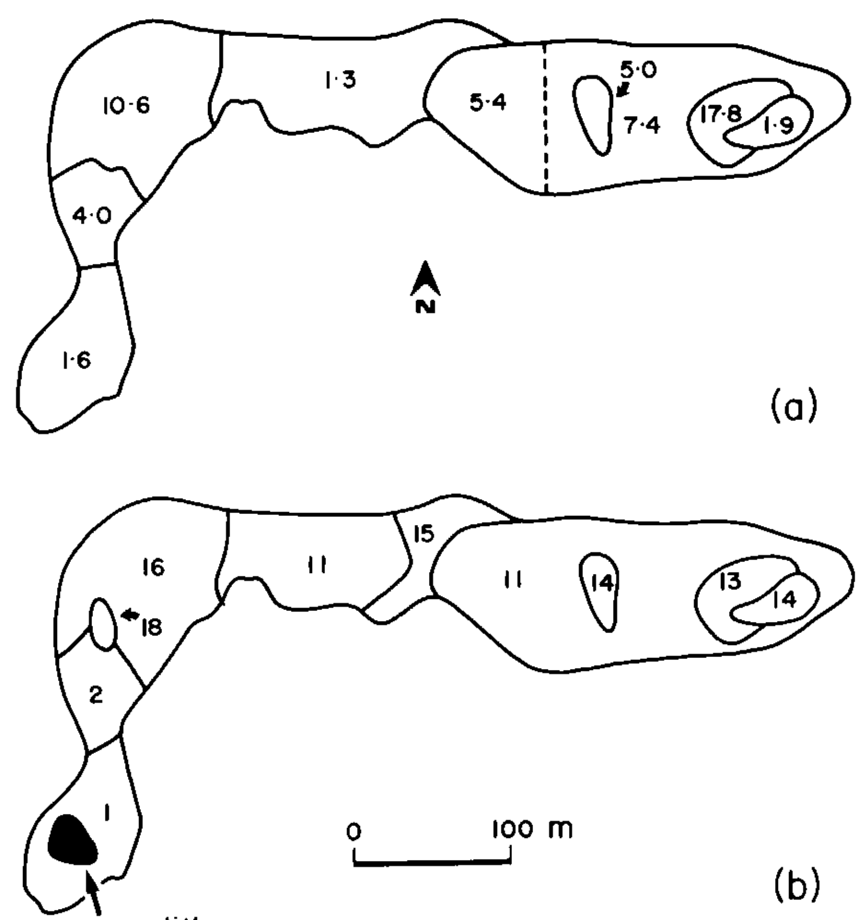

Figure 6 (**a**) *diamond grades (CM/100t) and* (**b**) *petrographic varieties of kimberlite (numbered) found at the 870m level of the Dutoitspan Mine, South Africa. (After Clement (1982)).*

south-central part of the intrusion. Clement (1982) notes that this intrusion also displays significant petrographic variation, especially with respect to the content of macrocrystal phlogopite, garnet, ilmenite and MARID-suite xenoliths. These variations undoubtedly reflect the operation of physical processes such as flow differentiation and/or batch mixing, which concentrate diamonds (and macrocrysts) in unpredictable portions of the magma during intrusion.

Diatreme facies group 1 kimberlites are relatively more uniform with respect to their diamond content than hypabyssal rocks (Clement, 1982). For example, the Dutoitspan (South Africa) D11 tuffisitic kimberlite breccia has a consistent low grade [5-8 CM/100t][2] over a vertical distance of 870 m. Consistent grades are also found at the Premier (South Africa) and Letseng (Lesotho) mines within individual volcaniclastic units. Relatively constant grades in diatreme facies rocks probably are a result of the mixing processes involved in diatreme formation, which lead to relatively uniform bodies of kimberlite.

It is commonly believed that the diamond content of kimberlite bodies decreases with depth. Modern studies by Clement (1982) have verified this hypothesis in only two instances, De Beers Mine (South Africa) and Dutoitspan. For example, grades in the hypabyssal root zone intrusion DB3 in the De Beers mine decrease from 61 CM/100t at the 595m level to 39 CM/100t at the 720m level. However, Clement notes that the hypothesis is not supported by studies of the majority of other kimberlites and thus has no universal applicability.

It is well known that group 1 kimberlite intrusions occur in distinct petrological provinces. Within these provinces, groups of intrusions define fields or clusters consisting of 1-20 individual kimberlites. Within a single cluster, all kimberlites appear to be either diamond-bearing or barren of macro-diamonds. Notwithstanding the variations in diamond grade within individual diatreme-root zone systems, there appear to be regional variations in diamond content. Thus, the Gibeon field in Namibia is barren, whilst the Kimberley field is relatively rich in diamonds. Similarly, in the Soviet Union, the Malo Butuobinsk field is diamondiferous, whilst the Lower Olenek field lacks diamond. It is particularly important to note that barren and diamondiferous kimberlites are not petrographically different with respect to their primary mineralogy or the composition of the macrocryst suite (Mitchell, 1987). The origins of these inter-province variations in diamond content are discussed below.

Within a given field, not all intrusions are necessarily economic. All intrusions will contain diamond, but the grade will vary widely. In the Kimberley cluster, the average grade reported ranges from 4 CM/100t to 56 CM/100t (Williams, 1932). The larger diatremes (De Beers, Wesselton, Kimberley, Dutoitspan) appear to be richest in diamond. However, the average grade must be regarded with caution, bearing in mind the comments above on intra-kimberlite diamond variation. Moreover, recovery grades do not necessarily reflect the actual diamond content of a kimberlite, as dilution by country rock may occur during mining.

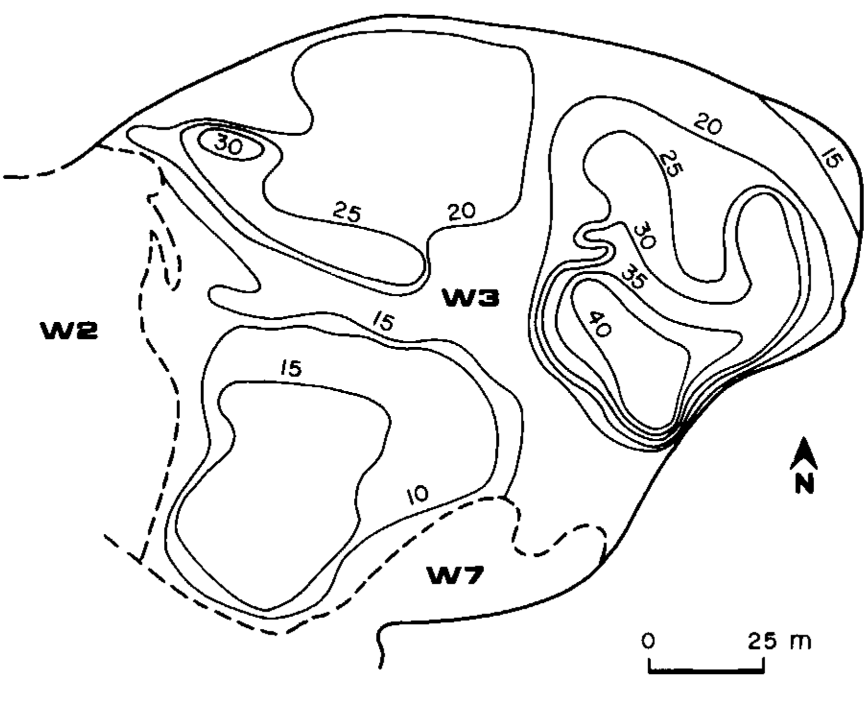

Figure 7 *Variations in diamond grade (CM/100t) within the W3 kimberlite, 930m level, Wesselton Mine, South Africa. (After Clement (1982)).*

A group 1 kimberlite does not have to be high grade (30-80 CM/100t) to be exploitable. Apart from normal economic considerations, other factors, such as the quality of the diamonds, also play a role in determining whether a body is payable. Thus, the relatively low average grade (<10 CM/100t) of the Letseng kimberlite is balanced by the high quality and size of the diamonds produced, despite the unfavourable location in the Maluti Mountains of Lesotho.

Group 2 Kimberlites.

Group 2 kimberlites are known so far only from southern Africa, where they form a petrological province that is older (200-110 Ma) than the bulk of the geographically associated group 1 kimberlites (<100 Ma).

Group 2 kimberlites have been inadequately characterized, and a formal definition has not yet been presented. Skinner (1989) and Mitchell and Meyer (1989) have noted that group 2 kimberlites consist principally of rounded olivine macrocrysts set in matrix which consists of macrocrysts and microphenocrysts of phlogopite and diopside, together with spinel (titanian magnesian chromite to ulvöspinel-magnetite), perovskite and calcite. K-Ba-V titanates belonging to the hollandite group are characteristic accessory phases. Considerable petrographic variation exists as a consequence of variations in the modal amount of olivine. In contrast to group 1 kimberlites, magnesium ulvöspinel and monticellite are absent and spinels and perovskite are relatively rare. The presence of K-Ba-V titanates and zirconium-bearing minerals such as kimzeyitic garnets in group 2 kimberlites further emphasizes the mineralogical distinctions between the two groups. The megacryst suite which is characteristic of group 1 kimberlites is notably absent from group 2 rocks. This difference is especially important with respect to prospecting programs which rely upon heavy mineral sampling as a means of locating kimberlites.

Mitchell and Meyer (1989) and Mitchell and Bergman (1990) have suggested that the mineralogy and isotopic differences between groups 1 and 2 are so great that they should not be regarded as members of the same petrological clan. The isotopic and other data indicate derivation from compositionally different sources, located at different depths in the mantle. Mitchell (1989) has suggested that, in order to highlight these differences, group 2 rocks should be referred to as orangeites rather than kimberlites, as they are not similar to the archetypal group 1 kimberlites.

Diamond deposits in group 2 kimberlites. Group 2 kimberlites have a similar magmatic style to group 1 rocks in that diatremes and hypabyssal facies rocks are known. The bulk of the occurrences are, however, thin (up to 2m), extensive (1-5 km),

[2] CM/100t = metric carats per 100 tonnes

hypabyssal dykes. In South Africa, these dykes are termed fissures. Typically, several sub-parallel or *en echelon* dykes comprise each occurrence. For example, the Swartruggens fissure system is composed of three *en echelon* dykes, only two of which contain diamonds. The dykes typically meander and split into smaller veins. They commonly pinch out entirely, but re-establish themselves at nearby unpredictable locations. Small pipes or blows, *e.g.*, the New Elands pipe or the Lion Hill pipe at the Star fissure system, may be developed upon these dykes. Many of the dykes are far richer in diamonds than group 1 diatreme or root zone kimberlites. Grades of 100-200 CM/100t are common and grades of up to 300 CM/100t have been reported (but not substantiated) from the Main fissure at Swartruggens. Note, in contrast, that hypabyssal dykes of group 1 kimberlite are typically uneconomic or barren of diamonds. Diamond distribution within the dykes of group 2 kimberlite is erratic and unpredictable. Many dykes are multiple, with each phase having a different diamond content. The narrowness of the dykes requires the use of labour-intensive mining techniques that contribute little waste to the ore.

Currently, there are two open pit mines working diatreme facies group 2 rocks. The Finsch (South Africa) mine consists of multiple intrusions of tuffisitic kimberlite breccias cut by thin dykes of hypabyssal kimberlite. The Dokolwayo (Swaziland) mine consists of at least 12 distinct intrusions of tuffisitic and hypabyssal kimberlite. Epiclastic kimberlite blocks can be found in one of the tuffisitic kimberlite breccias. Diamond grades are similar to those of diatreme facies group 1 kimberlites; thus, the Dokolwayo average recovery grade is currently about 20 CM/100t, and the run of the mine grade for Finsch in 1982 was approximately 100 CM/100t.

It is currently believed that the absence of group 2 rocks in other kimberlite provinces is related to the occurrence of a unique metasomatic event within the Kaapvaal lithosphere (Mitchell and Bergman, 1991). In many respects, group 2 rocks have compositional and mineralogical similarities to some lamproites. Lamproites are absent from the Kaapvaal craton and adjacent mobile belts and it is suggested by Mitchell and Bergman (1991) that group 2 rocks are the sole expression of peralkaline ultrapotassic magmatism in this craton. Group 2 kimberlites are absent from other cratons because the metasomatic events leading to the enrichment of lithospheric sources were different in character from those occurring in the Kaapvaal craton. Partial melting of these sources resulted in the formation of lamproites. Thus, group 2 rocks, on the basis of the current evidence, may represent a unique example of mantle-derived diamondiferous rocks.

LAMPROITES

Lamproites are ultrapotassic peralkaline rocks. Prior to the 1970s, lamproites were considered to be petrological curiosities of little scientific or economic importance. However, the discovery of very high-grade diamond-bearing lamproites in northwestern Australia, coupled with their unusual isotopic characteristics, has prompted re-appraisal of their nature (Scott Smith and Skinner, 1984; Mitchell, 1985, 1988; Bergman, 1987; Mitchell and Bergman, 1991). The three major provinces of lamproite magmatism are West Kimberley (Western Australia), the Leucite Hills (Wyoming) and Murcia-Almeria (southeast Spain). About 15 to 20 other smaller provinces are known throughout the world, although there is some uncertainty as to whether some of the inadequately characterized occurrences are actually lamproites (Mitchell and Bergman, 1991).

In common with group 1 kimberlites, it is impossible to devise a definition of lamproites based solely upon their modal mineralogy. Currently, the lamproite clan is defined on the basis of geochemical and mineralogical criteria and the rocks are described petrographically using Mitchell's (1985, 1988) modification of the Scott Smith and Skinner (1984) mineralogical classification scheme. Lamproites are a clan of rocks that exhibit an extremely wide range in modal mineralogy, as a consequence of the potentially large number of liquidus phases which crystallize from compositionally diverse lamproitic magmas, coupled with the mineralogical diversity resulting from differentiation. Consequently, olivine-rich lamproites bear little petrographic resemblance to leucite richterite lamproites or sanidine leucite lamproites, even though both rocks belong to the same clan. The old type locality nomenclature of lamproites, *e.g.*, wyomingite, wolgidite, *etc.*, has now been abandoned (Scott Smith and Skinner, 1984; Mitchell, 1985).

Rocks may be considered to be members of the lamproite clan if they are typically ultrapotassic (molar $K_2O/Na_2O>3$), peralkaline (molar $(K_2O+Na_2O)/Al_2O_3>1.0$) and rich in Ba (typically >2000 ppm, commonly >5000 ppm), Zr (>500 ppm), Sr (>1000 ppm), La (>200 ppm), and F (0.2-0.8 wt.%). Lamproites have Nd (Figure 3) and Pb isotopic compositions which indicate derivation of the magma from mantle sources that have undergone long-term enrichment in the light rare elements and depletion in U. The source regions may or may not have undergone Rb enrichment (Figure 3). Major element compositions are extremely variable. Two broad compositional groups are recognized, olivine/madupitic lamproites characterized by low SiO_2 (40-51 wt.%) contents, and phlogopite lamproites with 50-60 wt.% SiO_2. Olivine lamproites are silica deficient primarily because of the high modal amounts of olivine macrocrysts and/or xenocrysts. Primitive lamproite magmas are actually intermediate rocks with respect to their silica content and may be quartz and hypersthene normative. (Mitchell and Bergman, 1991).

Lamproites are characterized by the presence of widely varying amounts (5-90 vol.%) of the following primary phases: titanian (2-10 wt.% TiO_2) Al_2O_3-poor (5-12 wt.%) phenocrystal phlogopite, titanian (5-10 wt.% TiO_2) groundmass poikilitic tetraferriphlogopite, titanian (3-5 wt.% TiO_2) potassium (4-6 wt.% K_2O) richterite, forsteritic olivine, Al_2O_3-poor (<1 wt.%) Na_2O-poor (<1 wt.%) diopside, non-stoichiometric Fe-rich leucite (1-4 wt.% Fe_2O_3) and Fe-rich sanidine (1-5 wt.% Fe_2O_3). It should be particularly noted that the presence of all of the above phases in a rock is *not* required in order that it be classified as a lamproite. Any one mineral may be modally dominant and, in association with two or three others, determine the petrographic name, *e.g.*, leucite diopside lamproite, leucite phlogopite lamproite, sanidine richterite lamproite.

Minor and characteristic accessory phases include priderite [$(K,Ba)(Ti,Fe^{3+})_8O_{16}$], wadeite ($K_4Zr_2Si_6O_{18}$), apatite, perovskite, magnesiochromite, titanian magnesiochromite, and magnesian titaniferous magnetite. Less common, but nevertheless characteristic, accessories include jeppeite [$(K,Ba)_2(Ti,Fe^{3+})_6O_{13}$], armacolite, ilmenite, enstatite and shcherbakovite [$(Ba,K)(K,Na)Na(Ti,Fe,Nb,Zr)_2Si_4O_{14}$].

Minerals which are characteristically absent from lamproites include nepheline, sodalite, kalsilite, melilite, plagioclase, alkali feldspar, monticellite and melanite.

The presence of significant modal quantities of macrocrystal olivine gives some olivine lamproites a superficial petrographic similarity to group 1 and 2 kimberlites. They may be easily distinguished from these rocks on the basis of the chemical and mineralogical criteria outlined above. Detailed discussion of the mineralogical distinctions between the three groups of primary diamond-bearing rocks can be found in Mitchell and Bergman (1991).

Lamproites occur principally as extrusive, subvolcanic and hypabyssal rocks. In contrast to kimberlites, lavas and pyroclastic rocks are the characteristic manifestation of lamproitic igneous activity. Mitchell and Bergman (1991) consider that lava flow, crater, pyroclastic and hypabyssal facies of the lamproite clan may be recognized. Lamproitic volcanism is similar in style to that of common basaltic volcanism and consequently is not discussed in detail in this work. However, it is extremely important to note that *lamproites do not form diatremes or root zones analogous to those formed by kimberlites and lamproitic equivalents of tuffisitic kimberlite do not exist.*

Diamond deposits in lamproites. Diamonds have been recovered from the Argyle AK1 lamproite (Western Australia), the Ellendale lamproite province (Western Australia), and the Kapamba (Zambia), Majhgawan (India), Prairie Creek (USA), and Bobi (Ivory Coast) lamproites. The latter three localities were previously considered to be kimberlites, but are now recognized, on the basis of detailed mineralogical studies, to be members of the lamproite clan (Scott Smith and Skinner, 1984; Mitchell, 1985; Scott Smith, 1989).

In marked contrast to kimberlites, the majority of lamproite diamond deposits are found in pyroclastic rocks: only the Bobi occurrence is a hypabyssal dyke. Lamproite lavas appear to be devoid of diamonds. Diamond-bearing lamproite vents in Western Australia have been described in detail by Jaques *et al.* (1986), Boxer *et al.* (1989), and Smith and Lorenz (1989) and the discussion below is based primarily on these works.

Figure 8 illustrates the typical morphology of the best preserved lamproite vents found in the Ellendale field of the West Kimberley lamproite province. In this field, the vents range from 100 m to 1 km in diameter. Many are elongate in plan due to the coalescence of two or more craters. Exploration drilling has shown that the vents are shallow structures which flare out rapidly upward from a depth of about 300 m below the present erosion surface. Crater walls typically slope inward at an angle of about 30° toward a central feeder pipe of magmatic olivine lamproite. Commonly, this central conduit is less than 100 m in diameter and rapidly decreases in size with increasing depth. The vent has the shape of a champagne glass and is clearly unlike the carrot-shaped diatremes (Figure 4) formed by kimberlitic magmatism.

The vents contain pyroclastic and magmatic rocks. The bulk of the vent-filling material consists of well-bedded lapilli tuffs containing juvenile lapilli and lithic clasts. Structures within the tuffs indicate the presence of air-fall and base surge deposits. The tuffs are intruded by, and in some cases overlain by, magmatic hypabyssal olivine lamproite.

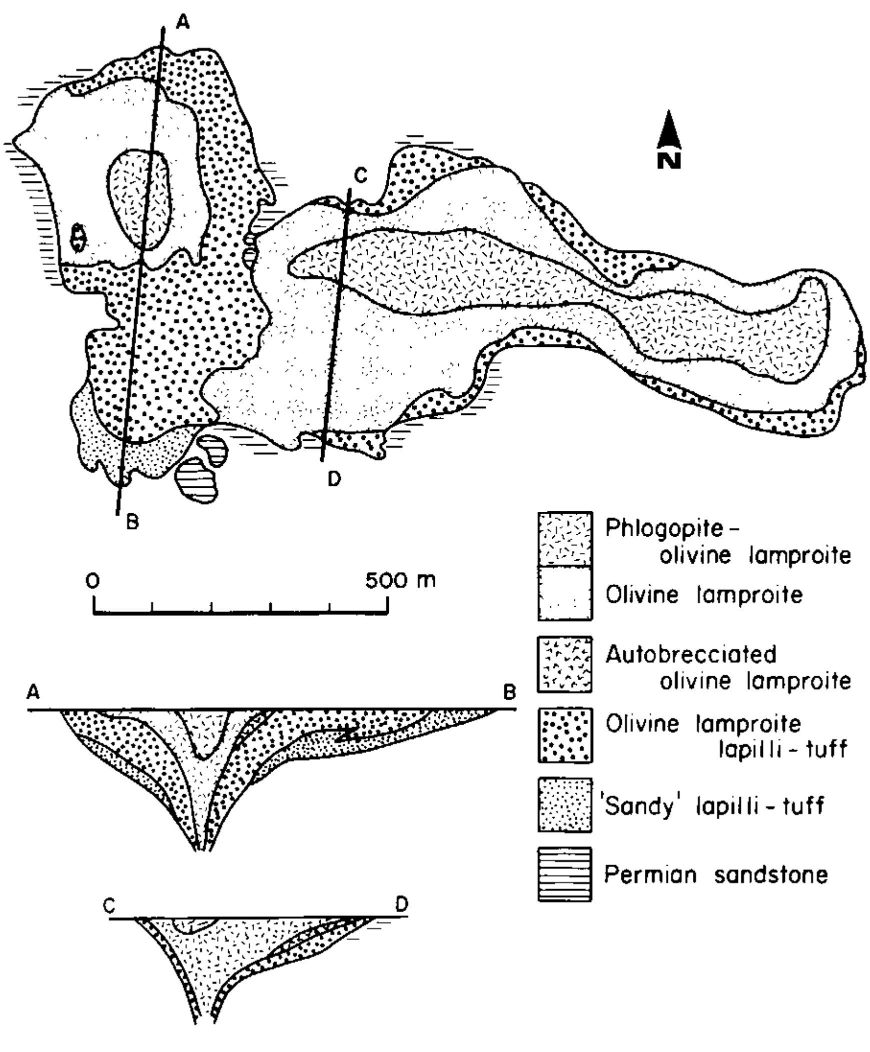

Figure 8 *Plan and cross-sections of the Ellendale 4 lamproite vent. Note the distinctly different morphology as compared with kimberlite diatremes (Figure 4) and the presence of hypabyssal magmatic rocks within the crater facies pyroclastic rocks.*

Smith and Lorenz (1989) have proposed that crater formation begins when rising lamproite magmas interact with water-bearing unconsolidated sands and sandstones. The ensuing hydrovolcanism produces a maar and tuff ring. Slumping of rim deposits and the formation of epiclastic deposits were followed by, and alternated with, base surge and other pyroclastic activity as the vent extended downward until dry country rocks were reached. At this time, pyroclastic activity ceased and lamproite magma intruded the crater deposits and in some instances formed lava lakes.

The Argyle vent (Boxer *et al.*, 1989) is broadly similar in character to the Ellendale vents, although the original shape has been modified by post-intrusional faulting and extensive erosion. The bulk of the vent (Figure 9) consists of lithic tuffs (sandy tuff) that may contain up to 60 modal % xenocrystal quartz grains in association with juvenile lapilli of olivine and leucite lamproite. Lapilli ash tuffs (non-sandy tuffs) containing juvenile lapilli of olivine lamproite form a minor part of the intrusion. The lithic tuffs are intruded by thin, highly altered olivine lamproite dykes.

Re-interpretation of the structure of the Prairie Creek vent (Mitchell and Bergman, 1991) is in accordance with the Ellendale model. The Majhgawan vent has the shape of an inverted cone (Scott Smith, 1989), but details of the structure are unknown.

The highest diamond grades in the Ellendale vents are associated with the earliest pyroclastic units. At Ellendale 4, the lamproite lapilli tuffs with few country rock clasts or quartz xenocrysts have grades that range from 3 to 30 CM/100t, while the xenocryst-bearing tuffs contain 1-4 CM/100t. Commonly, there is considerable intra-unit variation in diamond content. Thus, grades in the southern margin of the eastern lobe of Ellendale 4 range from 3.1 to 24.5 CM/100t, with an average of 14 CM/100t. The hypabyssal olivine lamproite core of the vent is poor in diamond and averages 0.5 CM/100t. A similar diamond distribution is found at Ellendale 9, where grades in pyroclastic rocks range from 3 to 8 CM/100t (average approx. 5 CM/100t) with the lower grades being associated with xenolith-rich tuffs. Hypabyssal olivine lamproites in the western and eastern lobes of the vent average 2.1 and 0.6 CM/100t, respectively. Other vents in the Ellendale field follow the same pattern of diamond distribution, but the average grades are much lower and not economically significant, *e.g.*, Ellendale 8, 12 and 18 pyroclastics average only 0.34, 0.12 and 0.25 CM/100t respectively (Jaques *et al.*, 1986).

The diamond distribution in the Argyle lamproite is the reverse of that determined for the Ellendale vents. Here, pyroclastic rocks rich in xenocrystal quartz (sandy tuffs) contain up to 680 CM/100t (Deakin and Boxer, 1989). In contrast, the tuffs rich in juvenile clasts are relatively diamond-poor, with grades of 100 CM/100t. Given the significant grade dilution factor for the sandy tuffs, it is apparent that the parental magma to the Argyle lamproite pyroclastics was extraordinarily rich in diamonds. No other primary diamond deposits approach these very high grades. Unfortunately, the quality of the Argyle diamonds is poor, with 95% of the production being industrial- or poor-quality gem stones. Exact details of the diamond distribution at Argyle remain proprietary information.

The Prairie Creek diamond deposit is broadly similar to the Ellendale vents. Although grades have not been determined, it is apparent that diamonds are concentrated in a volcaniclastic unit termed the "breccia phase", and are absent from the hypabyssal olivine lamproites. The diamond content of associated tuffs is unknown. The Majhgawan hyalo-olivine lamproite lapilli tuff has a grade of 8-15 CM/100 t (Scott Smith, 1989).

From the above, it is apparent that diamond deposits in lamproites are associated with the earliest pyroclastic eruptions of lamproite vents and that grades are typically relatively low (3-24 CM/100t) compared to those of most group 1 and 2 kimberlites. Diamond grades decrease as the eruption progresses. The Argyle deposit is clearly exceptional, and its very high grade has led to the impression that all diamond deposits associated with lamproites are of similar magnitude. However, Argyle may be unique, and result from the sampling of anomalously high concentrations of diamond in the mantle.

The relatively low grade of most lamproite vent-derived diamond deposits is balanced by their being amenable to open pit working, as the pyroclastic units involved lie near the surface and cover a wide area, *e.g.*, Ellendale 4 and 9 are 76 and 47 ha in area, respectively. Only crater facies kimberlites and the upper parts of kimberlite diatremes may be exploited by such methods, and most active mines in kimberlite employ subsurface methods of ore extraction.

Primary diamond occurrences in the Ivory Coast (Knopf, 1970) are thin dykes of highly altered olivine lamproite (Mitchell, 1985). Alluvial diamond deposits surrounding the dykes are clearly locally derived from these intrusions, as they follow their strike. The Toubabouko dyke is 1-3 m in width at the surface, but splits into small veins with depth. The Bobi occurrence consists of a diamond-bearing olivine lamproite dyke and two barren or very low-grade leucite phlogopite dykes. All three dykes are less than 1m in width. The absence of diamonds in the more evolved, phlogopite-rich dykes is in accordance with observations on diamond distribution in the West Kimberley field. The diamonds have similarities with the Argyle diamonds in being strongly resorbed and predominantly (66%) industrial and bort stones. Grades of the Ivory coast occurrences are not known. However, although of small volume, they are of apparently high grade, as significant alluvial diamond deposits surround the dykes. These are clearly locally derived, as they follow the strike of the intrusions.

TECTONIC SETTING OF KIMBERLITES

It has long been recognized that diamond-bearing kimberlites are located in ancient Archean cratons and that barren kimberlites occur in the adjacent mobile belts of younger Precambrian rocks (Clifford, 1966). Cratons which are covered by relatively undeformed Phanerozoic rocks provide particularly favorable environments for the preservation of kimberlite diatremes. Where these rocks have been stripped off the craton by erosion, the chances of discovering diatreme facies rocks are significantly reduced and many occurrences in this environment consist only of the root zones of diatremes and hypabyssal feeder dykes.

It is commonly stated that kimberlites are related to rift zones. This hypothesis arose because of the incorrect classification of lamprophyres associated with rifts as kimberlite. Actual kimberlite distributions show no relation to known rift zones (Mitchell, 1986). In the rare cases where an apparent relationship exists, it is evident that rifting has been superimposed upon a pre-existing kimberlite field, *e.g.*, Somerset Island (Canada), or represents an unsuccessful attempt by rift faults to penetrate a resistant craton, *e.g.*, the Singida field of the Tanzanian province.

The factors which control the location of kimberlite provinces in cratonized regions have not yet been satisfactorily determined. Several mutually contradictory hypotheses have been proposed. Typically, each hypothesis is introduced to explain the distribution of kimberlites within a given province and is not necessarily applicable to other provinces. However, a common theme of these hypotheses is that kimberlite fields are believed to lie upon linear or arcuate trends related to the presence of major crustal fracture zones termed lineaments or disjunctive zones. The zones of weakness provide channels for the ascent of mantle-derived mag-

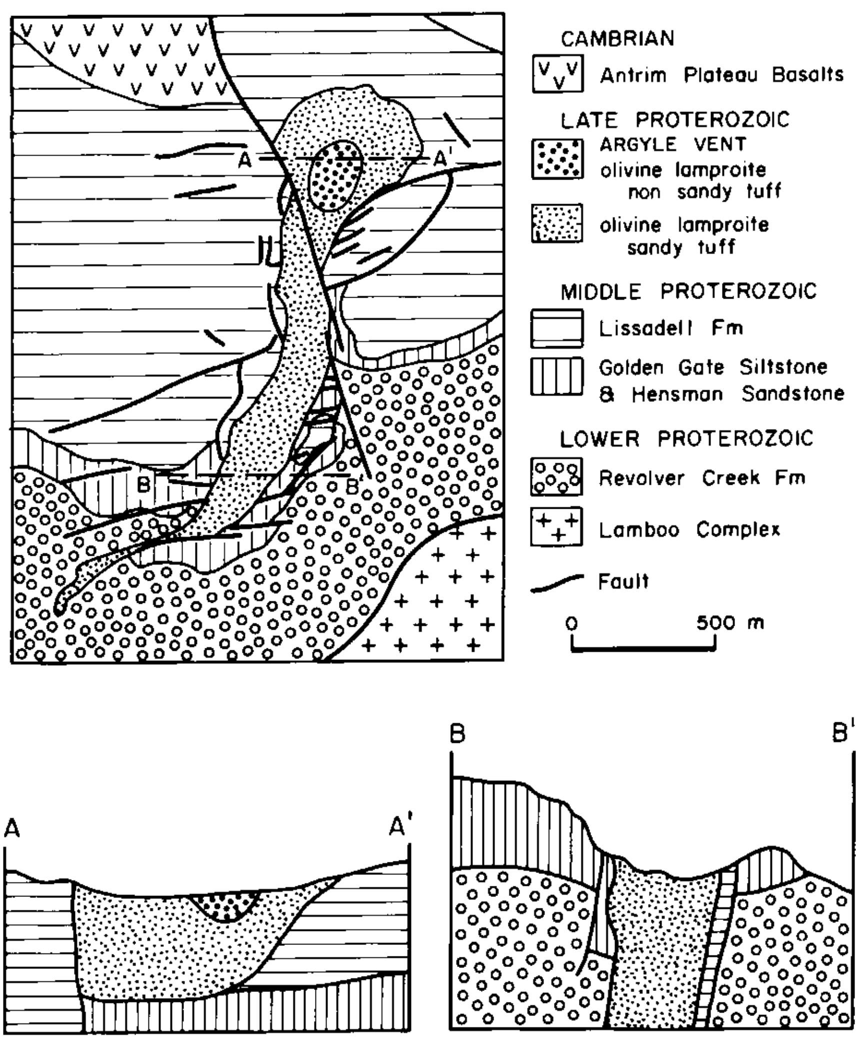

Figure 9 *Plan and cross-sections of the Argyle AK1 lamproite vent. After Jaques* et al. *(1986).*

mas through the lower lithosphere. Figure 10 illustrates the distribution of kimberlite fields in the Yakutian province (Arsenyev, 1962; Bardet, 1965) according to this model. In this example, the locations of kimberlite fields are used to determine the trends of lineaments beneath the cover of platform sedimentary rocks. Note that the trend of the lineaments does not necessarily correspond with the near surface structures that control the locations of individual diatremes within a given field. Disjunctive zones can be directly observed by satellite-based imagery in eroded Precambrian basement terrains. Haggerty (1982) has demonstrated the effectiveness of lineament analysis in explaining the distribution of Liberian kimberlites and in predicting possible sites for further exploration. The close association of kimberlites with linear features is particularly well demonstrated in Namibia (Marsh, 1973).

Models of the tectonic control of kimberlite emplacement differ mainly with respect to the mechanism suggested for the initiation of partial melting in the mantle and the ascent of magmas along the fracture zones. These include epeirogenic uplift, transform fault extension, and hot spot- and subduction-related magmatism. Discussion of these processes is beyond the scope of this work, and the reader is referred to Mitchell (1986) for a summary of the merits and drawbacks of these diverse propositions.

TECTONIC SETTING OF LAMPROITES

The emplacement of lamproites in a wide variety of tectonic settings has as yet precluded the development of a universal model explaining their temporal, geological and tectonic position (Mitchell and Bergman, 1991). These difficulties are compounded by the few *bona fide* lamproite provinces so far recognized.

Lamproites occur along the margins of cratons or in cratonized accreted mobile belts in regions of thick crust (>40-55 km) and thick lithosphere (>150-250 km). The lithosphere typically records multiple episodes of resurgent tectonic events, both extensional and compressional, some of which possess metamorphic ages coincident with Sr-Nd model ages inferred for the source regions of lamproites (Mitchell and Bergman, 1991). Their tectonic setting is well illustrated by the disposition of kimberlites and lamproites in Western Australia (Figure 11). Here, the Argyle (1150 Ma) and Ellendale (20 Ma) lamproites are found in the Proterozoic mobile belts surrounding the Archean Kimberley craton. In contrast, the 800 Ma kimberlites of the North and East Kimberley provinces are found within the craton (Jaques *et al.*, 1986). This example illustrates the general conclusion that lamproites do not occur on cratons and geographically and tectonically overlapping provinces of kimberlites and lamproites do not occur.

Many lamproites occur along continent-scale lineaments which parallel, *e.g.*, West Kimberley, or cross-cut Proterozoic mobile belts. Thus, the Prairie Creek lamproites (Arkansas) are approximately located at the intersection of the Reelfoot Rift and the Ouachita orogenic belt. The Reelfoot Rift is believed to have acted as a passive lineament as lamproites are not associated with zones of active rifting (Mitchell and Bergman, 1991).

Lamproites are not related to active subduction zones. However, trace element and isotopic studies suggest that the subducted materials found in paleo-Benioff zones are excellent candidates for the lithospheric sources of lamproites (Nelson *et al.*, 1986; Mitchell and Bergman, 1991). This hypothesis is important regarding the origin of diamonds in lamproites emplaced in mobile belts (see below).

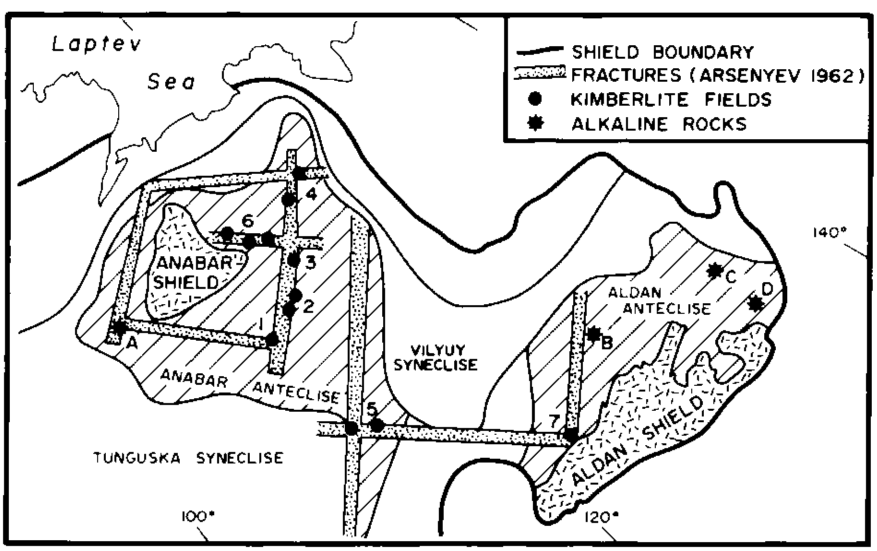

Figure 10 *Relationship of the Yakutian kimberlite fields and alkaline complexes to deep-seated lineaments or disjunctive zones as proposed by Arsenyev (1962) and Bardet (1965). Kimberlites:* ***1****, Daldyn-Alakit,* ***2****, Muna;* ***3****, Middle Olenek,* ***4****, Lower Olenek;* ***5****, Malo-Butuobinsk;* ***6****, Upper Anabar;* ***7****, Aldan. Alkaline complexes;* ***A****, Maimecha-Kotui;* ***B****, Tobukski;* ***C****, Ingili;* ***D****, Arbarastakh. After Mitchell (1986), reproduced by permission of Plenum Press.*

TECTONIC FACTORS CONTROLLING THE DIAMOND CONTENT OF KIMBERLITES AND LAMPROITES

The association of group 1 diamondiferous kimberlites with Archean cratons is important in that it suggests that the preservation of diamonds deep in the mantle is related to the long-term stability of these lithospheric regions. Boyd and Gurney (1986) have suggested that diamonds form and are preserved in the highly depleted ultramafic roots of cratons. This assumption is supported in the case of the Kaapvaal and Rhodesian cratons by the distribution of subcalcic Cr-pyrope (G10) garnets (Gurney, 1984; Boyd and Gurney, 1986). The term "G10 garnet" is derived from the Dawson and Stephens (1975) system of garnet classification and refers to garnets that fall within Gurney's (1984) high Cr-low Ca compositional field for chrome pyropes occurring as inclusions in diamonds. Such subcalcic Cr-pyrope garnets are only found within kimberlites occurring within the bounds of the Archean cratons in southern Africa, and their occurrence is positively correlated with the presence of diamondiferous group 1 and 2 kimberlites. Group 1 kimberlites in adjacent mobile belts lack such garnets and are barren of diamonds.

The diamond - subcalcic Cr-pyrope garnet association was initially recognized by Sobolev (1977) for kimberlites of the Anabar Shield. Unfortunately, detailed studies of this kimberlite province comparable with those of the southern Africa province are not available. Recent information regarding the association in Siberian kimberlites may be found in Verzhak *et al.* (1989).

It is important to note that the presence or absence of G10 garnets is not an infallible indicator of the presence of diamond in kimberlite. For example, Somerset Island (Canada) kimberlites contain diamonds (although the grade is low, *i.e.*, <1 CM/100t), but entirely lack G10 garnets (Jago and Mitchell, 1989). Although G10 garnets are present in some diamond-bearing Colorado-Wyoming kimberlites, they do not occur in significant quantities relative to common Cr pyrope (G9) garnets (Carlson and Marsh, 1989). The Skerring (Australia) and the Zero (South Africa) kimberlites contain G10 garnets, but lack diamond (Jaques *et al.*, 1986; Shee *et al.*, 1989).

The absence of G10 garnets in kimberlites or lamproites does not imply that diamond is not present in the roots of cratons, but only that garnet harzburgites analogous to those underlying the Anabar and Kaapvaal cratons have not been formed or sampled. Diamonds

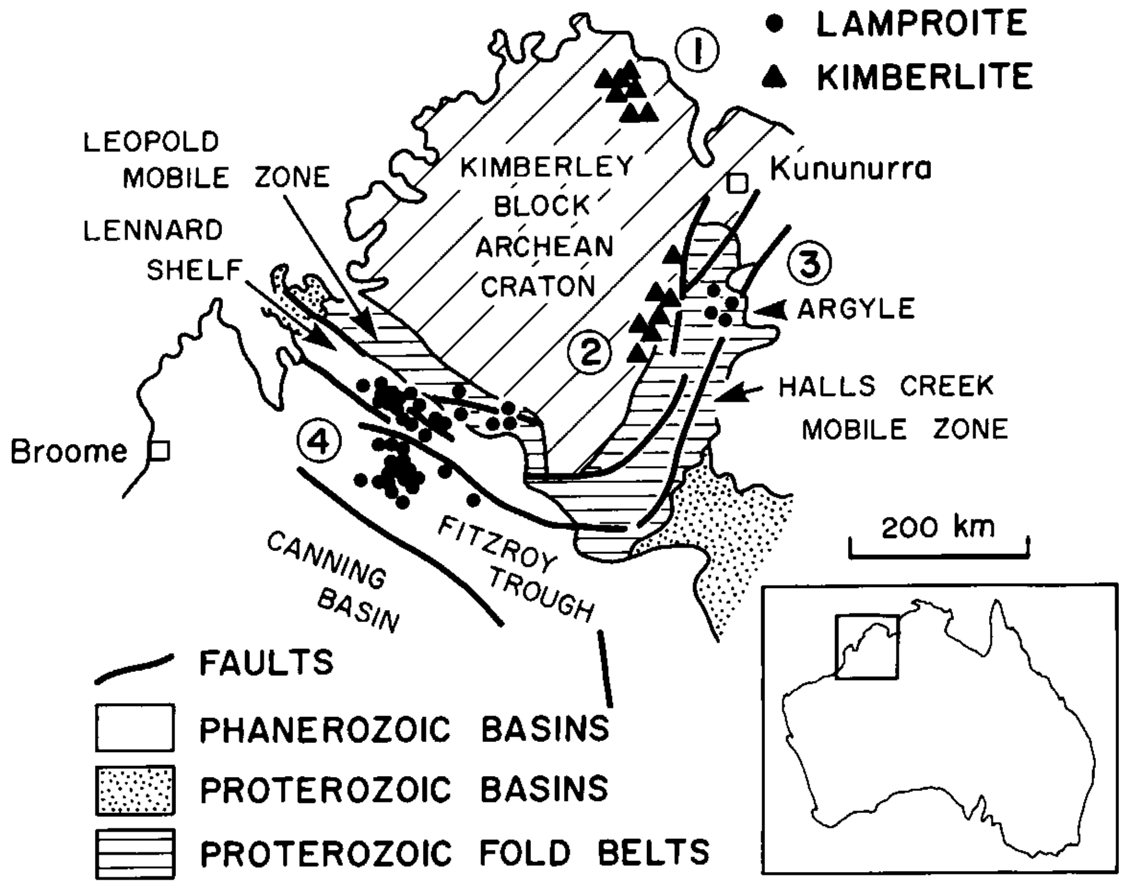

Figure 11 *Distribution of kimberlites and lamproites in West Australia showing their relationships to the Kimberley Archean craton and the adjacent Proterozoic cratonized mobile belts. (After Jaques* et al. *(1986)).*

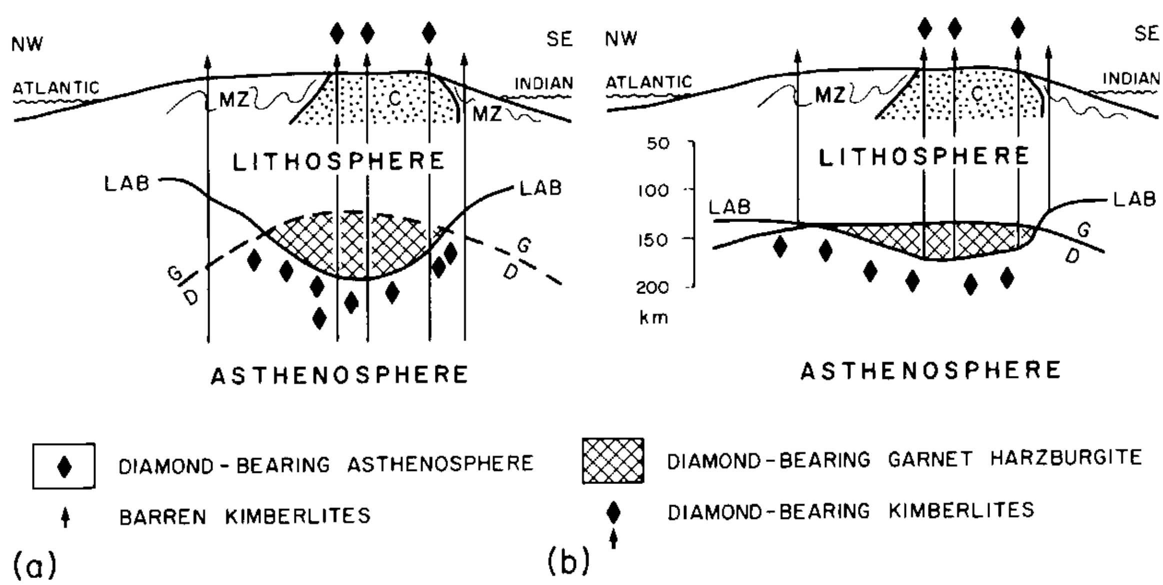

Figure 12 *Contrasting models illustrating why diamond-bearing group 1 kimberlites are restricted to within the bounds of the Kaapvaal craton and barren kimberlites are confined to the adjacent mobile belts.* **(a)** *Haggerty (1986) and Mitchell (1986, 1987) model. Kimberlites are derived from similar depths within the asthenosphere as a result of the partial melting of upwelling asthenospheric material. The graphite-diamond univariant curve (**D-G**) is convex toward the Earth's surface due to the low geothermal gradient of the Archean craton. Asthenospheric diamonds are formed by methane oxidation at the lithosphere-asthenosphere boundary (**LAB**) in the vicinity of the deepest parts of the craton root (Haggerty, 1986). Lithospheric diamonds occur in the highly depleted garnet harzburgite root of the craton. Only kimberlites which pass through the craton root regions traverse diamond-bearing horizons. Kimberlites which are emplaced in mobile belts do not pass through diamond-bearing regions of the asthenosphere or lithosphere.* **(b)** *Boyd and Gurney (1986) model. Kimberlites in this model are derived from different depths at the lithosphere-asthenosphere boundary (**LAB**). The location of this boundary is defined by the equilibration parameters of garnet lherzolite xenoliths found in kimberlites. In the mobile belt regions, the boundary is considered to lie within the graphite stability field. Diamonds are believed to be stable only within the deepest parts of the craton root. In this model, all diamonds are of lithospheric origin.*

are found associated with common garnet lherzolites and eclogites, and diamonds derived from such sources clearly will not be associated with G10 garnets. Thus, although the Gurney (1984) model appears to be valid for the majority of occurrences in southern Africa, one must be cautious in transferring it to other cratons as there is no *a priori* reason to expect that the long-term development of different continental cratonic nuclei and their adjacent mobile belts should be everywhere identical. It is especially important to note that mobile belts surrounding other cratons host diamond-bearing lamproites that lack G10 garnets (see below).

According to the southern Africa model, the difference between diamondiferous and barren kimberlites is simply related to the presence or absence of diamond-bearing rocks in the mantle traversed by kimberlites *en route* to the surface. Two hypotheses may be advanced to explain the distribution of diamonds. In one case, all kimberlites are considered to be derived from similar depths within the asthenosphere and there is no petrological difference between "on" and "off" craton kimberlites (Mitchell, 1986, 1987, 1989; Haggerty, 1986). Such kimberlites, on passing through the diamond-bearing harzburgitic root zones of the craton, will have the opportunity to incorporate diamonds as xenocrysts. Kimberlites ascending through mobile zones in which diamond-bearing horizons are absent will obviously never contain diamonds (Figure 12A). In this model, the equilibration pressures and temperatures of xenolith suites found in kimberlites are believed to represent kimberlite-mantle thermal interactions and/or random samples of mantle material derived from depths well above those at which the magma originally formed.

The alternate hypothesis assumes that kimberlites are derived from different depths at the asthenosphere-lithosphere boundary (Boyd and Gurney, 1986). This boundary extends into the stability field of diamond only at the deepest part of the continental root (Figure 12B). It follows from this hypothesis that the absence of diamond from kimberlites emplaced in mobile belts is related to their origins at depths shallower than those at which diamond is stable. Evidence for this hypothesis is based mainly upon a particular interpretation of the equilibration pressures and temperatures of xenolith suites found in "on" and "off" craton kimberlites. Boyd and Gurney (1986) believe that inflexions found in the distribution of xenolith equilibration parameters define the lithosphere-asthenosphere boundary. The maximum pressure and temperature recorded in a particular xenolith suite are believed to record the depth of origin of the host kimberlite.

Discussion of xenolith P-T equilibria is beyond the scope of this work and the reader is referred to Finnerty (1989) and Carswell and Gibb (1987) for summaries of the conflicting interpretations of these data.

Diamondiferous lamproites from Australia and Arkansas do not contain G10 garnets (Lucas *et al.*, 1989; Mitchell and Bergman, 1991). Thus, the diamond potential of lamproites cannot be assessed on the basis of the presence or absence of such garnets. Studies of mantle-derived lherzolites and xenocrysts in the Australian lamproites and kimberlites suggest that their lithospheric mantle sources are less refractory than the Kaapvaal cratonic roots, *i.e.*, highly depleted garnet harzburgites are absent (Jaques, 1989).

Studies of inclusions in diamonds from the Australian lamproites suggest that they are probably derived from eclogitic sources. These may represent subducted oceanic lithosphere that was eventually cratonized. This diamond-bearing paleo-Benioff zone material now comprises the deeper parts of mobile belts (Figure 1). The situation is distinctly different from that inferred for the Kaapvaal craton, where cratonized mobile belts appear to be diamond free. Importantly, if kimberlites were to intrude the mobile belts around the Kimberley craton, one might expect them to be diamond bearing, in contradiction to the southern Africa model.

Models have not yet been developed to explain the formation of group 2 kimberlites and their generally high diamond contents. Group 2 kimberlites define a linear trend of decreasing age from northeast to southwest across the Kaapvaal craton, suggesting that magmatism is related to hot spot or plume activity. These kimberlites are characterized by high contents of G10 garnets, but typically lack xenoliths of garnet lherzolite. Diamondiferous and other eclogite xenoliths are commonly present. Clearly, group 2 kimberlites must have sampled the same garnet harzburgitic sources of diamond as group 1 kimberlites (Figure 1). However, it is not known if these regions were also the source rocks of group 2 kimberlite magmas.

OTHER FACTORS CONTROLLING DIAMOND GRADES

Several other factors which control the diamond content of kimberlite and lamproite may be recognized. Unfortunately, none of these can be quantified and, compared with the tectonic factors, they are as yet poorly understood. The differences in diamond grade within and between kimberlite intrusions are related to: (1) heterogeneous diamond distribution in the source regions, (2) rate of release of diamond from entrained xenoliths, (3) sorting of xenocrystal diamonds during entrainment, flow and mixing of different batches of kimberlite magma, and (4) resorption of diamond in the ascending magma.

Diamond-bearing mantle xenoliths contain widely ranging diamond contents. The grades calculated for these typically small specimens are typically very high. Diamond-bearing eclogites may contain up to approximately 10^5 CM/100t. Two diamond-bearing garnet lherzolites from the Finsch mine contain 55 CM/t and 300 CM/t, that is 50 to 300 times as rich as the host kimberlite (Shee *et al.*, 1982). The grades of diamondiferous lherzolites occurring in the Argyle lamproite are even higher, with an average of 2200 CM/100t being reported by Jaques *et al.* (1990). Disaggregation of such source rocks could easily account for the diamond contents of kimberlites or lamproites. Interestingly, G10 garnets are absent from these lherzolites. However, the significance of the grade of these typically small specimens, which contain only small diamonds (<2 mm diameter), with respect to the overall diamond content of the mantle is uncertain. That very high grades (approx. 10^6 CM/100t) may be expected in some parts of the mantle is supported by the Beni Bousera (Morocco) pyroxenites, which contain up to 15% graphite that is considered to be pseudomorphic after diamond (Pearson *et al.*, 1989).

Further grade variations may arise as a result of mixing of diamonds derived from two or more sources located at different levels in the mantle. Different sources may be expected to release their diamonds to the transporting magma at different rates. Data derived from the study of inclusions in diamonds in individual kimberlites indicate that multiple sources are common. The ratio of peridotitic to eclogitic garnets varies widely between kimberlites, and several sources of eclogitic type diamonds are predicted to exist (Gurney, 1989). As yet, no pattern can be discerned in the distribution of diamonds containing different inclusion suites with respect to location of their host kimberlites in the craton. Multiple sources of diamonds are also seen in lamproites. Argyle diamonds have dominantly eclogitic inclusions, while Ellendale diamonds contain approximately equal proportions of peridotitic and eclogitic inclusions (Jaques, Hall *et al.*, 1989). Interestingly, the Argyle lamproites contain xenoliths of diamond-bearing peridotite, but not of eclogite (Jaques, 1989).

Source depletion effects may also be expected to occur. The initial batches of magma creating a pathway through the mantle probably contain the highest xenolith load. Formation of "swept conduits" as the eruption progresses will decrease the probability of entraining xenoliths and will result in gradually decreasing diamond content in the later batches of magma. This effect is undoubtedly responsible for the decreasing diamond tenor of the Ellendale volcaniclastic lamproites as the eruption proceeds. Source depletion effects may be the cause of the differing diamond contents of kimberlites found within a given intrusion (Clement, 1982).

No data are available on the rate of release of diamonds from their xenolith hosts. It is to be expected that xenoliths may disaggregate at any depth during transport. Small, euhedral, octahedral diamonds that have undergone little resorption in their kimberlite hosts are probably released during the later stages of transport at crustal levels, as Robinson *et al.* (1989) predict that such diamonds will be rapidly resorbed. Most diamonds show the effects of resorption. The process involves the conversion of primary growth forms to tetrahexahedral dissolution forms. Morphological variations in diamond populations reflect the depths at which different varieties of source rocks disaggregate. The amount of resorption also will depend (in part) on the size of the diamonds and their transport rate. Robinson *et al.* (1989) suggest that up to 60% of a macrodiamond crystal released from a xenolith at 150 km depth may be resorbed during transport. Asthenospheric microdiamonds formed by methane oxidation at the lithosphere-asthenosphere boundary (Haggerty, 1986) or small lithospheric macrodiamonds may be completely resorbed at depth.

If the oxygen fugacity of the magma increases during ascent, it is highly probable that diamond resorption will be enhanced, as they will be rapidly converted to CO_2. Diamonds in lamproites are characteristically strongly resorbed, suggesting that these water-rich magmas are more oxidizing than carbon dioxide-rich kimberlites.

Studies have not been undertaken on the sorting behavior of diamond in kimberlitic or lamproitic melts. Within kimberlites, grade variations as described above from Wesselton indicate that this process, operating within the root zone, can significantly influence the local diamond grade. Gravitational settling and mechanical concentration can obviously also occur in the mantle where batches of diamond-bearing magma pool. Re-mobilization of such diamond-rich "cumulates" by subsequent batches of magma will clearly result in the formation of hybrid magmas which will have enhanced diamond potential.

CONCLUSIONS

Primary diamonds are found in group 1 and 2 kimberlites and lamproites. All of these rocks have particular mineralogical characteristics that readily permit their identification and discrimination from non-diamondiferous rocks such as melilitites, alnoites and ultramafic lamprophyres. Primary diamond deposits in kimberlites are found principally in crater and diatreme facies rocks, whereas those in lamproites are primarily in pyroclastic crater facies rocks. Diamond is a xenocryst that is derived from the disaggregation of mantle material. Source lithologies include garnet harzburgite, garnet lherzolite and eclogite. Our knowledge of the vertical and lateral distribution of diamond in the upper mantle is inadequate.

Data for the Kaapvaal (South Africa) and Kimberley (West Australia) cratons indicate that models devised to illustrate the distribution of diamondiferous and barren rocks for any given craton and its adjacent cratonized mobile belts are not necessarily valid for another craton. Each craton appears to have evolved differently, and the type and distribution of diamondiferous source rocks must be different between cratons. The diamond potential of each craton, therefore, must be assessed on an empirical basis until sufficient data have been collected to develop a model for that particular craton. This conclusion is especially relevant to North America, where few kimberlite provinces are as yet known.

With the exception of the Yakutian and South African provinces, it is impossible to predict on the basis of tectonic setting or indicator mineral composition whether or not a newly discovered kimberlite or lamproite field will be diamondiferous. Only direct evaluation of new discoveries for their diamond content, accompanied by indicator mineral studies, will lead to predictive models for that particular field.

Despite considerable advances in our understanding of primary diamond deposits, we are still unable to explain the differing grades of individual kimberlites or even predict the grades of new discoveries within a well-studied field. It is unlikely that we shall ever be able to accomplish this without direct knowledge of the distribution of diamondiferous source rocks in the upper mantle.

ACKNOWLEDGEMENTS

This review summarizes the data produced and concepts formulated by many petrologists and mineralogists concerned with the origins of kimberlites, lamproites and diamonds. I wish to thank all of these scientists for their contributions to this topic. Research on the petrology of kimberlites and other alkaline rocks at Lakehead University is sponsored by the Natural Sciences and Engineering Research Council of Canada and Lakehead University.

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Accepted, as revised, 4 March 1991.