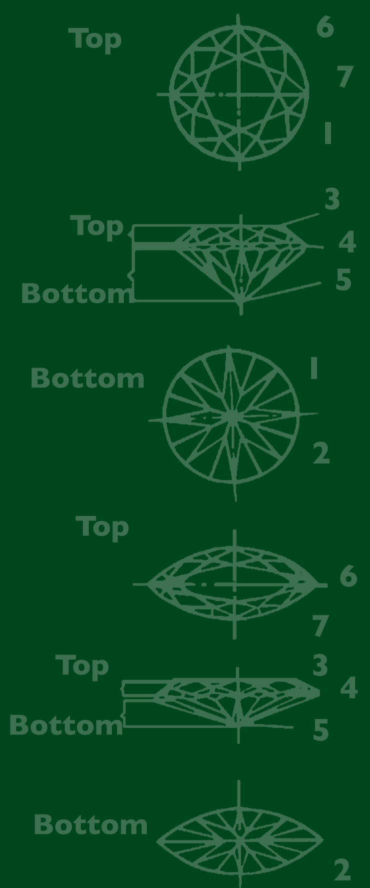
DIAMOND DEPOSITS

Origin, Exploration, and History of Discovery



EDWARD I. ERLICH
W. DAN HAUSEL

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Edward I. Erlich

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Preface

Diamonds have inspired dreams of wealth and power throughout history. Until modern times, most diamonds were insignias of royalty and were beyond the reach of the common person, who could only elicit visions of the astounding beauty and wealth brought forth by diamonds. It's no wonder that other gems and precious metals have historically taken a back seat to diamonds.

Some diamonds are so valuable that a person can literally carry a king's ransom in a pocket. A similar value in gold would mean one would have to have access to a forklift, as some of the most valuable diamonds in the world have been appraised for many thousands of times that of a similar weight in gold! It is no wonder that so many companies, geologists, and prospectors have invested time and money to locate and develop new diamond resources around the world.

Diamond deposits are not easily found. Diamonds occur in some of the rarest rock types on the surface of the earth, and when found, they are disseminated in trace amounts even in the richest deposits. The principal host rock, kimberlite, forms very small deposits. Being a relatively soft rock, kimberlite commonly erodes faster than the surrounding country rock and often is covered by thin layers of soil and regolith derived from adjacent rock outcrops. As a result, many kimberlites go unrecognized for decades. Thus it is necessary for the geologist and prospector to search for clues for the whereabouts of these hidden or partially hidden rocks. The clues are not easily found and require detailed detective work.

Kimberlite intrusives are generally exposed over relatively small surface areas. They occur in the form of pipes, dikes, and sills that may crop out over an area of a few hectares at the most. Other diamondiferous host rocks exhibit similar characteristics. For example, lamproite, another host rock, may occur in the form of dikes, sills, and small volcanoes. But lamproites are even rarer than kimberlite. Then there are other deposits that have been termed *unusual* or *unconventional*. Some of these are extremely rich in diamonds, and some will probably become prominent sources of diamond in the future.

If you are lucky enough to find an outcrop of kimberlite or lamproite, this alone is considered to be a significant discovery, as these rock types represent two of the rarest found on the earth. But this does not mean that you will find diamonds. Reported statistics suggest that only about 10% of all kimberlites in the world contain diamond, and less than 1% contain commercial amounts of diamond at grades less that 1 ppm (part per million) (Lampietti and Sutherland 1978). The chances of finding diamonds in lamproite are even less than in kimberlite. Thus, the odds of finding a commercial diamond deposit are not great. But through diligent exploration, one can increase the odds of finding a king's ransom.

Diamonds have value not only as gems, but also because their unusual physical properties are indispensable to industry. Industrial diamonds are required in drilling, metal cutting, and polishing of extremely hard surfaces. They also have many other industrial applications and are used in a variety of measuring instruments. But their real value comes in the form of gems, which are often valued at more than 100 times as much as an industrial diamond.

Most terrestrial diamonds begin their journey deep within the earth. To get diamonds to the surface, deep-seated magmas accidentally trap diamonds during their rise to the surface from depths as great as 200 km. Thus one might visualize diamondiferous pipes as a natural drill hole originating at depths inaccessible to commercial drilling and mining—the deepest mines in the world have reached depths of only a little more than 3.8 km.

Scientific curiosity related to diamonds has resulted in a mountain of publications during the past two to three decades. Each new discovery has meant a major influx of new scientific publications, such that the literature has more than doubled in volume over the past two decades.

Since the 1968 International Association of Volcanology and Chemistry of the Earth's Interior conference, which hosted a kimberlite symposium within the framework of the Oxford conference on volcanoes, there have been seven international kimberlite conferences held around the world including a special symposium in Cambridge in 1973. Each of these meetings resulted in major symposium volumes that contained volumes of abstracts and papers related to all aspects of diamonds and their host rocks.

To date, only kimberlite, lamproite, and some placers have produced commercial quantities of diamond. However, other diamondiferous host rocks have been identified. Examples of these so-called *unconventional* host rocks are less common than kimberlite and lamproite, and commercial varieties have yet to be found. But, some of these will undoubtedly become more prominent in the future when exploration groups place greater emphasis on the search for these seemingly unusual occurrences, as some varieties may potentially yield very high ore grades.

Some of the unconventional host rocks are minette, peridotite, pyroxenite, lamprophyre, alkaline basalts, some "normal" basalts, specific ring structures—astroblemes, and a few rare metamorphic rocks.

For many years, it was assumed that the only host rock for diamond was kimberlite. Unfortunately, this concept is still ingrained in the thinking of most exploration geologists and company chief executive officers, even though the discovery of a world-class commercial diamond deposit in lamproite in the Kimberley Block of northwestern Australia in late 1970s provided direct evidence for nonkimberlitic sources of economic diamond deposits. Following this discovery, it was realized that two of the first host rocks mined for diamonds were also lamproites—these were found at Murfreesboro, Arkansas, and Majhawan, India. The Majhawan pipe was mined for diamonds many years before kimberlite was identified in South Africa.

Acknowledgment of this new type of economic diamond deposit resulted in what can be called a "lamproitic revolution." This revolution resulted in discoveries of lamproite-related diamond deposits in Africa, Australia, Russia, and North America as well as a flurry of academic studies on lamproite-related issues.

The lamproite revolution has also led to other discoveries. Among these are diamonds associated with huge ring structures termed *astroblemes* or *geoblemes*, and diamonds in high-pressure metamorphic rocks. More and more diamonds are also being reported in other unusual rock types such as alnöite, minette, alpine-type ultramafic plutons, komatiites, and even in basalts. Each of these unconventional varieties exhibits a unique signature, which can be used in exploration. The significance of the discovery of graphite pseudomorphs after diamond in eclogite-ultramafic melanges similar to the Beni-Bousera massif in Morocco remains to be seen. But such deposits could become the new exploration targets of tomorrow.

It is the intention of this book to provide an overview for the geologist and prospector on how to find new diamond deposits, how to recognize them, how to evaluate them after they're discovered, and to provide some overviews on the history of some of the more significant diamond discoveries in the world.

The coauthors of this book have many years of experience in the exploration for diamond deposits. Edward Erlich spent a large part of his professional career searching and evaluating unconventional and conventional diamond deposits in the former Soviet Union, Alaska, Arkansas, and other localities. In the late 1950s, as chief geologist of a geological team of the Institute of Arctic Geology (USSR), he participated in the discovery of kimberlitic bodies and fields within the Olenek region of the northeastern Siberian Platform. At the same time he studied the regularities of spatial distribution of kimberlites, which resulted in series of publications and his Ph.D. dissertation.

Erlich was involved in the geological mapping of the Udja region, which resulted in a discovery of a new province of ultramafic-alkaline magmatism including a ring complex with a carbonatitic core known as the Tomtor massif and a new series of kimberlitic fields. While working for the Institute in Volcanology Academy of Science, USSR, in Kamchatka, he initiated and supervised mineralogical studies that resulted in the discovery of diamonds in basalts. After immigrating to the United States in early 1984, he worked as a consultant on kimberlites in Colorado and Montana and in the Arkansas lamproitic field. He also participated in diamond exploration in Alaska, California, and the Arkhangel'sk region of Russia.

W. Dan Hausel is an igneous petrologist and economic geologist, who is a well-known author on economic geology, gemstones, and kimberlites. He began his research on high-pressure mineral phases in basalts at the University of Utah. Later, while employed as the Senior Economic Geologist and head of research related to diamond deposits at the Wyoming Geological Survey, he conducted research related to kimberlites, lamproites, peridotites, komatiites, and lamprophyres. As an independent consultant, he developed exploration programs for various companies related to diamonds in kimberlite, lamproite, lamprophyre, and peridotite throughout the United States.

Since 1977, his work at the Wyoming Geological Survey has led to the discovery of several kimberlites in the State Line, Iron Mountain, and Middle Sybille Creek districts along the margin of the Wyoming craton of the western United States. He has also spent considerable time mapping and studying lamproites and lamprophyres in the craton and craton margin. His work also led to the discovery of several other gemstone occurrences, as well as some gold, platinum-group metal, and nickel occurrences. He has mapped several Archean greenstone terranes and related mining districts.

A widely published author with more than 400 publications, he has received special recognition in *Who's Who in Science and Engineering, 2000 Outstanding Scientists of the Twentieth Century, Who's Who in the West, Who's Who in America, Who's Who in the World, Who's Who in the 21st Century, 2000 Notable American Men,* and others.

It is our hope to provide our readers with a kind of guidebook on diamonds and diamondiferous host rocks and to lead them from historical data and crystal structure to diamond deposits, methods of exploration, and even to some marketing.

ACKNOWLEDGMENTS

The authors are indebted to Dr. T.N. Solovieva (Moscow, Russia), Dr. I.F. Gorina (St. Petersburg, Russia), and Dr. William Tafuri (Park City, Utah) for the collection of the latest Russian literature on the geology and petrology of the Kazakstan diamond deposits, Popigay astrobleme and diamonds, and the Russian lamproitic diamond deposits. Dr. Solovieva and Dr. Gorina not only compiled the appropriate publications but also took the time to deliver the information to the authors. Dr. A. Rub (Sacramento, California) obtained some of the newest Russian works. Mr. S. Sutley (U.S. Geological Survey, Denver) also provided invaluable assistance in obtaining the appropriate literature. Mr. M. Paukstis (Aurora, Colorado) provided computer assistance.

The authors also enjoyed the constant cooperation and help of Mr. John Haigh, director of Arkhangel'sk Diamonds Corporation for obtaining literature on diamond marketing. Employees of the Aurora Public Library and the University of Wyoming libraries were extremely helpful in obtaining reprints on the geology of diamond deposits. Critical comments from Wayne M. Sutherland and Prof. J. Kenneth were highly appreciated. We also appreciate the excellent editorial work provided by Mary-Margaret Coates, who greatly improved our manuscript.

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SECTION 1

Diamonds

CHAPTER 1

History of Diamond Discoveries

INTRODUCTION

Diamonds have been known since the most ancient times. They have an intrinsic value not only because they are rare, but also because of their extreme hardness and beauty as a gemstone. These exceptional properties have led to numerous legends attributing magical powers to the gem.

It is unknown when mankind first recognized the unique properties of diamond. Possibly the earliest reference to the precious stone is in the Old Testament. The prophet Ezekiel (3:9) writes, “As an adamant harder than flint have I made thy forehead.” *Adamant*, a word that implies extreme hardness, has been used throughout time to describe corundum as well as diamond. Modern gemologists and mineralogists refer to the brilliant luster of diamond as *adamantine*, a word having its roots in the word *adamant*. We do not know for certain if Ezekiel’s reference is to diamond. However, Exodus 28:18 states that a high priest’s breastplate was decorated with several gemstones including diamond. According to Miller (1995), this chapter of the Holy Bible may have been written around 1200 B.C.

Where these early diamonds came from we can only speculate. However, most of the earliest known diamonds were found in India, where the first diamond mines in the world were developed. Some famous diamonds, such as the Hope and Kohinoor, have their origins in the Indian mines.

HISTORY OF DIAMOND DISCOVERIES

India

The first significant diamond mining centers in the world were in India. The richest was the Golconda. The name became a symbol of wealth during the early history of mankind. Legend says that the first diamond found in India was found by a shepherd who sold the stone for a small amount of millet. The new owner sold the stone to another person, and in this manner the stone passed through several hands until it eventually came to someone knowledgeable.

The name *Golconda* came from the name of a fort near Khaiderabad, in south-central India. However, no diamond mines ever existed near this fort. Instead, Golconda

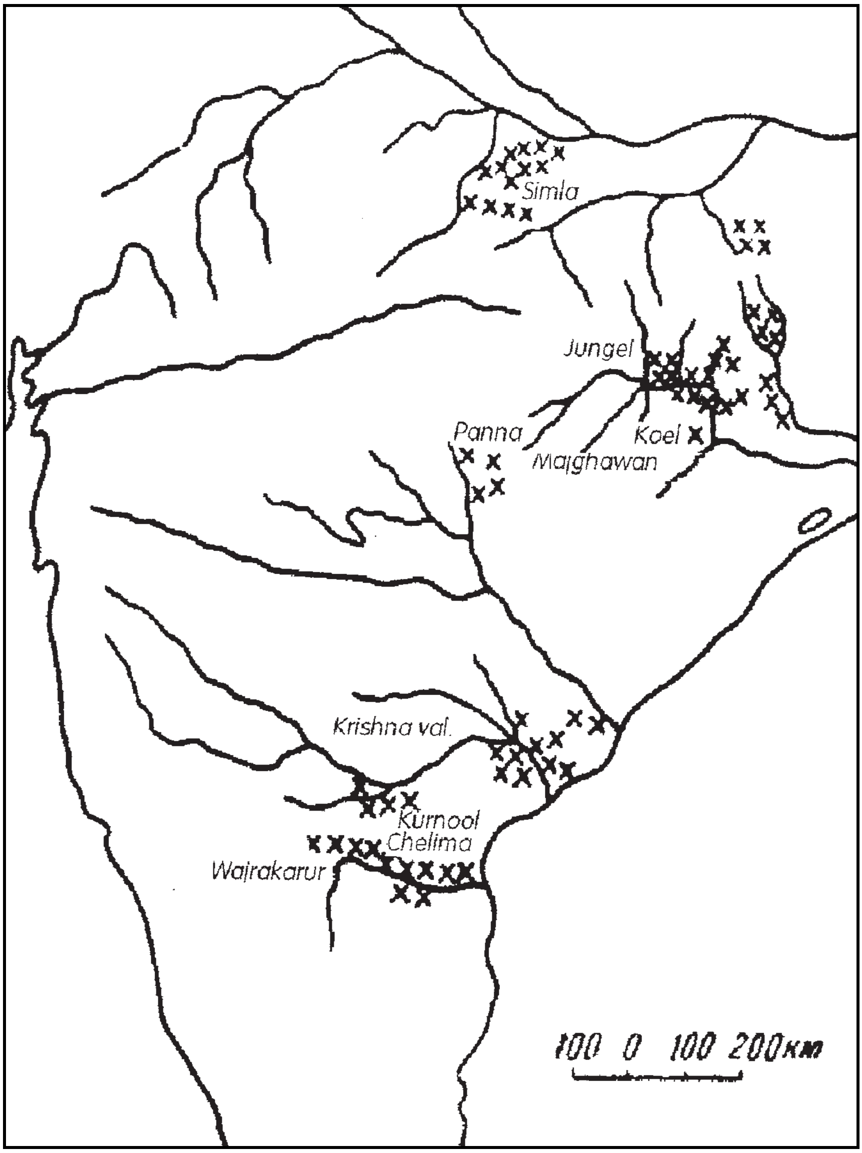

FIGURE 1.1 Diamond localities of India. Placer diamonds locations (modified from Shafranovsky 1964).

was only a marketplace where dealers came to buy and sell diamonds. The actual source of the ancient Indian diamonds were placers in the Krishna (or Kistna) Valley, in the Penner, Karnool, Godvari, and Makhnadi rivers, and possibly in the Panna diamond field to the north (Figure 1.1). Indian folklore and mythology are replete with diamonds. According to tradition, the diamond that later become known as Kohinoor (also spelled

Koh-i-Noor) was in the possession of King Karna, who ruled the Anga empire about 6,000 years ago in Mahabharata time (Mathur 1982).

In 1622, an Englishman named Metgold visited the Golconda region and described a miners' settlement of about 30,000 people. Several mines in the region were supported by a work force of about 100,000 people. Many diamonds were found in this region from medieval time to the nineteenth century that ended up in the royal treasuries of sultans and shahs in India and Persia. The total amount of diamonds mined from India may have been as much as 12 million carats (about 2.4 tonnes) (Milashev 1989).

Diamond exploration and mining in India went through several characteristic stages. The majority of the diamonds was found in placer (stream) deposits (Sakuntala and Brahman 1984). After kimberlite was described in South Africa in 1877 and kimberlite was institutionalized as the original source rock of diamond, intensive exploration in the ancient diamond-producing areas of India resulted in the discovery of kimberlites in areas adjacent to many of the placers (e.g., the Majhagawan and Hinota deposits near the Panna placer district and the Wajakurnur and other kimberlites in the Anantpur district). Many years later, following the discovery that lamproites also can host diamonds, petrographic studies of some Indian kimberlites confirmed that some were actually olivine lamproites (e.g., Majhagawan and Chelima) (Scott-Smith 1989; Middlemost and Paul 1984; Rock et al. 1992). The ore grade of these olivine lamproites was reported to be about 0.1 carats/tonne. Diamonds recovered from the pipes are mostly transparent and flawless, commonly in octahedral and dodecahedral forms. About 40% are gem quality.

According to Mitchell and Bergman (1991), there are several other lamproites, kimberlites, and peridotites in this same region. Rock et al. (1992) also report several lamprophyres (olivine lamproites and minettes) of potential economic interest in eastern India. Known kimberlites are primarily Proterozoic; they include some diamondiferous kimberlites in the Wajrakarur field in the Andhra Pradesh of the southern Kimberlite province and kimberlites in the Raipur field in southeastern Madhya Pradesh in the central province (Middlemost and Paul 1984).

By the nineteenth century, many of the Indian mines were depleted. However, by the mid-twentieth century, geologists had found new deposits there—in particular new placer areas in the Junkel region and Koel Valley and new discoveries in the Simla region near the Himalayas, which had originally been described in Sanskrit texts.

Brazil

The next great diamond rush occurred in Brazil during in the first half of eighteenth century. In 1725, Bernardo Francisco Lobo was attracted to some unusual chips used by local gamblers. He asked the gamblers for some of the chips and was offered them as a gift. Lobo, nevertheless, paid the gamblers for the chips and delivered them to the governor. The governor sent the chips to Europe requesting identification, and the answer was swift and clear: "These chips are diamonds!"

Bernardo returned to the area with some merchants to buy more stones, and the great Brazilian diamond rush began. People abandoned their homes, dug in their backyards, tore plaster from the walls of houses and fences, and began crushing and sieving the material for diamonds.

The diamond rush expanded into adjacent areas, and by the end of 1729 several diamond placers had been found in eastern Brazil in the region of Teune City, which was later renamed Diamantina ("diamond city"). Placers were also found along Rio Abaete

(Sao Francisco River), Bagagem (Parana River), Goyas (Rio Claro), and in other streams in southeastern Brazil.

In 1844, rich diamond deposits were found in another region of Brazil—the state of Bahia to the north. During the first 120 years of mining Brazilian deposits, a total of 10.17 million carats were recovered, including some diamonds greater than 100 carats in weight. The greatest production was in 1850, when 300,000 carats were mined.

The primary source of the diamonds has not yet been found, although it was assumed by many early diamond prospectors that itacolumite, a micaceous sandstone, was the source of the diamonds found in the nearby placers. This assumption was based on the presence of middle Proterozoic diamondiferous conglomerates that have supported extensive mining operations in the Diamantina area.

The extremely large number of diamonds found in the placers suggests that some major primary diamond deposits will someday be found in Brazil. Since 1967, a systematic exploration program reportedly found more than 300 kimberlitic intrusives. However, detailed petrography indicates that only a few of these are true kimberlites, and none of them contain economic amounts of diamond. Many of these are described as kimberlite, lamproite, kamafugite, melilitite, and olivine basalt (Bizzi et al. 1994; Meyer et al. 1994). The location of the principal kimberlites and lamproites in Brazil is shown in Figure 1.2.

South Africa

The next major discoveries were made in South Africa, which claims some of the richest diamond discoveries in history. Diamonds were initially reported from South Africa in the 1860s. The occasional diamonds found along the banks of the Orange River by local Boer farmers and natives in 1867 prompted a London diamond dealer to hire James Gregory of Great Britain to evaluate the region. Gregory traveled to South Africa in 1868 to investigate the source of the stones.

After visiting the Orange River, Gregory published a report in the *London Journal of Society of Arts* stating that he had made a lengthy examination of the new diamond fields. It was his conclusion that there was no possibility of diamond production. He explained that the gems already found had been "salted" by locals attempting to increase the value of their farms, or that many of the diamonds had been transported to the Orange River by ostriches. Where the ostriches had found the diamonds, Gregory did not say. As a result of Gregory's appraisal, the term *gregory* has become entrenched in the African lexicon in reference to any significant blunder (Thomas 1996; Hausel 2000).

And what a blunder Gregory's appraisal was. The Orange River became the greatest diamond-producing region in the world, where several significant and world-class diamond deposits were later discovered. Between 1870 and 1871 a great diamond rush occurred along the Orange River. This rush resulted in discovery of the South African diamond fields, and it ultimately led to the formation of the DeBeers diamond conglomerate (Wagner 1914).

The Orange River diamonds were traced to what the early diamond prospectors thought were dry placer deposits. Unknown to the prospectors, these dry placers represented eluvial detritus produced from the weathering of underlying kimberlite. It wasn't until 1876 that the mines were sunk deep enough to intersect the primary host rock. Up to that point, the diamonds were thought to be associated with dry placers, because the overlying weathered kimberlite contained common rounded megacrysts and xenoliths that gave the impression of a conglomerate. When "blue ground" (weathered kimberlite)

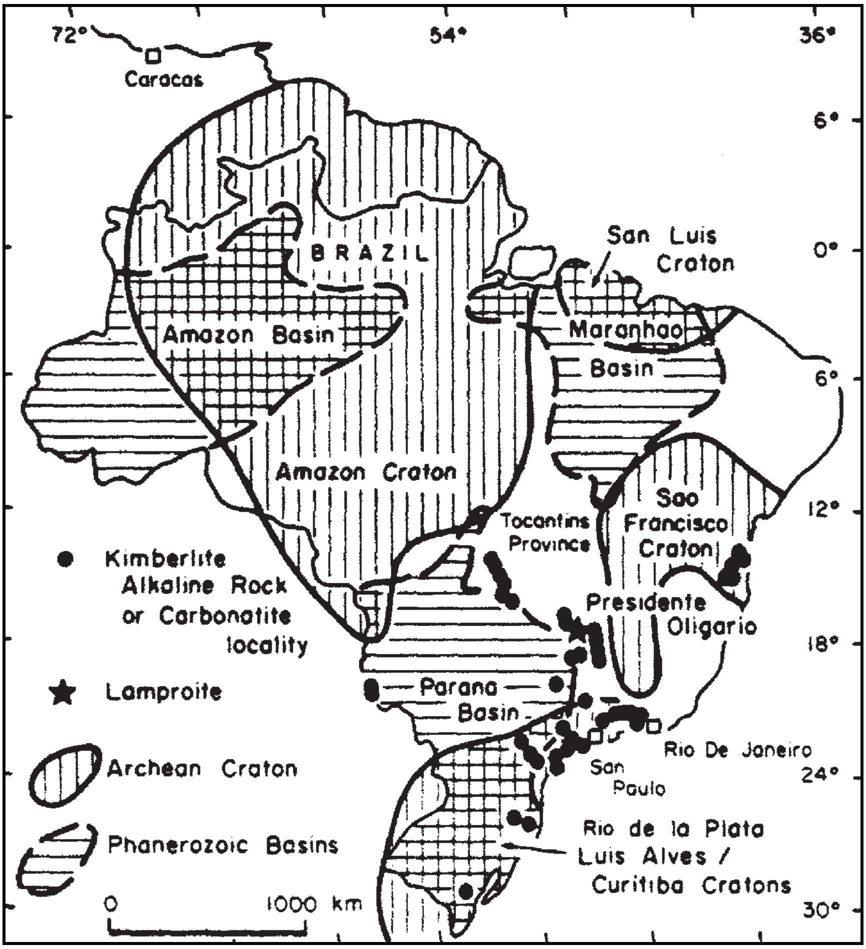

FIGURE 1.2 Kimberlite and lamproite fields in Brazil (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

was intersected at relatively shallow depths beneath the so-called dry placers, many miners mistakenly thought that the mines had hit the bottom of the pay dirt, and several sold out. In reality, the African mines had intersected the weathered clay-rich form of the diamondiferous host rock that was later christened "kimberlite" (Figure 1.3).

As mining continued, it became clear that the kimberlite was associated with pipelike bodies of volcanic origin. Understanding this fundamental fact provided geologists and prospectors with the clues needed to develop exploration methods and models to search for similar deposits elsewhere. These clues, and the later refining of the clues, resulted in all subsequent discoveries of diamond deposits in other regions of the world.

The recognition of kimberlite was an important step in developing effective exploration models to search for diamond deposits. Prior to the development of modern exploration models, diamond source rocks were thought to be restricted to placers. The placer concept was so ingrained in the thinking of diamond prospectors at the end

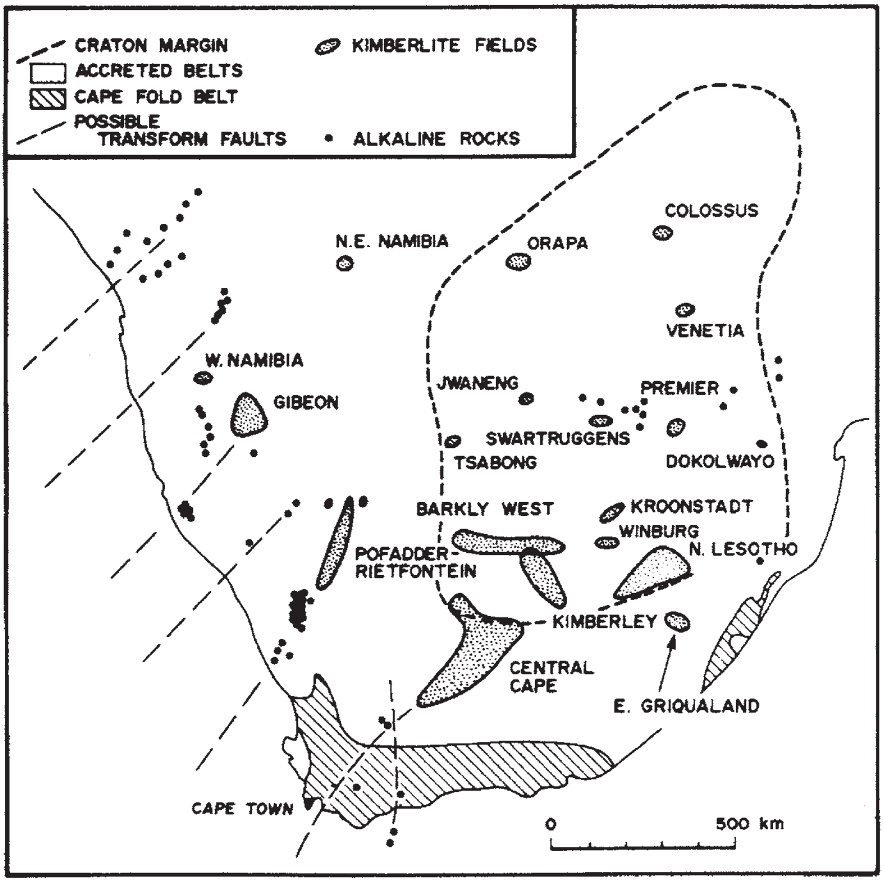

FIGURE 1.3 Kimberlite fields in South Africa (modified from Mitchell 1986)

of the nineteenth century that one author is reputed to have written: "I am sure that diamond host rocks are sedimentary conglomerates. If the opposite will be proven, I'll consider that my lifetime is wasted."

On the basis of extensive fieldwork in the African diamond fields, Wagner (1914) published his classic work providing field observations on the volcanogenic nature of lode diamond deposits. This work and the subsequent summaries of Williams (1932) provided the basis for the modern concepts that govern diamond exploration. The exploration models generated from these ideas led to many discoveries in the twentieth century in other parts of South Africa as well as in Tanganyika, Botswana, Angola, Namibia, West Africa, Australia, Canada, the United States, Russia, and Venezuela.

Russia

On the basis of the diamond exploration concepts developed in South Africa, several diamond deposits were later discovered in the former Soviet Union that resulted in Russia becoming a diamond-mining giant.

FIGURE 1.4 Major kimberlite sites in Russia (modified from U.S. Geological Survey 1997). * designate sites of early discoveries: western, Urals region; eastern, Enysey region. Boxed areas are shown in Figure 1.5 (Siberia) and Figure 1.6 (White Sea). Names of main diamond-mining cities in italic.

The first diamond in Russia was reported in 1829 in the vicinity of the Bisertsky gold mine in the Permsky region of the Urals Mountains. A 14-year-old boy named Pavel Popov took a diamond to the manager of the gold mine. Soon, two other small diamonds were found in the same area. The diamonds were presented to a famous German geographer, Alexander Humboldt, as proof of the immense Russian richness. Humboldt later presented the diamonds to the tsar's wife.

Another discovery was made close to the Promisel village, located on the western slope of the Urals, and a third discovery was made close to the Severnaya village within 12 km of the first discovery. Diamonds were later found in other gold placer deposits in the Urals Mountains (Figure 1.4). During the first 50 years following the initial diamond find, about 100 crystals were found; the largest weighed less than 2 carats.

Many years later, other diamonds were found in the black sands of Great Pit River (a tributary to the Enysey River). The first was found in 1898. Up to the time of the 1917 Revolution, about 250 crystals had been discovered, the largest of which weighed about 25 carats. The number of diamonds suggested that some primary diamond deposits would someday be discovered.

During the period of Russia's industrialization, the demand for diamonds increased. Starting in 1937, a highly detailed plan was developed for the exploration of placer diamonds along the western slope of the Middle Urals Mountains near the Bisertsky gold mine. Owing to these efforts, several small (both in grade and reserves) placers were discovered, and commercial mining of these placers started at the time of World War II.

During the course of geologic mapping in the early 1930s, the famous Russian polar geologist Nikolay Urvantsev described kimberlite-like rocks in the southern Taymir

Mountains north of the Central Siberian Plateau. Another geologist, Georgy Moor, described rocks similar to South African kimberlites within the northern part of the Siberian platform south of Urvantsev's discovery (Moor 1940, 1941). This description was the first direct indication of potential diamondiferous rocks in the Siberian platform. But because Moor was considered to be no more than a good descriptive petrographer, and Urvantsev spent only minor amounts of time in the field between serving terms in labor camps, neither observation was considered seriously by the socialist government. As it turned out, these rocks were later classified as alnöites.

In 1937, owing to the initiative of a distinguished Russian geologist A.P. Burov, a program was started to compare the geology of the known diamond fields of the world to the geology of the USSR. This program was initiated in the All-Union Geological Institute in Leningrad. Among its participants were petrologist V.S. Sobolev, the future academician who had just finished a study of Siberian Plateau basalts (Sobolev 1937). Sobolev's report to the Central State Planning Committee recommended that an extensive exploration program for diamond deposits be initiated within the Siberian platform.

World War II postponed the exploration for diamonds. Following the war, in 1946, the Ministry of Geology organized a special branch of exploration for optical minerals and diamonds. This special branch, like others created from time to time during Stalin's era, received the highest priority for funds, equipment, and personnel (including substantial personnel from forced labor camps). It was the ministry's practice to organize specially named expeditions, or working groups, for periods of several years. Each expedition had a relatively narrow goal and a fixed time frame.

In 1948 one expedition (organized in 1945 in Irkutsk under the leadership of Mikhail Odintsov, with Vladimir Belov and Grigory Fainstein as heads of geologic teams) led to the discovery of the first diamond in Siberia. The diamond was found in one of the tributaries of the Nizhnyaya-Tunguska River. A short time later, several small diamonds were found along other tributaries of the Tunguska River. Because no rich placer or lode had been discovered, it was decided to move the survey to the northeast and explore the Viluy River. On August 7, 1949, the geologic team led by Grigory Fainstein found the first diamond within the Sokolinaya kosa (spit) in the Viluy River system. By the end of the field season, 25 diamonds had been found in the black sands of the riverbank Sokolinaya on the Viluy River (Milashev 1989).

In 1950, another special expedition was established in the Nyurba settlement on the bank of Viluy River to coordinate all regional exploration for diamonds. But the diamond exploration efforts didn't go any further than the ordinary washing of black sands until two events led to the start of a bona fide exploration rush.

In the course of geologic mapping at a 1:1,000,000 scale, a geologic team led by Konstantin Zaburdin described an outcrop of dark green rock on the bank of a tributary of the Olenek River to the north of the Viluy. Zaburdin, who was a very experienced but not a well-educated geologist, knew that all types of limestones in this region belonged to the Cambrian Period. But this kind of rock was unusual so he submitted it for thin-section description to the Research Institute of Arctic Geology. The thin section was described as kimberlite by a talented (and hard drinking) petrographer named Vladimir Cherepanov. This identification was the first verification of kimberlite in this region.

The rock name was unknown to Zaburdin, so he sought the advice of the chief of the petrographic lab, Mikhail Ravich. Ravich not only recognized the name, but he also realized the implications of finding kimberlite. He exclaimed, "Do you know what the term kimberlite means? It means diamonds! Call these rocks eruptive tuffs related to flood basalts. I think that we should be very careful. Cherepanov was obviously all wet when he wrote this description."

Consequently, kimberlite was not identified on the State Geological Map of the USSR, sheet N-56-57, published in 1954. Instead this sheet shows a dark green spot in a sea of pink Cambrian sedimentary rocks to the east of the Anabar shield. This spot, necessarily exaggerated on the map, depicts the now famous "Leningrad" kimberlite pipe, one of two major exposed pipes in the Siberian platform. This event occurred 2 years before the official discovery of the platform's first kimberlite pipe, the Zarnitsa.

In the meantime, within an area now known to contain several major kimberlite pipes and most of the current diamond mines of Olenek-Viluy watershed, another expedition from the Research Institute of Arctic Geology mapped an area on a 1:200,000 scale. In the course of mapping, one of the expedition's mineralogists, Yagna Stakhevich, found a group of pyrope grains in black sands. Mikhail Rabkin, the scientific curator of the expedition, advised her to be cautious and to call the grains "garnets of almandine-pyrope association" and suggested that they could have originated in the Anabar crystalline shield (Erlich and Slonimsky 1986).

In the same area, a team of two women mineralogists from the All-Union Geological Institute in Leningrad, Natalya Sarsadskikh and Larisa Popugayeva, worked on heavy minerals in some soils of the Siberian platform. They worked with one of most prominent Russian mineralogists, Alexander Kukharenko. In the autumn of 1953, a boat with the two mineralogists visited the expedition camp on the Olenek River. Yagna Stakhevich enthusiastically told them about her pyrope discovery and provided a map of pyrope distribution. Over the winter, the pyropes from her samples were confirmed in the laboratory.

During the 1954 field season, Natalya Sarsadskikh remained in Leningrad on maternity leave. Her graduate student, mineralogist Larisa Popugayeva, went to the field armed with Stakhevich's map. From that point on, the discovery of kimberlite was just a technical problem. Pyrope garnets that had been found in heavy-mineral samples taken from the creek bed were traced up to the point where the mineral was no longer present in black sands. At that point, the black sands recovered from the soils in the bank of the creek were sampled. This resulted in the first acknowledged kimberlite pipe, which was named "Zarnitsa" ("summer lightning"). Rumors about Popugayeva's discovery traveled ahead of her. Even the chiefs of the Amakinskaya expedition helped her to obtain food and transported it to the field area.

They also reminded her of this when the time came to divide the glory of discovery. At the end of the 1954 field season, Popugayeva wrote her report on the season's work. Upon reviewing the report, the chiefs of the Amakinskaya expedition called in Popugayeva and gave her a choice: either sign an application (predated to the beginning of 1954 field season) to join the expedition, or never work in the area again. The student accepted the first choice, because she had hoped to use the discovery of the kimberlite in a thesis project.

However, this woman, who had survived the rigors of antiaircraft artillery in World War II and had worked harsh field seasons in Siberia, was not able to overcome the wiles of desk-bound empire builders, who now claimed for their organization the credit for the discovery. As a result, all bonuses and fame for finding the USSR's first kimberlite was given to the Amakinskaya Expedition. Popugayeva was relegated to the background and ignored by the expedition. Furthermore, she was fired from the All-Union Geological Institute, whose directors considered her as a traitor who robbed them at the last moment of credit for the discovery. Her only publication about the discovery was written with Sarsadskikh (Sarsadskikh and Popugayeva 1955). The names of Kukharenko and Stakhevich were never mentioned. But there is a belated happy ending to the story. A reporter heard about Popugayeva's role in the discovery, and she was later awarded the Order of Lenin for her accomplishment.

Following the discovery of kimberlite, diamond exploration began in earnest, guided by a clear concept of the source of diamonds in the region, and by the simple exploration technique of tracing pyrope in black sands. The chiefs of the Amakinskaya expedition, who previously had exhorted geologists on the forthcoming anniversary of the Revolution to redouble their efforts to discover and sample differentiated doleritic intrusions, which were considered potential sources of diamonds, now redirected their teams to seek only kimberlite.

During the next 30 years Russia became the second largest diamond producer in the world, and nearly all its diamond production came from mines within the northern Siberian platform (Figure 1.5). A series of mines was developed in Yakutiya, namely the Mir, Udachnaya, Aikhal, Sytykanskaya, and Internatsional'naya (Yakutalmaz 1999).

In 1954, the Mir pipe was discovered. Development began 3 years later in 1957 on the associated placers. Development was followed by open pit operations in the pipe itself. Operations ceased many years later at the depth of 340 m when the pit's bottom was flooded. The average ore grade was unusually high in the upper mine levels: 4.0 carats/tonne (400 carats/100 tonnes) but decreased near the pit's bottom to 1.5 to 2.0 carats/tonne (150–200 carats/100 tonnes). The Mir had a high gem content and was the source of several large gem diamonds, which included the largest Yakutian diamonds such as the Star of Yakutiya (232 carats) and the Diamond of Twenty-Sixth Party Congress (342.57 carats). Annual output from the mine was 6.0 million carats (Miller 1995).

In 1955 another famous pipe, the Udachnaya, was found. Mining began on the associated placers in 1957, which was shortly followed by open pit operations on the pipe. Udachnaya has been the most productive diamond mine in Russia (annual production, 12 million tonnes) since 1978. Total production from the pipe was 14.4 million carats of which 80% were gems.

In 1960, the Aikhal pipe was discovered. Mining began on the pipe in 1962 and ceased between 1981 and 1988, presumably owing to overproduction from other sources. However, production resumed after 1988, and by 1995 the pit reached a final depth of 240 m. Annual production at the peak of mining was 1,000 tonnes of ore from which 600,000 carats were recovered each year. The average grade was reported to be 1.0 carats/tonne (100 carats/100 tonnes) (Yakutalmaz 1999).

Sytykanskaya (sometimes spelled *Sytykan*) was discovered in 1955. Open pit mining began in 1979, and annual production reached 1,000 tonnes (600,000 carats) at an average grade of 0.6 carats/tonne (60 carats/100 tonnes). The mine is currently (2001) facing closure owing to exhausting existing ore reserves. (The reported annual production and average grade do not agree in the original reference.) (Yakutalmaz 1999).

The next important pipe to be discovered was the Internatsional'naya pipe. It was discovered in 1969, mining operations began in 1971, and the open pit was developed to a depth of 280 m, which was reached in 1980. After displacement of 70 to 110 m of overburden, mining continued in the open pit. The mine was planned as a replacement of the Udachnaya pipe. Construction of an underground mine is currently underway.

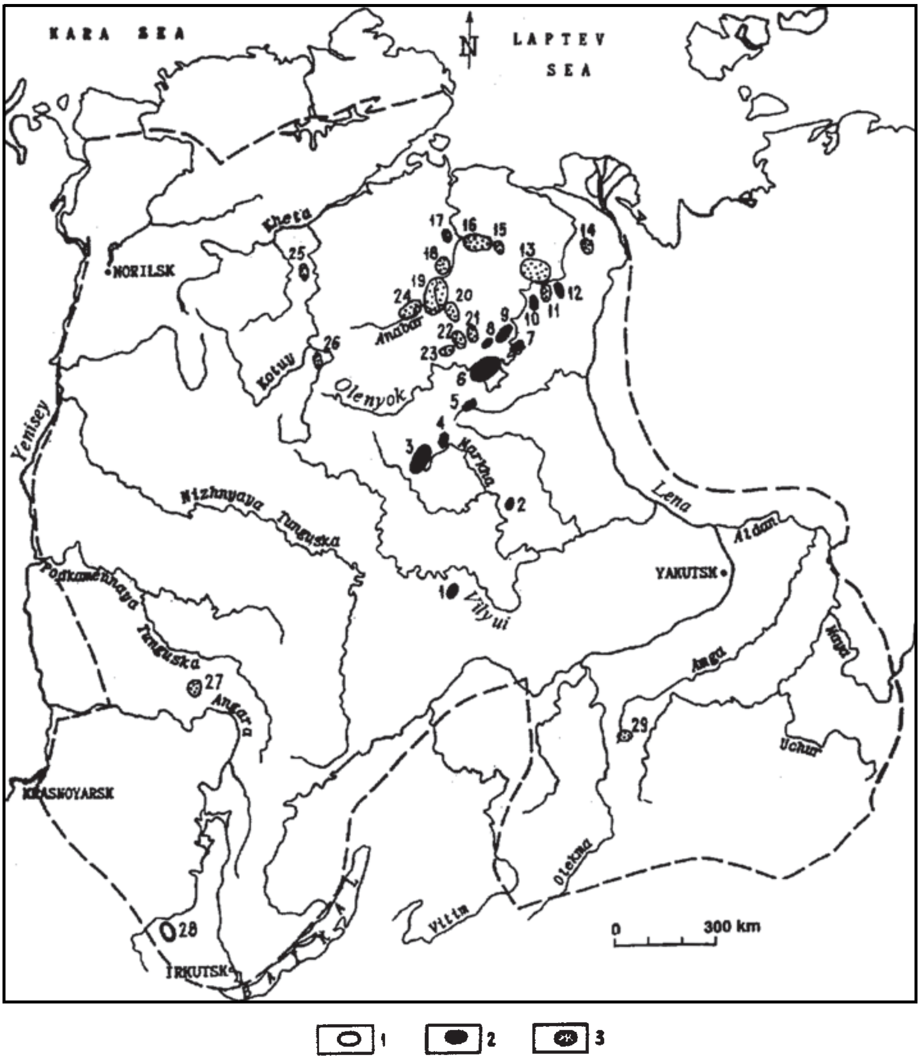

FIGURE 1.5 Main Russian diamond mine sites and kimberlite bodies within Siberian platform. 1 = Proterozoic, 2 = Middle Paleozoic, 3 = Mesozoic. Map: 1 = Mirny, 2 = Sredny Markha, 3 = Alakit, 4 = Daldyn, 5 = Verkhny Muna, 6 = Chomurdakh, 7 = Ogoner-Yuryakh, 8 = West Ukukit, 9 = East Ukukit, 10 = Merchimden, 11 = Molodo, 12 = Toluop, 13 = Kuoika, 14 = Khorbusuon, 15 = Tomtor, 16 = Ebelyakh, 17 = Orto-Yargin, 18 = Starorechenskoye, 19 = Ary-Mastakh, 20 = Dyuken, 21 = Luchakan, 22 = Birigindin, 23 = Kurannakh, 24 = Anabar, 25 = Kotuy, 26 = Kharamai, 27 = Chadobets, 28 = Ingashin, 29 = Tobuk-Chompolin. (Modified from Sobolev et al. 1995.)

The Yubileinaya (Jubilee) pipe was discovered in 1975. After stripping operations, which removed a 70 to 100-m-thick basalt overburden, open pit mining started. It has been suggested that Yubileinaya will replace production from the declining Udachnaya pipe.

The Twenty-Third Party Congress pipe was discovered in 1959. Mining began in 1966 on this deposit; the average ore grade was 6 carats/tonne (600 carats/100 tonnes). The open pit reached a final depth of 124 m after 15 years of operation.

Later, huge deposits of a new genetic type of industrial diamonds were found in association with the Popigay astrobleme and in metamorphic rocks in Kazakstan (see Section 3, Unconventional Source Rocks). These deposits remain to be fully evaluated.

During the 1970s diamondiferous kimberlites were discovered within the Russian platform. The location of the main kimberlite fields south of the White Sea is shown in Figure 1.6. At about the same time several kimberlitic pipes were discovered northeast of the city of Arkhangel'sk, which included the Lomonosov diamond deposit consisting of several pipes.

Venezuela

In the mid-twentieth century, years after the great discoveries in South Africa, diamond deposits were found north of Brazil in Venezuela. Near the end of the 1960s, a placer deposit was found on Caroni River in southeastern Venezuela. Two months following this discovery, a riverbank mining settlement had grown to a population of 8,000. To mid-1969, 1.3 million carats of diamonds had been mined; the largest stone weighed 12 carats. The location of diamond placers is shown on Figure 1.7. In September 1971, near the town of Salvacion in the state of Bolivar, another significant placer deposit was discovered. Within a short period, monthly diamond production from the Salvacion region reached 50,000 carats.

In total, Venezuelan deposits have yielded about 6 million carats. The production figures don't even consider the great quantity of diamonds that were smuggled out of the country. The primary source for the diamond deposits remains to be discovered.

United States

To date, diamond discoveries in the United States have been primarily of scientific interest; however, the presence of large Archean cratons and Early to Middle Proterozoic mobile belts in and around the Wyoming and Superior provinces makes us anticipate some significant discoveries in the future. Additionally, a deep subduction zone along the West Coast of the United States may also provide some favorable diamondiferous host rocks in that region.

The first diamond discoveries in the United States were made in association with placer gold mining in the Georgia, North Carolina, South Carolina, and Alabama gold fields. A short time later diamonds were found in the California gold fields and much later in the Great Lakes region. Diamonds were later found in place in Arkansas, Wyoming, Colorado, Michigan, and Wisconsin, and more recently in Montana. They have also been reported in other regions of the United States, although these diamonds remain to be verified. Diamonds have been reported in 27 of the 50 states (Figure 1.8). But in nearly every case, various researchers and writers write off the diamonds as being derived from some far distant place. Luckily, the United States does not have any native ostriches, and many conclusions rely on little or no scientific fact (see section above on South Africa).

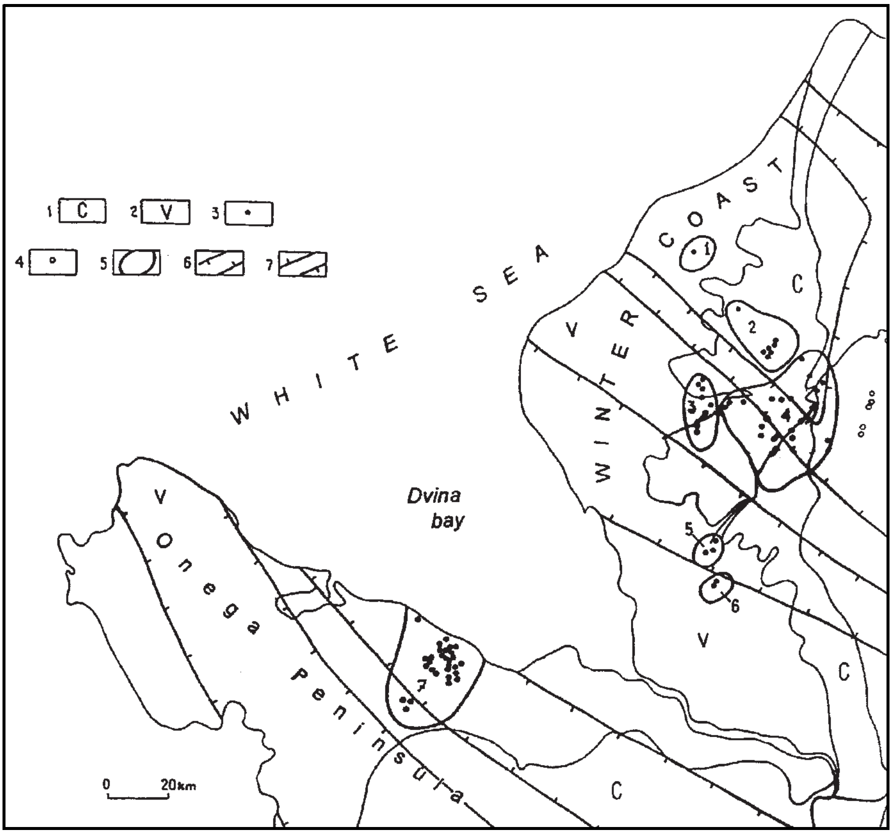

FIGURE 1.6 Kimberlite fields south of the White Sea. 1 = Carboniferous limestones and sandstones; 2 = Vendian sandstones and siltstones; 3 = kimberlites, picrites and olivine, melilites (pipes and sills); 4 = basaltic picrites, kimberlites and related rocks (1 = Mela; 2 = Verkhotina; 3 = Zolotitsa [Lomonosov deposit]; 4 = Kepino; 5 = Chidva, Izhmozero; 6 = Nenoksa; 7 = group of pipes at Onega Peninsula); 5 = contours of geological formations; 6 = crystalline basement uplifts; 7 = crystalline basement depressions. (Modified from Sobolev et al. 1995.)

Cratonic basement rocks underlie large portions of the United States, and portions of the West Coast are underlain by a deep subduction zone. These geologic environments are considered favorable terranes for the discovery of in situ deposits. Most past U.S. diamond finds were related to gold prospecting and mining, whereas many of the recent discoveries have been the result of modern scientific exploration in areas that had little to no history of gold prospecting.

Diamonds, having a relatively high specific gravity compared with quartz and calcite, are commonly trapped in sluices and gold pans along with gold and black sands. However, the specific gravity of diamond is relatively low compared with that of gold, which results in many diamonds reporting to the tailings in many placer mines and being lost.

Georgia. Diamonds were first reported in the Georgia gold fields in 1843. Rumors of diamonds weighing more than 100 carats from this region cannot be verified. However,

FIGURE 1.7 Diamond placers in Venezuela (modified from U.S. Geological Survey 1997)

smaller diamonds were found in Georgia and in neighboring Alabama, Kentucky, North Carolina, South Carolina, Tennessee, Virginia, and West Virginia (Sinkankas 1959). The diamonds from the Appalachian region ranged from 0.25 to 34.46 carats (Hausel 1995a, b).

Appalachians. The source of the diamonds found in the Appalachian region is unknown (Hausel 1998), but reports of kimberlitic indicator minerals in some regions suggest the presence of nearby undiscovered mantle source rocks. Some kimberlites were identified in eastern Kentucky during the late 1800s, and a shallow shaft was sunk on one of these in 1885. The area received sporadic interest, but it is not known if any diamonds were actually recovered, although local residents claim that three diamonds

FIGURE 1.8 Location of main diamond-bearing kimberlites, lamproites, and related rocks in the United States (modified from Hausel 1998)

were found by the Kentucky Kimberlite Diamond Mining Company in 1907. Another diamond that was later cut into a gem was reportedly found in a nearby drainage (James E. Bond, personal communication, 1994).

The geochemistry of the garnets from the Kentucky kimberlites suggests a very low grade diamond source, because most peridotitic garnets are calcic-chrome pyropes of lherzolitic affinity (see Chapter 15, Exploring for Diamond Deposits). However, a few subcalcic chrome (G10) pyropes have also been reported, which supports a minor contribution from the diamond stability field.

At about the same time, kimberlite was found in Kentucky, and one of the largest diamonds in the United States was discovered nearby in 1885. The Dewey diamond was found in the James River valley near Manchester in eastern Virginia. The site of the discovery is now part of the city of Richmond (Sweet 1997). On the basis of little evidence, Blank (1934) suggested that the diamond may have been transported by spring floodwaters of the James River from the Virginia gold fields.

Another large diamond was found in neighboring West Virginia in 1928. This diamond, the Punch Jones, weighed 34.46 carats and was found in Peterstown, near the Virginia state line (*Washington Post*, February 6, 1964).

The source of these and other Appalachian diamonds remains unknown. The Appalachians are underlain by relatively young Precambrian (Lower Proterozoic) basement of the Grenville platform, which is considered to have low potential for the discovery of diamondiferous kimberlite, lamproite, alnöite, or peridotite. Normally, the Appalachian country would not be considered high priority for diamond exploration. However, the discovery of several sizable (pebble size) diamonds suggests this region should be examined. The discovery of diamonds in alnöite in a similar terrane at Ile Bizard, Quebec, Canada (Raeside and Helmstaedt 1982), suggests that similar discoveries might be possible in the Appalachians. Some early reports claim diamonds were found in peridotite in New York and Maryland (see Hausel 1995a, b). These reports, however, have not been verified.

Even though there are potential host rocks for diamonds in the Appalachians, past researchers and writers have suggested that the diamonds originated from a very distant source and were transported from some unknown area by streams, glaciers, and even migrating birds from South America. Some have even suggested the possibility of the diamonds originating in Africa, before the breakup of the supercontinent of Pangea during the Cretaceous era. Other writers have suggested that the diamonds were the product of fraud, or were derived from nearby outcrops of itacolumite (micaceous sandstone) (Hausel and Bond 1994). We suspect that many of these suggestions are unfounded.

California. Following the early Appalachian discoveries, diamonds were found along the west coast of California, Oregon, and Washington. California, in particular, produced a relatively large number of diamonds.

Diamonds were discovered in 1849 near Placerville, California. Three years later, in 1852, diamonds were recovered from sluice boxes at the Cherokee hydraulic gold placer mine, 13.6 km north of Oroville and 96 to 128 km northeast of Placerville. In total, more than 600 diamonds were found in 15 counties in the state and diamonds have even been identified in beach sands south of San Francisco (Hutton, 1959). Most of the diamonds were found north of the 36°N latitude, and nearly all have been found in the Sierra Nevada and the Klamath Mountains downstream from serpentinized ophiolite complexes and melanges. Less than 1% of the diamonds were found in beach sands.

The most productive area was the hydraulic gold placer mine immediately north of Oroville in the Round Mountain area. Between 1853 and 1918, about 400 diamonds and 600,000 ounces of gold were recovered by mining operations in this area situated near the Feather River. In 1906, the U.S. Diamond Company sunk a 54.5-m-deep shaft in what was reported to be kimberlite. The shaft reportedly intersected a rich pocket of diamonds, although it is questionable whether any diamonds were recovered (Rosenhouse 1975). Kimberlite is not known in the area, but serpentinized amphibolite has been described (Heylum 1985).

The largest recorded diamond from in the Oroville area, the Moore diamond, was found in 1868. This diamond weighed 2.25 carats. However, local folklore claims that the largest diamond was a 6-carat stone found outside the entrance of the Spring Valley mine at Thompson Flat (Rosenhouse 1975; Hill 1972).

In recent years, five diamonds were recovered from Hayfork Creek, a tributary of the South Fork of the Trinity River in the Klamath Mountains of northern California. The largest was found in 1987 and named Doubledipity. The Doubledipity weighed 32.99 carats (Kopf, Hurlbut, and Koivula 1990). Four other relatively large diamonds were recovered from Hayfork Creek: the Enigma (17.83 carats), the Serendipity (14.33 carats), a poorly documented diamond discovered in the 1860s that was about a 2.5 cm in diameter and estimated to weigh 10 to 15 carats, and the Jeopardy diamond (3.9 carats).

These large stones exhibit an overgrowth of multiple tiny crystal faces that coat the underlying crystal or aggregate. The morphology of the crystals indicates that they were subjected to disequilibrium conditions within the diamond stability field, typical of diamonds thought to have originated from a subduction zone. The form of the diamonds also suggests minimal stream transport distance and supports a nearby source (Kopf, Hurlbut, and Koivula 1990).

In the mid-1990s, diamondiferous lamproite was reportedly discovered at Leek Springs near Placerville. This discovery has not been verified, and the source of the diamonds on the West Coast still remains unknown.

Great Lakes Region. Another region of the United States that has produced some notable diamonds is the Great Lakes area. Much of this region is underlain by the Superior platform, an Archean cratonic core that extends under much of Minnesota and eastern South and North Dakota, continuing north into Canada. Early to Middle Proterozoic cratonized basement rocks underlie the eastern and southern margin of the Superior Archon, which is bounded by Late Proterozoic rocks of the Grenville platform farther east. The platform extends into eastern Michigan and Indiana.

Famous diamonds from this region include the Dowagiac diamond. This diamond was found in Michigan near the Indiana border and weighed 10.875 carats (Sinkankas 1959). According to Blank (1934), this stone, as well as the Burlington and Saukville diamonds from Wisconsin, were recovered from the Lake Michigan moraine. Since 1876, approximately 25 diamonds have been found in seven regions of southern and central Wisconsin. All were found in Pleistocene glacial deposits or in Holocene river gravels.

Another diamond found in Wisconsin was the Terresa. This diamond was found near Kohlsville in southeastern Wisconsin in 1883, a few kilometers west of the Saukville diamond. The Terresa weighed 21.25 carats and was found near the Green Lake moraine (Cannon and Mudrey 1981).

Another large diamond was found in a 96-m well collared in glacial drift at Eagle, near central Wisconsin, in 1883 (Kunz, 1885). The diamond was slightly warm yellow and was recovered from hard, yellowish ground in the well. However, Sinkankas (1959) reported that the Eagle diamond was a dodecahedron weighing 15.37 carats that was found in the Kettle moraine in 1876. Two other diamonds were found in this area, each weighing less than 0.5 carat. Other diamonds were reported from the Indiana gold fields to the south.

Early investigators suggested that essentially all of the diamonds from the Great Lakes region had been transported from Canada by continental glaciers during the last ice age. But the discovery of several kimberlites in Michigan in the 1980s and other kimberlites in Wisconsin and Illinois in the 1990s suggests that the source of many of these diamonds may be nearby kimberlites or related mantle-derived intrusives. For instance, the *Northern Miner* (September 4, 1995) reported 26 kimberlites in Michigan, Wisconsin, and northern Illinois, eight of which contained subeconomic quantities of diamond. Eleven magnetic anomalies that are suggestive of buried kimberlite pipes were also identified in the region. Michigan also has some Paleozoic outliers that are completely surrounded by Proterozoic age rocks interpreted as cryptovolcanic structures possibly related to kimberlite (Carlson and Floodstrand 1994).

Another intrusive, the Six-Pak diatreme, was discovered in Wisconsin on the basis of an airborne magnetic anomaly. The rock forming the pipe is described as ultramafic lamprophyre (melnoite) with typical kimberlitic indicator minerals. The pyrope garnets that have been tested are calcic G9 lherzolitic garnets. Even so, some diamonds were recovered from the intrusive (Carlson and Adams 1997). Other diamond deposits will undoubtedly be found in the Superior platform in the future.

Ouachita Mountains. Diamonds were discovered in 1906 along the edge of the Ouachita Mountains in southwestern Arkansas near the mouth of Prairie Creek, 4 km southeast of the town of Murfreesboro. A farmer, John Wesley Huddleston, was panning for gold when he noticed an unusual white grain in the black sand concentrates in his pan. He took the crystal to the local bank where it was placed in the care of the bank president. The mineral was later shown to George F. Kuntz, vice president of Tiffany of New York, who identified it as diamond.

The diamond was traced to a weathered serpentinized outcrop. Discovery of the source rock led to a diamond rush that rivaled some of the western gold rushes. In one year, the Conway Hotel in Murfreesboro reportedly turned away more than 10,000 people. Many of the diamond seekers set up tents between Murfreesboro and the Prairie Creek diamond pipe, and they aptly named their tent city Kimberley.

A few years later, in 1911, a similar deposit was discovered 3.2 km northeast of the Prairie Creek pipe. This pipe was named the Kimberlite mine, on the basis of apparent similarities to South African kimberlites. In 1919 a geologist and mining engineer, Austin Millar, and his son Howard constructed the first diamond mill in North America with the financial backing of the Ozark Mining Company. More than 1,000 diamonds, many gem quality, were recovered during the first 3 months of operation. Mining operations continued until the plant burned to the ground. According to Miller (1976), the fire was described as an act of arson and coincidentally occurred at about the same time as the arrival of a DeBeers representative.

Following the fire, Stanley Zimmerman, the mine manager of the reorganized Arkansas Diamond Mining Company, reported that the company had been promised as much cash as needed from J.P. Morgan and Company to rebuild the mill. However, 3 months later, Zimmerman returned from a meeting with Sir Ernst Oppenheimer, head of DeBeers in South Africa, and within a short time he (Zimmerman) coauthored an article in the Engineering and Mining Journal condemning the Arkansas deposits as unprofitable. Some said that he had been paid to prove the deposits were uneconomic (Miller 1976). Interest was renewed in 1923 when Henry Ford offered $1,000,000 for it on the condition that he would receive title to the property. However, Austin Millar was unable to persuade all of his partners to sell (Wilcox and Young 1981).

In 1949 the site of old Ozark mine was leased by Millar, who opened the mine as a tourist attraction until 1972, when it was purchased by the Arkansas Department of Parks and Tourism for $750,000. Currently, the Arkansas Department of Parks and Tourism operates a park for tourists who pay a minimal fee to dig for diamonds.

The rock that forms the Prairie Creek, Arkansas, pipe has been described by various authors as peridotite and kimberlite, primarily because it contained diamonds. But detailed petrographic studies resulted in the reclassification of the rock as olivine lamproite (Mitchell and Bergman 1991). Five other lamproites have also been recognized in the area, located along the northeasterly trend of the Prairie Creek intrusive (Hausel 1998). The acknowledgment of lamproites in Arkansas is part of worldwide "lamproitic revolution" that followed the discovery of diamondiferous lamproites in Australia.

The Murfreesboro area is one of the two most productive diamond areas in the United States and has produced more than 90,000 diamonds. Diamonds from the Arkansas lamproite include 30% gems and 70% industrial stones. No attempt has ever been made to recover microdiamonds (Sinkankas 1959). The grade of the pipe is estimated at only 0.11 carats/tonne (11 carats/100 tonnes) with an average diamond size of 0.26 carat. For the most part, the deposit has not been worked commercially, and on the basis of the ore grade the deposit would be considered uneconomic to marginal, at best.

Some of the larger diamonds recovered from the lamproite include the largest verified diamond found in the United States, the Uncle Sam (40.42 carats). Other large diamonds include the Star of Murfreesboro (34.25 carats), the Amarillo Starlight (16.37 carats), and the Star of Arkansas (15.24 carats). Most of the diamonds are white, yellow, or brown, and the most common form is a distorted hexoctahedron with rounded faces (Bolivar 1984; Kidwell 1990).

Rocky Mountains Great Diamond Hoax. In 1871, the report of a major diamond discovery in the West attracted interest all over the continent. The location was not initially revealed; but hundreds of diamonds, rubies, sapphires, emeralds, and pyrope garnets were displayed to potential investors in New York by a couple of prospectors working near the Union Pacific railroad in the Wyoming Territory. Excitement generated by the diamond discovery resulted in the formation of more than 25 diamond companies. One of the companies, the Golconda Mining Company, was formed on October 31, 1871, and listed several affluent members on their board of directors—former Civil War generals, a U.S. senator, and two former presidential candidates. The discovery was significant enough that it led to the passing of Sargent's mining law, the precursor to the 1872 mining law.

After various agreements were signed and advance payments were made, the prospectors led a group of investors (the Roberts' Party) from the Golconda Mining Company to the diamond prospect at the base of a high mountain that is today known as Diamond Peak. Diamond Peak lies in the extreme northwestern corner of Colorado several kilometers south of the Union Pacific railroad. During the trek, hundreds of diamonds, rubies, sapphires, and pyrope garnets were collected by the investors. Because of the success of the trek, the prospectors were paid additional funds amounting to a few hundred thousand dollars for their discovery.

But the discovery was a fraud. The fraud was revealed by a group of geologists and surveyors from the Fortieth Parallel Survey. Members of the Fortieth Parallel Survey found the site on the basis of information they had obtained from the Roberts' Party. The survey members located the site and dug about 30 test pits around the "diamond discovery." After a thorough field investigation, it was concluded that the stones had been salted. At an emergency executive meeting of the Golconda Mining Company in 1872, Clarence King of the Fortieth Parallel Survey presented several facts to the Board of Directors of the Golconda Mining Company supporting fraud. The facts included the following: (1) The nearly uniform numerical ratio between rubies and diamonds found on the property was considered impossible. (2) Ninety percent of the gems were found on a sandstone outcrop. (3) The gems in the soils did not continue to bedrock. (4) Ruby Gulch, which drained Table Rock (the site of the discovery) down to Arnold Creek, was extremely rich in rubies at the drainage head, but the gemstones did not persist below the topsoil level or down drainage. (5) Microscopic examination of sandstone samples from Table Rock contained no gemstones. (6) Anthills in the immediate area containing gemstones were surrounded by footprints and often showed artificial holes suggesting that someone had pushed the gems into the anthill with a stick, whereas anthills outside the disturbed area contained no gemstones. (7) The variety of minerals included four types of diamonds, Oriental rubies, garnets, spinels, sapphires, emeralds, and amethysts, all occurring at one locality. (8) Prospect pits dug all around the Table Rock outside the disturbed area produced no gemstones.

By researching various records, it was discovered years later that the majority of the gemstones used to salt the property at Diamond Peak had been purchased in London.

Many other gemstones used in the salt were collected from a group of ultramafic breccias in the Navajo volcanic field near Fort Defiance, Arizona. It is an extraordinary coincidence that these diatremes contain pyrope garnet and chromian diopside that were also used in the salt, because there was no known relationship between such "kimberlitic indicator minerals" and diamond deposits at the time of the salting. In 1871, the South African diamond fields were just being discovered and kimberlite was unknown (Hausel and Stahl 1995). Several other extraordinary coincidences were noted. In particular, on the basis of today's exploration models, this region would be considered to have high potential for diamondiferous kimberlite and lamproite. Some other coincidences include the following:

- The site of the diamond hoax lies on a craton, a region considered to have high potential for diamond deposits on the basis of modern diamond exploration theories.
- The Bishop conglomerate capping the top of Diamond Peak actually contains some pyrope garnet and chromian diopside. However, the chemistry of these kimberlitic indicator minerals is slightly different from those found at the hoax site. Those kimberlitic indicator minerals found at the hoax site have chemical signatures identical to those found in the Navajo Volcanic Field (Tom McCandless, personal communication, 1996).
- Just a little more than 100 years later, in 1975, diamonds were discovered in kimberlite only 288 km east of the Great Diamond Hoax site.
- Lamproites were studied by one of the members of the Fortieth Parallel Survey (Sam Emmons) several years later, just 80 km north of the hoax site. One of the lamproitic mesas in the Leucite Hills bears his name. However, it wasn't until nearly a century later that these unusual rocks were classified as lamproites and were recognized as possible diamond targets.
- Diamondiferous breccia pipes were discovered just 96 km to the west of the hoax site in 1996.
- Widespread kimberlitic indicator mineral anomalies were identified a short distance west and west-northwest of Diamond Peak (see McCandless, Nash, and Hausel 1995).

Wyoming and Colorado. The most significant diamond discovery made in the United States to date was the accidental discovery of diamonds found in kimberlite in Wyoming in 1975. Prior to this discovery, serpentinized breccia had been found on the Sloan Ranch in the Front Range of Colorado in 1959, in what is now known as the State Line district.

The breccia was quarried for terrazzo and shipped to Cheyenne, Wyoming, for processing by the Wyoming Tile and Terrazzo Company for decorative tile. However, quarry operations ceased in 1960 when the tile company refused to purchase any more of the unusually hard rock. The company reported that their carborundum wheel had been damaged during the polishing of the rock (Frank Yaussi, personal communication, 1979). Carborundum, which has a hardness of 9.5, is harder than any natural mineral except diamond. Apparently, no one took note of this fact.

At about the same time, in 1960 and 1961, two small elliptical features containing out-of-place blocks of Paleozoic limestone and sandstone were found. These Paleozoic outliers were surrounded by 1.4 Ga granite in the Laramie Mountains a few kilometers north of the Colorado–Wyoming border and about 16 to 19 km north of the Sloan quarry. The outcrops presented a puzzle to geologists and geophysicists of the Rocky Mountain Association of Geologists Fourteenth Field Conference, because the features contained blocks of Cambrian, Ordovician, and Silurian sandstones and carbonates, which were otherwise unknown in the region. The outcrops, described as Early Paleozoic outliers at the 1963 conference (Chronic and Ferris 1963), were interpreted as down-dropped "piston grabens" on the basis of their geophysical signatures (Ray 1963) because they exhibited pipe-like characteristics.

In 1964, similar out-of-place lower Paleozoic rocks were identified in the Sloan Ranch quarry in Colorado about 19 km south on the earlier-discovered outliers. This outlier not only contained the Paleozoic xenoliths, but the Paleozoic rocks were closely associated with the serpentinized breccia that had been quarried for terrazzo. Samples of the serpentinite were collected for petrographic studies by M.E. McCallum at Colorado State University. The rock was later confirmed to be kimberlite (McCallum and Mabarak 1976).

There was also an unverified report that a representative from DeBeers, Arnold Waters Jr., offered to lease the Sloan property after visiting the terrazzo quarry and taking several hundred kilograms of material back to South Africa in the early 1970s (Frank Yaussi, personal communication, 1978). Even so, diamonds weren't "officially" discovered until a serpentinized garnet peridotite nodule was collected from the Schaffer group of kimberlites in Wyoming (section 16, T12N, R72W) in December 1974 by McCallum. The nodule was sent to the U.S. Geological Survey for thin section preparation and several deep scratches were cut in the carborundum wheel during polishing of the sample in June 1975. The nodule was digested in hydrofluoric acid and microdiamonds were recovered.

The discovery remained a curiosity until 2 years later in 1977 when Daniel N. Miller Jr., director of the Geological Survey of Wyoming, had the foresight to see a potentially new economic resource for Wyoming and spearheaded a campaign to have the kimberlites evaluated. Since the 1960s, approximately 40 kimberlites were found in this region along the Colorado–Wyoming border, and more than 130,000 diamonds were recovered from a small group of kimberlites in the district. In 1996, a low-grade diamond mine (Kelsey Lake) began production and more than 10,000 macrodiamonds were recovered, including stones up to 28.3 carats. As of 2001, exploration in the State Line district is still considered to be in the early stages because several geophysical targets remain untested, no placers had been evaluated, and several of the known diamondiferous pipes remain untested for ore grades.

Other discoveries followed. The second largest kimberlite district in the United States (second only to the State Line district) was discovered in the Iron Mountain region of the Laramie Range north of the State Line in 1971. Mapping by Smith (1977) recognized a small number of blows and dikes. More recent mapping by Hausel et al. (2000) resulted in the recognition of one of the largest kimberlite dike-sill-blow complexes in North America. Despite the recovery of significant amounts of diamond-stability garnets from the district, and the limited testing by Cominco American in the early 1980s that resulted in the recovery of one macrodiamond (Howard Coopersmith, personal communication), much of the district remains unexplored for diamonds.

In the late 1990s, another group of mantle-derived pipes were found in southwestern Wyoming. These intrusives, the Cedar Mountain pipes, reportedly yielded three diamonds from limited sampling (Hausel et al. 1999). Numerous other kimberlitic indicator mineral anomalies have also been identified in the Wyoming region (Hausel, Sutherland, and Gregory 1988; Hausel 1998).

Australia

Alluvial diamonds were reported in Australia in New South Wales as early as in 1861. Later, diamonds were reported in Queensland in 1887, in South Australia in 1894, and in Tasmania in 1899. Some diamonds were produced from the Copeton field in New South Wales from 1884 to 1922. In total, the recorded production was 167,548 carats (actual production, considering unrecorded stones, was undoubtedly greater). The largest reported stone was 8 carats and an average size of the diamonds was 0.3 carat. The number of diamonds produced in this area is considered significant.

The Copeton diamonds are associated with gravels within a valley buried by Tertiary basalts that are known as “deep leads.” The general tectonic environment of the New South Wales area is considered similar to that of the Urals and the California coast. Cliff Resources Pacific developed the idea that the deep diamond leads were not alluvial but rather the result of phreatomagmatic volcaniclastics and tuffs associated with adjacent lamprophyre pipes (Atkinson and Smith 1995).

Other diamonds were found in Western Australia. In 1936, a geologist named Wade had noted the presence of unusual volcanic rocks within the Canning Basin–Kimberley Region of the Northern Territories of Western Australia. Wade described these occurrences to Professor R.T. Prider at the University of Western Australia. Prider became interested in the rocks, termed lamproites, and noted their affinities to kimberlite.

Later diamond exploration in Western Australia led to the discovery of nine diamonds totaling 1.65 carats in weight in 1969. They were found in the Lennard River drainage in northwestern Australia. However, the source was not found. In 1973, kimberlitic indicator minerals were found in the west Kimberleys in the same general region, and diamonds were found in the north Kimberleys. In 1976, 300 diamonds were found in the King George area. Shortly thereafter, the diamondiferous Pteropus Creek kimberlite was discovered.

In late 1976, Ashton Exploration discovered small diamonds near Mt. Percy after following a trail of kimberlitic indicator minerals, which led to the discovery of a diamondiferous pipe. In August of 1979, diamonds were found in Smoke Creek, and in October of the same year, the AK1 (Argyle) pipe was discovered (Atkinson and Smith 1995).

After 25 years of prospecting within the Western Australian cratons, 12 kimberlite fields containing 360 pipes, 180 of which contain diamonds (Figure 1.9) had been identified. Some of the recently discovered kimberlites have yielded minor to significant diamond grades, but so far none of the kimberlites appear to be of ore grade (see Berryman et al. 1999).

Nearly all exploration efforts in Australia focused on kimberlitic sources of diamonds in ancient cratons until diamonds were found in Tertiary lamproite pipes in the Ellendale field in West Kimberley in 1976 and in the Proterozoic (1.2 Ga) Argyle lamproite in 1979. Both were located within Proterozoic mobile belts cratonized around 1.8 Ga and were tectonically active until Devonian or even younger time.

The general conviction that kimberlites are the single primary source of diamonds resulted in rejection of these sites by DeBeers. Australian geologists should be commended for persisting in the exploration programs that resulted in discoveries of new, very rich diamond deposits in olivine lamproites (Jaques et al. 1982, 1983; Atkinson, Smith, and Boxer 1984).

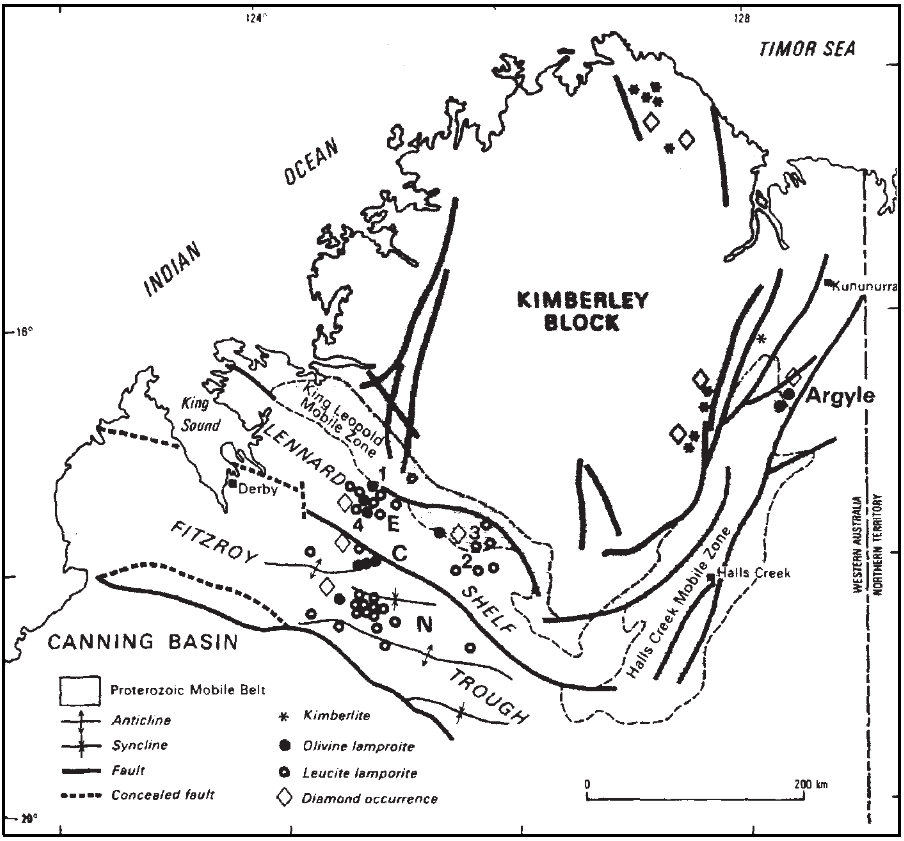

FIGURE 1.9 Kimberlites, lamproites, and diamonds in the Kimberley region, Western Australia. E = Ellendale Field, C = Calwynyardah Field, N = Noonkanbah Field (Skinner et al. 1985). Reprinted with permission of the Society of South Africa.

The mining of the single diamondiferous olivine lamproite at Argyle placed Australia among the leading diamond-producing countries of the world. At full production, the Argyle was the source of nearly 30% of the world's diamonds. Current activity has been centered on the mining of the Argyle lamproite and the evaluation of potentially commercial diamond deposits in two of the Ellendale lamproites in the West Kimberley platform and in two kimberlites in the North Kimberley platform (Shigley, Chapman, and Ellison 2001) (Figure 1.9).

Concerning the Ellendale field, Snowden Mining Industry Consultants estimated resources of 23 Mt at 8.8 carats /100 tonnes (> 2 million carats) to a depth of 140 m for a high-grade, near-surface enrichment zone in the Ellendale pipes 4 and 9 (reported by MiningNews.net on January 10, 2002). As of the end of 2001, some 3,500 diamonds, mainly good quality gemstones, had been recovered from limited exploration sampling at Ellendale pipe 9. The near-surface enrichment zone is a portion of the proposed mining target for 2002.

Canada

Canada had long been thought to be a potential source for diamonds. Many of the diamonds recovered in the Great Lakes region were assumed to have been carried by glaciers from Canada. Thus, it was just a matter of time before a source of diamonds was finally discovered in Canada.

Diamonds were reported in glacial tills in Ontario as early as 1863. Nearly a century later, in 1960, the Ontario Department of Mines located the first kimberlites known in Canada; they were described as narrow dikes. Kimberlites later found in various regions of Canada all proved to be uneconomic.

Some of the more important events reviewed by Miller (1995) are as follows:

- the discovery of kimberlitic indicator minerals in the Frenchman River basin in Saskatchewan in 1960
- the discovery of six kimberlite pipes in the Kirkland Lake district in 1960
- the discovery of kimberlite near Ile Bizard in 1964
- the recovery of a 7.3-tonne sample from a narrow kimberlite dike near Kirkland Lake in 1968 that yielded no diamonds
- the discovery of 19 kimberlite pipes at Somerset Island in the Canadian Arctic in the early 1970s; diamond grades were reported to be low and the composition of diamond indicator minerals were not promising
- the reported discovery by Monopros (a DeBeers subsidiary) of a kimberlite intrusive near Prince Albert in Saskatchewan in 1980; the "intrusive" is now considered to be a glacially transported block rather than an in situ pipe
- the discovery by Monopros of kimberlite pipes, some of which were diamondiferous, in the James Bay lowlands. The first Fort de Corne kimberlite pipe was discovered in Saskatchewan in 1989, 90 km east of Sturgeon Lake, by Uranez and Cameco. Ultimately, 70 pipe-like structures were found by drilling beneath about 100 m of overburden. Most samples yielded grades around 2 carats/100 tonnes.

Kimberlites have been known in Canada for many years. Some early reviews of kimberlite in Canada include Brummer (1978) and Dawson (1980). The North American craton is by far the largest craton in the world; it has an enormous Archean core, much of which extends under Canada. The Slave craton, which is a large segment of the North American craton, was considered to be favorable for hosting diamondiferous kimberlite. In order to take advantage of this enormous craton, the Geological Survey of Canada in 1986 funded research on indicator minerals and exploration for diamonds in Canada (Fipke, Gurney, and Moore 1995). This research provided the basic foundation that led to the Canadian diamond rush.

Chuck Fipke, who is credited with making the first major diamond discovery in Canada (as well as in North America), spent 20 years exploring Canada and tracing kimberlitic indicator-mineral trains (Krajick, 2001). Using the direction of glacial flow as a guide while following the diamond indicator minerals, Fipke traveled 1,200 km up ice to the Slave craton in the Northwest Territories. In 1990, he recovered kimberlitic indicator minerals from samples in the Lac de Gras region. One of several anomalies in the Lac de Gras region, Point Lake, was drilled by a Broken Hill Proprietary (BHP)/DiaMet joint venture. In the following year (1991), 81 microdiamonds were recovered from 59 kg of core sample.

In early 1992, a 160-tonne bulk sample from Lac de Gras was shipped to the Sloan diamond-recovery plant near Fort Collins, Colorado. Ninety carats of diamonds were recovered from this sample, of which 25% were gem quality. The diamonds included a few stones in the range of 3 to 5 carats. The results of the sampling led to additional exploration and to several more discoveries. The following chronological events highlight some of the activity:

- September 1992: BHP announced the discovery of nine pipes within the 853,000 acres of the joint venture's prospect area.
- March 1993: Kennecott began drilling the Tli Kwo Cho prospect 35 km southeast of the BHP/Diamet Point Lake discovery.
- March 1993: Lytton Minerals discovered the Lake Ranch pipe.
- April 1993: Microdiamonds were reported in samples from the Tli Kwo Cho prospect.
- September 1993: BHP announced that the Point Lake kimberlite was uneconomic. Kennecott began bulk sampling on the Tli Kwo Cho prospect.
- Winter 1993–1994: The diamond rush continued and the 45 millionth acre was staked in the Northwest Territories.
- April 1994: Aber Resources discovered the A-21 pipe; a 154.6-kg sample yielded 154 microdiamonds. BHP announced drilling results at its Panda Lake pipe. From 229.5 tonnes of kimberlite, 270 carats of diamonds were recovered (118 carats/100 tonnes). Lytton Resources reported that a 28.45-tonne sample from its Ranch Lake pipe yielded 112 stones weighing 5.38 carats equivalent to a diamond grade of 0.189 carats/tonne (18.92 carats/100 tonnes). Slightly more than 30% of the stones were gem quality.
- May 1994: The KWG/Spider joint venture with Ashton in the James Bay lowlands recovered 101 microdiamonds from a 106.6-kg sample, suggesting that this was the most important discovery outside the Northwest Territories.
- July 1994: BHP revealed diamond grades from the Koala and Fox pipes. As with samples from Point Lake, they were significantly lower than the grades announced a year before.
- July 1994: Kennecott/Aber announced results of sampling from A-154 discovery hole: 1,296 microdiamonds were recovered from 750.8 kg of kimberlite. Seven stones weighed more than 0.2 carats and 112 microdiamonds had diameters exceeding 1.0 mm. All the diamonds were gem quality. If the initial reports are confirmed by a following bulk sample, the A-154 pipe has the potential to be the richest kimberlite pipe ever found outside Siberia.
- August 1994: Bulk sampling results at Tli Kwo Cho are announced: 0.359 carats/tonne (35.9 carats/100 tonnes), and less than 30% are of gem quality.
- December 1994: BHP/Diamet reveal that the summer exploration program located 13 diamondiferous kimberlite pipes. Both companies remained optimistic. The initial production forecast was raised to 9,000 tonnes/day, and it will increase to 18,000 tonnes/day after 10 years.

The discovery was considered to be one of the greatest exploration successes in history, and it led to one of the most remarkable rushes, where regions the size of entire states were staked for diamonds. The first kimberlites found in the region were located under shallow lakes. The pipes were small and typically about 100 m across.

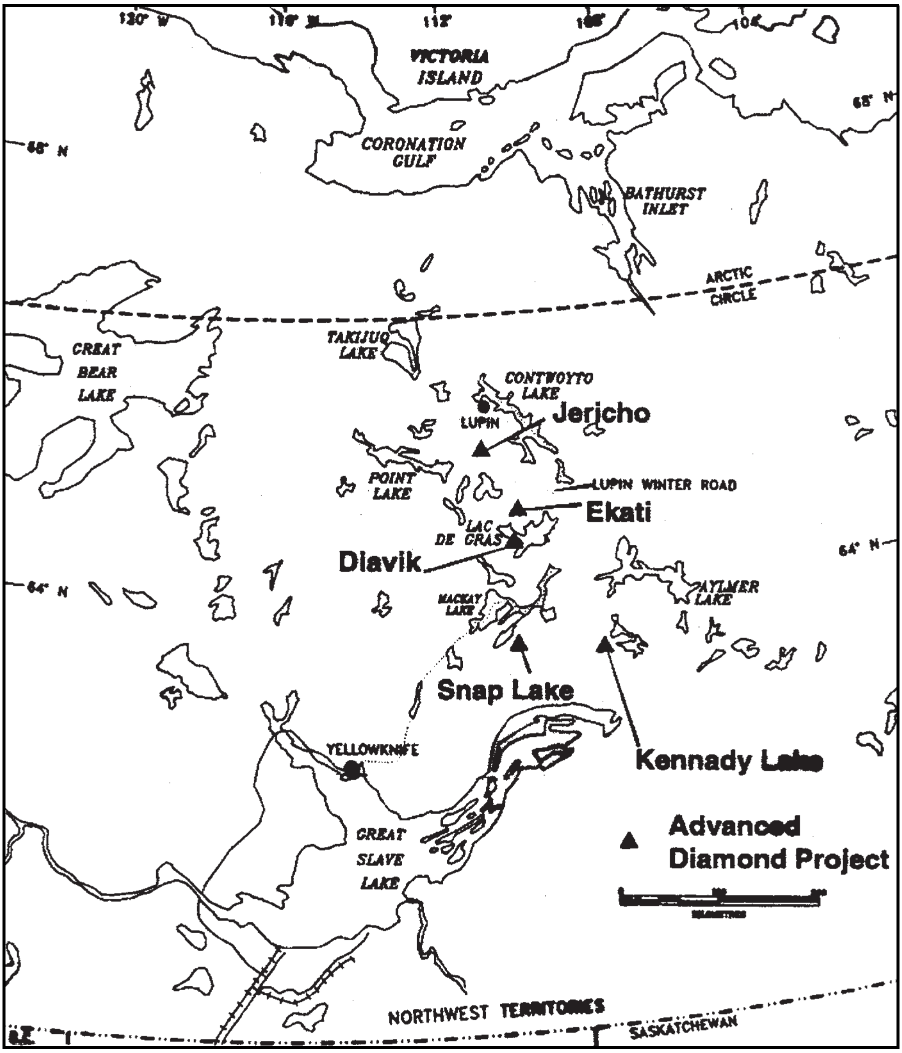

FIGURE 1.10 Important diamond localities in Canada (Goepel McDermid Securities 1999). Reprinted with permission from Robert W. Klassen. Taken from Canadian Diamond Companies, New Facets in a Profitable Industry.

The discovery, however, was significant, because some of the pipes were rich in high-quality diamonds. Continued exploration in the region led to the discovery of more than 100 other pipes. The number of kimberlite pipes and dikes that were discovered in Canada owing to the rush totals in the hundreds.

Canada's first diamond mine went into production in October 1998. The mine, the Ekati, lies about 300 km north of the town of Yellowknife. Proven reserves in 1999 were reported at 43 million tonnes of ore with an average grade of 1.21 carats/tonne (121 carats/100 tonnes) and probable reserves of 21.6 million tonnes at an average grade of 0.87 carats/tonne (87 carats/100 tonnes). It is estimated that this mine will produce between 53 and 71 million carats. The cost of building the mine in the Arctic north, however, was enormous. Capitalization amounted to more than $700 million, but the value of the diamonds from the mine during its 17-year life is estimated to be more than $4 billion. The locations of most important diamond sites in Canada are indicated on Figure 1.10.

Nearby, the Diavik project, 48 km east of Ekati, is anticipated to become Canada's second diamond mine. Diavik is anticipated to produce 5 to 6 million carats annually compared with Ekati's 3.5 to 4 million carats. In a short period, Canada became a major source of diamonds, and it is anticipated to provide 10% of the world's diamonds. Another mine in the Northwest Territories, the Snap Lake project, was recently proposed by Winspear Diamonds, and a fourth commercial pipe was rumored to have been found in Alberta. More than 250 pipes were identified to 1999 in the Northwest Territory (Goepel McDermid Securities 1999).

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CHAPTER 2

Diamond Morphology

GENERAL PROPERTIES AND INTERNAL STRUCTURE

Diamonds are formed of carbon. Native carbon may occur in nature as diamond, lonsdaleite, and graphite. In addition, chaoite, a high-temperature form of graphite, has been described in association with some astrobleme-type structures.

Diamonds are isometric. Many diamonds crystallize in an octahedral form or some modification of an octahedron. Others may have cubic to dodecahedral habit. Flattened and elongated crystals are common, and curved faces are frequently observed. Twins of the spinel type with a flattened morphology parallel to the twin plane, macles, are also common.

Two types of diamond are known, called type I and type II. Type I is by far the most common diamond. The electrical conductivity of type I is so poor that it is essentially a nonconductor, whereas type II diamond has good conductivity. The two types also differ in infrared and ultraviolet characteristics.

The result of more than a half of century of x-ray studies established that the crystalline cell of diamond can be approximated as a cube with sides of 3.56 Å (1Å = 10^{-8}cm) in length (Orlov 1977). The unit cell has carbon atoms located at the apexes of the cell and also in the centers of the faces. In addition, four carbon atoms are located inside the cell (Figure 2.1). The coordination of carbon atoms in diamond is tetrahedral and each atom is held to four others by strong covalent bonds, which produce the extreme hardness of diamond. The space lattice of diamond is considered as a face-centered cubic Bravais lattice with four additional atoms at regular positions inside (Figure 2.1A). Considering the orientation of tetrahedral atoms within diamonds' internal structure, it is usually presented as a combination of tetrahedrons (Figure 2.1B).

It is commonly accepted that the unusual physical properties of diamond directly reflect specific patterns of the mineral's internal structure. The strong bonds within the unit cell, which are the result of common electrons between atoms, determine several unusual properties such as extreme hardness, bright diamond luster, stability within wide range of temperature and pressure, and low electrical conductivity.

In the eighteenth and nineteenth centuries, information on the crystal system of diamond was based primarily on its external shape. There was little doubt that diamond was isometric, but the crystal class was a source of controversy (Orlov 1977). Occasional reports of tetrahedral diamond crystals led a several crystallographers to the conclusion that diamond belonged to the hexatetrahedral class (T_d) of the isometric system (Rose 1853; Fedorov 1899).

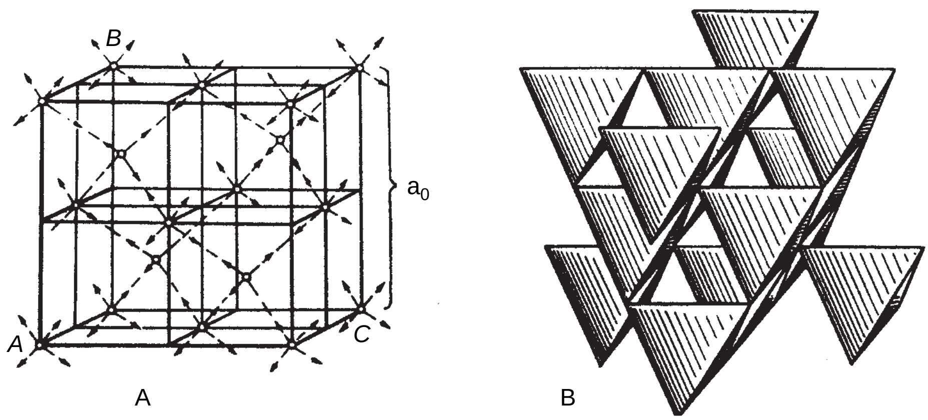

FIGURE 2.1 Structure of diamond. A = position of carbon atoms in lattice; B = polyhedral structure (Orlov 1977).

The presence of both octahedral and tetrahedral forms and data on internal crystalline structure produced two schools of thought, one of which accepted that diamond has octahedral symmetry and the other of which suggested that diamond has tetrahedral symmetry. Supporters of the latter considered diamond octahedra to be the result of intergrowth of penetration twins of two tetrahedra on the {100} plane according to the Mohs-Rose law (Orlov 1977).

While studying the symmetry of diamond and its piezoelectric properties, Van der Veen (1911) stated that it was erroneous to assume a hemihedral symmetry for diamond and that, instead, diamond should be assigned to holohedral O_h class. Almost simultaneously, Fersman and Goldschmidt (1911) upheld the idea of a hemihedral nature for diamond on the basis of diamonds with tetrahedral habit and on the reflection patterns of goniometer studies. Owing to inferior stability of one of tetrahedra twins according to the Mohs-Rose law (which was affected by rapid dissolution in nature or by artificial etching), it was concluded that the crystals are hemihedral and produce simple and composite twin forms according to the spinel law (twinning plane p) and the Mohs-Rose law (twinning plane d) (Orlov 1977).

A new approach to this problem investigated the d internal structure of diamond with x-rays and confirmed that diamond belongs to the hexoctahedral class of the isometric system (Raman 1944). The x-ray research showed the electron configuration of carbon atoms yielded tetrahedral symmetry. Raman concluded that there are four possible structural types of diamond, two with hexatetrahedral class symmetry (T_1 and T_2) and two with hexoctahedral symmetry (O_1 and O_2) (Figure 2.2).

Raman's studies showed that different types of diamonds differ insignificantly from each other. Different types of diamonds can occur within a single crystal as parallel-intergrown twinned plates. Such plates are regularly oriented parallel to the octahedral face.

Later, V. Nardov (1954) argued that the first two classes (T_1 and T_2) are identical and thus there are only three classes of structure: T, O_1, and O_2: the first is tetrahedral and the other two have octahedral symmetry. It is possible that this discussion ensued because Raman and his collaborators studied diamonds exclusively from Indian deposits, whereas the other scientists studied diamonds from South Africa and Brazil.

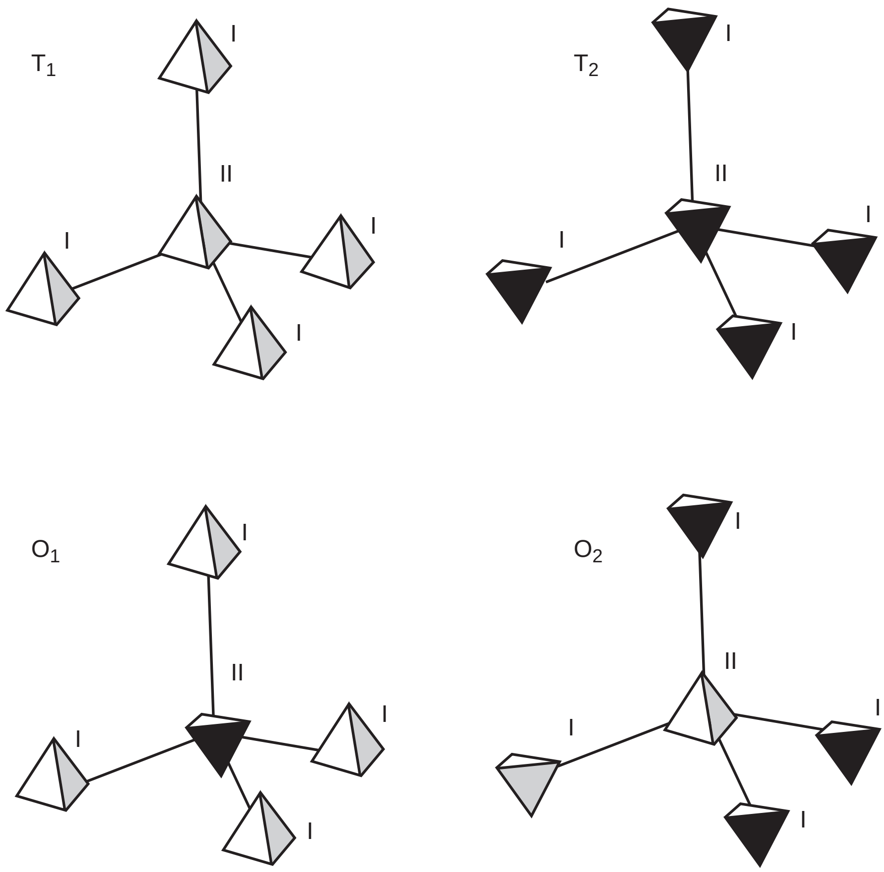

FIGURE 2.2 Theoretically possible types of diamond structure, after C.V. Raman (Shafranovsky 1964)

In contrast to diamond's lattice, the crystalline lattice of graphite is formed by a series of layers. Each layer consists of hexahedrons with carbon atoms in their apexes. The distance between layers is 3.39Å. The layers are consistently shifted in relation to one another. As a result only a half of atoms coincide with the centers of the cells, whereas the other half are projected into the centers of cells. The difference between crystal structures of diamond and graphite is visible in Figure 2.3.

Diamonds have perfect octahedral cleavage and possess an average hardness of 10 on the Mohs hardness scale. The hardness will vary slightly depending on the crystallographic direction of the crystal: the hardness parallel to the octahedral cleavage is much greater than that parallel to the cubic face. Diamond is considerably harder than any other mineral. Corundum, the next hardest mineral on the Mohs scale has a hardness of 9. However, the difference in hardness between corundum (H = 9) and diamond (H = 10) is three times as great as that between topaz (H = 8) and corundum. Diamond has an adamantine luster but uncut crystals have a characteristic greasy appearance.

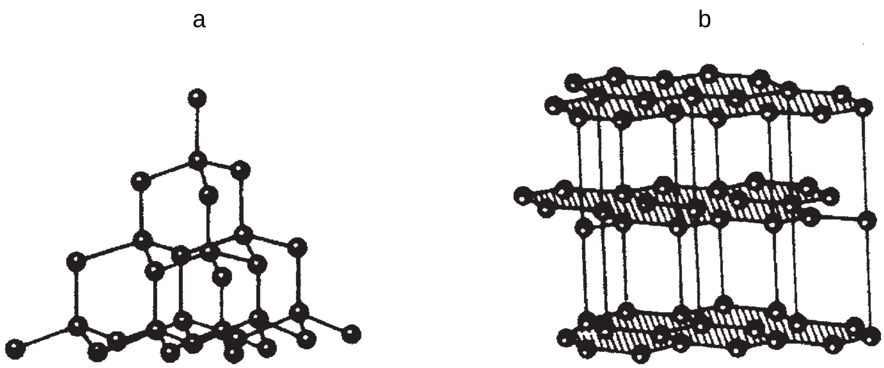

FIGURE 2.3 Atomic structures: a = diamond, b = graphite (Milashev 1989). Reprinted with permission of Publishing House Nedra.

The presence cleavage in diamond is partially due to an increase in interplanar spacing between the {111} planes in the diamond structure. In addition, there is only one bond per unit area between these planes, and as a result diamond has perfect octahedral cleavage (Kukharenko 1955; Orlov 1977).

Diamond crystals commonly enclose tiny minerals along their cleavage planes. These tiny mineral inclusions provide important data on the origin of diamond, and some are used to determine the age of the diamond.

Diamonds will fluoresce in long- and short-wavelength ultraviolet radiation. The fluorescence is usually greater in long-wavelength ultraviolet light and may be blue, green, yellow, and occasionally red. However, fluorescence is generally weak, and it may not always be readily apparent to the naked eye. Some diamonds will also phosphoresce.

Diamonds have an above-average specific gravity; it is 3.52 times as heavy as water. However, diamonds are hydrophobic (nonwettable) and will repel water, and they can actually be induced to float on water. Because they are hydrophobic, diamonds attract grease (grease will adhere to the surface of a diamond), providing an efficient method for extracting diamond from waste material.

Studies of diamond in ultraviolet and infrared light show that most diamonds have tetrahedral symmetry. Different types of luminescence in diamond are produced by its internal crystalline structure. It is also possible that the structural differences result in different abrasive properties for diamond.

It is apparent that the supporters for a tetrahedral and octahedral symmetry for diamond were both correct. Diamond crystals are usually formed by complex intergrowths of twinned plates.

Shafranovsky (1964) mentioned that factories using diamond powder often observed microscopic diamond plates characterized by considerably lower hardness. Such grains are probably associated with twinned plates in diamond crystals. Thus, the theoretical problem of the internal atomic structure of diamond is a matter of great technical interest. It is no wonder that the decoding of the internal atomic structure of diamond has been considered a great triumph of research.

The hexagonal modification of diamond, lonsdaleite, has long been considered a rarity associated with meteorites. It has a closer-packed arrangement of atoms and is denser than diamond or graphite. Only in the 1960s was it described as the main form of native carbon associated with ring structures known as astroblemes. It has also been described in metamorphic rocks, in particular in eclogites (Golovnya, Khvostov, and Makarov 1977). (For additional material on lonsdaleite, see Chapter 12, Specific Type of Ring Structures [Astroblemes]).

CRYSTAL HABIT

Diamonds are isometric, which implies that the mineral is highly symmetrical and equidimensional. Isometric crystals show the highest form of symmetry. The simplest form, the unit cell of the isometric system, is a cube.

Cube

Cubes, like all forms of the isometric system, possess three crystallographic axes of equal length that lie at right angles to one another. The intersection of these three imaginary axes lies at the center of the cube. From the cube's center, they project outward into the surrounding crystal faces. Because of the high symmetry of the cube, each of these axes intersects the respective crystal faces at the same distance from the center of the cube.

The crystal faces intersected by the crystallographic axes are symmetrical, or interchangeable; thus for reference, each axis is designated by the same letter with subscripts to identify the mirroring axis. The front and back of a cube are designated a_1; a_{-1}, the axis running from left to right is designated a_2; a_{-2}, and the vertical axis is a_3; a_{-3}. In mineralogy, it is convenient to use Miller Indices to designate the crystal faces. These indices are a set of numbers that describe where the three axes are intersected by the crystal planes. For example, the front crystal face of the cube intersects the a_1 portion of the axis at one unit from the center. Because the face is a vertical plane and does not cross the remaining two axes (a_2;a_{-2} and a_3;a_{-3}), it is designated {100}; the numbers refer to the axes (a_1, a_2, a_3). The parallel crystal face at the designated back of the cube would be identified as {−100}.

The cube is a six-sided form; because it possesses six sides, a cube is sometimes referred to as a hexahedron. Diamond cubes are relatively uncommon, and most cubes are of industrial quality.

Octahedron

Diamonds with octahedral habit, or some modification of an octahedron, are more common in nature than cubes. The term *octahedron* implies an eight-sided form. The octahedron consists of an eight-sided crystal that appears as two four-sided pyramids that are attached at their bases. Each pyramid contains four equilateral triangles that intersect the three crystallographic axes one unit from the center. Thus these octahedral faces are referred to as {111} faces.

In nature, the {111} octahedral face commonly contains either positive or negative trigons. These trigons are small equilateral triangles that either grow, or are etched, on the crystal surface. They represent the products of disequilibrium (changes in temperature and pressure) when the crystal was transported to the earth's surface from initial stable conditions at depth.

Partial resorption of octahedral diamonds can result in a rounded dodecahedron (12 sides) with rhombic faces. Many dodecahedrons develop ridges on the rhombic faces producing a 24-sided crystal, a trishexahedron. Four-sided tetrahedral diamonds are sometimes encountered; they are probably distorted octahedrons (Bruton 1979).

Tetrahedron

Along with octahedral and cubic diamonds are tetrahedral forms (see review in Orlov 1977; Shafranovsky 1964). A tetrahedron by definition is a four-faced polyhedron, in which each face forms a triangle (Bates and Jackson 1980). The crystal exhibits four faces, four edges, and six apexes.

VARIETIES OF DIAMOND

Diamond varieties were summarized and classified by Orlov (1977). Ten varieties, with different crystal morphology, internal structures, physical properties, and impurities, are described.

Varieties I–VI

Variety I. Orlov's variety I diamond includes plane-faced octahedral crystals with smooth and stepped faces. In the latter case, the octahedral habit can be strongly distorted owing to the uneven combination of surfaces instead of sharp edges, or modification of the corners by irregular surfaces instead of cube surfaces. Nitrogen atoms singly substituting for carbon and possessing paramagnetic properties compose an average of about 0.001% of the total impurities in this variety.

Quantitatively, variety I is predominant and composes 98% to 99% of all known diamond crystals. Sometimes crystals of this variety are coated and form crystals characteristic of variety IV.

Variety II. Diamonds that are plane faced and of cubic habit with characteristic amber-yellow and green color are variety II diamonds. Dissolution converts these crystals into cuboids, and progressive dissolution results in dodecahedrons with well-defined morphologic features. They are rich in nitrogen impurities. The relative abundance of variety II crystals among diamonds from different deposits varies, though they are usually rare.

Crystals of this variety are described from the Kimberley pipe and its surroundings, Aikhal pipe, some placer deposits of the Anabar and Lena Rivers, and the Murfreesboro, Arkansas, lamproites.

Variety III. Variety III forms are cubic crystals showing a combination of forms (octahedron and rhombic dodecahedron plus cube). All have large amounts of nitrogen impurities. Crystals of this variety are characterized by a transparent, colorless zone in the center of the outer portion of the diamond. This zone is characterized by microscopic inclusions that are responsible for the gray and dark color of the crystals. These crystals have been recognized among the diamonds from the Udachnaya and Mir pipes.

Variety IV. Variety IV diamonds possess zoning that is visible to the naked eye. Such crystals are also described as "coated." Cores of grains belonging to this variety are represented by transparent octahedral crystals typical of variety I.

Variety V. Variety V diamonds are characterized by an abundance of syngenetic graphitic inclusions in the outer zones of the crystal, whereas the central zones are usually transparent and colorless. The octahedral habit within the outer zone is saturated by graphitic inclusions. The shape of the diamond is rounded owing to development of curved-faced dissolution surfaces. This diamond type has been reported at Lena, Anabar, and in placers from the Urals.

Variety VI. Variety VI diamonds, also known as ballas, are formed by diamond spherulites with a radial fibrous structure. These diamonds are usually perfectly spherical although rare pear-shaped specimens are known. Owing to the fibrous structure, ballas grains are characterized by a specific aggregate's surface sculpture. The best known are the ballas diamonds from Brazil and South Africa's Cape Province.

Varieties VII–IX

Varieties VII through IX occur as various types of polycrystalline aggregates.

Variety VII. Variety VII comprises aggregates of diamond crystals that are generally yellowish and are rendered subtransparent by various defects (such as cracks and graphite inclusions).

Varieties VIII and IX. Varieties VIII and IX, bort, are granular aggregates that have the appearance of irregular lumps. Individual crystals of variety VIII show typical octahedral habit commonly with stepped development such that pseudorhombododecahedron forms are common.

Individual Variety IX diamonds show no regular crystalline forms although they are easily distinguishable. Lumps are opaque, dark gray, or completely black, and the structure is sometimes inequigranular. Both VIII and IX varieties of diamond occur in the Mir and Aikhal pipes, as in many non-Russian diamond fields.

Variety X (Carbonado)

By definition, carbonado is an "impure opaque aggregate composed of minute diamond particles forming a usually rounded mass with granular to compact structure and displaying a superior toughness as a result of cryptocrystalline character and lack of cleavage planes" (Bates and Jackson 1980).

Brazilian miners first coined the term as early as 1843 to designate the opaque, black or gray, polycrystalline diamonds that were found mainly in highlands of Bahia, Brazil. Specimens were also found in smaller amounts in Venezuela (Fettke and Sturges 1933; Kerr, Graff, and Ball 1946). Many diamond dealers have applied the term *carbonado* exclusively to Brazilian stones. The same mineral found in Central Africa was termed *carbon*.

The original source of carbonado was unknown until recently. Carbonado is abundant in Brazil, Central African Republic, and Australia, and it has recently been found in placer deposits in the Urals, Ukraine, northern Yakutiya, and Sikhote-Alin' (the Russian Far East), where it is usually associated with placers and intergrown with granitic minerals.

Carbonado is described as a variety X diamond (Orlov 1977). However, the term *carbonado* has been adopted in the mineralogic literature (Dana 1892). It is represented by distinct crystal polyhedra usually found as imperfectly shaped cubes, octahedra, and rhombic dodecahedra of the isometric system. The material is black and hard, occurs mainly as irregular porous concretions and dendritic aggregates of minute octahedra, and sometimes forms regular, globular concretions. Carbonado's carbon isotope composition and structure are so different from those of other diamonds (Vinogradov et al. 1966) that they seem to indicate an independent source of carbon and specific crystallization conditions. Minerals associated with carbonado support the latter conclusion.

Minerals associated with carbonado in Brazil include orthoclase, hematite, allanite, perovskite, rutile, corundum, cassiterite, chloritoid, pseudomalachite, covellite, rosasite, parisite, anhydrite, graphite, chalcedony, and quartz. In Ubangi, the following assemblage is commonly reported with carbonado: Al-serpentine, high-Al chlorite, chromite, florencite, weinschenkite, ilmenite, magnetite, melilite, pyrophyllite, perovskite, quartz, olivine, and rutile. In Venezuela, quartz is reported as an associated mineral.

In Yakutiya, almandine, pyrope, staurolite, zircon, ilmenite, graphite, lonsdaleite, and anatase are commonly found with carbonado (Smith and Dawson 1985). With the exception of pyrope garnet and sometimes lonsdaleite in Yakutiya, there is no indication of any other high-pressure mineral phases associated with carbonado. Smith and Dawson (1985) noted that there is no indication whether the pyrope garnets found in Yakutiya with carbonado are the upper mantle, Cr-rich variety. The presence of florencite, which is closely associated with the alunite group, is thought to have felsic affinity. The presence of weinschenkite, which is commonly associated with gypsum in volcanic regions, indicates replacement of calcium by rare earths (Trueb and DeWys 1971). Pyrophyllite is typical of schistose rocks and is found in abundance in both the Ubangi and Bahia localities in Brazil.

Carbonado's unusual strength is probably explained by its internal polycrystalline microstructure. Brandenberger (1930) postulated that carbonado is an aggregate of randomly oriented diamond crystals. These crystallites show no evidence of a coating or cementing agent such as amorphous carbon or silica, an observation confirmed by Trueb and Bitterman (1968).

Carbonado is characterized by large aggregates. Aggregates have an average diameter of 8 to 12 mm and weigh approximately 20 carats. Larger stones up to several hundred carats are by no means rare. The largest stone of "carbon" from Ubangi, weighing 740.25 carats, is exhibited in the Smithsonian Institution (Trueb and DeWys 1971).

Carbonado has a typical porous texture, prevalence of light isotopes of carbon (PDB)—$\delta^{13}C/^{12}C < -25‰$), association with placers, some intergrowths with granitic minerals, enrichment in titanium, and an absence of high-pressure chromium minerals. Carbonado usually forms porous coke-shaped grains of gray, dark-gray, and black with some violet shades.

Carbonado's density is less than that of normal diamond, and it varies from 3.13 to 3.46 gm/cm^3. The static load required to destroy grains in the size range 0.20 to 0.25 mm derived from crushed samples of carbonado depends upon the sample's porosity; it varies from 2.4 to 3.7 kg for densities of 3.42 and 3.46, respectively (Kaminsky 1992; Kaminsky et al. 1979).

The isotopic composition of two samples of carbonado was reported as −29.2 and -30.6^{13}C PDB (Galimov, Kaminsky, and Kodina 1985). Luminescence of carbonado is heterogenic, and nonluminescent areas tend to alternate producing areas of prevailing orange-red, yellow, yellowish-green, and yellowish-orange luminescence.

Thermal treatment of carbonado will change its luminescence considerably. Heating in air at 600°C will increase the intensity of luminescence. As a result, color will change owing to the redistribution of the relative emanation intensity of some centers, or because of the partial (up to 4%) destruction of samples that have a loose powderlike mass structure. The general feature of carbonado's luminescence is the widening of nonphonon lines (up to 15–17 MeV) in comparison with similar lines in normal diamonds, where semiwidth usually equals several MeV at 80 K. This widening probably indicates tension in the crystalline lattice of carbonado close to optical centers, which disturbs and depletes their energy levels.

Carbonado in the Sikhote-Alin' placers in the Russian Far East is found in an area underlain by Paleozoic and Mesozoic sediments that are cut by Jurassic meymechite-picrites, including some Neogene and Quaternary micaceous alkaline-basalt pipes. Along with the grains of carbonado are found crystals of gold; normal diamonds; large (up to 20 mm) crystals of gem-quality corundum, zircon, chrome-diopside, enstatite, Cr-Al-spinel, and manganese-rich picroilmenite; and small grains of rutile and anatase.

The Sikhote-Alin' carbonado is represented by coarse-grained (7–8 mm) crystals that are characterized by abundant rounded pores; these pores are commonly interconnected by pipe-like channels as viewed with a scanning electron microscope. Laser spectral analysis established that the carbonado contains inclusions of xenotime enriched in La, Zr, and Cr. Anomalous amounts of Si, Al, Fe, Ti, Sr, Ba, Y, La, Zr, Cr, P, Cl, and S have been detected in areas having increased concentrations of Y, Zr, Sr, and Ba. Other inclusions include anatase, rutile, corundum, ilmenite, and graphite.

After carbonado is burned, the residue contains a bluish-gray carcass aggregate of anatase (up to 15%–20% by mass). The carbon isotope composition is quite typical for this type of carbonado (−26.5‰ to −9.0‰ ± 0.03‰), whereas transparent diamond crystals from the same placer have PBD −6.4‰ ± 0.02‰.

Two turbid (cloudy) gray and greenish diamond crystals, represented by dodecahedrons with orange-red luminescence, show a contrast in isotopic zoning: the outer zones (5–10 microns) are characterized by isotopic means of −16.7‰ to −21.7‰, whereas the inner zones are characterized by isotopic means of −10.4‰ to −11.1‰.

Such sharp zoning can be explained by the overgrowth of normal, isotopically heavy diamonds by isotopically light rims under the influence of gas (CO, CH_4). Such a mechanism was suggested by Marakushev (1993) for the oxidation and recrystallization of primary isotopically heavy diamonds through a complete transition to carbonado. The isotope exchange in this process is enhanced by low, 600° to 700°C, temperatures (the approximate temperature of the rutile-anatase transformation) and low pressures (which promotes crystallization in pores) (Scheka, Ignat'ev, and Vrzhosek 1995). These conditions are metastable for the growth of primary diamond, but they become possible with the presence of diamond embryos. The presence of carcasses of anatase crystals within carbonado grains may be an indicator of the catalytic role played by titanium in this process.

Brazilian carbonado is abundant in areas containing phyllites. Possibly, some of these phyllites are the products of oxidization of kimberlites by hydrothermal fluids derived from granites.

Research by Scheka, Ignat'ev, and Vrzhosek (1995) suggested a new direction for the origin of carbonado. This work demanded additional data on diamonds from impact-related structures similar to the Popigay depression and a discussion about the origin of diamonds in metamorphic rocks.

Although carbonado had been found in placers in Brazil and Russia, it was not until 1993 to 1995 that it was found in situ. Twenty-six grains of carbonado ranging in size from 0.1 to 1 mm were recovered from a 150-kg sample taken from avachites (a specific type of basalt from the Avacha volcano of eastern Kamchatka) (see avachites in Chapter 10, Meteorites, Ultramafic, and Mafic Igneous Rocks).

The visual identification was confirmed by x-ray and electron microscope analyses (Smyshlyaev 1999). Trace minerals associated with the carbonado included small grains of red corundum, platinoids, some grains of native gold, and native silicon. The mineral train of the avachites led to the glacier-filled explosive funnel of the Kozel'sky volcano and confirmed a possible connection of carbonado with high-pressure explosive processes. During later exploration, grains of carbonado were found in the black sands taken from areas of Kamchatka in a region of ultramafic plutons (Baykov, Anikin, and Dunin-Barkovsky 1995).

Variety XI

In 1967, specific types of carbonado intergrown with lonsdaleite (a shock-related hexagonal variety of diamond) were found in placers in northern Yakutiya, not far from the Popigay ring volcano-tectonic structure. These carbonados have been termed "yakutite." The intergrowths were later described by Frantsesson and Kaminsky (1974) as a specific type of polycrystalline diamond aggregate identified as Diamond XI or yakutite.

In contrast to the irregular and rounded Brazilian carbonado, grains of Diamond XI have a relict hexagonal form. X-ray analysis established the presence of a lonsdaleite admixture in these grains. This admixture is characterized by a distribution of atoms oriented toward the cubic crystal lattice of normal diamond. Thus it is possible to mistake this admixture for the octahedral {111} layers of diamond.

In contrast to normal carbonado, which is characterized by an irregular splice of crystals and the absence of any orientation, grains of Diamond XI are characterized by the presence of a porous texture. The angle of dispersion of samples from the Yakutian placers is 16° to 25°.

Another specific feature of Diamond XI in comparison with typical carbonado is the small size of the grains. Grains of typical Brazilian carbonado range from 0.5 to 80 microns, whereas Diamond XI grains are usually less than a micron.

Because of its small grain size, Diamond XI is interpreted to form during rapid crystallization under conditions of strong oversaturation by carbon. The presence of a lonsdaleite admixture is a direct indication of short-lived shock waves required to produce Diamond XI. On the basis of these characteristics, Diamond XI aggregates are similar to intergrowths of diamond and lonsdaleite described in the association with astroblemes (Masaitis, Futergendler, and Gnevushev 1972).

Owing to its polycrystalline habit, this material is extremely tough and is in great demand for use in drill bits designed to pierce the hardest rocks and for cutting metals. As Brazilian deposits were being played out at the beginning of the century, the price of carbonado rose dramatically, but this situation reversed around 1925, when the rich Central African deposits were put into production. Today only a few thousand carats of drilling-grade carbonado are mined annually in Brazil. Most of the demand for this type of diamond is being met by "carbon" from the Ubangi area in the Central African Republic. Mining in the latter region is concentrated in two regions located 600 km apart: Berbezati, Carnot, and Nola in the west (West Ubangi) and Ouadda and N'Dele in the east (East Ubangi). In the east Ubangi region, the "carbons" are usually small and form 6% to 7% of the total diamond production, whereas in the west Ubangi region they are quite a bit larger and they form up to 30% of total production. In this latter area, large, white, monocrystalline diamonds are relatively rare.

All of the Central African deposits are associated with Cretaceous and younger alluvial sandstones and conglomerates with no apparent affiliation with kimberlite. Despite a great amount of geologic prospecting and surveying, no kimberlite pipes, nor any kimberlitic minerals, have ever been found in Central Africa (Trueb and DeWys 1971). Neither carbonado nor lonsdaleite have ever been seen in kimberlites.

Lonsdaleite

The existence of a hexagonal polymorph of diamond was initially established by Bundy and Kasper (1967), who synthesized the hexagonal form at temperatures greater than 1,000°C and under static pressures exceeding 130 kbar. Another laboratory (DuPont deNemours and Co. 1965; Netherlands Patent 65063695) obtained the same transformation of diamond by intense shock compression and thermal quenching. The name of the new polymorph was chosen to honor distinguished British crystallographer Kathleen Lonsdale. The name was approved by the International Association of Mineralogy (Fleischer 1967).

Almost simultaneously the mineral was discovered in meteorites from Diablo Canyon, Arizona, (Frondel and Marvin 1967) and Goalpara (Hanneman, Strong, and Bundy 1967). Lonsdaleite was next discovered in the North Haig and Dingo Pup meteorites in Western Australia (Vdovykin 1970). Owing to the fact that natural lonsdaleite was identified in meteorites, it was suggested that the origin of this mineral was exclusively connected with shock metamorphism. This concept has even been incorporated into the standard description of lonsdaleite (Bates and Jackson 1980).

Controversy surrounded lonsdaleite's origin from the beginning. It was Kathleen Lonsdale (1971) who indicated the possibility of a nonshock-related origin for the mineral and who pointed out that lonsdaleite might be generated under conditions of static pressure.

Two years later, probable nonmeteoritic lonsdaleite was reported in titanium placers of the Ukrainian shield on schistose bedrock. The flat and irregular grains ranged from 0.05 to 0.3 mm (Sokhor et al. 1973). Four years later, lonsdaleite was described in eclogites in Sal'niye Tundra, Kola Peninsula, and in the Urals (Golovnya et al. 1977). In both deposits, lonsdaleite was closely associated with moissanite and graphite.

The latest discovery of lonsdaleite in diamond placers in Yakutiya (Kaminsky et al. 1985) indicates that this mineral can occur as a result of phase transformation in normal (cubic) diamonds. This possibility was experimentally confirmed: lonsdaleite was obtained during the process of dynamic synthesis of diamonds (Sozin, Belyankina, and Svirid 1983). Finally, a detailed geochemical study of diamond crystals established that natural diamonds contain inclusions of lonsdaleite enriched in carbonate and hydrogen ions (Miliuvina 1977). These data indicate that the presence of lonsdaleite does not necessarily mean meteor impact.

Lonsdaleite remained an obscure mineral until it was found in crystalloclastic rocks within the Popigay ring structure in Siberia and later in many other similar structures all over the world. Besides the widespread occurrence of lonsdaleite within in-crater deposits of Popigay-type ring structures, it is also dispersed hundreds of kilometers from the crater—possibly because it was entrained with pyroclastics thrown from the crater (Vishnevsky et al. 1997).

The presence of lonsdaleite within these ring structures stimulated arguments over the possible origin of the structures. Supporters of a terrestrial origin used essential geologic data (Erlich and Slonimsky 1986), whereas geochemical and mineralogic data have been used as proof of their meteoritic impact origin.

Detailed mineralogic descriptions of lonsdaleite were reported by Vishnevsky et al. (1997). The authors considered all existing data on lonsdaleite as decisive proof of meteoritic impact origin for this mineral. The presence of lonsdaleite in metamorphic rocks was explained as a result of Precambrian impact events.

Lonsdaleite is described here separately owing to the great genetic significance assigned to it by supporters of a meteoritic origin for astroblemes—ring structures where diamonds are present in hexagonal form. When examined with an electron microprobe, the mineral was found to consist of pure carbon. It belongs to the hexagonal class $6/m^2/m^2/m$ and the space group P6 3/mmc, 2:4. Lonsdaleite is pale brownish yellow under a microscope and faintly birefringent with *n* slightly higher than 2.404 (Roberts, Rapp, and Weber 1975). Its strongest diffraction lines are 2.19 (100) and 1.26 (75).

Its crystal structure is characterized by unit cell dimensions of 2.52 and 4.12 angstroms. The main spacing of lonsdaleite on the x-ray pattern is equal to that of cubic diamonds (d/n, nm): 0.206, 0.126, and 0.176 with the addition of several new lines, such as (d/n, nm): 0.218, 0.183, 0.150, 0.116, and 0.109. In structural aspect, lonsdaleite is a würzite-like polymorph of carbon.

The mineral was found in meteorites as cubes and cubo-octahedrons up to 0.7 mm diameter. The theoretical density of lonsdaleite is the same as that of cubic diamond (3.51 g/cm^3). The mineral is probably formed by the transformation of diamond to a hexagonal form.

In paramorphoses in astroblemes, cubic diamonds and lonsdaleite show twinning by the spinel law (Masaitis 1996). Most diamond morphologies found in the Popigay ring structure are characterized by lonsdaleite, although some are represented by a cubic phase. Minor impurities of chaoite and graphite are sometimes found in these diamonds (Valter et al. 1992).

Lonsdaleite from in-crater crystals within the Popigay structure have carbon isotope compositions similar to those of parental graphite, and they lie within the range of $-10\ \Delta^{13}C$ up to $-20\ \Delta^{13}C$ similar to C isotope compositions characteristic for carbonado (see Vishnevsky et al. 1997).

DISSOLUTION PHENOMENA

A characteristic feature of natural diamond crystals is that they constantly bear traces of dissolution. In distinct contrast to crystals of most other minerals, diamond crystals usually possess convex distorted faces (Shafranovsky 1964). Curve-faced surfaces on rounded diamond crystals are generally considered as dissolution surfaces because they cut the octahedral-plane stratification of many type I diamond crystals.

Kukharenko (1955) described diamond crystals that he termed “Brazilian type” and “Urals type.” These forms result from intensive dissolution of sharp-edged plane-faced crystals or more complex forms when the initial structural features of the faces have already been obliterated by dissolution.

Curve-faced surfaces of rounded diamond crystals cut across the octahedral-plane stratification of many type I diamond crystals (Lindley 1937; Kukharenko 1954). The origin of rounded surfaces is particularly evident in crystals of coated diamonds with yellow, dark-green, and other coloration. The coats are always evenly developed over all sides of the crystals when the crystal has plane faces. However, if rounded surfaces are prevalent, the relevant parts of the coat are much thinner than on the residual {111} faces or at the corner made by the L_3 axes. The dissolution origin of curved faces on diamond crystals is provided by the development of etch channels, changes of direction and branching of edges near etch channels, and the relative orientation and position of curved-channel surfaces in regard to rounded borders of etch channels.

Polycrystalline aggregates of diamond also show traces of dissolution: many specimens of carbonado have rounded angles producing "smoothed" shiny surfaces.

In addition to the general form of curved-face diamond crystals, there are other characteristics of dissolution. They include negative triangular pyramids (trigons), and rhomboidal cracks around the trigons. The two schools of thought on the origin of these features hold that they are a result either of diamond growth or of diamond dissolution. Most evidence supports the latter.

Dissolution may also be indicated by crystal etching. The triangular depressions on {111} faces typically are 10 microns across, but they may vary throughout a considerable range and be as much as 100 microns wide. Their depth varies from 0.9 to 2.0 microns. Corrosive etching of the faces of rounded crystals is usually accompanied by surface cracks with a rhombic network. These cracks may reflect increased temperatures at lower pressures. For instance, Titova (1960), artificially reproduced rhombic fractures by melting diamond crystals in air at 700°C for 2 hr.

Etch figures on the {111} and {100} crystal faces of diamond indicate that the theoretical rounded dissolution form of the diamond crystal should be a dodecahedroid. This observation and the Goldschmidt-Wright law (Goldschmidt and Wright 1904) support the interpretation that rounded crystals are dissolution forms. A dissolution origin for the triangular depressions on {111} crystal faces were first established in 1911 (Fersman and Goldschmidt 1911).

It has been suggested that oxygen released by thermal dissolution from etching will react with the carbon of diamond to produce CO_2. A series of experimental studies was designed to reproduce conditions needed to produce etchings. Various agents were used in these experiments including kimberlite melts, NaOH and KOH, Na and K niter, silicate melts in air, $KClO_4$ and $NaClO_4$, OH, Cl, and H + OH.

The selection of these kinds of dissolution agents reflects a prevailing conviction that although the exact composition of the dissolution agent can differ, generally they are oxysalts. If this conviction is true, it assumes an evolution in time of an oxidizing–reducing gas media, and it reflects the change in gas composition from the original CH_4 to CO_2 at the final stages of kimberlite emplacement. An explanation in physicochemical terms for such an evolution has been provided by Portnov (1982), who introduced the term *self-oxidation*.

Infrared examination of coated natural diamonds indicates that there are various amounts of carbonate and water in the concentric rings of the coated region, suggesting that the diamond grew in a magma containing carbonate and water (Chrenko, McDonald, and Darrow 1967). These results reflect the increasing role of CO_2 at the latest stages of the diamond's growth.

Experimental data vary on temperature ranges at which dissolution takes place. Mitchell (1961) used kimberlite melt at 1,450°C. In another experiment, etching was achieved at 380°C using $KClO_3$ as an oxidation agent for 181 hr (Patel and Ramanathan 1962). Mineral inclusions in diamond, such as olivine, pyrope, and chrome-spinel, do not show their usual plane-faced form but also appear as rounded irregular particles with resorbed surfaces, indicating that they have been affected by dissolution.

It is also noteworthy that diamonds found in eclogitic xenoliths show octahedral forms. Crystals with rounded habit have not been found among these diamonds.

These seemingly minute details of diamond crystallography have important implications for the assessment of diamond deposits. By examining diamond morphologies, one can assess the physicochemical conditions of the last stages of kimberlite formation to provide information on the degree of diamond preservation. In some cases, the conditions during kimberlite emplacement were favorable for the preservation of diamonds, but in other cases, partial or complete dissolution of diamonds may have occurred. The degree of crystal preservation depends on the conditions of formation of the original diamond crystals and a quantitative ratio of flat plane-faced crystals to curve-faced crystals. Transitional forms can be used as an indicator of diamond resorption (Milashev 1972, 1989).

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CHAPTER 3

Age and Origin of Diamonds: An Overview

METHODS OF DATING

Diamonds, or more precisely some minerals enclosed by diamonds, can be dated by radiometric-decay methods. Some inclusions (garnet, pyroxene, olivine, sulfides, and chromite) can be extracted and dated. The assumption is that a diamond and its mineral inclusions are essentially the same age.

Diamond itself cannot be dated because diamonds are composed only of carbon. The carbon does not contain radiometric decay elements useful in age dating, and ^{14}C dating is useless because the method depends on organic carbon that has been subjected to the recent carbon cycle at the earth's surface.

The methods used to date the mineral inclusions in diamond include the $^{145}Nd/^{144}Nd$, $^{87}Sr/^{86}Sr$, $^{40}Ar/^{39}Ar$, and Rb-Sr isochron methods. The data that are obtained by $^{40}Ar/^{39}Ar$ methods are considered to reflect the age of kimberlite emplacement, whereas data obtained from $^{145}Nd/^{144}Nd$ isotopes are considered to most reliably reflect the time of diamond formation.

AGES DETERMINED

Nearly every diamond examined to date (2001) has yielded a Precambrian age (3.3 Ga to 990 Ma). The one notable exception is a diamond from an unconventional source rock along the margin of a subduction zone in New South Wales, Australia, which yielded a date of 300 Ma.

The ages of mineral inclusions within diamonds show that most diamonds are significantly older than their host lamproite or kimberlite. Thus diamonds are xenocrysts, "foreign crystals," that were accidentally picked up by the kimberlite or lamproite magma during ascent to the earth's surface.

For example, most lamproites and kimberlites yield much younger ages (1.6 Ga to 1.0 Ma) than the diamonds themselves (Helmstaedt 1993). Again, mineral inclusions in peridotitic (P-type) diamonds from the Finsch and Kimberley pipes in Africa yield ages of 3.3 Ga, whereas the kimberlite host rock for these two pipes is only about 100 Ma. The Argyle olivine lamproite contains 1.58 Ga eclogitic (E-type) diamonds, whereas the lamproite itself is dated at 1.1 to 1.2 Ga (Kirkley, Gurney, and Levinson 1991). Sm-Nd model ages of the E-type diamond inclusions (1.58 Ga) from the Argyle lamproite postdate orogenesis (1.8 Ga) but predate the eruption of the lamproite (1.2 Ga). This relationship may suggest that these diamonds formed during postorogenic cooling of the metasomatized mantle.

E-type and P-type diamonds often yield different ages even within the same pipe. For example, E-type diamonds from the Finsch pipe are dated at 1.58 Ga, whereas the P-type diamonds are considerably older (3.3 Ga). The dates on E-type diamonds range from 2.7 Ga (Roberts Victor pipe, South Africa) to 990 Ma (for the 100 Ma Orapa pipe, Botswana). Typically, the P-type diamonds yield Archean ages, and many E-type diamonds often yield a wide range of ages that are predominately Proterozoic.

On the basis of geology and radiometric dating, the Orapa and Finsch kimberlites, South Africa, are Cretaceous. U-Pb dating of zircon from Orapa yielded a 93 Ma date. A 118 Ma Rb-Sr isochron was determined for the Finsch pipe. The Orapa diamonds are dominated by inclusions of the eclogitic suite, such as orange garnet, bluish-green clinopyroxene, and sulfides, which compose more than 85% of all inclusions, whereas in the Finsch they form less than 2% (Richardson et al. 1990). The Orapa eclogitic deep-orange garnet/clinopyroxene diamond inclusion pairs define a two-point isochron age of 990 ± 50 Ma. Similarly, the Finsch deep orange garnet/clinopyroxene pairs provide two-point isochron ages of 1580 ± 50 Ma. A single large garnet inclusion from Finsch analyzed by Smith et al. (1983) lies on the 1580 Ma isochron.

$^{40}Ar/^{39}Ar$ ages for eclogitic clinopyroxene inclusions in Orapa are close to the age of host kimberlite. It is assumed that in contrast to Sm, Nd, Rb, Sr, and K cations, which remain within the crystal structures of inclusion minerals, the noble gas Ar seems to be able to accumulate at inclusion/diamond interfaces, which effectively separates the radiogenic argon from parent K and resets the K-Ar clock within the clinopyroxene (Richardson et al. 1984). It is concluded that the $^{40}Ar/^{39}Ar$ technique will provide ages that are more indicative of the kimberlite emplacement event than of the diamond crystallization event.

At the Premier kimberlite, $^{40}Ar/^{39}Ar$ dates are similar for diamond crystallization and emplacement of the host kimberlite. Nevertheless, these diamonds have nitrogen aggregation characteristics that require a minimum period of 1 to 10 million years of mantle storage. The data imply that the diamonds have a xenocrystic relationship with the host kimberlite.

It has been suggested that kimberlitic magmas are stored in the mantle for as long as 1 billion years (see, for example, Kirkley, Gurney, and Levinson 1991; Richardson et al. 1990; Richardson et al. 1984). This long storage period is necessary if one accepts the idea that the time of diamond formation is related to multiple episodes of Proterozoic or Archean kimberlitic volcanism. To some degree the concept of several stages of kimberlite formation (and emplacement) relates to Davidson's (1967) concept without using his extremes of temperature of kimberlite emplacement.

It is assumed that the Proterozoic dates determined for the eclogitic suite of diamond inclusions are related to the transformation of the crust leading up to cratonization. It is also suggested that the Archean dates determined for the peridotitic suite of mineral inclusions reflect that these encapsulated minerals were formed at the same time as their diamond host.

CONDITIONS OF FORMATION

Diamonds are thought to exist within the earth's upper mantle, in unique host rocks that include eclogite, pyrope-harzburgite peridotite, chromite-harzburgite peridotite, and some pyrope-lherzolite peridotites. Equilibrium subcontinental geotherms require depths of 150 to 200 km at 45 to 55 kbar and 1,050° to 1,200°C to stabilize diamond within such cratonic environments. However, diamond may also possibly form in a cold subducting slab at depths of only 80 to 90 km at 22 to 25 kbar and 200° to 400°C. Many alkaline rocks originate at depths >100 km and could potentially entrain and preserve diamond in xenoliths or xenocrysts if ascent were rapid.

In many deposits, diamonds are partially resorbed en route to the earth's surface. This process is sufficiently common and significant that many diamonds lose their original octahedral form and assume a rounded dodecahedral morphology. In order to produce a dodecahedral form from an octahedral crystal, a weight loss of at least 45% is required. Under such conditions, any microdiamonds should be completely resorbed. This possibility has led to a proposal that the microdiamonds found in many kimberlites may represent a separate population of diamonds that crystallized before a diatreme was emplaced.

Diamonds containing inclusions that are characteristic of peridotite appear to have a somewhat different origin than diamonds containing inclusions characteristic of eclogite.

Peridotitic Diamonds

Most P-type diamonds are found in association with garnet or chromite harzburgite and less commonly in garnet lherzolite. These rocks, especially the harzburgites, have Mg-rich mineral components consistent with an origin as residua from a major mantle melting event involving the extensive eruption of ultramafic komatiites.

The tiny mineral inclusions found within the crystal structure of P-type diamonds are similar to those found in peridotite. This assemblage includes olivine, orthopyroxene, clinopyroxene, garnet, chromite, sulfides, and diamond. On the basis of geothermometry and geobarometry, P-type diamonds are interpreted to form at depths of 150 to 200 km.

The source of carbon in the peridotites, with few exceptions, is characterized by the narrow carbon isotope range of $\delta^{13}C$ values between −2 and −9. This narrow range suggests that the carbon source originated in a homogeneous convective zone within the upper mantle (asthenosphere). The carbon may have been one of the original components of the primitive Earth (Kirkley, Gurney, and Levinson 1991). P-type diamonds appear to be more abundant than E-type diamonds.

Eclogitic Diamonds

Unlike the case with P-type diamonds, organic carbon is implicated in the formation of some E-type diamonds. E-type diamonds display a wide range of carbon isotope compositions ($-34 < \delta^{13}C < +3$). The delta values of the carbon in carbonate and hydrocarbons found at the Earth's surface also span the range of +3 to −34, a range identical with values in E-type diamonds. The identical values provide evidence that the carbon in E-type diamonds may have originated at the earth's surface and was transported by a subducting plate to depths of greater than 150 km (Kirkley, Gurney, and Levinson 1991).

As a result, Kirkley, Gurney, and Levinson (1991) suggest that E-type diamonds originated either from an upper mantle that retained a primary heterogeneity or from crustal sources subducted into the mantle. Because most of the isotope ranges can be found within a single locality, it would be hard to explain this range of isotope values unless some E-type diamonds are related to recycling processes active during subduction events and plate movements.

The mineral inclusion assemblage in E-type diamonds is similar to those that form eclogite. These include orthopyroxene, clinopyroxene, garnet, chromite, sulfides, diamond, rutile, kyanite, corundum, and coesite. On the basis of geothermometry and geobarometry, E-type diamonds are interpreted to have formed at depths possibly greater than 300 km (Kirkley, Gurney, and Levinson 1991).

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CHAPTER 4

Gemstone and Synthetic Diamonds

GEMOLOGY

The primary monetary value of diamonds is as gemstones. Many gem diamonds are worth many times that of equivalent weights of gold, platinum, palladium, or rhodium. Rough gemstone diamonds additionally have values that are as much as 100 or more times that of industrial diamonds. After the diamonds are cut and polished, the value of the gem can increase another 10 to 100 fold.

Diamonds are the most valuable gemstones on earth. For example, some Argyle pink diamonds have sold for as much as $1 million (Australian) per carat. The extreme value of diamond as a gemstone is due to its mystique, rarity, extreme hardness, and high refractive index and dispersion that can result in brilliant gems with distinctive "fire" when they are faceted and polished.

Four general types of commercial natural diamonds are recognized in the diamond trade. These are *gem* (well-crystallized and transparent), *bort* (poorly crystallized, gray, brown translucent to opaque), *ballas* (spherical aggregates of many small diamonds), and *carbonado* (opaque, black to gray, tough, and compact; a variety of industrial diamond). Gem diamonds are sometimes further subdivided into *gem* and *near-gem* (low-quality gemstones).

Two types of diamond with different conductive properties have also been recognized. These are Type I and Type II diamonds. Unlike the Type II diamond, the electrical conductivity of Type I is poor. Diamond conductivity is a useful property in identifying diamonds. For instance, the Gemological Institute of America uses user-friendly, relatively inexpensive GEM® testers that measure the surface conductivity of minerals and reliably separates diamond from other gemstones.

Diamond colors vary. The color of diamonds may range from colorless to black and include shades of pale yellow, pink, red, orange, green, blue, and brown. Deeper shades are rare, but when such stones are found they are termed *fancy*.

The gemology of diamonds is an important part of the science of diamonds, because the fashioning of diamonds into gemstones takes into consideration many facts to produce gems that will yield the greatest light reflectance, "fire," or brilliance. In general, there are six steps in fashioning a diamond crystal ("rough") into a finished gem. These are marking, grooving, cleaving, sawing, girdling, and faceting (Hurlbut and Switzer 1979). Whether or not all of these steps will be used in the manufacture of a given gem depends on the size, shape, and quality of the rough stone.

Marking determines how the rough should be cut to produce the best gemstone. In this planning stage, the size and shape of the rough, the number and location of imperfections, and the direction of the cleavage (referred to as "grain" by gem cutters) are considered. If the diamond is large, the stone may be cleaved. Large diamonds may be preshaped by cleaving them into pieces that are suitable for sawing. When this is done, the octahedral cleavage selected for cleaving is marked on the diamond with a pen, and a small groove is cut into the octahedral plane using a sharp-edged diamond chip. The diamond mounted on a dop (holder), with a steel knife placed in the groove, is then cleaved by striking the back of the steel knife with a mallet.

The cleavage is a weak point in the diamond; thus, if the cleavage is properly identified and marked on the diamond, the stone should break along a smooth cleavage surface. However, if the cleavage is improperly identified, then the diamond can shatter into several pieces producing fragments with cleavage breaks and shards with conchoidal fracture.

Most primary shaping is completed by cutting the stone with a diamond saw. After the stone is placed in a clamp and the cleavage is marked, the diamond is either cut parallel to the cube or parallel to the dodecahedron with a rapidly rotating blade impregnated with diamond powder. Typically, it takes 4 to 8 hr to complete a cut through a 1-carat diamond (Hurlbut and Switzer 1979).

Girdling (bruting and rounding) is completed by placing the diamond on a lathe. While the diamond rotates on the lathe, a second diamond attached to a dop is pushed against the rotating diamond in order to round it into a cone. The diamond dust produced during the operation is collected for later use in saws and laps.

Faceting is accomplished by grinding and polishing, and it is completed on a revolving horizontal lap impregnated with diamond powder. The diamond being processed is held at various angles with a mechanical holder, or on a dop, and polished. In a standard, round, brilliant diamond, as many as 58 facets may be cut and polished.

Any given diamond possesses significant variations in hardness that are considered during polishing. The optimum directions for polishing are those that parallel the crystallographic axes. Because the cubic faces of the diamond are parallel to the crystallographic axes, they are easiest to polish. Not all facets are parallel to a crystallographic axis, but the most favorable direction for polishing is the one that most nearly lies parallel to an axis.

The dodecahedral faces lie parallel to one crystallographic axis, and each dodecahedral face has one optimum polishing direction. The octahedral face is the hardest face on the diamond, as it lies at the greatest angle from the crystallographic axes. The octahedral faces are equally inclined to three crystallographic axes and are the most difficult to polish. It is almost impossible to polish a facet parallel to an octahedral face and almost impossible to saw a diamond if the plane of the cut varies more than a few degrees from that of a cubic face. Diamonds are then cut so that the table (the top) of the cut diamond is parallel to the face of a cube, octahedron, or dodecahedron.

The value of finished gem diamonds is judged by what are termed "the four Cs"—cut, clarity, carat weight, and color. The cut of a diamond can increase its value tremendously, and the better proportioned, polished, and faceted the stone, the greater its value. When the girdle (base) and table of the diamond are proportioned correctly, the diamond will exhibit greater fire and brilliance.

The color of the diamond can greatly affect its value. Diamonds include fancy diamonds and white (colorless) diamonds. The colorless diamonds range from colorless (white) and blue-white to pale yellow (Bruton 1978). One of the more common systems for evaluating diamonds is the Gemological Institute of America's color grading system, which ranges from D (colorless) to X (light yellow) (Hurlbut and Switzer 1979). Each letter of the alphabet from D to X shows a slight increase in yellow tinge—not apparent to the untrained eye.

The color grading may be visual or instrumental. The visual appraisal is done in a well-lighted room using natural north light through windows. Such appraisals are usually done by comparison with a master set of instrument-graded diamonds. The instrument used in color grading is a colorimeter, which measures the degree of yellowness of a diamond quantitatively (Hurlbut and Switzer 1979).

The fancy diamonds are separated from the colorless diamonds. The brown and green diamonds are first separated from the other fancies, and the other stones are separated in groups including full-bodied yellow stones termed "golden fancies." The other full-bodied fancies are separated into deep blues, pinks, and ambers (Bruton 1978).

Clarity is determined by the presence or absence of blemishes, flaws, and inclusions. Many of the grading systems in use have descriptive terms such as *flawless* or *imperfect*, and terms denoting intermediate grades such as *VVS* (very, very slightly imperfect). The scale used by the Gemological Institute ranges from Fl (flawless) to I_3 (imperfect). The intermediate grades include VVS_1, VVS_2, VS_1, VS_2, SI_1, SI_2, I_1, and I_2.

TYPES OF CUTTING

It was recognized in ancient India that rubbing the surface of diamond would polish it and sharply increase the luster. In most ancient times in India and later in Europe (Italy, Belgium, and France) diamonds were cut by grinding, using natural octahedral diamonds.

Legend says that such diamonds decorated the mantle of the French king St. Louis (1214–1270). The first European to learn how to polish diamonds was Ludwig Burkem. In 1454, he cut the first diamond, later called Sancy (see *Famous Diamonds*). After Burkem's death the secret of cutting was lost, and the method of polishing had to be reinvented.

The cleaving of the diamond requires detailed studies of the diamond's crystallography and internal defects. Diamonds have been sawed since the start of the seventeenth century. In this process, the saw blade is impregnated with small diamonds. Sawing is time-consuming and costly. For example, sawing the Regent diamond, which weighed 410 carats, took about two years. In the twentieth century ultrasonic diamond-cutting instruments were developed.

The final phase of diamond fashioning is polishing to create surfaces with regularly arranged facets of specific form and mirror-smooth surfaces. The optimum direction for polishing is parallel to a crystallographic axis of the diamond, because the hardness of the diamond varies depending on its crystallographic orientation. The direction parallel to the octahedral faces are equally inclined to three crystallographic axes and are the most difficult—almost impossible—to polish (Hurlbut and Switzer 1979).

Some early styles of diamond polishing merely polished the natural crystal faces. After the importance of refraction and reflection of light in the gemstone became understood, the diamonds were faceted. For example, when light passes into a gemstone and strikes the back facet at an angle that is greater than the critical angle, it will be totally reflected. Thus, if the angles of the back facets are correct for a given refractive index, a

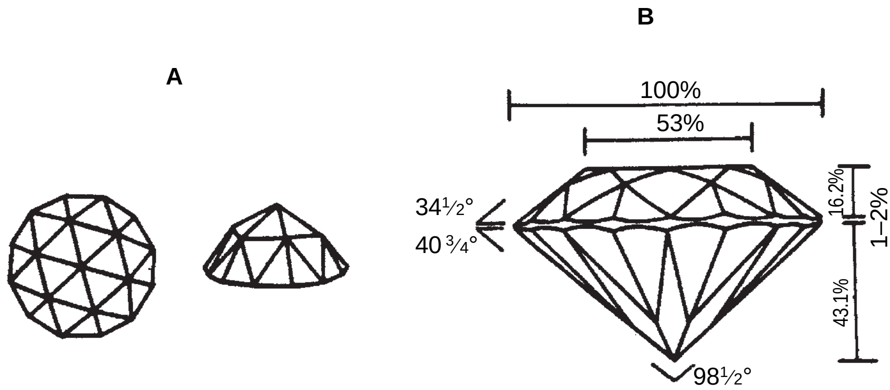

FIGURE 4.1 The rose cut of diamond (A) and the geometry of an ideal brilliant round cut (B). Reprinted with permission from Publishing House Nedra. Taken from Milashev 1989.

large portion of the light entering the top of the stone will be reflected twice before being directed to the eye of the observer. As such, both brilliance and color will be improved and striking. The difference in the indices of violet and red light in diamond is five times as great as those in quartz and two times as great as those in the best types of glass.

Through the ages, several different faceted cuts have been developed for diamonds. Possibly the earliest cut was that of the point cut. In this cut, an octahedral diamond was left in its natural form. Grinding to remove the top point of the octahedron to produce a flat table surface came later. The table cut took advantage of the octahedral shape by faceting one of the octahedral points parallel to the cube to produce a flat table, while keeping the remaining octahedral faces.

For this purpose facets are usually arranged in belts. Facets within each belt are usually inclined and come together forming apexes, concentrating light in one point. In some types of cuts, belts are divided by a flat zone to allow additional light to be gained.

There are three traditional types of cuts (Milashev 1989): the step-like cut, the rose cut, and the brilliant cut.

In the step-like cut, facets of adjacent belts are parallel and located exactly one over another. The facets are trapezoidal and triangular in form. The top is polygonal with acute or obtuse angles.

One of the earliest cuts, the rose cut (Figure 4.1) was probably developed more than 500 years ago. The rose-cut stone typically has a flat base, whereas the upper portion is convex and consists of 6, 8, 12, 24, or 31 facets. Stones with 12 facets or fewer are termed “rose d’Anver.” Those with more than 12 facets are termed “crown roses.” Both the Kohinoor and the Great Mogul diamond were shaped into roses in the 1500s (Hurlbut and Switzer 1979).

Many other types of cuts were developed (Figure 4.2). Some of the more common include the step or emerald cut, which is sometimes referred to as the trap or cushion octagon cut. Many emeralds have been fashioned with this cut. It is characterized by a large prominent table facet that is flanked by parallel rows of trapezoid-shaped facets. The pavilion (base) for this cut has similar rows of parallel facets that decrease in steepness toward the basal facet, or culet. The outline of step cut stones may be square, rectangular,

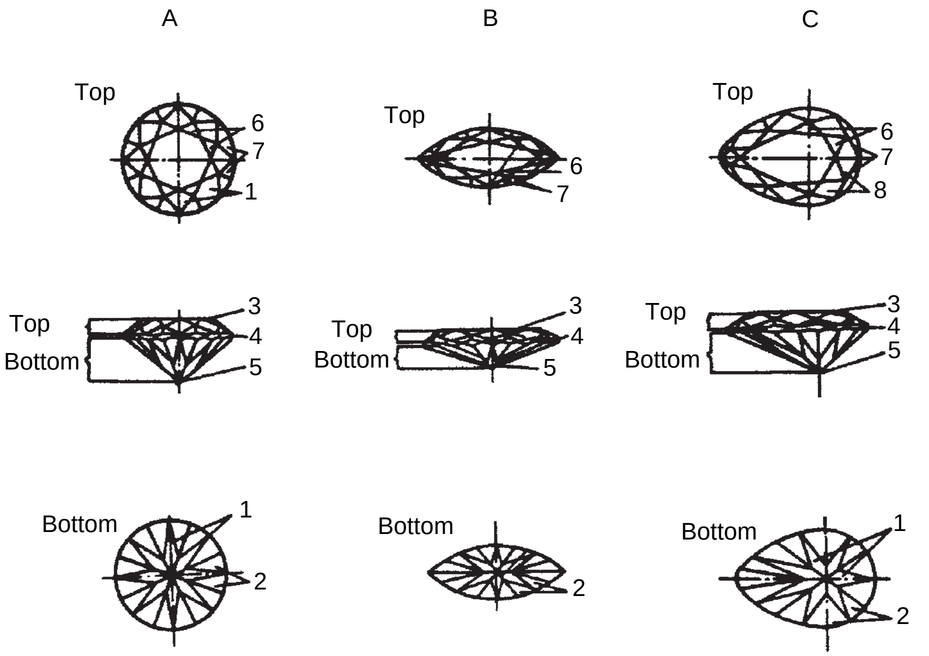

FIGURE 4.2 Complete brilliant cut of diamonds. A = round brilliant (57 facets), B = marquise (55 facets), C = pear-like (56 facets). 1 = crystal faces; 2 = wedges; 3 = plane; 4 = rundist; 5 = calette; 6 = upper wedges; 7 = lower wedges; 8 = middle wedges. Reprinted with permission from Publishing House Nedra. Taken from Milashev 1989.

triangular, kite shaped, keystone shaped, lozenge shaped, or other shapes. Small step-cut stones with a long rectangular shapes are baguettes.

The brilliant cut is the most frequently used and has become the most common in modern times. The standard brilliant cut consists of 58 facets, 33 on the top or crown of the stone and 25 on the base. Except for the table or culet, all other facets are either triangular or kite shaped. Some modifications of the standard round brilliant are the pear, oval, and marquise cuts.

Light gain in brilliant-cut stones is much greater than in rose-cut stones. Thus for stones of the same size, the color and clearness of a brilliant cut is much greater.

In the 1960s, a Belgian cutter, M. Westreich, proposed a new cut consisting of 73 facets that significantly increase the light gain. This cut, referred to as a "highlight cut" is usually used on stones greater than 1 carat.

Massimo Elbe also proposed another type of cut, the "impariant." While traditional cutting is based on symmetry of eight-faceted stones, the impariant cut uses an uneven number of facets. In the impariant cut, light is reflected out of the stone in a wider spectrum thus producing more beautiful stones. The luster of impariant-cut stones is 25% to 30% higher than that of brilliant-cut stones. The increased luster and light gain greatly improve the color of the stones.

FAMOUS DIAMONDS

The list of famous diamonds includes many rare and spectacular gems that are considered to be priceless (see Table 4.1). Many of these fabulous diamonds were found in India, Africa, and Russia. Some gems have dramatic stories surrounding their discovery and ownership, and many are interlaced with historical events; they may bear names that reflect their association with great historical figures or with ideological concepts of then-current bureaucracies.

It is commonly believed that many of the greatest diamonds that were found in the seventeenth century came from the famous Golconda mine in India. Other well-known and famous diamonds were found at the end of nineteenth century and during the first half of the twentieth century and were the products of mining at the Premier and Jagersfontein mines in South Africa. Most of the great Soviet diamonds were found in the second half of the twentieth century.

In recent years, a new variety of spectacular diamonds, "Argyle Pinks," has reached the status of "famous diamonds," principally because of their rare color rather than for their size. These diamonds have been produced from the Argyle mine in northern Australia since the mid-twentieth century. Other fabulous diamonds will be found in other places of the world and may also gain prominence in history. In particular, we expect to see some diamonds from the Canadian Northwest Territories making their mark.

Diamond weights are recorded as carat weight. The carat weight formerly varied in different countries. However, the metric carat of 200 milligrams (0.2 grams) was adopted in the United States in 1913, and the carat is now standardized throughout much of the world (Gaal 1977). The historic difference in the carat weight has resulted in a discrepancy of diamond weights reported by various sources in past years. Other differences in the carat weight may be due to the later recutting of the stone.

Baumgold Rough

The Baumgold Rough diamond was recovered from the Wesselton mine, South Africa, in 1922 and weighed 609.25 carats. This bluish-colored diamond was cut into 14 stones. The largest were two 50-carat, pear-shaped gems (Gaal 1977).

Cullinan

The most famous diamond in history was named the Cullinan. This diamond is famous not only because it was the largest diamond ever found, but also because it had such high clarity. It was cut into many stones, several of which are placed in the crown jewels in England.

The Cullinan diamond was recovered from the pit wall of the Premier mine in the Transvaal, South Africa, by the mine's manager, Fredrick "Daddy" Wells, in 1905. The diamond was named the Cullinan after the mine's owner and discoverer, Sir Thomas Cullinan. When he found the diamond, Wells was on an inspection tour and noticed a colorless stone protruding from kimberlite exposed in the pit wall. He walked over to the pebble and dug out the 10×6.35 cm diamond with a penknife. The stone weighed nearly 0.6 kg (3,106 carats), considerably more than any other diamond ever found. It was an irregularly shaped diamond with a flat, broken surface. The flat surface indicated that a sizable portion of the diamond had broken off the Cullinan; it was never found (Arnold Waters Jr., personal communication, 1978). It is possible that this portion of the

TABLE 4.1 Famous diamonds and brilliants of the world (excluding Russian diamonds) (after Milashev 1989 with corrections by Copeland 1973)

Mass (carats)	Name of Stone	Place and Year of Discovery	Comments
3,106.0	Cullinan*	South Africa, 1905	Produced 105 brilliants; the largest was the Star of Africa (530.2 carats); total weight of all gems produced from the diamond was 1,063 carats.
995.2	Excelsior*	South Africa, 1893	Prepared 21 brilliants ranging from less than a carat to 70 carats.
968.9	Great Mogul*		Produced a brilliant that weighs 279 carats.
961.0	The Star of Sierra Leone		The diamond was valued at $12 million.
787.0	The Diamond of Victory		
726.6	President (Presidente) Vargas		Value of diamond was $600,000 in 1939. Produced 29 brilliants, 16 of which had a weight of 10 to 48.6 carats.
726.0	Jonker*	South Africa, 1934	Purchased for 146,000 pounds. Produced 12 brilliants with a weight of 5.3 to 142.9 carats; total weight of all brilliants was 370.87 carats.
650.8	Jubilee*	South Africa, 1895	Produced two brilliants. The weight of the largest is 245.3 carats. Donated to Smithsonian Institution.
616.0	DuToitspan (Oppenheimer)	South Africa, 1964	Large pale-yellow octahedron that measured 0.2 x 1.25 in., with a few black inclusions. Donated to the Smithsonian Institution.
609.25	Baumgold	South Africa, 1923	
601.25	Lesotho Brown	Lesotho, 1967	Sold for $302,400. Produced 18 brilliants, weight of the largest was 71.73 carats.
600.0	Goyas	Brazil, 1906	Broken in process of determination.
599.0	Centennial	South Africa, 1987	Named after centennial anniversary of DeBeers.
527.0	Name unknown	Lesotho, 1965	
511.25	Venter	South Africa, 1951	
503.0	Kimberley	South Africa, 1900	Flawless 70-carat emerald-cut champagne-colored diamond; was recut in a modern shape in 1921. Valued at $500,000.
469.0	Victoria 1884 (Imperial, Great, Great White Diamond)	South Africa, 1884	Produced two brilliants, an oval that weighs 185 carats and a round with a weight of 277 carats.
460.0	Darsu Vargas	Brazil, 1939	
440.0	Nizam	India, 1835	Weighed 340 to 440 carats but was broken during Indian mutiny in 1857. A broken piece was sold to an Indian broker for £200,000. Cutting reduced weight to 277 carats.
434.0	Torch of the World	West Africa, 1969	
428.5	Victoria 1880	South Africa, 1880	Produced a brilliant that weighs 228.5 carats.
428.5	DeBeers	South Africa, 1888	Produced a brilliant that weighs 234.5 carats.
426.5	Ice Queen (Niarkhos)	South Africa, 1954	Produced three brilliants (197, 40, and 30 carats). Largest bought by Greek shipbuilder for $2 million.

(*Table continues on next page*)

TABLE 4.1 Famous diamonds and brilliants of the world (excluding Russian diamonds) (after Milashev 1989 with corrections by Copeland 1973) (continued)

Mass (carats)	Name of Stone	Place and Year of Discovery	Comments
416.25	Berglen	South Africa, 1924	
412.5	Broderik	South Africa, 1928	
410.0	Regent (Pitt)*		Produced a brilliant that weighs 136.9 carats.
409.0	President Dutra	Brazil, 1949	Produced 16 brilliants with total weight of 136 carats.
400.65	Koromandel	Brazil, 1941	
384.0	Ark	South Africa, 1948	
380.0	Red Cross	South Africa, 1948	Produced a brilliant that weighs 205 carats.
367.0	Raja Maltansky*	Borneo, 1787	
353.9	Blue Diamond	South Africa	Produced three large and two small brilliants. The largest, the pear-shaped “Rosa Premier” (137.02 carats), was valued at $5.4 million in 1977. Now in a private U.S. collection.
350.0	Black diamond from Baii	Brazil	Displayed at the 1851 exhibition in London.
296.0	Stuart	South Africa, 1872	Produced a brilliant that weighs 123 carats.
287.42	Tiffany (African Star)*	South Africa, 1978	Produced a brilliant that weighs 128.51 carats. Valued at $500,000.
254.0	Southern Star*	Brazil, 1853	
250.0	White Tavernier	India, 1500s	Great plate-like diamond.
244.0	DuToit	South Africa	
240.0	Transvaal	South Africa	Total weight of brilliants produced after first cutting, 75 carats. After second cutting, 68.79 carats.
228.0	Koromandel 3	Brazil, 1936	
200.0	Kruger	South Africa	An alluvial diamond presented to President of Transvaal by an African tribal chief.
189.62	Orlov (Darya-I-noor)*	India, 1600s	
186.0	Koh-i-Noor*	India, 1300s	Weight after recutting, 106 carats.
183.0	Moon Diamond (Moon)	South Africa	Sold at auction in London in 1942.
180.0	Koromandel 4	Brazil, 1934	
175.0	The Star of Minais	Brazil, 1911	
172.5	Minas Gerais	Brazil	
160.0	Beauty Helen	South-West Africa, 1951	Produced three brilliants with a total weight of 70.49 carats.
158.79	Chan Lin	China, 1977	
155.0	Liberator	Venezuela, 1944	Named after Simon Bolivar, Liberator of Venezuela. Produced four brilliants. Purchased in 1943 by Harry Winston of New York and cleaved into two pieces weighing 115 and 40 carats. These were fashioned into four stones; 56% of original weight was lost in process of cutting.

(*Table continues on next page*)

TABLE 4.1 Famous diamonds and brilliants of the world (excluding Russian diamonds) (after Milashev 1989 with corrections by Copeland 1973) (continued)

Mass (carats)	Name of Stone	Place and Year of Discovery	Comments
154.50	Winston	South Africa, 1952	Produced a brilliant that weighs 62.05 carats.
150.0	Porter Rhodes	South Africa, 1880	
150.0	Portuguese Diamond	Unknown	Weight after cutting, 127.0 carats.
146.0	Taj-e-Mah	India	A sister stone of Darya-I-Noor. Now among crown jewels in Tehran.
141.0	Koromandel 5	Brazil, 1936	
137.27	Florentine (Great Duke of Tuscany, Austrian Diamond)	India, 1300s	Great yellow diamond of Medici family. Light greenish-yellow in color and fashioned in the form of irregular nine-sided 126-facet double rose cut. Belonged to Charles the Bold of Burgundy. Property of Austrian Imperial Court. Was taken to the United States in 1920s; present location unknown.
133.0	Colenso	Unknown	Presented in 1887 to British Museum of Natural History by John Ruskin in honor of his friend John Williams Colenso, a distinguished mathematician and the first bishop of Natal.
133.0	Golden Diamond	South Africa, 1913	Produced a brilliant that weighs 61.5 carats.
130.0	Cartier	Africa, 1974	Produced a brilliant that weighs 107 carats. Insured for $5 million.
124.27	Chen-Fu-II	China, 1981	
Original weight unknown	Dresden Green	India, unknown	Striking apple-green, pear-shaped stone weighing 41 carats. The largest diamond of this color in existence. Was sold to Frederick Augustus II of Saxony for $150,000 in 1743. Was confiscated by the Soviet Trophies' Organization after World War II. In 1958 was returned to East Germany. Now on display at Dresden Historical Museum.
119.01	Mun-Shan-I	China, 1983	
119.0	Shah Akbar	India, 1618	The stone sold for about $175,000 to the Gaekwar of Baroda. Present location unknown.
112.25	Blue Tavernier	India, 1676	Produced a brilliant that weighs 67.5 carats.
112.0	Jagersfontein	South Africa, 1891	Produced a brilliant that weighs 56.6 carats.
102.0	Ashberg	Unknown	Brought from Russia to Sweden after 1917 revolution. Displayed at exhibition in Amsterdam in 1949. Offered for sale in 1959.
101.2	Gastings	India, 1500s	
100.2	Jacob	India	Sold in 1956 for $280,000.

* More detailed explanation in the text.

diamond never reached the surface of the earth during transportation in the kimberlite, was found and smuggled out of the mine, or was missed and sent to the crushers during dressing of the ore.

As a private buyer for the priceless stone could not be found, the diamond remained in the possession of the mine owner for 2 years until it was purchased by the Transvaal government and presented to King Edward the VII of England on his birthday (Bruton 1978). The stone contained fractures and impurities, and it was decided to cut the stone into several gems to eliminate the impurities. The best cutter in Europe at the time was Joseph Asscher, who studied the stone for 6 months before he established the proper place to "open" it. After a deep kerf was scratched into the stone, the first attempt to cleave the diamond ended up in breaking the cleaving chisel. The stone was successfully cleaved on the second attempt, at which time Asscher promptly fainted (Harlow 1998).

The fashioning and cutting of the two major fragments of the Cullinan took 2 years and produced two large brilliants. One of these is the largest fashioned diamond in the world, the Cullinan I, or Great Star of Africa. This gem is a pear-shaped jewel weighing 530.2 carats. The second largest fashioned diamond in the world, the Cullinan II (Lesser Star of Africa), is a 317.4-carat cushion cut. Other stones cut from the original Cullinan are the Cullinan III, a pear-shaped diamond weighing 94.4 carats, the Cullinan IV, a 63.6-carat square brilliant cut, and five other large diamonds and 96 smaller stones (Gaal 1977). The completion of the work was celebrated in 1912 by a great banquet.

The total weight of gems cut from the original Cullinan was 1,063.65 carats, or about 34.25% of its original weight. The nine largest diamonds remain in the possession of the British royal family.

The Premier mine has been a source of many large diamonds. In 1905, two other large stones weighing 334 carats and 600 carats were found at the mine. In 1919, a fragment with a total weight of 500 carats was recovered, and in 1954 another large stone of a total weight of 426.5 carats was found. The latter diamond, the Ice Queen, was a flawless diamond with exceptionally fine color that ended up producing three brilliant gems weighing 197 (Gaal [1977] reported this gem to weigh only 128.25 carats), 40, and 30 carats. In 1957, the Greek shipbuilder Stavros Niarchos purchased the largest diamond for $2 million.

The Premier mine has annually produced 15 to 20 stones in the range of 100 to 200 carats. About half of them have been of gem quality and typically valued at more than $4,000 per carat. In 1934, another large diamond weighing 726 carats was found in alluvium close to the Premier mine on a farm owned by Jacobus Jonker. The diamond was of unusually fine color and purity and was named the Jonker (Milashev 1989).

Daryainur

The Daryainur diamond is considered to be the most valuable diamond in the Iranian crown jewels and one of the first diamonds found by man. The cut diamond is estimated to weigh 176 carats and is a crudely fashioned pale pink stone measuring $3.8 \times 2.5 \times 0.66$ cm. *Daryainur* means Sea of Light.

DeBeers

The DeBeers diamond was recovered from the DeBeers mine in South Africa in 1888 and sold to an Indian prince. The yellow octahedron weighed 428.5 carats in the rough and was cut to a 234.5-carat gem (Gaal 1977).

Dewey

The Dewey diamond is a well-known diamond that was found in the United States, although its size and color would have made it insignificant had it been found in the Russian or African diamond fields. The diamond was discovered in 1885 in the James River valley near Manchester in eastern Virginia. The site of the discovery is now part of the city of Richmond (Sweet 1997). The source rock for the diamond has not been found. The diamond consisted of a slightly rounded, trigonal trisoctahedron weighing 23.75 carats that was cut into an 11.69 carat gemstone. The stone had a faint greenish-white color with perfect transparency (Kunz 1885).

Doubledipity

The Doubledipity diamond is of scientific curiosity only because it was found in the United States. The diamond was recovered from the Trinity River of northern California in 1987 and weighed 32.99 carats. The stone was a yellowish-brown diamond that lacked the typical adamantine luster commonly associated with diamond. It was opaque except on the edges where it was translucent and nonfluorescent. It was an aggregate of seven interpenetrating cubes in random crystallographic orientation (Kopf, Hurlbut, and Koivula 1990).

Excelsior

The Excelsior is the second largest diamond found in the world. This diamond was recovered from the Jagersfontein mine in South Africa in 1893 and weighed 995.2 carats (about 7 oz). The diamond was recovered by a worker who spotted the diamond while he was digging gravel. He concealed the diamond from the overseer until he had the opportunity to deliver it directly to the mine manager. In addition to a cash reward, he was also given a horse, saddle, and bridle for his find (Gaal 1977).

The stone was irregularly shaped with a cleavage face, and it had high clarity and a blue-white color. Following its discovery, the priceless stone remained for 10 years in the possession of a London syndicate that was unable to sell it. Finally, the stone was sent to Asscher's diamond company in Amsterdam and from it 21 gems were fashioned that had an aggregate weight of 373.75 carats representing 37.5% of the original weight of the diamond. The largest was a 69.8-carat marquise (Bruton 1978). Gaal (1977) reported that five pear cuts were produced weighing 47.15, 47.03, 34.97, 18.0, and 16.81 carats. Four marquise cuts weighed 40.36, 28.55, 26.37, and 24.38 carats. Eleven brilliant cuts with a total weight of 20.33 carats were produced. Tiffany and Company marketed some of the stones.

Great Mogul

The Great Mogul diamond was found in India in the middle seventeenth century. In the rough, the diamond weighed 787.5 carats. It was one of the treasures of the Shah Jehan, builder of the Taj Mahal and owner of the Kohinoor diamond. The diamond was cut into a rose cut of 280 carats. This diamond may be the same as the Orlov, which is described in the following text.

Hope

One of the world's most famous and celebrated diamonds is the Hope diamond. Few other diamonds have had such an intriguing history or have exhibited such a distinct blue color. The Hope diamond is believed to have originated from the Tavernier Blue diamond (110.5 carats), which was found in India. The Tavernier Blue was cut producing the 69.03-carat French Blue diamond; it was stolen from the French Royal Treasury in 1792 and never recovered. In 1830, a diamond of similar color but smaller size appeared on the London market and was purchased by Henry Philip Hope for $90,000.

The diamond developed a reputation for bad luck, as it was associated with many violent deaths and with disasters in two royal families. It was finally purchased by Harry Winston, a New York gemologist, for about $180,000 and donated to the Smithsonian Institution. In 1975 the diamond yielded a weight of 45.52 carats. It is still assumed that the stone was cut from the French Blue. The other stone believed to have been produced from cutting the French Blue was a 13.75-carat diamond that became known as the Brunswick Blue (Gaal 1977).

Jonker

The Jonker diamond, found in 1934 in alluvium on a farm owned by Jacobus Jonker near Pretoria, South Africa, weighed 726 carats. The diamond was of unusually fine color and purity. After cutting, the stone produced a marquise and 11 emerald cuts. The largest of the fashioned stones retained its namesake and was fashioned into a 142.9-carat emerald cut that was recut in 1937, producing a 125.65-carat gem. The diamond was purchased by King Farouk of Egypt, who later sold it to Queen Ratna of Nepal after Farouk was exiled from Egypt (Gaal 1977).

Jubilee

The Jubilee diamond was recovered from the Jagersfontein mine in 1895. The stone weighed 650.8 carats and was the third largest diamond found on Earth. It was an irregularly shaped stone of approximate octahedral shape and exceptional white color, clarity, symmetry, and brilliance. The diamond was originally named the Rietz after the president of the Orange Free State. A cushion-shaped gem of 245.35 carats cut from it was named the Jubilee, because it was cut in 1897 during Queen Victoria's diamond jubilee year (Bruton 1978). A second pear-shaped gem weighing 13.35 carats was cut from the original stone (Gaal 1977).

Julius Pam

This diamond was recovered from the Jagersfontein mine in 1889 and was reported by Gaal (1977) to have weighed 246 carats. Milashev (1989) reported the diamond to weigh 373.5 carats. The diamond was cut producing a 123-carat gem.

Kelsey Lake Diamonds

Some attractive diamonds (<1–28.3 carats) have been recovered from the Kelsey Lake diamond mine along the Colorado–Wyoming border in the United States. Individual names of the diamonds are unknown.

One attractive stone was a 28.18-carat diamond that was cut into a 16.3-carat canary-yellow gem, which was sold to a collector for about $300,000. Another stone of 28.3 carats was cut to a small 5.39-carat gem and sold for $87,000. Another attractive and flawless 14.2-carat octahedron was also recovered from the operation (Hausel and Sutherland 2000).

Kohinoor

The Kohinoor diamond was first mentioned in 1304 when the Sultan Alladin Kili stole the stone from the Sultan of Malwa (Malwa was a large territory in India) and took it to Dehli. Two centuries later (about 1526), the diamond was in the possession of the Sultan Baber, the first of the Mogul emperors who had conquered India. His son Khamayun found several precious stones in the Agra fortress, including this diamond. In the beginning of eighteenth century, the Persian Shah Nadir saw this shining stone in the Sultan's crown and exclaimed: "This is the real Koh-i-Noor!" ("Mountain of Light"). The diamond retained this name.

When Nadir conquered Delhi in 1739, the Sultan Muhammed Gurkhan tried to protect the diamond from the conqueror by hiding the stone in his turban. However, Nadir suggested that the two sultans exchange their hats as a sign of friendship. So, Nadir gave the Sultan his leather hat in exchange for the turban containing the Kohinoor. In 1747, Nadir was assassinated by his own guard. General Abdali, the king's chief guard, took the stone and escaped to Afghanistan and proclaimed himself king in Kabul.

In 1813, the King of Lahore, Ranjit Sing, returned the diamond to India. But in 1849, the army of the British East India Company burst into Lahore and forced the 12-year-old heir to the throne to sign an agreement by which he abdicated the throne as well as the Kohinoor. The British East India Company then presented the diamond to Queen Victoria of England. In 1852, Queen Victoria had the stone polished under the guidance of distinguished Amsterdam jeweler Vorsanger. The polishing continued for 38 days at a cost of 3,200 kg; the result was disastrous. The great stone of 186 carats was diminished to 106 carats (Milashev 1989). Gaal (1977) reports the final weight of the stone as 108.93 carats.

In 1911, the Kohinoor was placed in the so-called Small King's crown made for Queen Mary. In 1976, the Pakistani government appealed to the English government to return the Kohinoor. However, the Kohinoor remains as one of the most famous jewels in the English crown (Milashev 1989).

Lesotho

The brownish Lesotho diamond, discovered in 1967 in the Letseng-le-Draai diggings in Lesotho, South Africa, weighed 601.25 carats. The diamond was sold for $303,400 at auction to a South African diamond dealer; later it was purchased by Harry Winston of New York, who had it cut into 18 stones totaling 242.5 carats. The largest gem cut from the diamond was a 71.73-carat emerald cut. Two other large rough diamonds were named the Lesotho B (527 carats) and the Lesotho C (338 carats) (Gaal 1977).

Orlov Diamond

The Orlov (Orloff) diamond was a fragment of a larger diamond thought to originally weigh about 400 carats. The stone was a beautiful, water-transparent stone with slight greenish to bluish tinge that was cut to a 189.62-carat gem measuring $22 \times 32 \times 35$ mm (Gaal 1977).

The diamond is believed to have been found at the first part of seventeenth century in the Kollur mines near Golconda, India. Fersman (1955) confused this diamond with Daryainur, and other writers have suggested that it may be the same diamond as the Great Mogul diamond. However, the Daryainur resides in the Iranian crown jewels, and the Orlov is in the Russian diamond treasury (Bruton 1978).

The diamond was cut into three- and four-angular facets of beautiful Indian craft. The original owner of the diamond, Nadir Shah, ordered the stone recut in the form of a "high rose" to preserve its original shape. However, because the diamond lost some weight during polishing, the diamond cutter was not paid for his work and his belongings were confiscated to cover the damage to the stone.

The diamond may have been part of the 1739 Nadir Shah plunder that was taken during the pillage of Delhi (Gaal 1977). This diamond and a similar gem served as the eyes of the gigantic Brahma statue in the Srirangem temple in Trichinopoly in Southern India. Legend holds that a French soldier who had deserted during the Carnatic wars disguised himself as a Hindu convert and stole the Daryainur in 1747.

Later, the diamond was smuggled across the Caucasus Mountains by a shepherd who placed it in a wound that was made for this purpose on his leg. Eventually, the diamond was acquired by an Armenian salesman, Grigory Safras, who placed it in the Amsterdam Bank. In 1772, the diamond was sold to Safras's wife's nephew, a Russian court jeweler named Ivan Lazarev. Lazarev sold it to Grigory Orlov for 400,000 rubles ($450,000). Orlov presented the diamond to the empress Catherine the Great on her birthday on November 25, 1773. Since 1784, it has been located in the uppermost part of the scepter of the Russian Tsars, which is in the Kremlin Museum (Gaal 1977).

Punch Jones

The Punch Jones diamond was found in West Virginia in 1928. The diamond weighed 34.46 carats and was found in Peterstown, near the Virginia state line. It is the second largest verified diamond found in the United States (Hausel 1995). Even though the diamond was found in 1928, it was not identified until 1943 (Sinkankas 1959). The stone was a hexoctahedron with slightly greenish-gray color that exhibited several possible impact features that some researchers suggest might be the result of extensive transport (Holden 1944; Sinkankas 1959). The diamond was placed in the United States National Museum along with some other American diamonds.

Queen Elizabeth

This diamond is the Williamson diamond; it originally weighed 54.5 carats and subsequently was cut to a 23.6-carat gem. This diamond was found in 1948 in Mwadui mine, Tanzania, and was presented to then-Princess Elizabeth of England on the occasion of her marriage. Its namesakes are Queen Elizabeth and John Williamson, the discoverer of the Tanganyika diamond deposits.

The diamond was cut to a round brilliant of true rose color, and it is considered one of the greatest known diamonds of this color in the world. It is also one of the most valuable, at £500,000 sterling.

Red Cross

The canary-yellow Red Cross diamond was found in South Africa. In the rough it had a weight of 375 carats. The diamond was fashioned into a 205-carat square cut, which produced a "Maltese cross effect" that became visible through the table of the stone after cutting. The diamond was later presented to the British Red Cross in 1918 by the London Diamond Syndicate (Gaal 1977).

Regent

The Regent diamond was originally named the Pitt. It was found in 1701 in the Parteal mines on the Krishna River in India, near Golconda. According to folklore, the diamond, which weighed 410 carats, was stolen by a slave who smuggled it out of the mine by placing it in bandages covering a self-inflicted wound. After stealing the diamond, the slave conspired with a captain of a sea vessel to smuggle the diamond out of the country. The Captain, however, kept the diamond and drowned the slave.

The diamond was sold to the English governor, Thomas Pitt, for approximately \$100,000. Pitt arranged for the diamond to be cut in England. After 2 years, the diamond was cut into a 140.5-carat cushion-shaped brilliant that is considered one of the more perfect diamonds in the world. Fragments generated from the cutting were sold for 144,000 marks.

Pitt sold the processed gem for 2.5 million francs (\$500,000) to the French Regent Phillip of Orleans, and the diamond was renamed the Regent. It was set in the crown of the French king, Louis XV, but was later stolen in 1792 along with other crown treasures of France, and was found several months later. Later, at a time when money was needed to help finance the French army, the French Republic pawned the diamond to a Berlin jeweler. Napoleon I of France redeemed the diamond from the jeweler and placed it in the handle of his sword.

In the mid-nineteenth century, the French government sold part of its crown jewels at auction. At this time, the Regent was valued at 6 million francs. During the German occupation of France during World War II, the diamond was hidden in a marble panel of the Chambord Castle (Gaal 1977).

Russian Diamonds

Many large diamonds have been recovered from kimberlites in the Yakutia region of Russia. Many of these bear names related to the ideology of the Soviet regime, such as Twenty-Sixth Congress of Communist Party of the Soviet Union, which weighed 342.57 carats; Star of Yakutiya (232.1 carats) (the primary name for this stone was Fifty Years of Aeroflot in reference to the Soviet airline that was generally acknowledged outside Russia as the worst airline the world); Revolutionary Ivan Babushkin (171.15 carats); Volgogradsky (162.0 carats); Great Initiative (135.12 carats); Allende, named in reference to the socialist President of Chile (125 carats); Sixty Years of Great October (121 carats); and Ursa Major (114.3 carats).

Some other large diamonds recovered during the Soviet regime include the Diamond of the Twenty-Fourth Congress of Communist party of the Soviet Union, Fifty Years of the USSR, Unita (named after the newspaper of the Italian Communists), and Pravda (named after the Central Soviet Communist newspaper). Two other large diamonds were named after the Russian astronauts Yuri Gagarin and Valentina Tereshkova, and one large diamond was even named after Samantha Smith, an American girl who wrote a letter to Nikita Khruschev.

Sancy

The Sancy diamond is a very clear and transparent 53.5-carat diamond of Indian origin that was cut into a pear shape. The diamond belonged to Duke of Burgundy, Charles the Bold. According to legend, he placed it in his helmet because he believed that the diamond would protect him from injury or death. However, he was killed in a battle near Nancy.

The diamond was found by a soldier who sold the stone to a local priest for 1 gulden. The priest then sold it for a great profit at the price of three guldens. In the middle of sixteenth century, the diamond belonged to the King of Portugal, Anton, who was in great need of money and sold it to a French salesman for 100,000 francs. The salesman resold it to French Baron of Sancy, the namesake of the diamond.

In 1589, King Henry III of France asked Sancy to send him the stone to help finance an army. The diamond was sent to the king with one servant, who was attacked by robbers and murdered on the road. Because the diamond didn't appear at the market, the Baron ordered the grave of his servant to be exhumed, and the diamond was found in his remains: the faithful servant swallowed the stone before his death to protect it from being stolen.

Sancy later sold the diamond to James I, and the diamond remained in England until 1669. It was later sold by James II to King Louis XIV of France. During the French Revolution, the stone was stolen from the Royal treasury. It reappeared in 1828. In 1830, it was in the possession of the Duchess of Berry. The famous Russian industrialist P. Demidov purchased the stone for 500,000 francs and presented it to his bride, Aurora Shernval (Milashev 1989).

Shah

The Shah diamond is crystallographically represented by a great octahedron that is strongly elongated along one edge. The diamond, thought to have originated from the Golconda mines in India, is a white water color with a yellowish-brown shade. The polished stone weighs 88.7 carats. The elongation of the stone is thought to be related to deformation in a high-stress environment. A similar morphology has been recognized in diamonds recovered from basalts in Kamchatka.

During the process of polishing the diamond, part of its natural facets were preserved. On three of the polished surfaces, inscriptions were engraved in Persian. The inscriptions trace the history of this stone. The first inscription, "Burham-Nizam-Shah II, 1000," refers to the ruler of the Indian platform of Ahmendagar. The date is that of the Moslem calendar and corresponds to 1591. The second inscription, "The son of Jehangar Shah—Jehan Shah, 1051," refers to the grandson of Akbar, one of the Great Moguls, who had the Taj Mahal constructed. The date corresponds to 1641. The third inscription reads "Great Sovereign Kadjar-Fatkh-ali-Shah, 1242," a reference to the Persian Shah of the Kadjar dynasty, 1824.

In 1829, the diamond was presented to the Russian Tsar, Nicholas I, to appease the Tsar for the assassination of the Russian ambassador Alexander Griboyedov, a famous Russian poet killed by a mob of fanatics in Tehran. The diamond remains in the collection of the prized Russian Treasury of Diamonds and Precious Stones.

Star of Sierra Leone

The third largest diamond ever found was recovered in 1972 from the Diminco alluvial mine in Sierra Leone. The diamond weighed 969.8 carats. The stone was purchased by Harry Winston and was cut into several diamonds. The largest was a 143.2-carat emerald cut that was found to be flawed and was later recut into a flawless 32.52-carat emerald-cut gemstone. Seventeen diamonds with a total weight of 238.48 carats were cut from the diamond (Gaal 1977).

Terresa

The Terresa diamond was found at Kohlsville in 1883, in southeastern Wisconsin. A few kilometers to the east, another relatively large diamond known as the Saukville diamond was discovered.

The Terresa diamond weighed 21.25 carats and was found near the Green Lake moraine. It was nearly spherical with a flaw or cleavage that separated the diamond into a colorless crystal on one side and a cream-yellow crystal on the other side (Cannon and Mudrey 1981). The diamond was cut in 1918 and produced 9.27 carats of finished stones. The largest weighed only 1.48 carats (Sinkankas 1959).

Unnamed 616

In 1974, a 616-carat, poor-quality, yellow octahedron was discovered at the Dutoitspan mine in South Africa. At the time of the discovery, it was the ninth largest diamond found in the world. The diamond was placed on display in the DeBeers Hall of the Open Mine Museum in Kimberley (Gaal 1977).

Vargas

Weighing 726.6 carats in the rough, the President Vargas diamond was discovered in 1938 in the San Antonio River, Minas Gerais, Brazil. Harry Winston purchased the stone in 1939 for approximately $600,000. The stone was cut into 29 gems; the largest was a 48.26-carat emerald cut (Gaal 1977).

Woyie River

The Woyie River diamond was the second-largest alluvial diamond found and weighed 770 carats in the rough. The diamond was recovered from gravels in the Woyie River in Sierra Leone. The diamond was colorless with very high clarity. Thirty gems were cut from the stone in 1953, the largest of which is 31.35 carats (Gaal 1977). It has also been called the Victory diamond.

SYNTHETIC DIAMONDS

It is thought that synthetic diamonds were simultaneously produced in 1893 by a French physicist, A. Moiassan, and a Russian scientist, K.D. Khruschev, using different laboratory methods. In these experiments, Moiassan used an iron melt saturated with carbon (graphite) under a temperature of 2,000° to 3,000°C. The crust produced at the surface of the ingot enabled a high-pressure atmosphere to develop inside the ingot. After dissolving the ingot by acids, some small (0.7 mm) crystals were obtained from the residue, which were burned in oxygen and formed carbon dioxide (Milashev 1989). K.D. Khruschev, a mineralogy professor of the Russian Military-Medical Academy, crystallized from a silver melt grains of transparent dark matter that scratched the surface of corundum and also formed carbon dioxide when it was burned (Shafranovsky 1964).

Various scientists repeated both experiments with identical results, and it was decided that the problem of obtaining synthetic diamonds was solved. However, doubts arose during the early twentieth century when it was postulated that the crystals produced in the experiments were not diamonds, but rather a combination of carbon with metal (so-called carbides).

Later attempts to synthesize diamonds used decomposing carbon and condensation of carbon vapor. All these experiments failed owing to a lack of control of the high pressure and temperature, which is necessary to stabilize diamond. Instead, the lower temperatures and pressures used in the experiments were within the graphite stability field, rather than the diamond stability field.

In 1955, four physicists from the General Electric Company (Francis Bundy, Tracy Hall, Herbert Strong, and Robert Wentorf) obtained synthetic diamonds in autoclaves, which sustained pressures up to 10,000 MPa at a temperature of 2,700°C. Under a pressure of 5,300 MPa within the temperature interval of 1,300° to 2,200°C, sustained for 16 hr, small octahedral diamond crystals up to 1.2 mm were formed. Similar results were later obtained in Sweden, Holland, and Japan. In 1961, the president of Russian Academy of Science, Mstislav Keldish, reported to the Twenty-Second Party Congress that the Soviets had also created synthetic diamonds.

In 1990, the United States produced 100,000,000 carats of synthetic diamonds—one fourth of world production and twice as much as all of Europe, the Soviet Union, Asia combined (Austin 1990). The price of synthetic diamonds is expected to continue to decrease, or at least remain constant, owing to production increases.

The main application of synthetic diamonds is for drill bits for the gas, oil, and mineral resources industries. They also have many other uses in medical instruments, diamond tools for industry, high-precision instruments that require high-quality crystals, high-precision optics, diamond heat-deflectors, high-temperature materials, and radiation-resistant diamond electronics. Currently, the synthetic-diamond industry is in the process of developing efficient technology to produce gem-quality diamonds.

In 2001, the principal producers of synthetic diamonds were DeBeers, Sumitomo Electric Company, and General Electric. DeBeers and the Sumitomo Electric Company produce a comparatively large assortment of crystals, and General Electric plans to produce a similar variety of synthetic diamonds in the near future.

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SECTION 2

Conventional Host Rocks

CHAPTER 5

Overview

Most modern diamond exploration programs are designed to search for conventional diamondiferous host rocks (i.e., kimberlite and lamproite) or for placer deposits presumably derived from these rocks. Thus, it is important to become familiar with the characteristics of these potential host rocks. It is also important to become familiar with "unconventional" host rocks—because some of those deposits are enriched in diamond and could become important economic deposits in the future. However, in this portion of the book, we will concentrate on the characteristics of lamproite and kimberlite. Unconventional source rocks are described in Section 3.

Developing an all-encompassing physical, mineralogical, and geochemical description of diamondiferous host rocks is difficult because of the variety of potential hosts. In addition, many potential host rocks are hybrids that show dramatic variations from locality to locality, adding to the difficulty of description. Mitchell and Bergman (1991) report that the total volume of known kimberlite in the world is on the order of >5,000 km^3; for lamproite, it is only <100 km^3. Kimberlite is already considered to be one of the rarest rock types in the world, and thus lamproite is considerably rarer.

The difficulty in providing characteristic descriptions of diamondiferous host rocks has resulted in confusion in the literature. As an example, diamondiferous deposits in Murfreesboro, Arkansas, were initially classified as peridotite. They were later classified as kimberlite, because the thinking at that time was that all diamondiferous host rocks were kimberlites. However, detailed petrographic studies even later showed that these rocks were instead olivine lamproite (Mitchell and Bergman 1991). This example is not just an isolated case. Several diamond pipes, or rocks containing mantle-derived xenoliths and xenocrysts, that were initially described as kimberlite have been reclassified following more detailed research.

Prior to the discovery of the diamondiferous kimberlite in South Africa, diamonds were mined at Mhajawahn, India, as early as 1827. Because the host rock contained diamond, it was later assumed to be kimberlite and was classified as such in the 1900s. However, petrographic studies have now provided evidence that this rock is olivine lamproite. Much of the problem with the classification of such rocks is not merely the hybrid nature of some diamondiferous host rocks, but that many researchers originally considered diamond to be a primary accessory mineral restricted to kimberlite. But in recent years, most diamonds have been shown to be xenocrysts that were accidentally trapped in kimberlite and other rare magmas that originated at depth within the upper mantle.

Kimberlites and lamproites tend to occur in clusters. They may be found in small clusters of more than one, or in large clusters of more than 100, or any size in between. They are generally found in relatively small districts covering a few to several square kilometers of area. Structural control is often thought to be an important factor in the emplacement of the intrusives; however, several structural orientations are often recognized in each district.

Diamond is formed under very high pressure and temperature under extreme conditions within the deep keels of cratons. However, diamonds can also grow at shallower depth under tectonic pressure, such as along a subduction zone or deep shear zones, or they may grow metastably in unique geologic environments.

On the basis of unique mineral phases in the kimberlite magmas, the minimum depth of origin for diamondiferous kimberlites in cratonic environments is estimated at 150 km. The maximum depth is thought to be 300 km because olivine inverts to beta-olivine at 320 km depth and, so far, beta-olivine has not been observed in kimberlite.

Equilibrium subcontinental geotherms require 45 to 55 kbar and 1,050° to 1,200°C to stabilize diamond. These conditions would be equivalent to depths of 150 to 200 km. However, it has been suggested that diamond may form in a cold subducting slab at depths of only 80 to 90 km at 22 to 25 kbar and 200° to 400°C. It has also been suggested that some alkaline rocks, which originate at depths >100 km, could potentially entrain and preserve diamond in xenoliths or xenocrysts if ascent is rapid. However, most alkaline rocks appear to originate from too shallow a depth to trap diamond.

Cratonic environments provide favorable terranes for emplacement of diamondiferous host rocks, in particular kimberlite, lamproite, and lamprophyre. The temperature of the mantle beneath the cratons is so low that melting is unlikely except at considerable depths.

Cratonic terranes initially possessed steep geothermal gradients, as is evident by the abundance of Precambrian komatiitic and tholeiitic metavolcanics in them. These gradients decreased through geologic time to the point that partial fusion of the peridotitic mantle may have occurred at depths of around 200 km producing kimberlite melts (Hornlocker 1978).

According to Hornlocker (1978), a seismic low-velocity zone coincides with this zone of melting in cratonic environments. Many mobile belts and oceanic areas have low-velocity zones that occur at much shallower depths, which is unfavorable for the generation of kimberlites magmas but instead favors the production of calc-alkaline and tholeitic magmas.

Seismic data show a shallow low-velocity zone (30 to 50 km) beneath the Mesozoic-Cenozoic mobile belts in the off-craton, western United States. In this area, Pn (compressional seismic wave) is generally less than 8 km/s and reflects partial melting at shallow depths, which is also supported by high heat flow. Thus, these off-craton environments are not considered to be favorable terranes for diamondiferous kimberlite or lamproite.

Hornlocker (1978) suggested that the 8 km/s contour represented the outer edge of the stabilized craton of the United States. The transition from shallow to deep low-velocity zones (i.e., zones of melting) is continuous albeit abrupt.

The time–space distribution of kimberlitic magmatisim roughly follows the development of the continental crust (Goldich et al. 1966). Kimberlites appear to postdate crustal development by at least 0.5 billion years.

The upper mantle at depth in the region of kimberlitic, lamproitic, lamprophyric magma genesis appears to be inhomogeneous but dominantly peridotitic with zones of eclogite. This concept is based on the presence of peridotite and eclogite nodules in many kimberlites and less commonly in lamproites and some in lamprophyres. Some of these nodules are endowed with diamonds. As such, some of these diamondiferous nodules are considered to be the primary host rock for diamond.

Nevertheless, a major problem that continues to persist from both a theoretical and a pure exploration point of view is why diamondiferous nodules and xenoliths are constrained to kimberlite. This problem remains outstanding. But the fact is that kimberlites and, in more recent years, olivine lamproites are considered to be the primary exploration targets for diamond.

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CHAPTER 6

Kimberlites

INTRODUCTION

A source rock for terrestrial diamonds was discovered after alluvial diamonds mined from gravel on the Orange and Vaal Rivers of South Africa led to the discovery of "dry diggings" at Jagersfontein and Dutoitspan in 1870. The dry diggings were discovered in "yellow ground" and "blue ground," deeply weathered, oxidized kimberlite that consisted of carbonated montmorillonite with commonly interspersed rounded boulders. The significance of the dry diggings was not recognized at first, and many miners had concluded that the dry diggings were dry river beds (paleoplacers) formed when the Vaal River had overflowed its banks at an earlier time.

As the diamond mines reached greater depths, the operations entered into "hardebank," or fresh kimberlite. It was later determined that the rounded boulders were partially assimilated country rock and crustal and mantle xenoliths, along with cognate mantle nodules. The rounding occurred during the ascent of the magma to the earth's surface.

Cohen (1872) recognized that the cylindrically shaped areas composed of yellow ground and blue ground were volcanic pipes. H.C. Lewis introduced the name *kimberlite* in 1887, after the type locality at Kimberley, South Africa, and the rock was defined as a porphyritic mica-bearing peridotite. But the complex problems related to the mineralogy, chemical composition, and texture resulted in heated discussions on the definition of the term *kimberlite,* which persist to the present.

In 1914, Wagner divided kimberlites into "basaltic" (olivine-rich rocks with <5% phenocrystal mica) and "lamprophyric" (olivine-rich rocks containing mica phenocrysts in a groundmass with >5% mica). This classification has been widely recognized, although the term *basaltic* is now considered petrographically misleading. Feldspar (an important component in basalt) is absent in kimberlites, and kimberlites are not mineralogically or genetically related to basalts.

According to various researchers, kimberlite is a volatile-rich, potassic, ultrabasic igneous rock that is enriched in some incompatible constituents (Sr, Zr, Hf, Nb, and rare-earth elements) and some compatible lithophile constituents (Ni, Cr, and Co), along with alkalis. The mineralogy of kimberlites is variable and complex. They occur as dikes, sills, blows (small, lense-shaped dike enlargements in the root zone of diatremes), and pipes (diatremes). The pipes range from several hundred meters across to the Camafuca-Camazambo pipe in Angola that is reported to cover an area of 3.06×0.2 km. Kimberlite lavas include both tuffisitic and epiclastic facies. Kimberlites, which are essentially carbonated alkali peridotites, exsolve CO_2 rapidly during ascent and often produce diatremes with considerable brecciation and dissolution-rounded xenoliths and cognate nodules.

Clement and Skinner (1979) and Clement, Skinner, and Scott-Smith (1984) define kimberlite as

> *a volatile-rich potassic ultrabasic igneous rock which occurs as small pipes, dikes and sills. It has a distinctive 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 following minerals: phlogopite, carbonate (commonly calcite), serpentine, clinopyroxene (commonly diopside), monticellite, apatite, spinels, perovskite, and ilmenite. The macrocrysts are anhedral mantle-derived ferromagnesian minerals which include olivine, phlogopite, picroilmenite, chromian spinel, magnesian garnet, clinopyroxene (commonly chromian diopside), and orthopyroxene (commonly enstatite). Olivine is extremely abundant relative to other macrocrysts, all of which are not necessarily present. The macrocrysts and relatively early-formed matrix minerals are commonly altered by deuteric processes, mainly serpentinization and carbonatization. Kimberlite commonly contains inclusions of upper mantle-derived ultramafic rocks. Variable quantities of crustal xenoliths and xenocrysts may also be present. Kimberlites may contain diamonds but only as a very rare constituent.*

Clement, Skinner, and Scott-Smith (1984) defined kimberlite on the basis of its mineralogy. However, some minerals in kimberlite are unrelated to kimberlite. This paradox is eliminated if many of the macrocrystal phases are considered cognate. Mitchell (1970) worked out a precise mineralogic definition of kimberlite that excluded from the definition the mode of intrusion and the presence of xenoliths and instead emphasized petrologic characteristics. He also suggested that kimberlite originated from an alkalic peridotitic magma at depth.

Mitchell (1986) provided the following definition for kimberlite:

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

TYPES AND PETROGRAPHIC FEATURES

Types

Wagner (1914) divided kimberlites into basaltic and lamprophyric varieties. This terminology was introduced to describe the macroscopic appearance of the rocks and had no genetic significance. Wagner's lamprophyric kimberlite was further subdivided according to the presence of clinopyroxene.

However, Mitchell (1970) proposed that three mineralogic varieties of kimberlite are recognizable on the basis of the amounts of olivine, phlogopite, and calcite present. They include kimberlite (equivalent to Wagner's basaltic kimberlite), micaceous kimberlite (equivalent to Wagner's lamprophyric kimberlite), and calcite or calcareous kimberlite.

The latter category was introduced in recognition of primary magmatic calcite in kimberlite, although many authors consider calcite to be a product of autometamorphism. The terms *basaltic kimberlite* and *lamprophyric kimberlite* are now considered archaic and have been replaced by the terms *Group I kimberlite* and *Group II mica kimberlite*.

Group I and Group II kimberlites can be distinguished by their isotopic compositions ($^{87}Sr/^{86}Sr$ vs $^{143}Nd/^{144}Nd$) (Smith 1983). Group I kimberlites are derived from a relatively primitive mantle source, probably the asthenosphere, and the Group II kimberlites are thought to originate from metasomatically enriched parts of the lithosphere.

Group I kimberlite. Group I kimberlites are found worldwide. They range from Lower Proterozoic to Eocene and include olivine-rich monticellite ($CaMgSiO_4$)-serpentine-calcite kimberlites containing

- upper mantle xenoliths (i.e., peridotites and eclogites)
- megacrystal minerals or a discrete nodule suite derived from disaggregation of mantle xenoliths. They include rounded, anhedral grains of Mg-ilmenite (picroilmenite), Cr-Ti-pyrope, olivine, Cr-poor clinopyroxene, phlogopite, enstatite, and Ti-poor chromite. (Many are part of the kimberlitic indicator-mineral suite.)
- olivine-free nodules that contain mica (phlogopite), amphibole (potassium richterite), rutile, ilmenite and diopside (abbreviated to the “MARID” suite of inclusions); primary phenocryst and groundmass minerals olivine, phlogopite, perovskite, spinel, monticellite, apatite, and calcite; and primary serpentine.

Group II kimberlite. Group II kimberlites (micaceous kimberlites) consist of rounded olivine macrocrysts in a matrix of phlogopite and diopside with spinel, perovskite, and calcite. In contrast to Group I kimberlites, they lack magnesian ulvospinel and monticellite; spinels and perovskite are relatively rare. Group II kimberlites are currently known only from southern Africa, where they were initially recognized near the Orange River. As a result, Mitchell (1995) suggested that the Group II kimberlites be renamed “Orangeites.”

Petrographic Features

Kimberlites are hybrid rocks that exhibit several distinct elements—variations that arise from the differentiation process and include mantle-derived xenoliths and xenocrysts, a megacrystal suite of mineral inclusions, and a MARID suite of inclusions. Probably the most informative from the point of view of deciphering of petrologic processes at the early stages of magma evolution is a suite of mineral inclusions in the diamond itself.

Mineralogically, kimberlite, which is a porphyritic (or porphyroclastic) volcanic rock, consists of olivine phenocrysts (two types of olivine are present, i.e., olivine phenocrysts [olivine I] and olivine megacrysts [olivine II]), pyroxene (usually diopside), phlogopite, and perovskite. Olivine is extremely abundant relative to other macrocrysts. The olivine macrocrysts are set in a groundmass of anhedral, mantle-derived, ferromagnesian minerals including olivine, phlogopite, picroilmenite, chromium spinel, magnesian garnet, clinopyroxene (commonly chromium diopside), and orthopyroxene (commonly enstatite). The macrocrysts and relatively early formed matrix minerals are commonly altered by deuteric processes such as serpentinization and carbonatization.

Many of the problems related to the kimberlite groundmass are rooted in ideas applied to the definition of kimberlite that originated in the 1930s. For instance, Shand (1934) and Holmes (1936) concluded that kimberlites were merely altered melilite basalts or equivalents of melilite basalts with "emanations" (such as H_2O, CO_2) and fragmental lherzolites, a variety of peridotite. The calcite in kimberlite was considered to be a secondary mineral derived from the breakdown of melilite. This concept of secondary calcite (after melilite) was widely accepted by pertrologists of the Soviet school (Milashev 1963; Frantsesson 1968).

However, many Western researchers consider much of the calcite to be primary. For instance, Mitchell (1986) has clearly shown that carbonate-rich kimberlitic rocks (i.e., the carbonate-rich residue poor in silicates, oxides, and the megacryst/macrocryst suite found at the Premier mine, as well as some carbonate-rich ocellar dikes) are characterized by a specific mineralogy and geochemistry and should be considered calcite kimberlites. In addition, several kimberlites mapped in the Iron Mountain district of the Wyoming craton are carbonate-rich breccias that are intricately associated with classic kimberlite porphyries, that possibly are calcareous kimberlites or carbonatites.

Mineralogically these calcareous rocks can be defined as carbonatites, although their mineralogy differs from that of true carbonatites. They have a distinctive inequigranular texture resulting from macrocrysts set in a fine-grained matrix. The mineralogy consists of primary phenocrystal and/or groundmass constituents including olivine and several other minerals such as phlogopite, carbonate (commonly calcite), serpentine, clinopyroxene (commonly diopside), monticellite, apatite, spinel, perovskite and ilmenite.

The acceptance of magmatic (primary) calcite in kimberlite is based on the presence of veins filled solely with calcite that are often formed during the final stages of kimberlite emplacement. These veins led Mitchell to recognize specific types of carbonatites on the basis of the petrographic classification of Heinrich (1966). These carbonatites, however, have very different mineralogy than the antecedent and comagmatic carbonatites of alkaline rock complexes, and they are better termed *calcite-kimberlites* to reflect petrogenetic differences (Mitchell 1979).

Mitchell (1986) introduced the concept of a kimberlitic clan of rocks that reflects differentiation of the kimberlite magma. The three members of this clan are as follows:

- kimberlites as defined above, with a characteristic megacryst/macrocryst suite
- kimberlites as described above, but poor in or lacking the heavy mineral and megacryst/macrocryst assemblages. These minerals are thought to have been lost during differentiation.
- carbonate-rich residua, poor in silicates and oxides and minerals of a megacryst/macrocryst suite.

ASSOCIATED XENOLITHS AND MANTLE NODULES

The first kimberlites discovered in South Africa were termed *dry diggings*. They commonly contained rounded boulders and cobbles scattered on the surface and appeared similar to paleoplacers, or dry streambeds. Instead of being water-worn and transported in streams, these rounded boulders and cobbles were actually xenoliths and cognate nodules that had been trapped in kimberlite magma and transported to the earth's surface. Others were xenoliths downdropped into the kimberlite magmas from strata at a higher stratigraphic position. The rounding is thought to be due to chemical dissolution (Kirkley, Gurney, and Levinson 1991).

Mantle-derived breccia pipes in southwestern Wyoming were discovered because of a similar morphologic anomaly (Hausel et al. 1999). The hidden breccia pipes were covered with common rounded boulders and cobbles of quartzite restricted to small, elliptical areas. The country rock in the surrounding terrain was primarily Tertiary siltstone and shale that contained no quartzite. Trenching the apparent paleoplacer uncovered boulder-filled breccia pipes several centimeters beneath the soil (Richard Kucera, personal communication, 1995). Even though these mantle-derived rocks look similar to rocks in the initial South African discoveries, the host rock for the xenoliths does not appear to be kimberlite and has a distinct mafic signature (Hausel et al. 1999). The pipes also host considerable kimberlitic indicator minerals and some very small eclogite and metabasite xenoliths (Kuehner and Irving 1999). Thus, any mantle-derived rock can host similar rounded xenoliths and cognate nodules.

Some xenoliths and nodules found in kimberlite, lamproite, and lamprophyre, as well as in some basalts, represent material that was captured at the point of incipient melting and thus represent the birthplace of the magma. Other xenoliths were captured at shallower depth as the magma rose to the surface. These xenoliths provide a record of the magma's journey to the earth's surface.

Some of the cognate nodules found in the magmas have been interpreted as the primary source rock of diamonds. When these rocks break up and disaggregate, their diamond load is usually disseminated within the magma, whether it is kimberlitic, lamproitic, or lamprophyric.

The diamonds are a primary mineral of some of the peridotite and eclogite nodules rather than the kimberlite, lamproite, or lamprophyre. In most cases, the diamonds are by-products of the disaggregation of the nodules. Thus most diamonds are secondary, or accidental (xenocrysts), in these magmas.

Kimberlite may host a wide variety of xenoliths and nodules. Crustal xenoliths vary in size from 0.04 to 0.5 cm to 50 to 300 m across and usually have an angular shape, unlike the typically rounded cognate mantle nodules. The largest xenoliths (weighing several tonnes) found in some kimberlites have been termed "floating reefs." The rocks that are most resistant to disruption typically produce the largest xenoliths. For example, the largest xenoliths found in the Premier mine of South Africa are composed of competent quartzite; at the Voorspoed mine, the largest xenoliths are basalts.

Megaxenoliths show all gradations from massive unbrecciated types with no kimberlitic matrix to varieties veined and disrupted by kimberlite (Clement 1982). Generally, xenoliths derived from a variety of crustal levels are juxtaposed with mantle-derived nodules and xenoliths within the kimberlite or lamproite intrusive. However, there is a very rough zonation in the distribution of different sizes and compositions of xenoliths within kimberlite pipes. It is important to realize that the xenoliths moved in both directions in the pipes: some xenoliths were transported upward from depth, and a considerable number of xenoliths sank.

In the Kimberley mine (South Africa), sunken fragments of shale that originated near the present surface were found on the 762 m level. In the Siberian kimberlite fields, lower Paleozoic and Mesozoic xenoliths derived from rocks formerly present in the region are now found sunken in the pipes, and lie adjacent to in situ lower Paleozoic, Upper Proterozoic and Archean rock sequences. In a few kimberlite pipes in Colorado, Montana, and Wyoming, Ordovician, Silurian, and Cretaceous to Eocene sedimentary xenoliths are found in pipes emplaced in Precambrian crystalline rocks (Hearn 1968; McCallum, Eggler, and Burns 1975). Some of the Colorado–Wyoming kimberlites were initially discovered because abundant Ordovician and Silurian sedimentary xenoliths were exposed in the kimberlite at the surface, in a region where such sedimentary rocks were not known to exist. The distance through which some of the fragments descended may be measured in hundreds of meters—in some cases, more than 1 km.

The largest of the detached and sunken xenoliths (such as those at the Premier mine) are mainly but not entirely peripherally located in the upper portions of pipes (DuToit 1906; Williams 1932; Hawthorne 1975). Smaller blocks persist to considerable depths in some diatremes, e.g., Finsch (Clement 1982) and Kimberley (Wagner 1914), but they are notably absent from the root zones of hypabyssal facies kimberlites, although a relatively large granitic block was discovered in the root zone of the Sloan 2 kimberlite, Colorado. Locally derived xenoliths decrease with depth in the intrusives and tend to concentrate at the diatreme margins, whereas lower crustal and mantle-derived xenoliths tend to lie in the central part of kimberlite pipes (Novikov and Slobodskoy 1979; Dawson 1962). However, locally derived xenoliths in the Ferris 1 and Aultman 1 kimberlites in Wyoming were found in the diatreme margins as well as the central portion of the exposed pipes.

Essentially no trace of alteration can be found in most xenoliths along the exocontacts. Bituminous shales and carbonized wood fragments extracted from kimberlite show little evidence of alteration and indicate that temperatures of emplacement were relatively cool during the emplacement of the diatreme (Wagner 1914; Williams 1932; Novikov and Slobodskoy 1979). However, some limestone xenoliths in the Yakutian kimberlites developed rims of andradite-magnetite-pyroxene skarn. Such differences indicate that a wide range of temperatures existed at different levels in kimberlites during their emplacement.

Lower-crustal nodules that have been identified in kimberlite and lamproite included hypersthene granulites, charnockitic granulites, augite granulites, garnet kyanite granulites, pyroxenite, altered basalt, and intensely carbonated and fenitized rocks of unknown genesis. These nodules are well rounded and commonly rimmed by thin rings of finely crystalline serpentine-carbonate mixtures. Most of the nodules show some degree of alteration (usually intense to pervasive), and they may include serpentinization, carbonatization, silicification, and/or fenitization. From the mineral chemistry of some deep-seated nodules, the pressure and temperature of formation can be estimated.

The following elements and minerals have been used in geobarometry estimates: aluminum substitution in orthopyroxene (enstatite) coexisting with garnet, potassium substitution in clinopyroxene, and sodium substitution in garnet. In all of these cases, elevated amounts of Al, K, and Na indicate high pressure. Geothermometry is typically determined on the basis of partitioning (relative proportions) of Ca and Mg of coexisting orthopyroxene (enstatite) and clinopyroxene (diopside), relative abundance of Fe and Mg in these same pyroxenes, and the relationship of Fe and Mg in coexisting garnet and orthopyroxene (Kirkley, Gurney, and Levinson 1991).

Geothermometry, geobarometry, and seismic data may be used to construct a model of crustal thickness. For example, in the State Line district, Colorado–Wyoming, a crustal thickness was taken to be about 50 km on the basis of seismic evidence. Upper crustal rocks cropping out at the surface include Precambrian granites and gneisses, and they are also found as xenoliths in the kimberlite. On the basis of the xenoliths in the kimberlite, the lower crust consists of granulites, pyroxenites, and basaltic units. During the time of one of the major episodes of kimberlite emplacement in the State Line district (early Devonian), the upper mantle consisted of spinel and garnet peridotite. It is suggested that this garnet peridotite, through episodes of partial melting, had been depleted of less refractory elements and enriched in the more refractory elements such as magnesium and chromium (McCallum and Mabarak 1976).

A characteristic feature of kimberlite is the presence of mantle-derived xenoliths. Most xenoliths are rounded, especially those from depth, by chemical dissolution at the margins of the rock fragments. It is important to recognize that the xenoliths in kimberlite and lamproite do not always represent with fidelity the mineralogy and pressure–temperature (P–T) conditions of the mantle in which they formed, because extensive modification may take place as they react chemically with fluids in the transporting magmas.

Deep-seated, mantle-derived nodules that are found in some kimberlites are often grouped into broad categories that include the peridotite-pyroxenite suite, the eclogite suite, along with megacrysts, metasomatized peridotites, glimmerites (ultrabasic rocks formed almost entirely of phlogopite and/or biotite mica), and the MARID suite of rocks.

More specifically, nodules found in kimberlite may include garnet lherzolite and spinel lherzolite, garnet clinopyroxenite, garnet websterite, spinel websterite, harzburgite, dunite, carbonatite (sövite), eclogite, discrete mineral nodules and megacrysts, glimmerite, and MARID nodules.

Diamonds have been found in a variety of xenoliths, the most important of which are garnet-bearing peridotite and eclogite. Some of these nodules are very rich in diamond. For example, a hand specimen of eclogite collected from the Sloan 1 diatreme in the Colorado–Wyoming State Line district yielded an estimated ore grade of 21,000 carats per tonne.

The megacryst assemblage (Cr-rich and Cr-poor orthopyroxene, clinopyroxene, garnet, and associated ilmenite) indicates depths of 165 to 200 km, and the Cr-deficient crystals came from the greatest depths (McCallum and Mabarak 1976). Boyd and Nixon (1973) suggested that the Cr-poor varieties originate in a relatively undepleted deeper mantle peridotite. Megacrysts from the deeper depths exhibit sheared textures.

To summarize, kimberlite and lamproite magmas began their journey at great depths within the earth's upper mantle and incorporated a variety of rock types from various depths. Some of these rocks may contain diamond. In particular, some may be enriched in diamonds. During transport in the magma, some of the eclogite and peridotite were assimilated and disaggregated. The breakdown of these diamondiferous host-rock xenoliths disseminated diamonds throughout the magma.

Peridotite-Pyroxenite Xenoliths

Ultramafic nodules of peridotite and pyroxenite are most abundant in mantle-derived xenoliths found in kimberlite. They usually compose less than 2% of the diatreme. The xenoliths are ovoid and average less than 30 cm in diameter, and they rarely are longer than about 1 m. Shearing is common in large specimens, suggesting that its absence in smaller specimens is an effect of scale. In garnet-bearing rocks there is a complete modal gradation from garnet dunite through garnet harzburgite and garnet lherzolite to garnet websterite, garnet pyroxenite, and garnet orthopyroxenite. Garnet lherzolites have been observed to grade into eclogitic garnet-pyroxenite segregations in xenoliths from the Obnazhennaya pipe, and dunite has been observed to grade into chromite harzburgites (Sobolev and Sobolev 1964).

Mineralogically, these peridotites consist of various combinations of Mg-olivine, Mg-orthopyroxene, green chrome-diopside, and red to purple Mg-garnet, Mg-spinel,

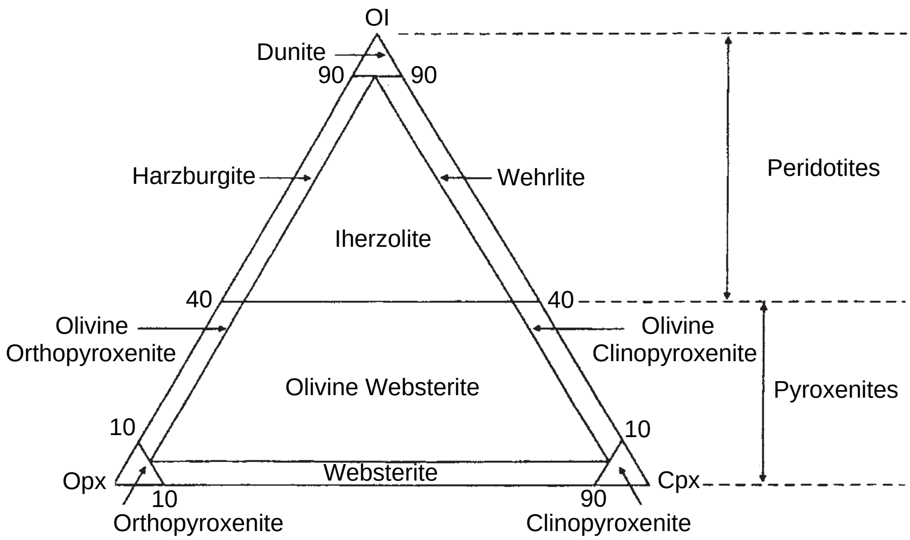

FIGURE 6.1 Peridotite-pyroxenite relations

phlogopite, and rarely pargasitic amphibole. Accessory minerals are diamond, graphite, moissanite, various sulfides (of pentlandite-chalcopyrite-pyrrhotite association), rutile, ilmenite, and zircon.

Peridotites (>40% olivine) are Cr-rich ultramafic rocks containing olivine and pyroxene with minor plagioclase and are subdivided on the basis of their pyroxenes. For example, lherzolites are peridotites with clinopyroxene and orthopyroxene. Wehrlites contain clinopyroxene but lack orthopyroxene. Harzburgites have orthopyroxene but no clinopyroxene (typically, contain 40% to 90% olivine, 5% to 60% orthopyroxene, <5% clinopyroxene) (Figure 6.1). Garnet lherzolites are the most common type of mantle-derived xenolith in most kimberlites. Of the peridotites, pyrope harzburgites are the most important xenoliths because many contain diamond.

Spinel peridotite is restricted to the uppermost mantle at 50 to 65 km depth, whereas the Mg- and Cr-enriched garnet peridotite extends from a depth of about 65 to 180 km. Rocks of the garnet websterite group (garnet clinopyroxenites, garnet websterites, garnet olivine websterite, and chemically related garnet lherzolite) occur as lenses or pockets in peridotite at 50 to 75 km deep. This depth is suggested by the presence of a few peridotite nodules in small lenses (up to 1 cm thick) of garnet pyroxenite aggregates that have a composition similar to websterite.

In the Colorado–Wyoming State Line district, depleted peridotite nodules collected from some kimberlites have been interpreted to originate at depths of less than 180 km. Many of these nodules exhibit pervasively sheared textures characteristic of deep undepleted peridotite. Unfortunately, they are intensely altered and the original mineralogy has been entirely replaced by serpentine, hematite, calcite, and quartz (McCallum and Mabarak 1976).

Several diamondiferous peridotite nodules have been recovered from kimberlite. The richest peridotite xenoliths have a diamond content that is a magnitude lower than that of the richest eclogites. The relative importance of peridotite nodules with respect to diamond content are pyrope harzburgite > chromite harzburgite > pyrope lherzolite

(Gurney 1984). Diamonds found in peridotite have mineral inclusions that are similar to their host rock peridotites. For instance, mineral inclusions found in peridotitic (P-type) diamonds include olivine, orthopyroxene, clinopyroxene, chromite, sulfides, and garnet.

Some peridotitic garnets have unique chemistries that are thought to indicate the diamond stability field. Because garnets are many times more abundant than diamond in kimberlite, they are often used in the exploration for diamond. Such peridotitic (pyrope) garnets are designated as G10. They have chemical affinities with subcalcic harzburgite and have relatively low Ca/Cr ratios compared with lherzolitic pyropes, and they are interpreted to have been derived from the diamond stability field (Gurney 1989).

Similar garnets found in kimberlite, designated as G9, are derived from the calcic lherzolites. They have lherzolitic affinity, and the majority of these are derived from the graphite stability field (Gurney 1984).

On the basis of the relative abundance of peridotite in kimberlite, peridotite is thought to be the most common rock type in the upper mantle in the keels of cratons. But what is surprising is that diamondiferous peridotite xenoliths are rare. Diamondiferous lherzolites are even rarer, and only a handful of samples have been found in kimberlite. Few have even been reported in alkali basalt.

Ore grades in the peridotite xenoliths are as much as three orders of magnitude higher than the host kimberlites, but they are still two to four orders of magnitude lower than in some eclogite xenoliths.

Eclogites and Griquaites

Bimineral, garnet-pyroxene xenoliths from South African kimberlites originally were described under the name *griquaite* in order to express the belief that they were cognate to the host kimberlites and different from metamorphic eclogite. Eclogitic xenoliths are usually ovoid or discoid in shape owing to their banded texture. Eclogites are predominant among xenoliths in some pipes (Roberts Victor, South Africa, and Zagadochnaya, Yakutiya). This coarse-grained rock is dominated by garnet and clinopyroxene.

Eclogite nodules are relatively common xenoliths in some kimberlites and show a range of accessory minerals, chemical compositions, and equilibration temperatures. They are interpreted to have originated from pockets within a peridotite mantle at depths of 75 to 130 km, or possibly from a separate layer at greater depth than the peridotite.

Eclogite forms a dense, coarse-grained mafic rock consisting of almandine-pyrope garnet (with trace sodium and titanium) and jadeitic clinopyroxene (grass green omphacite). Minor rutile, kyanite, corundum, and coesite may be present. Eclogites form rounded nodules in kimberlite and are interpreted as xenoliths unrelated to kimberlite. They are assumed to be either cumulates from garnet peridotite melts or subducted oceanic crust.

Some ecologitic garnets have unique chemistries that are used in diamond exploration. Mantle garnets of eclogitic parentage have been designated as either Group I or Group II. All diamondiferous eclogites belong to Group I. The Group I eclogitic pyropes have Na_2O contents of ≥0.07%, low levels of Cr_2O_3 (typically <0.05 wt. % Cr_2O_3), and CaO contents in the range of 3.5 to 20 wt. %. Group II eclogitic pyropes are similar to the Group I garnets but have <0.07% Na_2O. Low-Cr garnets with less than 3.5% CaO are probably derived from crustal rocks (Helmstaedt 1993).

A transition from bimineral eclogites to kyanite-bearing or corundum-bearing eclogites has been observed in some kimberlitic nodules. Garnets in kimberlitic eclogites are richer in pyrope molecules, and clinopyroxenes have a relatively high diopside/jadeite ratio. Fe-Mg partitioning between coexisting garnets and clinopyroxene in kimberlites equilibrated at comparatively high temperature (Banno 1970). Certain bimineral aggregates appear to have crystallized from the kimberlitic magma, which is why the term *griquaite* has been revived as a synonym for kimberlitic eclogites (Nixon, VonKnorring, and Bouller 1963; Nixon and Boyd 1973).

The existence of an eclogitic melt during diamond crystallization has been confirmed by identification of "melted inclusions" in diamond from an eclogitic xenolith from the Mir pipe (Bulanova, Novgorodov, and Pavlova 1988). Melted inclusions refer to inclusions of the original melt found within a diamond.

Some eclogites represent a primary host rock for diamond. This conclusion is supported not only by some very rich eclogites with upward of 0.7% diamond (some hand specimens of eclogite recovered from kimberlite have reported grades of 20,000 to 30,000 carats per/tonne), but also by the common occurrence of eclogitic (E-type) diamonds with eclogitic mineral inclusions, mainly sodic clinopyroxene and garnet with rutile, kyanite, corundum, and coesite. A genetic relationship between eclogite and the inclusion minerals is corroborated by elevated sodium and potassium contents respectively, in garnets and clinopyroxenes. The diamonds recovered from eclogites also provide carbon isotope signatures typical of organic carbon, supporting a probable subduction genesis (Gurney 1989; Kirkley, Gurney, and Levinson 1991).

Megacryst Suites

A characteristic feature of some kimberlites is the presence of a suite of megacrysts of single, large (5–20-cm) grains of olivine, Cr-poor pyrope, Ti-rich pyrope, Mg-ilmenite (picroilmenite), Cr-diopside, enstatite, and zircon. Many diamonds may also fit in this category. The megacrysts are primarily monomineralic and more rarely include polycrystalline aggregates. The large size of the megacrysts, which is considerably larger than the average grain size of mantle xenoliths, suggests that a possible origin in upper mantle xenoliths is unlikely, because those xenoliths have considerably smaller grain sizes (average size 2–4 mm).

Although all researchers recognize the existence of these minerals, different names have been applied. Nixon and Boyd (1973) refer to them as the "discrete nodule association." Dawson (1980) used the term *megacryst association.* Fragments of minerals that are members of the megacryst association are common in kimberlites, for which Mitchell (1986) applied the term *macrocrysts*.

The megacryst assemblage (Cr-rich and Cr-poor orthopyroxene, clinopyroxene, garnet, and ilmenite) indicate a depth of origin of 165 to 200 km, and the Cr-deficient grains were derived from the greatest depths (McCallum and Mabarak 1976). Boyd and Nixon (1973) suggested that the Cr-poor varieties originate in a relatively undepleted deeper mantle peridotite. The megacrysts from the greater depths also exhibit sheared textures. According to Boyd and Nixon (1973, 1975), the megacrysts formed throughout a wide range of temperature (900°–1,400°C) but at limited pressures equivalent to only 50 km depth. The upper depth limit of the megacrysts is believed to coincide with the lithosphere/asthenosphere boundary. It is thought that the crystal-mush magma generated at the megacryst site is interstitial to deformed lherzolites at a depth of around 50 km.

Megacrysts are typically rounded. They are usually surrounded by a thin crust or reaction rim produced by alteration (kelyphyitic rim surrounding pyropes) or by a rim of a micrograined kimberlite. Many of pyroxene megacrysts may also be enclosed by a calcium-carbonate rim or crust. The compositional variations and relationships suggest that all members of this association were derived from a common source. The amount of megacrysts varies widely in kimberlites from different pipes.

Points of view on the origin of this mineral association differ widely—xenocrystal origin or cognate inclusions—but neither hypothesis is totally accepted. Thus, Mitchell (1986) applies a genetically neutral term to the association, which he termed the *megacryst/macrocryst suite*.

The importance and uniqueness of this mineral association led some petrologists to include it as a major component of the definition of kimberlite (Koval'sky 1963; Milashev 1965; Mitchell 1970). Other petrologists rightly questioned including this suite in the definition of kimberlite, because the megacryst/macrocryst suite cannot be considered to be the result of crystallization of parental kimberlite magma.

A phenocrystic origin for the megacryst suite is supported by the following observations:

- the common presence of megacrysts that are compositionally distinct from those in kimberlite, and which are found in some other rock types, e.g., in minette (Ehrenberg 1982) and alnöite (Nixon and Boyd 1973)
- experimental confirmation by Eggler and Wendelandt (1979) that the megacryst suite could represent high-pressure liquids derived from a variety of magmas
- experimental data showing that megacrystal ilmenite and garnet could be derived from high-pressure liquidus phases in a variety of magmas (Green and Sobolev 1975; Mysen and Böetcher 1975)
- the presence of calcite and phlogopite inclusions trapped in kimberlite liquids (Rawlinson and Dawson 1979; Schulze 1981).

However, other characteristics of the megacryst suite suggest that it was originally in disequilibrium with their host kimberlite magma. As a result, it has been proposed that a series of hybrid rocks was formed consisting of several distinct suites: kimberlites themselves and variations arising from differentiation, resorption, and reaction of megacrysts with transporting hybrid magmas (Pasteris 1980; Hunter and Taylor 1984).

Geochemically, two types of megacrysts are recognized: Cr poor and Cr rich. So far, the Cr-rich suite has been described in North America only from kimberlites in the Colorado–Wyoming province, Pennsylvania, and Kentucky, and it is apparently absent in other deposits.

The average size of the megacrysts in kimberlites from Fayette County, Pennsylvania, and Elliott County, Kentucky, range from 0.5 to 3.0 cm. These grains are considerably larger than those found in the mantle nodules. A similar uneven distribution is characteristic of diamonds in different kimberlites. Moreover, a significant discovery from the exploration standpoint is that diamond appears to correlate with specific features of the mineralogy of the megacryst association.

Probably the strongest support for phenocrystal origin of the megacrysts is the fact that some megacrysts appear to have crystallized from a liquid phase, rather than being incorporated into the kimberlite magma as a solid xenolith. In some ilmenite nodules in the Ukukit group of kimberlites, the inner side of the olivine crystals is imprinted in ilmenite nodules, which indicates that the ilmenite was in a liquid state during the crystallization of olivine. The outer portion of the olivine is terminated by the edge of ilmenite nodules. Thus, the ilmenite was plastically deformed by the olivine. For instance, one side of an olivine inclusion found in the ilmenite nodule preserves its idiomorphic habit. However, the side adjacent to the edge of the nodule is deformed along with the host ilmenite. All relations between ilmenite nodules and the kimberlite host resemble relations between chromite nodules and dunite hosts in chromite deposits, which are commonly explained by the liquation (separation into immiscible fractions) of two magmas—silicate and ore (Erlich 1959).

Ilmenite and pyroxene (low-Cr diopside) form regular intergrowths. In these intergrowths, fine bars of ilmenite grow parallel to the c-axis of the host pyroxene and may form large single crystal intergrowths up to 20 cm long. Gurney, Fesq, and Kable (1973) on the basis of geochemical evidence favor a eutectic origin rather than exsolution from a preexisting high-temperature phase. This process has been experimentally reproduced by Wyatt (1977). Diopside-ilmenite intergrowths in rock from the Weltverden mine have been suggested to form from a crystallizing magma of nonequilibrated ilmenite-pyroxene rock (Rawlinson and Dawson 1979).

MARID Suite

Genetically similar to the megacryst suite are nodules dominated by the presence of phlogopite, known as glimmerites, or mica-amphibole-rutile-ilmenite-diopside (MARID)-suite xenoliths. These xenoliths are usually dominated by phlogopite, but amphibole, clinopyroxene, ilmenite, rutile, and apatite are also present in various amounts. The formation of the MARID suite is considered to be a reflection of the kimberlite-magma formation. Two facts are notable:

- Radiometric ages of the MARID xenoliths are much older than the time of the host kimberlite emplacement, implying a long period of magma genesis at depth (see Chapter 14, Temporal Distribution of Diamond Deposits).
- The characteristics of amphibole in the MARID-suite xenoliths are the same as those of amphibole in the kimberlite groundmass (Dawson 1980).

The MARID suite of xenoliths is interpreted to have crystallized under oxidizing conditions from a magma chemically similar to kimberlite within the upper mantle: the presence of amphibole restrains the depth of genesis to less than 100 km (Dawson and Smith 1977). However, other mica- and amphibole-rich xenoliths containing olivine and orthopyroxene are possibly metasomatites, perhaps representing wall rocks of the magma from which the MARID suite of rocks crystallized. They are possibly transitional and metasomatic peridotites.

KIMBERLITE FACIES

Kimberlites are separated into facies according to the pressure related to their origin and to their mode of geologic emplacement. The term *facies* is applied to the both conditions. In order to avoid confusion in terminology, in this section we use two different terms: *subfacies,* applied to the pressure conditions, or depth of origin, and *facies,* applied to the mode of emplacement.

The presence of high-pressure mineral associations in kimberlite permits one to recognize a series of subfacies generated under different pressures. Generally three subfacies are recognized by the presence of different inclusions of high-pressure phases: the diamond subfacies, pyrope subfacies, and picritic subfacies (without any nodules). Thermobarometric estimates on the basis of mineral inclusions in diamond generally fall within the range of 900° to 1,200°C and 40 to 50 kbar of lithostatic pressure. These

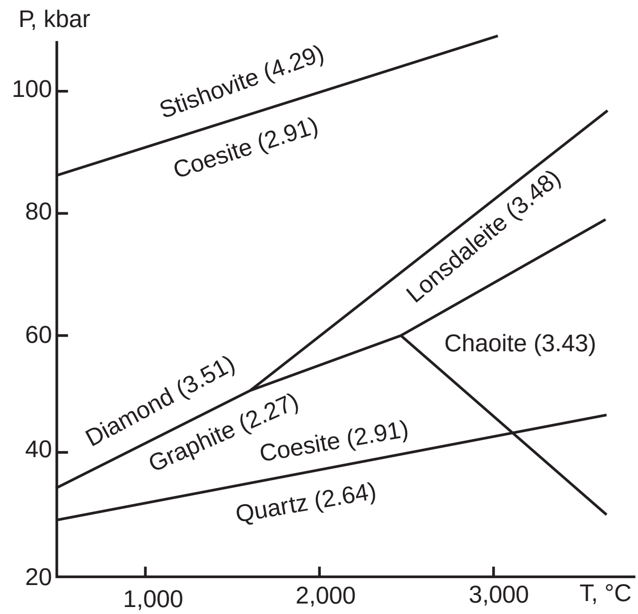

FIGURE 6.2 P–T stability fields for carbon and silicon minerals. Figures in parentheses = mineral density (g/cm³) in the temperature field higher than 1,500°C (from Marakushev et al. 1995). Reprinted with permission of A.A. Marakushev.

parameters correspond with the depth 150 to 200 km. The difference in the P–T conditions for diamond is well illustrated by the paragenesis of modifications of carbon and silicon (Figure 6.2). The thermodynamic conditions of diamond subfacies are characterized by P–T conditions characteristic of diamond itself. Diamond is almost always formed within the coesite field. An exception was noted for diamond recovered from a placer in San Luis (Brazil) that contained inclusions of stishovite and iron-rich periclase with dense perovskite-type structure modifications of $CaSiO_3$ and $MgSiO_3$. The perovskite structure is usually thought to be characteristic of the lower and middle mantle. The low Ca content in garnet and the high Ca content of Ni-bearing iron periclase is characteristic of peridotitic inclusions.

Rocks of the pyrope subfacies were formed at 1,320° to 1,370°C and 18 kbar. In experiments, these rocks do not form at pressures less than 13.5 kbar. Rocks of the picritic subfacies are characterized by the absence of high-pressure inclusions. Picritic basalts occur within this subfacies. The pressure and temperature dynamics related to kimberlite magma evolution and the corresponding thermodynamical parameters of both subfacies and rocks of the picritic facies are shown on Figure 6.3.

Various geologic conditions of kimberlite emplacement are reflected by different facies. Two groups of facies are recognized: an extrusive group and a hypabyssal group. They are interconnected by diatremes. A general model of the distribution of facies in relation to different types of kimberlite bodies is presented at Figure 6.4.

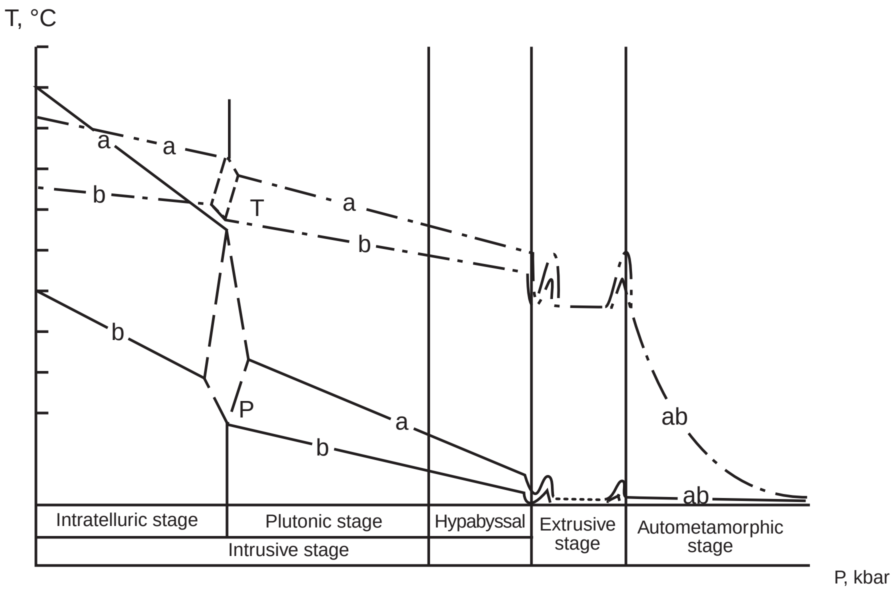

FIGURE 6.3 General character of dynamics of temperature T and pressure P during formation of kimberlites of the diamond (a) and pyrope (b) subfacies. The field below line b characterizes the thermodynamic parameters of porphyric ultramafic and ultramafic-alkaline rocks of the picritic facies. One unit along the y-axis = 200°C; one unit along x-axis = 7.5 kbar. Labels at the base of the diagram are stages of kimberlite development. We here change the term "explosive stage" used by Milashev to "extrusive stage," which in our opinion more precisely reflects the geological processes at this stage (modified from Milashev 1974).

Diatremes

Appearance. The recognition of diatreme facies kimberlite is important in exploration, because such facies could indicate the presence of a fairly sizable tonnage of material associated with a pipe. Diatreme facies is represented by volcaniclastic kimberlite that is sometimes referred to as tuffisitic kimberlite or tuffisitic kimberlite breccia (Clement 1979; Clement and Skinner 1979; Mitchell 1986).

In general, diatreme facies kimberlite is recognized by its fragmental nature (ash-fall tuffs and pyroclastic flows are nonexistent). The nature of volcanoclastic rocks is described in accordance with the term introduced by Fisher (1961), who considered them as rocks formed by any process of fragmentation, dispersed by any kind of transporting agent, and deposited in any kind of environment. Depending on the size and amount of fragments, these rocks may be divided into tuffisitic breccias and tuffisitic kimberlites. Tuffisitic breccias contain abundant country-rock fragments that range in size from few centimeters down to microscopic particles. Autoliths occur in lesser amounts and are formed of hypabyssal facies that represent an earlier magma phase. Tuffisitic breccias may include lense-like bodies of tuffisitic kimberlite, which differ petrographically only in clast content. They intrude earlier breccias or form narrow screens in the margins of the pipe.

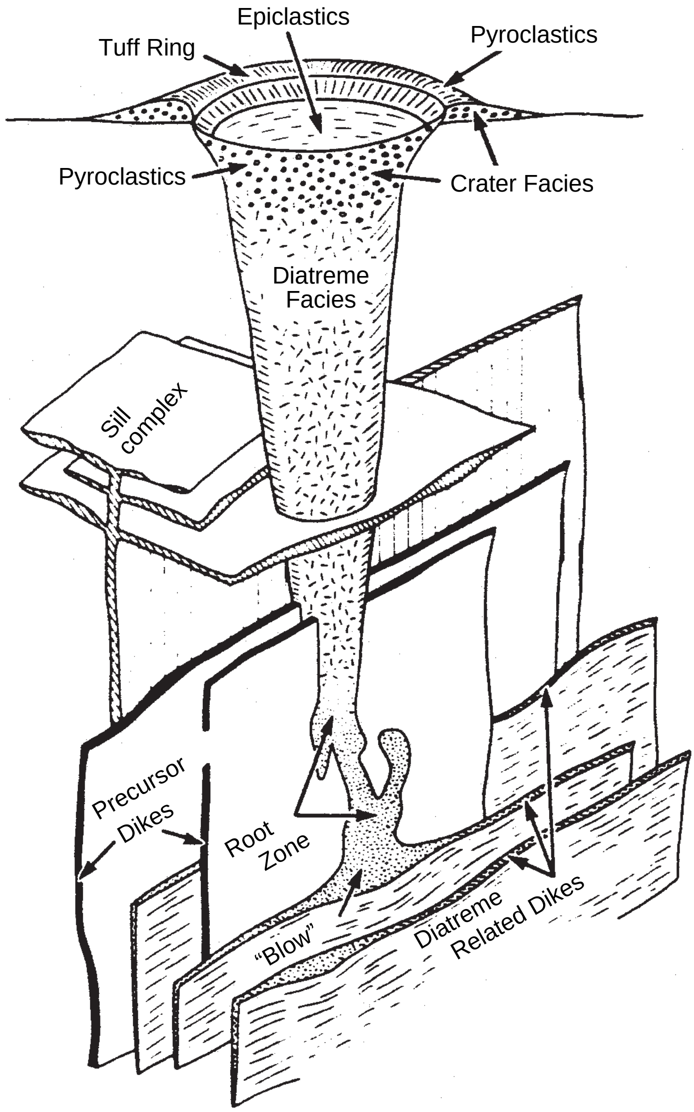

FIGURE 6.4 Model of idealized kimberlite magmatic system, illustrating the relationships between crater, diatreme, and hypabyssal facies. Hypabyssal facies rocks include sills, dikes, root zones, and "blow" (Mitchell 1986). Reprinted with permission from Kluwer Academic/ Plenum Publisher.

Diatreme facies kimberlite consists of abundant subrounded to angular rock fragments with serpentinized macrocrysts of olivine and lesser enstatite, Cr-rich and Cr-poor pyrope garnet, diopside, picroilmenite, clinopyroxene-ilmenite intergrowths, and phlogopite in a finely crystalline matrix of serpentine, carbonate, olivine, diopside, picroilmenite, phlogopite, perovskite, magnetite, Cr-rich spinel, hematite, apatite, and zircon (Rogers 1985; McCallum 1991). These facies are recognized by their fragmental and/or tuffaceous texture and abundant angular to rounded rock inclusions, most of which are small. These lapilli-sized clasts of kimberlite have been termed kimberlite pellets (Clement 1973); they give the rock a distinctive tuffaceous appearance. Rarely, some crater facies kimberlite fragments may occur in the diatreme facies.

The diatreme facies grades into hypabyssal facies kimberlite near the root zone of the diatreme. Diatreme facies kimberlite extends upward from the feeder dike complex and forms a pipelike structure that is referred to as a diatreme, or pipe. The diatreme facies kimberlite within this funnel is represented by volcaniclastic breccias with clasts of country rock, fragments of hypabyssal kimberlite, and pellet lapilli in a matrix of serpentine and diopside. Primary calcite, common in hypabyssal kimberlite, is not generally found in diatreme facies (although the diatreme facies contains considerable secondary calcite). More than one episode of diatreme facies may be recognized in a pipe, representing multiple intrusives with distinct vertical columns. As an example, McCallum (1991) mapped six different kimberlite facies in the Sloan 1 kimberlite diatreme and the adjacent Sloan 2 blind diatreme in Colorado. Each facies, if diamondiferous, may exhibit different ore grades.

In general, diatremes form subvertical to vertical pipes that taper down at depth, or form steeply inclined cylindrical bodies. The average angle of inclination of the walls of various pipes in the Kimberley region of South Africa (Wesselton, DeBeers, Kimberley, and Dutoitspan) is 82° to 85°. Ideally, the pipes produce isometric or ellipsoidal cross sections in a horizontal plane and are filled with kimberlitic tuff or tuff-breccia. However, the Kelsey Lake diatreme in Colorado was found to be atypical, because the intrusive appears to have stopped under the Proterozoic granite.

Most pipes continue from the surface to depths as great as 2 to 3 km to where they pinch down to narrow root zones with a feeder dike. This depth is deduced from two independent aspects:

- As lithostatic pressures decrease, kimberlites erupt at the surface as explosive magmas, highly charged with water vapor and CO_2. The water and CO_2 vapor increases enormously at pressures of 0.4 to 0.8 kbar, the equivalent to 1.5 to 3.0 km depth (Dawson 1971).
- Deep mining traced the Kimberley and Bullfontein pipes from roughly circular diatremes at the surface to narrow fissures at depth. The root zones of the larger pipes in the Kimberley area have persisted for about 500 m.

The outer borders of diatremes typically exhibit sharply defined crosscutting contacts with the country rock. The pipe's wall rocks may be grooved, striated, and slickensided. Many of these features are vertically oriented but other attitudes may also occur (DuToit 1906; Williams 1932).

Kimberlite adjacent to the country rock contact may be highly sheared with mylonitic zones and calcite veins (Williams 1932; Bloomer and Nixon 1973). In the State Line district in Wyoming, trenching of the Schaffer pipes revealed sheared and permeable granite country rock at the contact with kimberlite. The sheared contact supported considerable ground water flow. The endocontact breccia zones in many kimberlite pipes may or may not contain tuffisitic kimberlitic cement in which blocks and fragments of wall rock are incorporated. The breccias generally do not extend laterally or vertically more than a few tens of meters (Figure 6.5).

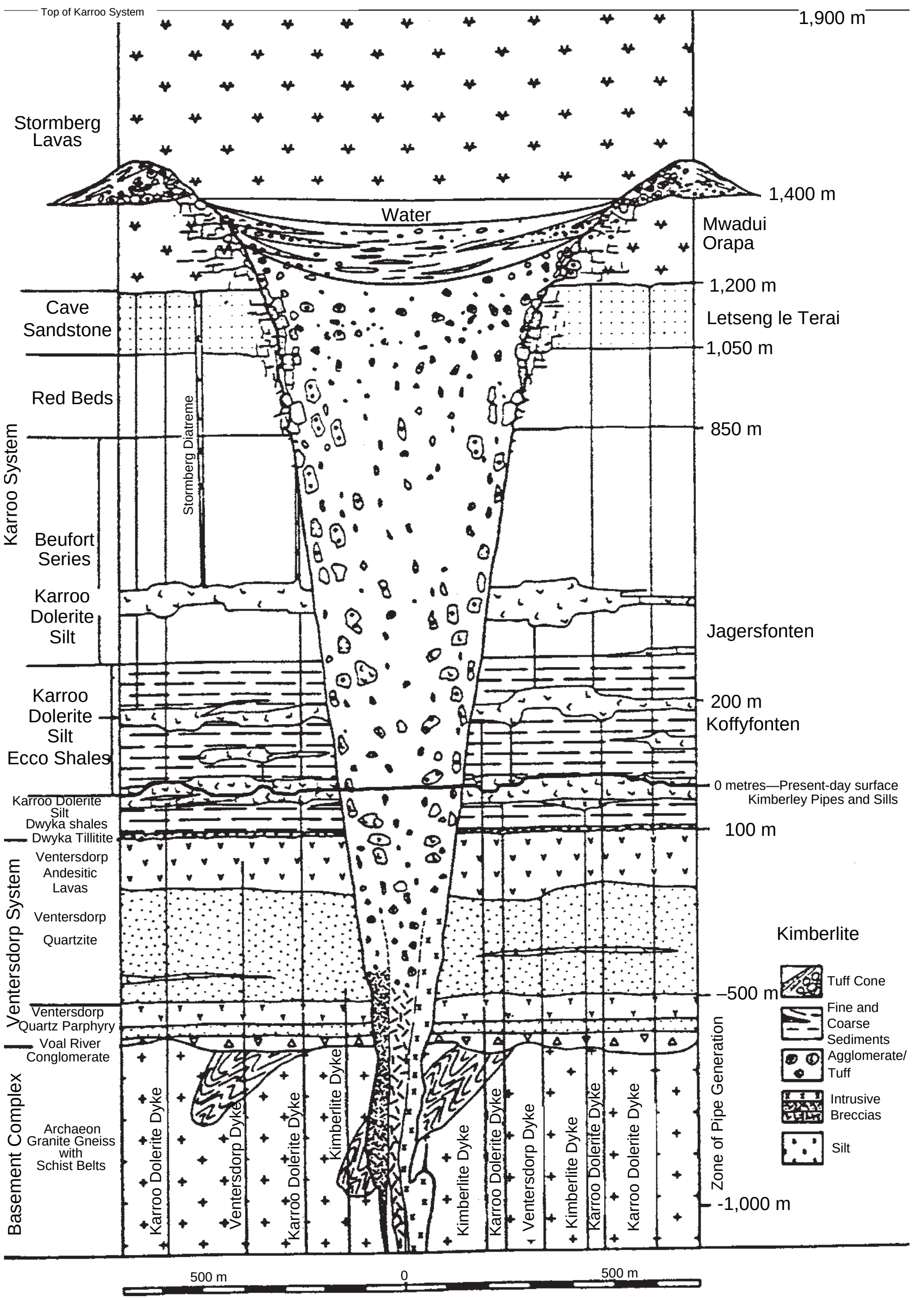

FIGURE 6.5 Model of a kimberlite pipe (Hawthorne 1975)

Models of emplacement. The classic model of a kimberlite pipe is based on mined-out kimberlite pipes in Kimberley, South Africa. This model includes a carrot-shaped pipe filled with diatreme facies kimberlite that is capped by a thin blanket of crater facies. The morphology of kimberlite pipes was described in detail by Hawthorne (1975) on the basis of mining in South Africa. He indicated that at depths of more than 600 m below the surface at the Wesselton, DeBeers, Kimberley, and Dutoitspan mines that the pipes contracted into root-like or dike-like bodies known as the root zone. This transition apparently has not been encountered in the Bullfontein pipe.

This morphology is readily apparent in the Kimberley mine. Between the present-day surface and depths of 300 to 600 m, the Kimberley pipe is generally regular in form. Kimberlite tuff or agglomerate generally predominated over the intrusive breccia within the pipe, and separate columns of intrusive breccia were discernible. From this depth to the surface, the pipe walls adopted an outward slope of 80° to 85°, irrespective of wall-rock type.

The Kimberley pipe, which was mined out by 1915, about 20 years after its discovery, contracted sharply with depth. At the lowest level of mining (1,083 m), it was no longer pipe shaped but rather had the appearance of three intersecting fractures with a slight enlargement at their intersection. The fractures were filled with kimberlite (Kennedy and Nordlie 1968). Combined with the estimated 1,600 m of eroson since the time of emplacement, the depth to the original point of expansion is probably 2.4 km.

Near the present-day surface large masses of slumped wall rock ("floating reefs") were encountered. These reefs do not persist at depth in any of the South African pipes, but they extend to depths of 500 m below surface in the Jagersfontein pipe.

The distribution of these large xenoliths is illustrated in a general model of a pipe in Figure 6.5. An example of a small floating reef was also encountered in the mine rib of the Sloan 2 kimberlite in Colorado (Figure 6.6).

Many of the floating reefs consist of igneous rock types. They tend to be peripheral in single vents but when central may indicate the coalescence of more than one diatreme as, for example, in the Premier pipe, whose plan view suggests such coalescence. In some instances these reefs are known to persist in high proportions below their source level. Cognate xenoliths of crater facies kimberlite were also apparently circulated at depth in the Kelsey Lake kimberlite in Colorado. Breccia fragments of the crater facies kimberlite were reported in the diatreme facies kimberlite at the Kelsey Lake mine (Howard Coopersmith, personal communication, 1997) *(*see Figure 6.7).

Above the present-day surface of most pipes, the material that formed the pipes is presumed to have consisted almost entirely of kimberlite tuff and agglomerate with a high degree of uniformity. A few late-stage kimberlite dikes may have intruded these rocks. There may also exist a rough vertical zonation, reflecting the stratigraphic sequence of the country rocks, although xenoliths are displaced downward relative to their original position. Kimberlite magmas are interpreted to erupt from depth within the Earth's upper mantle and travel to the Earth's surface in a matter of hours, on the basis of phase equilibrium studies (O'Hara, Richardson, and Wilson 1971) and some physical evidence.

Both mining of diatremes to deep levels and geophysical data (magnetic data in particular) do not indicate the existence of any intermediate magma chambers from which the diatremes initiated. Sometimes the inner structure of the diatremes is characterized by subvertical columns composed of different types of tuffisitic kimberlite. The contacts between these columns are sharp, and it is suggested that the various columns reflect existence of several episodes of kimberlite emplacement.

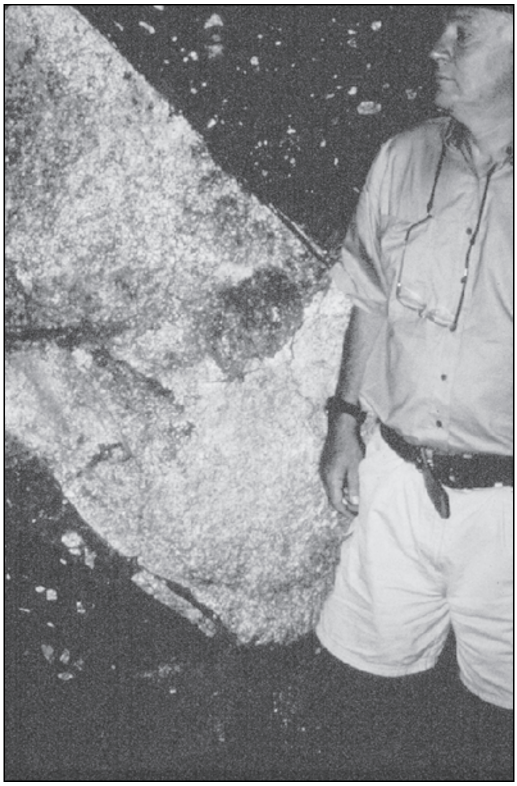

FIGURE 6.6 Floating reef of granitic country rock in the Sloan 2 kimberlite, Colorado. Bernie Free poses for scale (photo by W.D. Hausel)

From the surface, most pipes project downward to a narrow root zone formed of hypabyssal facies kimberlite. The root of the pipe, which taps a feeder dike, typically lies 2 to 3 km below the surface and may lie at the intersection of two fractures. The feeder dike has the appearance of a dike complex occupied by one or more intrusive phases of hypabyssal facies kimberlite. An excellent exposure of a feeder-dike complex, mapped in the Iron Mountain district of Wyoming, shows anastomosing kimberlite dikes with periodic blows and rare sills exposed at the present erosional surface. The dikes are formed of massive, porphyritic, hypabyssal facies kimberlite in places containing enlargements filled with hypybyssal facies kimberlite breccias and rare diatreme facies kimberlite.

How kimberlite magma ascended to the surface is largely speculative. It is thought that migration to the surface began in the Earth's upper mantle after small amounts of volatile-rich ultramafic liquid concentrated along grain boundaries and began to migrate to zones of lower pressure. (Gases such as CO_2 may result from the breakdown of carbonate-rich zones.) The early stages of fracture growth may have been controlled by the chemical reaction between liquid and solid in the fracture. Certain specific features of geochemical data suggests that the initial phase of kimberlitic magma in the upper mantle started to ascend as a diapiric upwelling (Green and Gueguen 1974).

The speed of ascent of the kimberlite magma must be sufficiently rapid to preclude sedimentation of high-density peridotite and eclogite xenoliths and to prevent total

FIGURE 6.7 High wall at Kelsey Lake mine in Colorado with kimberlite breccia lying under a thin crust of granite

resorption or inversion of diamonds (Dawson 1971). As the kimberlite magma rises, diamonds should survive as long as the magma is not oxidizing and is emplaced rapidly, because diamond is considered to be unstable at shallow depths and at high temperatures within an oxidizing environment. Such oxidizing conditions may cause diamond to partially resorb, revert to graphite and/or burn to CO_2.

At depth, the segregated magma is expected to rise rapidly, possibly at 6 to 18 mi/hr (10 to 30 km/hr), in order to transport high-density ultramafic xenoliths. Within the last few kilometers of the surface, emplacement rates would increase dramatically to several hundred kilometers per hour. Such velocities would bring diamonds from the mantle to the surface in less than a day.

McGretchin (1968) estimated that the speed of the fluidized material near the surface increased to as much as 870 mi/hr (400 m/s), or about the speed of sound (Mach 1 or 331 m/s). Some estimates have suggested kimberlite emplacement at the earth's surface may have achieved velocities exceeding Mach 3 (Hughes 1982), although such emplacement velocities are not accepted by all researchers. Dawson (1971) suggested that the characteristic rounding and polishing of xenoliths and xenocrysts were due to the sandblasting effect of the this fluidization process.

The breakthrough at the earth's surface would have been accompanied by a rapid drop in pressure with an explosive release of gas, which in part is probably the cause of

FIGURE 6.8 Exposed contact between the Schaffer kimberlite, Wyoming, and host Sherman granite shows no evidence of thermal baking indicating that the emplacement of the kimberlite magma was relatively cool (Hausel et al. 1979)

the extensive serpentinization of olivine in the kimberlite diatreme. Additionally, the eruption of the kimberlite into groundwater may also be partially responsible for the pervasive serpentinization.

The temperature of the magma at the point of eruption is relatively cool compared with that of other magmas. Watson (1967) suggested a temperature of less than 600°C on the basis of the coking effects (i.e., the volatiles driven off of a low-grade coal by heat, such that the fixed carbon and ash are fused together) on coal intruded by kimberlite. A low temperature of emplacement is also supported by the absence of any visible thermal effects adjacent to most kimberlite contacts with the intruded country rock (Figure 6.8). Davidson (1967) suggested the temperature of emplacement may have been as low as 200°C on the basis of the retention of argon. Hughes (1982) pointed out that the near-surface temperatures of the gas-charged kimberlite melt may even be as low as 0°C owing to the adiabatic expansion of CO_2 gas as the kimberlite erupted at the surface.

By contrast, hypabyssal facies kimberlite in root zones dikes and sills show thermal metamorphism of wall rocks and xenoliths, chilled margins, trachytic and flow differentiation features, and crystallization of high-temperature dendritic calcite. These features are consistent with the kimberlite matrix being a hot liquid at the time of emplacement (Dawson 1971). In the root zones of kimberlites at Iron Mountain, Wyoming, some silicification is apparent in the wall rocks (Smith 1977). The silicification is presumably an indication of higher magma temperatures at depth.

Initially, it was assumed that kimberlite pipes were formed as a result of explosive boring (Wagner 1914). However, certain well-documented observations led to the rejection of this hypothesis:

- the absence of any trace of explosive chambers at the root zones
- zonal distribution of xenoliths in the pipes
- vertically consistent subcylindrical form.

The most recent models for kimberlite emplacement have sought to explain two basic facts:

- the high amount of gases in the kimberlitic magma at the time of emplacement
- low temperatures of kimberlitic magma, inferred from the almost complete absence of contact alteration in either the wall rocks or xenoliths.

The first problem was approached by investigating two hypotheses.

- The fluidization model of Dawson (1962). In this model fragmental material is transported to the surface by a fast-moving gas stream. This process was originally postulated by Cloos (1941) to explain similar features of tuff pipes of Schwabia. The speed of ascent must have been sufficiently rapid to preclude sedimentation of high-density peridotite and eclogite xenoliths and also to preclude total resorption or inversion of diamonds.
- Hydrovolcanism, on the basis of the model presented by Sheridan and Wohletz (1983) and supported by Mitchell (1986). The concept of hydrovolcanism requires the ascending magmatic column to extract water from a meteoritic source to produce a diatreme. However, hydrovolcanism does not adequately explain that kimberlite pipes have been traced down to feeder dikes at depths of more than 1 km, and potentially to 2 to 3 km. An additional problem is that the depth of active water circulation generally may not exceed 200 to 250 m.

The complex structure of kimberlite diatremes can be explained by a combination of processes. Diatremes are probably initiated by subsurface brecciation (Clement 1979, 1982). Along joints or fractures where access to the surface is easiest for the magma, groundwater percolation could be deeper (possibly on the order of 2 to 3 km according to Clement 1979). Thus the mechanism of diatreme formation could combine fluidization and hydrovolcanic activity.

Any mechanism of emplacement for kimberlite pipes needs to explain the following:

- From a depth of about 2,000 m below the surface, a vertical, steep, yet fairly regular carrot-shaped perforation of the Earth's crust was made.
- The type of crustal material penetrated appears to have little bearing on the slope of the cone except at the surface.
- Within these pipes, large masses of wall rock weighing thousands of tons slumped peripherally to great depths.
- Wall-rock xenoliths weighing more than a few tons do not appear to have been transported upward to any extent.
- Small inclusions ranging from extremely dense to relatively light material were fairly evenly distributed throughout the tuffaceous or agglomeratic kimberlite that filled the vent.
- Some of these inclusions were transported upward from extreme depths.

The tuffaceous kimberlite is highly uniform in texture. Variations in composition are related to changes in the wall rock along the length of the pipe. It can be deduced from the composition of kimberlitic agglomerates and tuffs that 50% to 60% of the crustal material that originally occupied the vent was not permanently ejected but became incorporated as inclusions in the consolidated kimberlite that ultimately filled the vent.

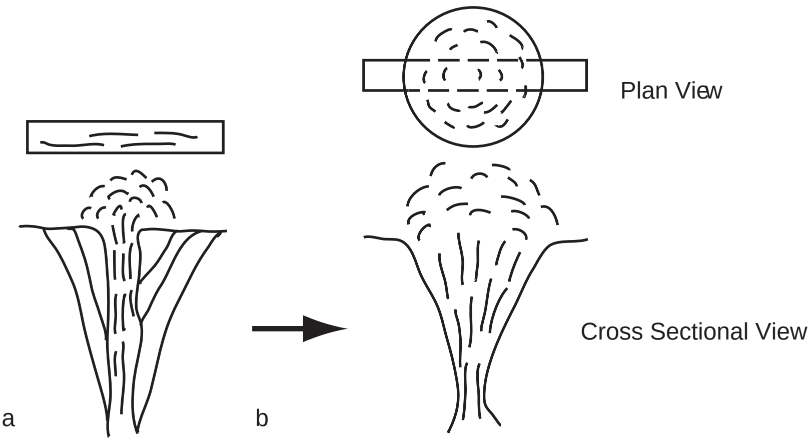

FIGURE 6.9 Conversion of a slit-like conduit (a) into a conical diatreme with a funnel-like extension (b) (modified from Novikov and Slobodskoy 1979)

Kobets and Komarov (1958) and Nikitin (1982) described crescent-shaped faults associated with some pipes in Yakutia, along which country rocks subsided. It was suggested that these faults reflected the positive topographic expression of domal structures created in the country rocks above the pipes. The assumed diameter of such uplifts probably was four times the pipe's diameter. At least in one case—the Leningrad pipe exposed on the Omonoos river, Yakutia—Cambrian limestones in contact with the kimberlite form an anticline around the diatreme and dip outward at an angle of 45° to 60° (Milashev et al. 1963).

At the intersection with the surface, and owing to rapid gas expansion, a funnel-like structure is also formed with consequent formation of extrusive bodies (either tuff cones or lava flows). Novikov and Slobodskoy (1979) wrote,

> *the most natural reason for the appearance of a funnel in the near-mouth part of a diatreme is the expansion of the gas jet at the outlet from the vertical conduit. The development of the horizontal component in the velocity vector of the gas flow and the particles transported by it will convert the near-mouth marginal upraise, formed by the intersection of the conduit walls with the ground surface, into the obstacle, which will also be subjected to intense disruption. The result of this should be smoothing of the walls in the near-mouth sector until their configuration coincides with the directions of the vector velocities at each point in the near-wall portion of the flow.*

The scheme of suggested formation of diatreme and funnel-like extension is presented in Figure 6.9.

There is no suggestion of any direct relationship between the size of any pipe at the present-day surface and the depth to which it has eroded. It is suggested that at whatever depth a pipe was generated from a dike or feeder, and whatever size it may have attained, it resembled the model in terms of wall-rock slopes, distribution of inclusions, near-surface flaring, and overall form. Thus, pipes with rapidly flaring walls found to contain predominantly agglomerate kimberlite may be inferred to be not too deeply eroded.

Extrusive Facies

Hawthorne (1975) cited personal communication with E. Gerryts, who described the only example of a preserved ejecta cone associatied with a tuff ring of kimberlite in Mali (near the village of Kasama, 140 km south of Kayes). Lava described at Igwisi Hills, Tanzania (Reid et al. 1975), is associated with an 250 m diameter diatreme and is surrounded by remnants of lateritized kimberlite tuff 125 m across. The tuff is only 3 to 4 m thick and is associated with a restricted vesiculated lava flow.

In all African regions characterized by erosion that exposes deep regions of the diatreme, the kimberlitic diatremes are surrounded by tuff cones, which form a relatively low structure in relation to the diameter of the crater; the diatremes resemble the classic maar volcanoes.

Reid et al. (1975) described three small tuff cones and a small vesicular lava flow in the Igwasi Hills, Tanzania. The lava consists of rounded olivine phenocrysts and peridotitic microxenoliths (some containing pyrope, chrome-diopside, and phlogopite) in a matrix of carbonate, serpentine, spinel, and perovskite; a trachytic texture reflects alignment of calcite plates.

The groundmass spinel belongs to the category of magnesian ulvospinel–ulvospinel-magnetite, which is similar to Ti-rich spinel in evolved kimberlite magmas. Petrographically, the lava is similar to some macrocrystalline kimberlites and kimberlites from sills described at Bullfontein (Reid et al. 1975).

A similar situation occurs at the Gross Bukaros volcano, Namibia. Mitchell (1986), referring to unpublished notes, stated that microbreccias of this volcano were probably deposited by highly inflated surges produced by the breakthrough stages of an eruption of a carbonatite-melilite-nephelinite volcano.

Crater facies. In regions characterized by little erosion since the emplacement of kimberlite pipes, diatremes may be surrounded by tuff rings—relatively small structures in relation to the diameter of the crater—that resembled a maar volcano. These rings are the only indication that volcanic edifices existed around kimberlitic pipes. Such a pyroclastic ring surrounding a maar is shown in Figure 6.5.

Although seldom preserved, the orifice of a kimberlite pipe is marked by crater facies kimberlite. The crater facies includes pyroclastic and epiclastic kimberlitic material (pyroclastics, tuffs, and lapilli pyroclastics) in a ring-tuff surrounding the pipe. The crater facies has been preserved only in pipes that have undergone little or no erosion since their emplacement.

Brecciation of the near-surface wall rock is associated with kimberlite pipes. The McKeawn's breccia found at the margins of the Mwadui pipe provides one example (Edwards and Hawkins 1966). At the surface, the pipe is surrounded by a tuff ring that resembles a maar-type volcano with a comparatively low height compared with its diameter.

The cone (or tuff ring) has large ejected blocks of kimberlite and wall rock concentrated close to the crater rim and finer material, possibly with rudimentary layering, deposited at the outer slopes. Owing to some erosion of the cone, epiclastic kimberlite was deposited. Near the center of the basin (maar), alternating coarse and fine sedimentary layers contain predominantly fine-grained material. The layers were produced by the reworking and sedimentation of kimberlitic material and country rock material in a crater lake. The depth of sediments is 130 m, and the diameter of the pipe at the original surface was about 1,000 m.

The tuffs in the crater facies kimberlite are vesicular, highly altered, and carbonatized. They are similar to the Igwisi lava and contain olivine macrocrysts in a fine-grained serpentine and calcite-rich groundmass. The well-stratified tuffs consist of alternating layers of coarse lapilli-sized tuffs and laminae of finer ash-sized tuffs. The layers consist of serpentine pseudomorphs after olivine, together with phlogopite, garnet, and ilmenite macrocrysts set in a matrix of serpentine, clays, calcite, and chlorite. These crystal tuffs grade into lithic tuffs with increasing quartz, feldspar, and biotite derived from the country rock. Underlying the well-stratified tuffs are poorly stratified tuffs and tuff breccias containing fragments of kimberlite, country rock, and mantle-derived xenoliths enclosed by pyroclastic material similar to the overlying tuffs. These tuffs may be overlain by epiclastic material washed in from the surrounding country rock and by kimberlite tuffs (Mitchell 1986).

Janse (1975) and Ferguson et al. (1975) described the Gross Bukaros volcano on the Hatzium farm in Namibia as having kimberlite affinity. This feature has a cryptovolcanic dome cut by kimberlite dikes (McIver and Ferguson 1979).

Other sites originally described as extrusive kimberlites were later discredited on the basis of petrography. They included Moroto (Uganda), Lesorogai (Kenya), and the Lashimi (Tanzania) volcanoes. The rocks were found to be nephelinitic, ankaramitic, and carbonatitic tuffs and are merely pyroclastic components of carbonatite-nepheline volcanism associated with the East African Rift; they are unrelated to kimberlite.

A similar situation occurs at the Gross Bukaros volcano, Namibia. Mitchell (1986), referring to unpublished notes, stated that microbreccias of this volcano were probably deposited by highly inflated surges produced by the breakthrough stages of an eruption of a carbonatite-melilite-nephelinite volcano.

Some researchers have suggested that ore grades of diamond may decrease with depth in kimberlite pipes. However, this decrease is not universal, because many kimberlites contain similar ore grades in both the diatreme and root zones. However, ore grades in crater facies kimberlite have been significantly higher than in the diatreme facies. Unfortunately, crater facies kimberlite is rarely preserved, and erosion of the crater facies may be a source of nearby rich diamond placers.

Epiclastic facies. Epiclastic facies are formed by rewashing and resedimentation of original kimberlites within a basin (similar to a crater lake) located above a pipe. Several such lakes were studied at Mwadui pipe (Tanzania), Orapa pipe (Botswana), Kao pipe (Lesotho), and also in the Kimberley region and in Kansas (United States) (Mitchell 1986). The terminology for such volcano-sedimentary rock is based on the grain size of particles.

Spatial distribution of epiclastic kimberlites in Africa depends on the depth of erosion. With few exceptions they are practically absent south and east of line that runs through eastern Zambia, northern Zimbabwe, southern Botswana, and the northern Cape Province and meets the coast in the southwestern Cape Province (Figure 6.10). Lying on some kimberlitic pipes in Arkhangel'sk region of Russia are similar formations that Russian geologists described as crater facies kimberlite.

Hypabyssal Facies

Hypabyssal facies kimberlite is a massive, porphyritic kimberlite that formed at depth in dikes, sills, and small plug-like bodies that represent root zones of deeply eroded pipes. The mineralogy of the hypabyssal kimberlite is essentially identical to that of the

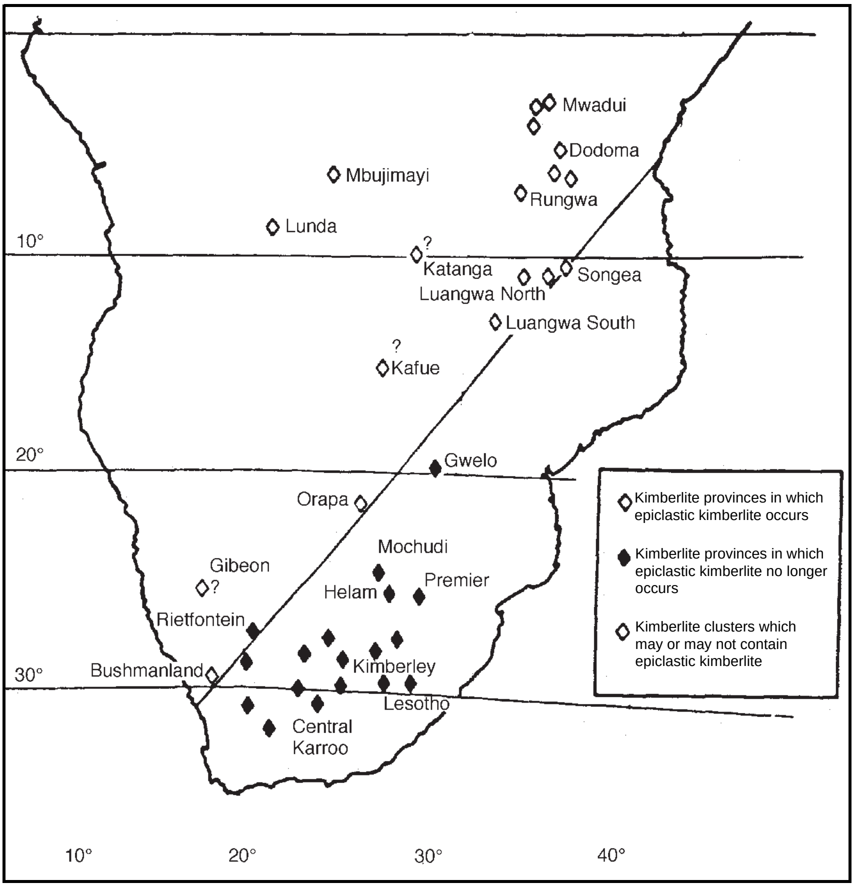

FIGURE 6.10 Spatial distribution of epiclastic facies of kimberlites in central and southern Africa (Hawthorne 1975)

diatreme facies; however, globular, emulsion-like segregations of serpentine and calcite are more abundant in the hypabyssal facies (Rogers 1985). In addition, xenolithic fragments rarely exceed a few percent in the hypabyssal facies (unlike the diatreme facies), and where present, they are typically moderately well-rounded upper mantle nodules and lower crustal xenoliths.

According to Mitchell (1986), hypabyssal facies are rocks formed by crystallization of volatile-rich kimberlite magma. Such rocks possess all features of igneous textures and show the effects of magmatic differentiation; pyroclastic fragments and textures are absent.

Dawson (1980) recognized two textural types of hypabyssal facies. Typically, hypabyssal facies consists of massive kimberlite with few or no xenoliths of country rock, because it experienced comparatively gentle intrusion of kimberlite melts into dikes, sills, and cavities cleared of fragmental debris by gas explosions. The textures show the effects of chilled margins of magmatic bodies, trachytic textures in the central part of dikes, and a median concentration of xenoliths and megacrysts. At Benfontein, some sills were described as having cross-lamination and graded-bedding washout structures attributed to magmatic sedimentation (Dawson and Hawthorne 1973). Kimberlite breccia (also termed autolithic breccia) contains different types of fragments immersed in a cement of massive kimberlite. Fragments can compose up to 20% of rock's total volume.

Dikes. Dikes are tabular, subvertical to vertical masses, which formed when open fissures were filled by hypabyssal facies magma. They may be 0.3 to 1.5 m wide or may be as much as 61 m wide. Their length is sometimes unusually great: one dike in the Winburg District in South Africa is reported to extend 65 km. Another dike southwest of Kimberley persists for 30 km (Dawson 1980).

Wagner (1914) described a 1-m-wide dike in the Bullfontein mine, whose width showed little variation downward for 350 m. Dikes are usually composed of massive kimberlite, and the contacts with the country rock are sharp with little to no evidence of alteration of the wall rock. The dike and diatreme relations are quite complicated owing to repeated episodes of kimberlite intrusion. Thus within the same region one may find feeder dikes for kimberlite pipes as well as pipes cutting earlier dikes. Wagner (1914), who first described such relationships, termed the dikes that formed during these subsequent episodes "antecedent" dikes. Several types of kimberlite dikes were described by Zubarev (1989).

Dikes that have no relationships with pipes show no relationship with any known pipe. Dikes of this type are characterized by great strike lengths, the absence of any direct connection with any kimberlite pipes, and a consistently high diamond content (3–5 carats/tonne).

Examples of this type of dike are found in the Rustenberg region, South Africa, northwest of the Pretoria kimberlite field. Two suites of dikes were traced for several kilometers. Within the Swartruggens region, the dikes were traced more than 5.25 km, and individual dikes were found to vary from 0.02 to 2.1 m thick. At depth, the dikes widen and some blind dikes have been found that don't continue to the surface. South of Swartruggens, the Mallin dike is 2.7 km long and 0.3 to 0.9 m wide.

Dikes that serve as feeders for pipes show a clear relationship with kimberlite pipes and include dikes that connect to diatremes at depths of 240 to 1,000 m below the surface. Such dikes are usually observed in the central part of a province. They typically vary from two to several tens of meters wide, and several pipes may be located along the dikes. These dikes often contain lens-like swells up to several tens of meters across.

In some areas, depending on the depth of erosion, the former pipes may for the most part have been completely removed by erosion. However, kimberlite blows (enlargments) along the dike complex may be an indication of a former pipe, especially where there is some evidence of brecciation. Such areas, particularly where the dikes are shown to be diamondiferous (regardless of grade) may be the source of some nearby, rich diamond placers.

Of economic interest are the swells within these dikes. The swells form zones of great length, from several kilometers (Barclay–West, Winburg region), to tens of kilometers (Lesotho). The width of the dikes within these zones range from 0.45 to 8 m and the swells (blows) reach up to tens of meters wide. The diamond content of these dikes is lower in comparison with dikes of the first type. When diamondiferous, they may average 0.45 carats/tonne, and reserves may be considerable.

Apparent kimberlitic feeder dikes with chemistry dissimilar to that of pipes in the region (for example, those associated with the Mir, Aikhal, Tayezhnaya, and Tayezhnaya No. 1 pipes within the Malo–Batuobinsky region of Yakutiya) are characterized by a well-developed fluidal texture expressed by the orientation of the long axis of phenocrysts parallel to the apparent magma flow. The mineralogic and chemical composition of these dikes is clearly distinct from those of kimberlite in the pipes. In addition, the pyrope content in the dikes associated with the Mir pipe is seven times as great as that of the kimberlite in the pipe.

The pyrope and chromite content of the kimberlitic dikes associated with the Aikhal pipe also show differences compared with the pipe. For instance, the dike is characterized by abundant picroilmenite, which is absent in the pipe. Satellite minerals in the dikes are characterized by an increased amount of the Ti-group minerals (picroilmenite and orange garnet) and by a decreased amount of the Cr-bearing minerals (Cr-garnets and chromites). The kimberlite of the dike is also characterized by increased amounts of Ti and P.

The diamond content of these dikes is lower than that of the associated pipe, and the morphology of the diamonds is different. The diamonds in the dikes are dominated by crystals with rhombododecahedral habit and by polycrystalline intergrowths with clear traces of plastic deformation. The difference in the chemistry and mineralogy of the kimberlite in the pipes and in the dikes does not support that the two belong to simultaneous and similar episodes of formation (Kharkiv 1975).

Some dikes are located near known pipes. An extensive system of dikes along the margin of the Lesotho basin occurs within terrigenous sediments of the Beaufort Series (Triassic–Jurassic). Thirty pipes and more than 200 dikes with 21 blows have been identified (Figure 6.11). In addition, a complex of kimberlitic dikes with blows in the Winburg area east of Kimberley has been traced more than 16 km. The main dike is more than 1 m wide. To the north, within the Kronstadt field, is a 6-km-long and 1.2-m-wide dike (Figure 6.11).

Other dikes developed within fracture zones. Dikes of this type are represented by bodies located within concentric fissure systems that were formed during the formation of the pipe. As an example, a 3-m dike girdles the Bullfontein pipe (Kimberley field) at 215-m depth and lies within 15 cm of the pipe's northern contact. Kimberlites, which form the pipe and associated dikes, are of identical composition.

Dikes emplaced after crystallization of the pipe are found in the Premier and Dutoitspan pipes (Kimberley field), which cut the kimberlite in the pipes. The kimberlite forming the dikes of this type consists of serpentine, magnetite, and carbonate with traces of the other typical kimberlitic minerals. Diamonds are usually absent in this type of dike. Similar dikes also cut kimberlite of the DeBeers pipe at a depth of 500 to 700 m.

Dikes are often labeled by diamond exploration companies as poor targets. However, Nixon (1995) indicates that dikes may contain mineable ore grades. Economic diamond mineralization was found in dikes at Elands, the Star-Theunissen-Kroonstad area, Bellsband, and Swartruggens, South Africa. Diamondiferous kimberlite dikes also are found in Guinea, Sierra Leone, Mali, and Liberia and are the source of some important placer deposits. The diamondiferous dikes at George Creek, Colorado, yielded significant ore grades during feasibility tests in the 1980s. Bulk samples from the dikes ranged from 0.18 carats/tonne to 1.35 carats/tonne (McCallum and Waldman 1991). Thus, kimberlite dikes should not be overlooked as potential targets or as potential sources of placer diamond deposits.

Sills. Kimberlite sills are comparatively rare. In accordance with their definition, they are subhorizontal intrusive bodies emplaced along fracture zones and contacts

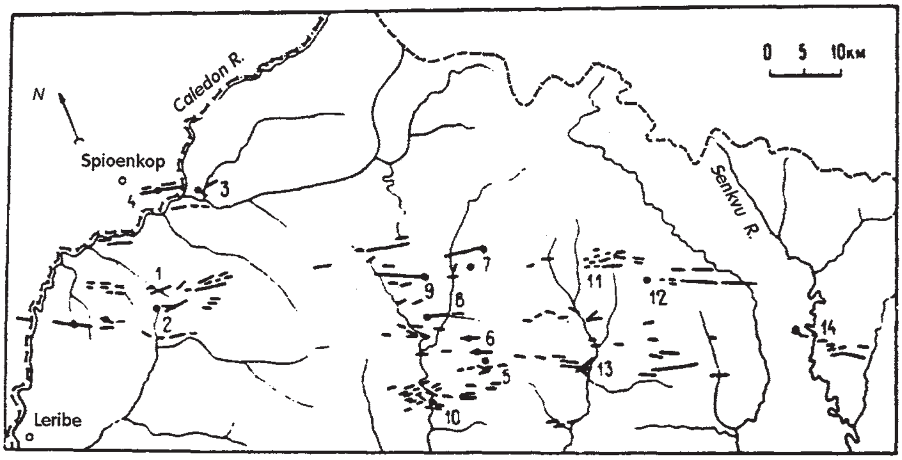

FIGURE 6.11 Lesotho dike swarm (Zubarev 1989). Reprinted with permission from Publishing House Nedra.

between consolidated rocks. Kimberlite sills were first described by Wagner (1914), who considered them as an integral part of a pipe-dike-sill system.

A summary of the geology of sills is provided by Hawthorne (1968), who described six kimberlite sills within a 64-km radius of Kimberley, South Africa. All of the sills intruded sedimentary rocks near the base of Karroo system. The sills were intruded more or less contemporaneously with neighboring dikes and pipes. The sills within the Kimberley area range in thickness from 0.9 to 1.2 m but have great lateral extent—about 3.2 km^2. Although rare, sills can be up to several hundreds of meters thick, and some show transitions from sills to laccoliths.

The Wesselton Floors sill in the Kimberley area is about 45 m thick in its central part but dies out marginally, continuing into the country rock as thin veinlets (Hawthorne 1968) (Figure 6.12). The Wesselton Floors sill actually forms a doubly convex laccolith with an east–west extension of 312 m and an inferred breadth of 190 m. The central part of this body is at least 44 m thick.

The rock in the kimberlite sills usually resembles rock found in the dikes and pipes. However, two sills of magmatic, nonbrecciated kimberlite have been described in the Benfontein and Karolusdrift area. These sills are composed of massive kimberlite enriched in CO_2.

Many sills are developed along contacts between rocks exhibiting different degrees of consolidation. For example, the Benfontein and Karolusdrift kimberlites were emplaced along the contact between basement rocks and overlying unconsolidated Mesozoic sediments. A similar position is occupied by the Mela sill in the Arkhangel'sk area. This sill was emplaced between different suites of Paleozoic country rocks. However, a sill mapped in the Iron Mountain district of Wyoming is relatively narrow (about 1.8 m thick) and was emplaced in the granite country rock, which forms both the footwall and hanging wall for the sill (W.D. Hausel, personal field notes, 1998).

The section through some of the kimberlite diatremes in the Bakwanga area of Zaire show flat-lying tabular bodies up to several hundred meters thick that extend radially

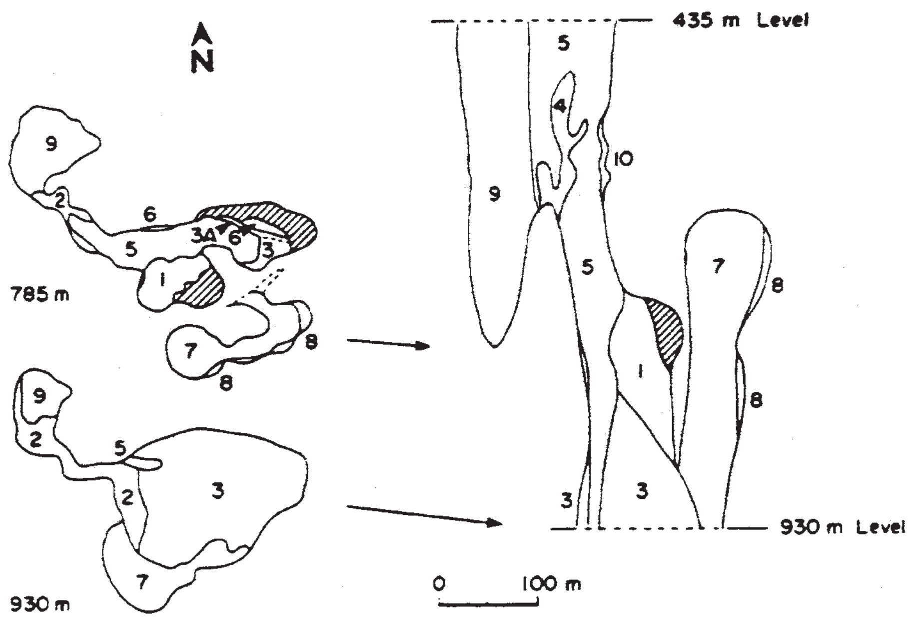

FIGURE 6.12 Vertical and horizontal cross section of the root zone of the Wesselton diatreme (Mitchell 1986). Reprinted with permission from Kluwer Academic/Plenum Publisher.

several thousand meters beyond the rims of the true kimberlite body. Wagner (1914) described a kimberlite dike as a feature from which minor sills projected into the well-stratified shale wall rock in the Kamferdam mine, Kimberley. Hawthorne (1968) showed that the Saltpeterpan sill south of Kimberley extends from north-south–striking dike.

Some kimberlite pipes in Bakwanga area (Democratic Republic of Congo) have some sill-like features near the present-day surface. These sills are represented by flat-lying tabular bodies up to several hundred meters thick that extend a few thousand meters from the pipe. These sills were intruded near the base of the Mesozoic sediments, which overlie rocks of Precambrian age. Hawthorne (1968) cited work of Meyer de Stadelhofen, who described them as sills developed during the Mesozoic period as a kimberlite was squeezed horizontally into recently deposited, partly consolidated sediments.

Hawthorne (1968) also reported that sills were exposed at the present-day land surface. Below this surface, major pipes such as Kimberley and Wesselton were mined to a depth of more than 700 m before the feeder dikes were reached. Such facts indicate that the feeder dikes for these sills penetrated to a much higher level than the levels typical for diatremes.

Hawthorne (1968) also reported that “no kimberlite sills have been found to constitute an ore body, although most do contain diamonds.” Diamondiferous kimberlite sills were discovered in the Quebrado Grande, a tributary to the Guaniamo River of Venezuela (Kaminsky et al. 1997). Diamonds recovered from the sills and nearby placers are identical.

Root zone. The root of a diatreme extends upward from a feeder dike grading into the diatreme (pipe). The rock in the root typically consists of massive, hypabyssal facies kimberlite with some brecciation.

The dike complex at Iron Mountain, Wyoming, represents one of North America's better exposures of a deeply eroded feeder-dike complex. In its present configuration, the complex appears as an anastomosing dike complex with sporadic enlargements containing zones of brecciation. One of these enlargements contains lower Paleozoic xenoliths indicating that this kimberlite originally erupted at the Paleozoic surface, and much of the diatreme was later removed by erosion (Smith 1977). At some other enlargements, kimberlite breccias are suggestive of root zones of blind diatremes that lost gas pressure when other diatremes along the complex reached the Paleozoic surface. These blows contain abundant calcium carbonate and granite xenoliths but no evidence of Paleozoic xenoliths, indicating that they never reached the surface. Other dike enlargements exhibit no evidence of brecciation. They probably represent only dike enlargements and are not blows of diatremes.

Roots zones are also described by Shee (1979), Shee, Gurney, and Robinson (1982), and Pasteris (1983) from the DeBeers and Wesselton pipes, and by Clement (1979, 1982) in Kimberley. The cross section of the root zones changes rapidly within a relatively short depth. The short axis of the St. Augustine mine contracts from a circular diatreme at the surface to a narrow fissure within a depth of about 250 m. The root zone was encountered 100 m below the present surface. The root zones of the larger pipes in the Kimberley area persisted for depths up to 500 m. Root zones are characterized by the dominance of hypabyssal facies over diatreme facies kimberlite (Hawthorne 1975; Clement 1979, 1982).

Nixon (1995) notes that dikes can be used as pathfinders to pipes or blows. From a feeder dike, the base of a pipe extends upward. The blow is irregular to lens shaped and produces an irregular enlargement along the dike. The kimberlite forming the blow consists of one or more hypabyssal facies kimberlite and kimberlite breccia. The hypabyssal facies is a porphyritic rock formed primarily of rounded olivine megacrysts in a highly serpentinized, massive, fine-grained matrix. The classic kimberlitic indicator minerals and megacrysts are less common in the hypabyssal facies kimberlite, and mantle xenoliths may also be less abundant.

On the basis of the characteristic elongation of the root-zone diatreme compared with the circular morphology of the upper portion of the diatreme, McCallum and Mabarak (1976) and Mitchell (1986) concluded that elongated kimberlite diatremes resulted from erosion of the upper parts of kimberlite pipes.

Offshoots of main pipes that are roofed by country rocks are termed blind pipes (McCallum 1976). They are filled with hypabyssal kimberlite and have contact breccias at their upper margins. Whitelock (1973) described a blind extension composed of xenolith-free hypabyssal kimberlite within the Monastery pipe. The core is surrounded by kimberlite with xenoliths of sedimentary rock that increase in quantity upward until only undisturbed sedimentary rocks are found. Outside this region the sediments have been injected by a stockwork of kimberlite veins. McCallum (1976) and Clement (1979, 1982) consider blind pipes to represent incipient diatremes that have become isolated from the main pipe owing to surface breakthrough of other portions of the system.

ORIGIN OF KIMBERLITE

The origin of kimberlite as well as diamond is closely interconnected with the origin of diamondiferous nodules (eclogites and peridotites). Peridotitic and eclogitic nodules found in kimberlite may be the result of immiscibility of melts (mafic and ultramafic) under conditions of high pressure and fluid saturation. This conclusion is suggested by the different compositional trends of the minerals found in these contrasting rock types (Figure 6.13). On the basis of the compositional variations of clinopyroxene and garnet, it has been suggested that the earth's mantle has separated into layers of calcium- and sodium-rich eclogite and calcium- and sodium-impoverished peridotite (Marakushev et al. 1995). It has also been suggested that the peridotitic and eclogitic nodules are sampled by the kimberlite magma and brought to the surface. These nodules are then regarded as fragments of the original ultramafic mantle (peridotite) and subducted oceanic material (eclogite).

Some of the most informative data on the origin of kimberlite is derived from a suite of mineral inclusions in diamond that provide clues to the initial stages of magma generation and differentiation. These inclusions belong to two types of paragenesis that correspond with peridotites on one hand and pyroxenites-eclogites on the other. The role of this suite of inclusions and discussions about their origin have been presented in works by Sobolev (1977), Gurney (1984), Sobolev et al. (1989), Garanin et al. (1991), Kirkley et al. (1991), and Bulanova et al. (1993) among others.

One of the basic facts about this suite is that the two types of mineral associations are rarely mixed (Meyer and Tsai 1976; Sobolev et al. 1980). Combinations of these two mineral associations as inclusions within a single diamond crystal are rare (Gurney 1986; Richardson et al. 1990; Garanin et al. 1991).

Figure 6.14 shows the different compositions of garnet and clinopyroxene diamond inclusions and the established paragenetic groups. Groups I through III represent diamonds of ultramafic-peridotitic paragenesis, whereas field III corresponds to the composition of ultramafic pegmatites (megacrysts) associated with very large diamond crystals. The regularities of the compositional variations in the garnets and pyroxenes are determined by the early crystallization of chromium-rich pyrope with a high (15%–20%) knorringitic component in the core of the mineral.

These changes are typical of magmatic crystallization—early extraction of chromium from the magma and incorporation into chromite and pyrope and the impoverishing of the magma in these components as reflected in the rims of garnet and diopside. Owing to calcium accumulation in the residual melt, calcium is sharply increased in the outer rims of garnet. With the subsequent crystallization of clinopyroxene, calcium is further extracted from the melt. Garnet and the associated clinopyroxenes are consistently impoverished in calcium, which corresponds to the transition from Group II to Group III (curve C-R, Figure 6.14). Such a transition is in accordance with other mineral inclusions (Bulanova et al. 1993): Ni-rich pyrrhotite, chromite, enstatite, and chrome-diopside in the core (C); and ilmenite, diopside, and potassium minerals (phlogopite and jerfisherite) within the rim (R) of diamond crystals.

Within the upper mantle, it is thought that there is a bimodal segregation of the rock-forming elements into two magmas—a relatively reduced ultramafic component with relatively high hydrogen content, and a relatively oxidized mafic component with a relatively low water content. This difference in deep-seated magma chambers provides contrasting environments for the formation of diamond (Marakushev et al. 1995).

Analysis of the isotopic equilibrium of different gases such as CO, CH_4, and CO_2 show that increasing oxidation increases the concentration of heavy carbon isotopes in the fluids CO, CH_4, and CO_2. It is suggested that through the following gas reactions, the carbon isotopes will become heavier during the process of diamond formation: $CO + H_2 = C_{(diamond)} + H_2O$; $CH_4 + 2CO = 3C + 2H_2O$; and $CH_4 + CO_2 = 2C + 2H_2O$ (Marakushev et al. 1995).

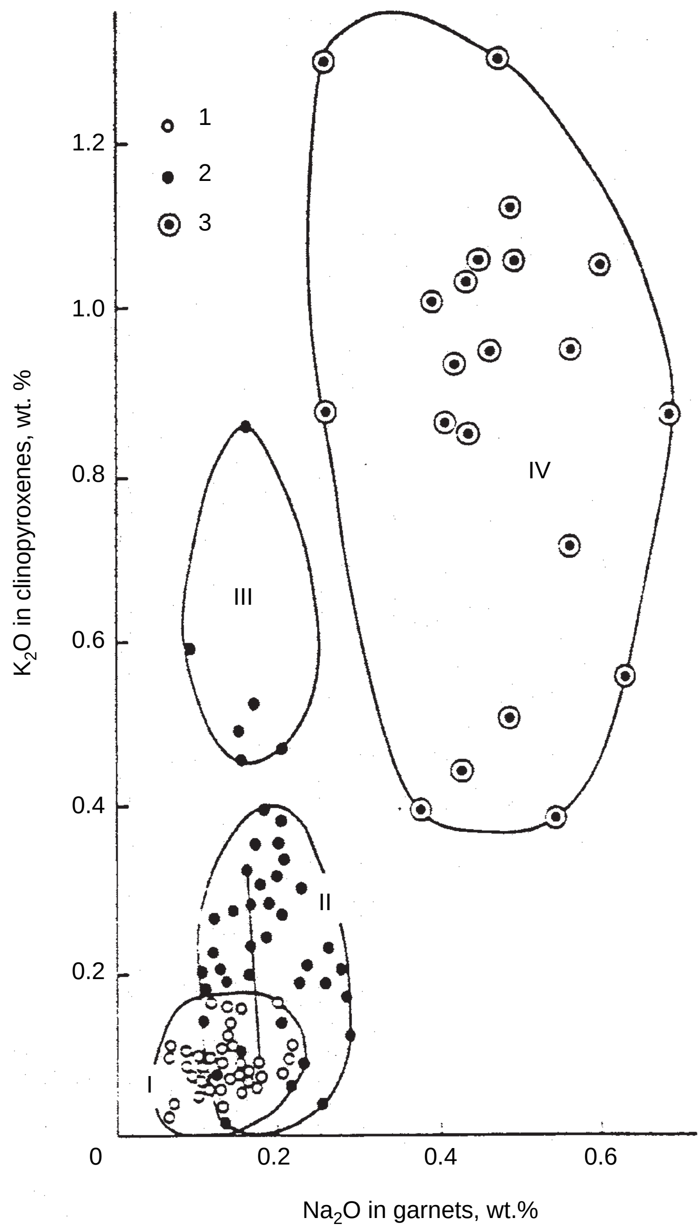

FIGURE 6.13 Characteristics of diamond-bearing xenoliths by content of admixtures of alkaline elements in their rock-forming minerals. Bimodal distribution of mineral associations is a result of suggested initial magma's layering into eclogite and peridotite. Rings I, II, and III = kimberlite pipes: 1 = diamond-bearing eclogites, 2 = mineral inclusions in diamonds; Ring IV = lamproitic pipes: 3 = mineral inclusions in diamonds. Straight line (bottom) shows typical relationship between the composition of minerals in diamond-bearing eclogite and composition of mineral inclusions in diamonds (Marakushev et al. 1995). Reprinted with permission of A.A. Marakushev.

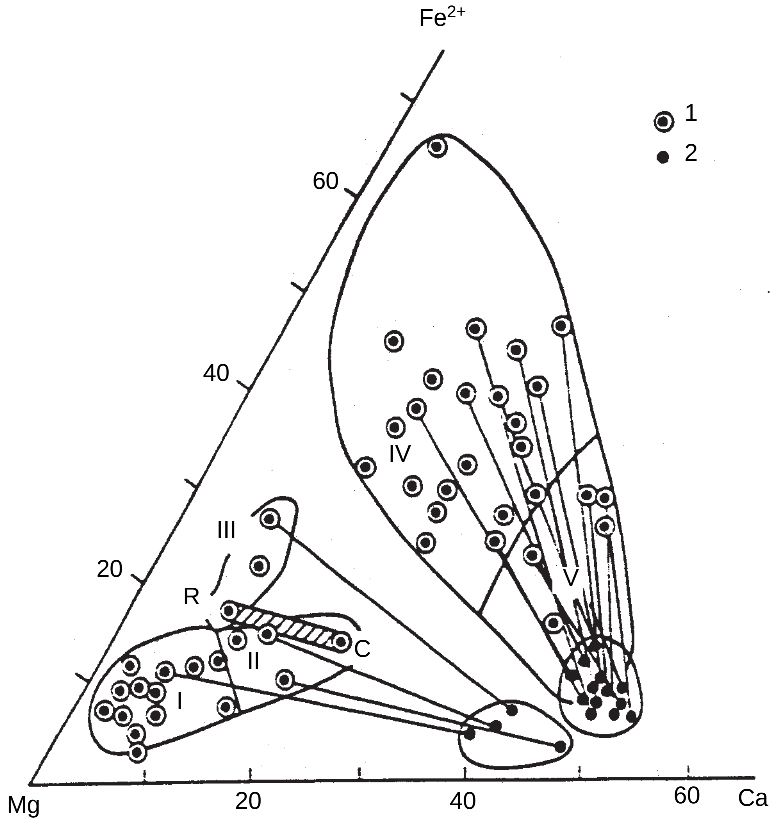

FIGURE 6.14 Types of diamond recognized by garnet and clinopyroxene inclusions, which characterize composition of original rocks: Ring I = dunite-harzburgite, Ring II = lherzolite-wehrlites, Ring III = pegmatite (recognized by garnets from megacrysts), Ring IV = clinopyroxene-eclogitic, Ring V = kyanite eclogites. Shaded band = changing of composition within zonal garnet from the core (C) toward the rim (R), which characterizes the transition from peridotites to enstatite pegmatites. Legend: 1 = garnet, 2 = clinopyroxene (Marakushev et al. 1995). Reprinted with permission of A.A. Marakushev.

Discrete changes in the isotopic compositions reached ^{13}C several per mil (6–10‰) as has been shown by the study of diamond-in-diamond structure from the Udachnaya pipe (Galimov et al. 1990). The natural evolution of diamond in eclogitic systems resulted in positive values of $^{13}C = 2‰$, but in peridotites it is limited to $^{13}C = -1‰$. This difference supports a higher-oxidation environment for fluids in diamonds of eclogitic origin in comparison with diamonds of peridotitic origin.

Constant and regular combination of peridotitic and eclogitic magmatic rocks along with systematic variations of composition of mineral phases led to a hypothesis of a immiscibility of ultramafic and mafic melts under high pressure and considerable fluid saturation (Marakushev et al. 1995).

Marakushev et al. (1995) cited the presence of widespread eclogite (griquaitic), peridotite massifs within the Bohemian Precambrian massif. Layers of Cr-rich dunites with pyrope containing as much as 8% Cr_2O_3 indicated that the early stage of crystallization of this mineral occurred within the diamond stability field. Another example of a potentially diamondiferous peridotite massif is the layered massif of Beni Bousera, Morocco, which is concordant with the original layering of Paleozoic gneisses and crystalline schists, along with which it forms an overturned anticline. This massif is composed of peridotite interlayered with garnet pyroxenite and eclogite containing graphitic pseudomorphs after octahedral diamond. The former diamondiferous layers contain omphacite (15%–18% jadeitic component) and almandine-pyrope garnet (48%–54% pyrope) and also contain olivine, enstatite, ilmenite, and spinel (see Section 3, Unconventional Source Rocks).

In any discussion of the genesis of kimberlite, it is important to recognize the magmatic differentiates of kimberlite magma. It is necessary to stress two basic facts. The first is the sedimentation of phenocrysts observed in kimberlitic sills, which produces layers enriched in olivine. The second is the squeezing of interstitial carbonate into adjacent layers (Dawson and Hawthorne 1973).

A question that arises is whether or not the carbonate in these rocks is an actual carbonatite. This question is important for any discussion of relationship between carbonatites and kimberlites (see Mitchell 1986). For the purpose of discussion, it is necessary that the amount of interstitial carbonate in kimberlites is proportional to the amount of "precipitated" heavy minerals.

A review of data on the petrologic evolution of kimberlite suggests that it can be approximated by a series of immiscible melts. Liquation is accompanied by sedimentation of comparatively heavy silicate minerals and the squeezing out of some interstitial liquid. At the earliest stage of magma generation, the formation of independent (i.e., immiscibile) eclogitic and peridotitic melts is expressed in two silicate melts (mafic and ultramafic). The next stage is characterized by separation of ore-enriched magma in the melt (ilmenitic nodules, apatite-magnetite melts in some ultramafic-alkaline melts). During the latest stage of evolution, liquids enriched by magnesium-rich minerals separate, and the carbonatitic melt is squeezed out.

An important insight in the evolution of kimberlitic magma is provided by ilmenite. It is of importance because it potentially permits one to trace the evolution of the effects of oxygen, the evolution of certain heavy metals including Ti and Cr, and the effects of diamond mineralization.

The chemical evolution of ilmenite in kimberlite has been described by Garanin, Kudryavtseva, and Soshkina (1983). In total, five chemically distinct groups of ilmenite have been recognized (Figures 6.15 and 6.16). Ilmenite of Group I includes discrete grains of more than 100 μm in diameter. This group includes most ilmenite nodules as well as grains from ultramafic rocks that are characterized by a systematic change in the hematite component from harzburgite to lherzolite to diopsidite. This systematic trend is interpreted as reflecting changes in pressure and temperature.

Ilmenite of Group II consists of grains that are enriched in Cr_2O_3 and are relatively high in MgO but very low in Al_2O_3. A partial overlap of ilmenite compositions of Group I and Group II is explained by Group I ilmenites with rims having compositions similar to those in Group II.

Group III ilmenite is characterized by a low Cr_2O_3 and high $FeTiO_3$ component (75–100 mole %). Ilmenite of this group is subdivided into two subgroups with clear differences in the hematite and manganese content. Subgroup IIIa includes ilmenite from

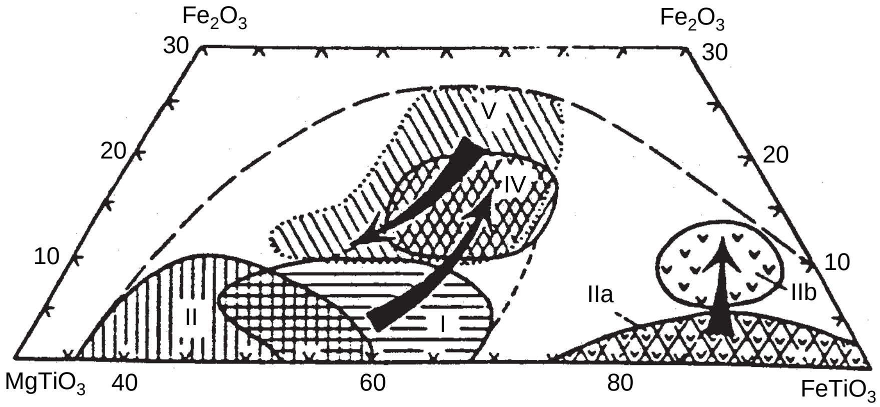

FIGURE 6.15 Compositions and main crystallization trends (arrows) of ilmenite from kimberlite bodies of the Yakutian diamond province (Garanin et al. 1983). Reprinted with permission from the Doklady Russian Academy of Science.

eclogitic xenoliths. A detailed study of the composition of ilmenite from diamondiferous kimberlites of the Malaya Batuobiya field shows that Group III ilmenites occur only in isolated grains. However sizable concentrations of ilmenite of this group were found in the kimberlite matrix in diamond-free pipes of the Lower and the Middle Kuonamka fields.

Ilmenite high in $FeTiO_3$ (>10 mol %) that is ferromagnetic at room temperature can be clearly subdivided into two groups (IV and V). Group IV includes ilmenite with a low Cr_2O_3 content (rarely exceeding 0.5 wt. %) and variable hematite component (10–19 mol %). Ilmenites of group IV have not been found as inclusions of plutonic, ultramafic and mafic rocks or among small ilmenite crystals of the kimberlite matrix.

In the composition field for group V ilmenite are all analyses from ilmenite nodules or fragments in kimberlite genetically related to garnet-ilmenite intergrowths.

The formation of ilmenite-bearing rocks can't be considered separately from the process of generation and evolution of kimberlite as a whole. Established systematic change in ilmenite composition in ilmenite-bearing rocks from the Mir pipe as well as the composition of silicate minerals in these rocks reflects the evolution of ultramafic and mafic magma. The evolution is in accordance with the mechanism of kimberlite pipe formation proposed by Marakushev (1982).

It is important to recognize that ferromagnetic ilmenites high in admixed Cr_2O_3 and Group V commonly show exsolution structure. This structure is also characteristic of ilmenite from diamond-poor kimberlite bodies, whereas the same mineral in pipes with higher diamond content is mainly homogenous.

The difference can be explained by different degrees of diamond preservation that are related to the rate of kimberlite magma ascent to the surface. Slowly rising magma will oxidize (burn) diamond to CO_2, whereas a rapid rise will preserve diamonds (as long as the magma originated within the diamond stability field). Therefore, the presence of ilmenite with conspicuous exsolution texture in kimberlite bodies or nearby placers may be regarded as a negative test for diamond preservation.

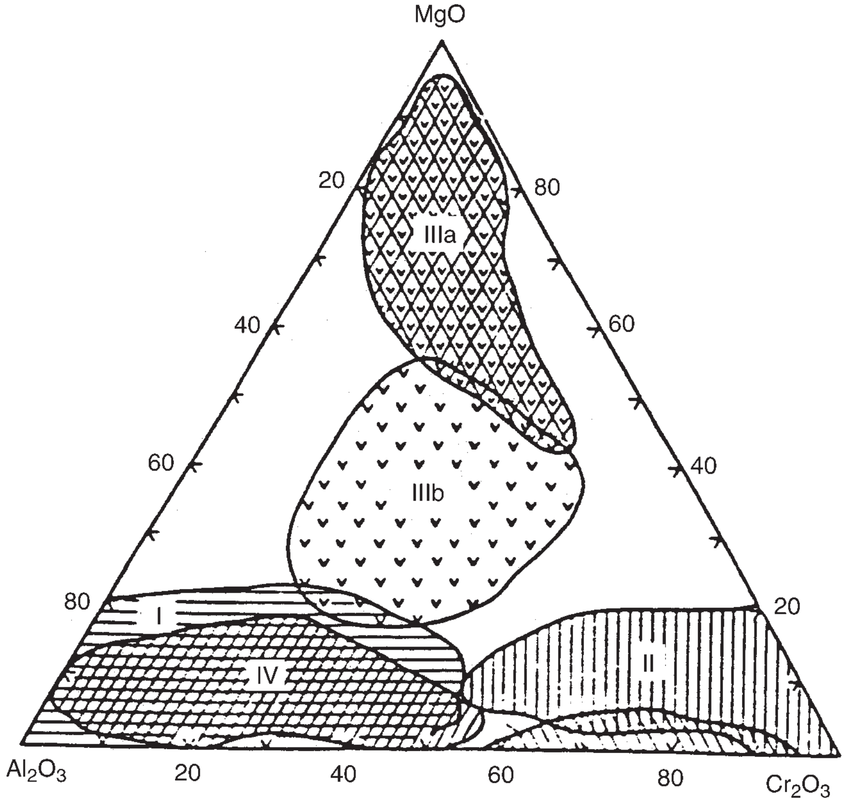

FIGURE 6.16 Ratio of Cr_2O_3, Al_2O_3, and MgO in ilmenite from kimberlite bodies of the Yakutian diamond province (Garanin et al. 1983). Reprinted with permission from the Doklady Russian Academy of Science.

Probably the most important change in physicochemical conditions in an ascending kimberlite column is related to self-oxidation of the volatile "bubble." It is commonly accepted that kimberlite magmas initiate by a process of fluid diapirism. This fluid is chiefly composed of hydrogen and methane that were accumulated in the mantle under platform plates. During ascent of the magmatic column, if the pressure of the magmatic system drops below 20 kbar, the oxygen fugacity will increase resulting in the oxidation of the fluid to CO, CO_2, and H_2O (Portnov 1982; Figure 6.17). However, rapid emplacement of the magma at pressures lower than 20 kbar may preserve diamonds.

Near the surface, water is absorbed by the ultramafic magma during serpentinization. Owing to hydrogen and methane oxidation, the volume of the parent material may decrease.

It appears that the main process in the evolution of kimberlite magma is related to magma layering. At different stages the layering takes on different chemical forms, and the magma is separated into two immiscible melts (liquation). Magmatic sedimentation of heavy crystals and squeezing will separate a portion of magmatic liquid from the earlier-crystallized layer. Deep in the upper mantle, the stratification of melts into two parts with different silica contents resulted in eclogite and peridotite layers. In the previous section

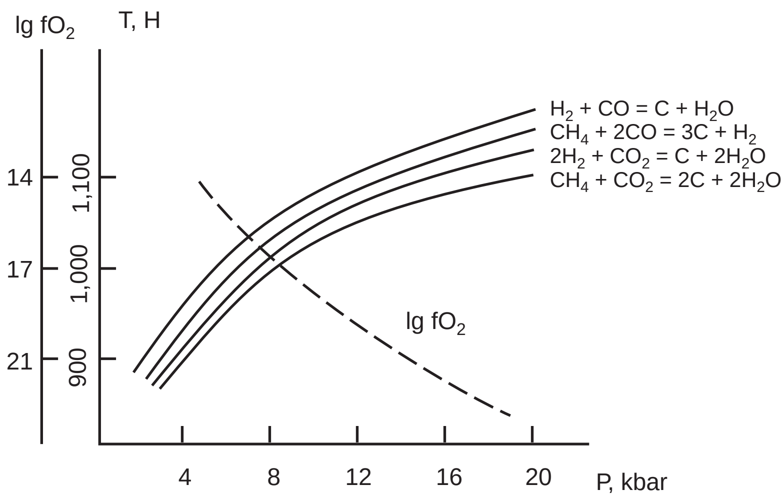

FIGURE 6.17 Increasing O_2 activity with decrease of pressure and approximate parameters of diamond formation within the system C-H-O-S (Portnov 1982). Reprinted with permission from the Doklady Russian Academy of Science.

we attempted to show that these two rock types were probably formed as a result of immiscibility at the intratelluric stage of magma evolution (Marakushev et al. 1995).

The separation of the rock-forming elements in the original ultramafic melt is accompanied by contrasting distribution of the fluid components in two different types of diamond crystals, eclogitic and peridotitic (Figure 6.18). The next stage of magma layering probably occurs at plutonic stage. The primary process at this stage is a separation of an ilmenite-rich magma from the silicate melt. The third and final step of kimberlitic magma evolution occurs during the hypabyssal stage. At this stage, following the sedimentation of the heavy mineral phases, a carbonate-rich liquid is squeezed and separated, possibly resulting in both kimberlite and carbonatite magmas.

ORIGIN OF DIAMOND

In considering the origin of diamond, one must determine the primary source of diamonds in kimberlite. Do they have a xenolithic origin, or are they the result of crystallization of kimberlitic magma itself? Diamonds are obviously present in both eclogitic and peridotitic suites of xenoliths, but as it is properly argued by Dawson (1980), an unusually large proportion of diamondiferous xenoliths would have had to be preferentially fragmented to provide all diamonds that occur in kimberlite.

Several aspects must be considered, including the following:

- the diamond distribution in kimberlite
- features of diamond morphology
- the connection between diamond's crystallographic features and the mineral inclusions in the diamonds

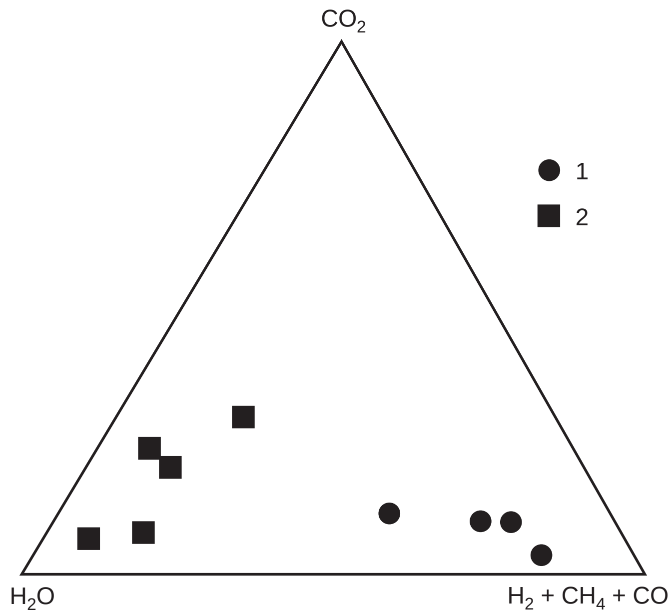

FIGURE 6.18 Compositions of fluids typical of eclogitic (1) and peridotitic (2) diamonds (Marakushev et al. 1995). Reprinted with permission from the Doklady Russian Academy of Science.

- existence (or the absence) of correlation between diamond content and the composition of the megacryst/macrocryst suite
- the correlation (or the absence of correlation) between diamond content and the chemical composition of kimberlite.

Sobolev (1960), in his review of diamond deposits, stressed that there is no tendency for the diamond content to decrease at depth within a diatreme. Additionally, the size of diamonds, on average, is greater than the size of the microcrystals in the kimberlitic groundmass, which reflect the latest stage of kimberlitic magma crystallization.

Dawson (1980) listed several features that support a phenocrystal origin for diamond:

- the delicate growth features on diamond crystals, which could have been derived only from crystallization from a liquid. It is acknowledged that delicate growth features have also been found in diamonds from eclogite xenoliths; they are due to growth of diamond from a much older liquid unconnected with the kimberlite event.
- the presence of submicroscopic liquid inclusions of picritic composition in diamonds
- a disproportionate amount of diamondiferous peridotitic and eclogitic xenoliths in comparison with nondiamondiferous xenoliths and the diamond content in host kimberlites. Both imply preferential treatment of diamondiferous xenoliths.

Plastically deformed diamonds in some kimberlites provide evidence that diamond has had a long and complex history. This fact, together with evidence that most diamonds yield a much older radiometric age than their host kimberlite, supports the concept that the majority of, if not all, diamonds are xenocrysts rather than phenocrysts.

The mineral paragenesis of diamond inclusions are clearly divided into two groups, which correspond with pyroxenite-eclogites and peridotites (Meyer and Tsai 1976; Sobolev et al. 1980).

Harte, Gurney, and Harris (1980) cited unpublished work of Hawthorne showing that diamonds from a given kimberlite may display a marked predominance of inclusions of one suite that does not correlate with the predominant type of xenolith recovered from the same locality. Whereas peridotitic-suite inclusions are more abundant than eclogite-suite inclusions, peridotitic xenoliths containing diamonds are rare and diamondiferous eclogites are relatively common.

The peridotitic suite of mineral inclusions in diamond show a trend that is marked by lower Ca, higher Cr (and lower Al), and Mg/Mg + Fe ratios in the higher part of the range shown by nodules.

In addition, any hypothesis that proposes that diamond crystallized from kimberlite magma must explain the absence of inclusions or intergrowths of a high-Ti megacryst suite in diamonds or, conversely, it must deny that megacryst suite phases are connected with the kimberlite event.

It is necessary to stress the importance of establishing a dependence of diamonds on the chemical composition of kimberlites and mineral indicators of the megacryst/macrocryst suite. For example, specific kimberlitic mineral indicators with favorable geochemistry (high-Cr, low-Ca pyropes and specific types of picroilmenite) can be used as direct exploration indicators of diamond. As a result, electron-microprobe analyses of these minerals from panned samples are commonly obtained in the early stages of exploration and assessment of regions for their diamond potential (Dawson 1980; Mitchell 1986).

Along the same lines lies the idea of correlating the kimberlite composition with diamond content (Milashev 1965). Milashev suggested that the polymorph transformation of diamond to graphite within a melt is a form of catalytic process. The catalyst is related to metatitanium acid ions, which are constantly present in kimberlitic magma and later result in the formation of titanium acid salts (titanates): ilmenite ($FeTiO_3$) and perovskite ($CaTiO_3$). The catalyst follows this chemical reaction:

$$\begin{matrix}\text{diamond} \\ C + [Ti^{4+} + O_3]^{2-} \rightleftarrows CO + [Ti^{3+}O_2]^{-} + e \\ \text{graphite}\end{matrix} \quad \textbf{(EQ 6.1)}$$

The action of the catalyst tears carbon atoms from diamond and returns them back to the diamond. If the pressure in the melt remains high and diamond is stable, the anions of the metatitanic acid may support the growth of large crystals. But when the pressure drops, diamond is unstable and the carbon atoms will tear off under the influence of metatitanic acid $[Ti^{4+} + O_3]^{2-}$ and will form graphite, which is stable under the lower P–T conditions.

Milashev (1965) worked out a series of formulas using a diamond potential coefficient. They are based on the idea that the diamondiferous potential (DBP) of a kimberlite results from two interrelated factors: the chemical potential of diamond formation (CPD), and the degree of preservation of the crystals (DPC). (For a further discussion of these formulas see Chapter 16, Predicting Diamond Content).

In this study, the chemical composition of various kimberlites was compared with the sampling results of different pipes for diamond content, and the chemistry showed a correlation with diamond content. The dependence between these two parameters was noted, and it can be calculated using the Ti, Fe, Al, and K content. The data permits one to determine the degree of a melt's favorability for diamond's crystallization and consequent preservation of diamond crystals.

$$\mathrm{CPD} = \frac{\mathrm{Fe/Ti}}{\log(\mathrm{Fe} + \mathrm{Ti}) + 0.5\ \log(\mathrm{Al} + \mathrm{K})} \quad \textbf{(EQ 6.2)}$$

Another formula worked out by Milashev provides a measure of volatiles in the diamond-formation process. It is based on the idea that diamond forms (shapes) depend on a combination of two processes: growth and dissolution. If the degree of the diamond's crystal preservation depends on the P–T conditions of the formation of kimberlite, a quantitative proportion of flat-facet transitional and curved-facet crystals can serve as a indicator of the P–T conditions of diamond formation. In order to calculate the DPC using a simplified formula, it is sufficient to know the ratio between octahedral, dodecahedral, and transitional forms of diamonds in the deposit under consideration.

The resulting formula is as follows:

$$\mathrm{DPC} = o + 0.45od + \frac{0.45d}{1 + \log(d)} \quad \textbf{(EQ 6.3)}$$

Where o = percentage of flat-faced octahedral crystals; d = percentage of rhombic dodecahedra; od = percentage of transitional forms, and log(d) = percentage of dodecahedral diamonds in logarithmic form. The diamond-bearing potential is then expressed by the formula DBP = CPD × DPC.

These formulas generated different reactions: on one hand they were welcomed as a major achievement that directly and on a quantitative basis connected the kimberlite chemistry and the degree of diamond mineralization (Dawson 1980). On the other hand, Mitchell (1986) cautiously suggested that the formula in EQ 6.3 is proven only for Siberian pipes and has to be confirmed using a worldwide database. A practical problem for the formula is simply that a very large bulk sample of kimberlite must be processed in order to acquire a sufficient number of diamonds to adequately determine the ratio of the forms of diamond, and each phase of kimberlite must be tested separately.

Another characteristic feature of the crystallization of kimberlite magma is the abundance of breccia textures. Some breccias were apparently formed within the diamond stability field, and they contain fragments of diamond as well as kimberlitic indicator minerals within larger diamonds. The diamond fragments in such breccias are always characterized by lighter isotopic compositions in comparison with mineral inclusions. All of them indicate a series of breccia and explosive events during the earliest stages of magma evolution (Galimov 1984; Galimov et al. 1990).

It is signficant that a diamond recovered from the Mir pipe was described as containing a fragmented pyrope crystal (Garanin et al. 1991). This breccia texture is interpreted to be the result of explosive processes in the magma chamber at depth within the diamond stability field (Marakushev et al. 1995).

The concept of stratification of the initial parental magma into two immiscible melts, one peridotitic and the other eclogitic, should provide clues that are needed to resolve most contradictions related to the diamond-kimberlite connection, and it should also explain the existence of two suites of diamonds (E-type and P-type) that have different mineral inclusions and different ages. On the basis of the model, both types of melts are directly related to kimberlite and developed simultaneously at the intratelluric phase of kimberlite evolution.

However, there is the possibility that when the two melts separated, one initially crystallized while the other remained at depth. This possibility can explain the difference in radiometric age between E-type and P-type of diamonds. Kimberlitic magmas originating from extreme depths, probably within the middle mantle with fluids originating at the core/mantle boundary, make it possible for melts to exist at depth for a long time (up to 1.2 billion years—a period that covers the suggested age difference between diamonds and their host kimberlites).

Conversely, some researchers advocate that the E-type diamonds are derived from the compaction of organic material subducted to mantle depths during movement of tectonic plates, a concept that is supported by carbon isotope signatures, and that the host eclogites represent a high P–T transformation of oceanic basalts. P-type diamonds, which provide a consistent and narrow range of ages and carbon isotopes, are associated with peridotites believed to have been stored in cratonic keels that have not be actively involved in plate tectonic motions for at least 1.5 billion years.

In summary, diamonds have formed during a considerable length of geologic time (<1 billion years to 3.3 billion years) and typically are older than the host kimberlites. P-type diamonds provide older ages than E-type diamonds. The two schools of thought indicate that diamonds are either phenocrysts and formed within the kimberlite magma, or they are xenocrysts and were accidentally trapped in the kimberlite magma. In most cases, the age difference between the diamonds and the host kimberlite support that latter, and only at the Premier pipe in South Africa do the kimberlite and diamonds have contemporaneous ages (Kirkley et al. 1991).

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CHAPTER 7

Lamproites

INTRODUCTION

Interest in lamproites intensified following the discovery of a world-class diamond deposit in olivine lamproite in 1979, in the Kimberley region at Argyle, Western Australia. This discovery led to the recognition of other diamondiferous lamproites in Australia, Zambia, Ivory Coast, India, Russia, and the United States, some of which had been originally classified as kimberlite. More recently, a rich source of microdiamonds was reported in a dike resembling lamproite west of Hudson Bay, Canada. Other lamproites have been identified in Spain and Antarctica. In the United States, lamproites have been identified in Wyoming, Montana, Utah, Kansas, and Arkansas (Mitchell and Bergman 1991; Hausel, Love, and Sutherland 1995; Hausel, Sutherland, and Gregory 1995, 1998). Other lamproites have been reported in Colorado, California, and South Carolina, although these latter occurrences have not been verified. Of the known lamproites, the olivine lamproites offer the highest potential for economic concentrations of diamond. In 2001 about 40 lamproite fields were known in the world, but only a small number were diamondiferous lamproites.

As early as 1967, altered diamondiferous leucite lamproite (fitzroyite) had been described near Seguela, Ivory Coast, by Dawson (1967), but little attention was given to it. In addition, diamonds were recovered from the Majhgawan lamproite in India as early as 1827. It was only recently that this vent was shown to be hyalo-olivine lamproite lapilli tuff (Scott-Smith 1989). Currently there are seven known diamondiferous lamproite provinces, fields, or deposits in the world. These include the Ellendale field and Argyle lamproite in Western Australia; the Prairie Creek, American, and Kimberlite lamproites in the Murfreesboro, Arkansas, areas of the United States; the Bobi lamproites near Seguela of the Ivory Coast, Africa; the Kapamba lamproite field in Eastern Zambia, Africa; the Majhgawan vent in India; and the Koromandel lamproites in the Minas Gerais region of Brazil. They range in age from Proterozoic (1.2 Ga) to Miocene (20–22 Ma) (Mitchell and Bergman 1991).

Early studies of lamproites recognized their unique mineralogy and chemistry. In 1871, Samuel F. Emmons of the Fortieth Parallel Survey described unusual leucite-bearing rocks in the northern Cascades and in the Leucite Hills of Wyoming (Emmons, 1877). Other descriptions followed, including a discussion of similar potassium-rich rocks in Spain (Osann 1906).

The term lamproite, meaning "glistening lamprophyre," was introduced by Niggli (1923) to describe this clan of unusual leucite-bearing lamprophyric rocks containing very high (>0.8) Niggli mg and k numbers, high K_2O content, and high MgO, and that were known only in Wyoming and Spain at that time. Since Niggli's work, lamproites have been found in more than 25 provinces or fields in the world (Mitchell and Bergman 1991).

Lamproites led an obscure life in academic studies until the discovery of diamondiferous lamproites in Western Australia in the 1970s, even though many years earlier Wade and Prider (1940) had suggested that there was a genetic link between lamproites and diamondiferous kimberlites. This link was also suggested in the later works by Prider (1960) and Carmichael (1967). Carmichael also implied a genetic link between kimberlite and lamproites on the basis of his research in the Leucite Hills of Wyoming. A small group of Australian geologists persisted in exploration programs that ultimately led to the discovery of new diamond deposits that were not associated with kimberlite, but which were later proven to be lamproite (Jaques et al. 1982, 1983; Atkinson, Smith, and Boxer 1984; Jaques, Lewis, and Smith 1986; and Boxer, Lorenz, and Smith 1989). These discoveries led to the development of exploration techniques specifically adjusted to search for this newly recognized host rock.

In the United States, Hausel persued a similar venture in the Leucite Hills in 1985 on the basis of Carmichael's (1967) study, but he was unsuccessful in finding diamonds in those lamproites (Hausel, Love, and Sutherland 1995). However, Hausel, Sutherland, and Gregory (1995) noted an increasing incidence of olivine-bearing lamproites in the northeastern portion of the Leucite Hills and proposed the possibility of hidden olivine lamproites along a major east-west–trending shear that cuts through the northern portion of the field. Although no diamonds have been recovered from the field, in 1998, more than 13,000 carats of detrital peridot and some industrial olivine were recovered from anthills along the northeastern edge of the Leucite Hills (Hausel and Sutherland 2000).

Because of its conviction that kimberlites were the single primary source rock of diamond, DeBeers rejected the Australian lamproites. The stunning success of the Australian geologists and companies led to the discovery of diamondiferous olivine lamproite in the Ellendale field and later led to the discovery a diamond deposit in olivine lamproitic tuffs at Argyle. That at Argyle is far richer than any known kimberlite; the average grade is 680 carats/100 tonnes, and some bulk samples contain as much as 2,000 carats/100 tonnes. Compared with the typical 20 to 200 carats/100 tonnes of many commercial kimberlites, the Argyle lamproite is a world-class diamond deposit.

The discovery of lamproite-related diamond deposits in tectonically deformed belts led to the assessment of other potentially new types of diamond deposits. The old rule that diamonds are restricted to kimberlites in stable cratons was completely shaken, and now serious exploration ventures began in geologic terranes and nonkimberlitic rocks that formerly were not considered for diamonds. Initially, assessments primarily reviewed localities with known diamondiferous rocks that traditionally had been referred to as kimberlite related. The best example was the reevaluation of the diamondiferous intrusives at Prairie Creek, Arkansas.

Diamonds were first discovered near Murfreesboro, Arkansas, in 1840 (Powell, 1842). Some unusual intrusives found in the area were referred as peridotites (Branner and Brackett, 1889). Diamonds were found in place in one of the intrusives in 1906, and the affinity of these rocks with kimberlites was suggested by Miser and Ross (1923). Another intrusive was even named the "American Kimberlite," because of the presence of diamonds. Only after the discovery of diamondiferous lamproites in Western Australia did Scott-Smith and Skinner (1982) undertake a detailed examination of those rocks to confirm their lamproitic character. Scott-Smith and Skinner (1984a, b) also showed that the tuffaceous rocks at Prairie Creek, Arkansas, were analogous to those at the Ellendale lamproite field in Australia. Later exploration programs resulted in legitimizing lamproite as a source of in situ diamond deposits (Smith-Scott and Skinner 1984a, b).

Intense exploration in Russia at this time resulted in several major discoveries within the East European craton. Detailed analysis of different types of diamond deposits within the Urals and adjacent to the Timan region led to the conclusion that some intrusives within these regions, and at similar sites in Karelia, and the Kola Peninsula, previously referred as kimberlite-related rocks could instead be related to lamproite. The exploration led to the discovery of diamond deposits with huge reserves in the Arkhangel'sk diamond province (Stankovsky, Verichev, and Grib 1979; Makhotkin and Skinner 1998). Lamproitic bodies were also found in the Urals (Luk'yanova et al. 1992, 1993; Ribal'chenko et al. 1997), Timan (Smirnov and Smirnova 1995), Karelia (Makhotkin 1998), and the Kola Peninsula (Arzamastsev, Polyakov, and Kalinkin 1995).

The recognition of lamproites as another potential host for commercial diamond deposits led to a flurry of academic studies intended to work out a more precise classification of lamproites and to establish clear boundaries between lamproites and kimberlites. This activity resulted in important research by Barton (1979), Scott-Smith and Skinner (1982, 1984a, b) and Mitchell (1985). Eventually, a nomenclature was developed for these rocks (Foley et al. 1987; Kogarko, et al. 1995). Unfortunately, research in Russia was not well coordinated with research in the western scientific world (Bogatikov, Makhotkin, and Kononova 1985).

Using the most rigorous definition of the lamproitic clan, the nature of some previously known diamond localities was reassessed and confirmed as lamproite. These localities included not only the intrusives at Prairie Creek, Arkansas, but also the first in situ diamond deposit found by mankind at Majhawan, India. The existence of lamproite has now been established on nearly every continent, and it is found within ancient cratons, along cratonized boundaries within orogenic zones (i.e., Rocky Mountains, Alps, Betic Cordillera, Urals), and within zones of buried aulacogens (deep-seated transform fault zones).

Some recently reported lamproitic rocks found in Pamir, Tajikistan (Dmitriev 1974), and the southwestern Kazakstan and Kyrgyzstan regions of Asia (Abdulkabirova and Zayachkovsky 1996) have received only cursory petrographic studies, but they are considered to be "lamproite-like" with lamproite affinity. They form diatremes, dikes, and rarely sills. On the basis of their chemistry they have leucite lamproite affinities, and they have proven to be diamondiferous. Their tectonic location is quite similar to that of lamproitic rocks within the western United States.

GEOCHEMISTRY AND MINERALOGY

Lamproites exhibit a very unusual mineralogy that may include diopside, phlogopite, K-Ti richterite, leucite, sanidine, wadeite, priderite, and/or olivine, with minor apatite, perovskite, ilmenite, and spinel (Mitchell and Bergman 1991). The typical "kimberlitic indicator" minerals (pyrope garnet, chromian diopside, and picroilmenite) are uncommon in lamproite. As stated by Mitchell and Bergman (1991), "The mineralogy of the lamproite clan is perhaps one of the most exotic of all alkaline rocks. The perpotassic and peralkaline traits of the magmas lead to a distinctive mineral assemblage that is virtually unparalleled." Nevertheless, an overlap in the chemical compositions between micaceous kimberlites (Group II) and lamproites has been noted.

Geochemistry

In general, lamproites are mafic, peralkaline, ultrapotassic igneous rocks enriched in the trace elements Zr, Nb, Sr, Ba, and Rb relative to kimberlite (Kirkley, Gurney, and Levinson 1991). Lamproites are silica poor (although some may be quartz normative)—SiO_2 (43%–55%). Relative to other mafic alkaline igneous rocks, they are distinguished by their high K_2O (3%–13%), TiO_2 (1%–7%), P_2O_5, MgO, and F, and low Al_2O_3 (<12%), CaO (<13%), and CO_2 (<1%). These rocks exhibit molar K_2O/Na_2O >5, which is ultrapotassic; molar K_2O/Al_2O_3 >6.8 (commonly >1; perpotassic); molar K_2O + Na_2O/Al_2O_3 >1 (peralkaline), and they are typically enriched in light rare earth elements (LREE) and incompatible elements such as Ba (>5,000 ppm), Rb (>150 ppm), Sr (>1,000 ppm), Zr (>500 ppm), and La (>200 ppm), as well as the compatible elements such as Cr and Ni (Kirkley, Gurney, and Levinson 1991; Smith-Scott 1996). Enrichment in Ti, rare earth elements (REE), Rb, Sr, Ba, and Zr coupled with high Ni, Cr, Co, and Sc contents imparts a distinctive geochemical signature.

Lamproites have high potassic content compared with Group I kimberlites (0.6%–2%) and Group II kimberlites (5%). They are considered hybrid mixtures of magmatic crystallization products and upper mantle xenoliths and xenocrysts. Lamproites show extreme enrichment in titanium compared to kimberlite. Kimberlites tend to have lower SiO_2, TiO_2, and alkalis and higher MgO than lamproites. In sharp contrast to kimberlite, lamproites contain little evidence of significant CO_2 content. Upper-mantle–derived xenocrysts or xenoliths (the megacryst series of kimberlites) may or may not be present.

Lamproites have elevated compatible elements Ni, Cr, Co, V, Sc, and incompatible Rb, Sr, Ba, Zr, Hr, Ti, P, Nb, REE, Y, Th, U relative to average crustal rocks. In a seeming paradox, these rocks have silica contents that range from ultrabasic to intermediate, but they are also enriched in K, Rb, and Ba elements normally concentrated in crustal-derived igneous rocks. Ni and Cr, which are depleted in crustal igneous rocks, occur in high concentrations in lamproites, whose chemistry is similar to the chemistry of many other mantle-derived rocks (Scott-Smith and Skinner 1982). Zr/Nb ratios in lamproites are comparable to those in oceanic basalts.

Although all mantle-derived alkaline magmas have been subjected to some type of metasomatic event, lamproites are unique in their extreme enrichment in elements added during metasomatic alteration, their extreme fractionation of K and Rb relative to geochemically similar Na (Sr and Ba to Ca), and their extreme radiogenic isotope (Nd-Sm) ratios, reflecting variable time-integrated enrichments with Rb/Sr > bulk earth and Nd/Sm < bulk earth (Mitchell 1985).

Lamproites are distinguished from similar rocks (such as minettes, leucitites, and kamafugites) on the basis of high weight percent K_2O/Na_2O (>5), high atomic K/Al (>0.8), and low CaO (<6 %). High amounts of Ti, Zr, and Nb produce a variety of accessory minerals such as perovskite, priderite, shcherbakovite, and wadeite, which are considered diagnostic of lamproite (Mitchell 1985).

Scott-Smith (1996) subdivided lamproites into two general groups: phlogopite-leucite lamproites (60% SiO_2), and olivine lamproites with abundant serpentine pseudomorphs after olivine: MgO (>20%), SiO_2 (35%–45%), and K_2O (7%). Both may be found in contact in the same vent. Glass in rapidly chilled lamproites can have SiO_2 >60%; quartz may even occur in lamproite melts crystallizing at high mantle pressures.

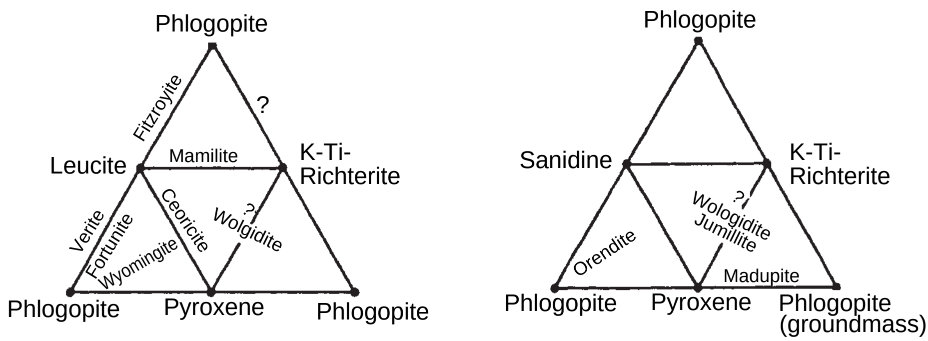

FIGURE 7.1 Nomenclature of lamproitic rocks (Mitchell 1985). Reprinted with permission of Geological Society of South Africa.

Mineralogy

Kimberlites and lamproites are distinctly different rock types and have petrographic and petrologic differences. They also have different indicator-mineral assemblages and mineralogy, and they experienced significant differences in their mode of near-surface emplacement as compared with kimberlite. Lamproites are quartz, acmite, hypersthene, and amphibole normative, and their extreme enrichment in titanium is expressed in the major and minor mineralogy.

Lamproites exhibit an extreme range in modal mineralogy. While having several major (olivine, diopside, phlogopite) and minor (enstatite, apatite, perovskite, ilmenite, and spinel) mineral phases in common with kimberlite, lamproite has a number of minerals that distinguish it from kimberlite. The most important are amphibole (K-Ti-richterite), leucite, sanidine, wadeite ($K_2ZrSi_3O_9$), and priderite [$(K,Ba)(Ti,Fe)_8O_{16}$]. Another characteristic of lamproite is the presence of matrix glass and relatively low $CaCO_3$ content. The phlogopite in lamproite occurs as phenocrysts and/or poikilitic groundmass grains and includes titanium phlogopite and titanium tetra-ferriphlogopite.

Minor and accessory mineral phases include enstatite, priderite, apatite, wadeite, manganochromite, Ti-Mg-chromite, scherbakovite, armalcolite, perovskite, ilmenite, jeppeite, and/or olivine. Analcime is common as a secondary mineral replacing leucite and/or sanidine. Other secondary phases include barite, carbonite, chlorite, and zeolites. Lamproites have not been found to contain kalsilite, melilite, nepheline, or plagioclase (Mitchell and Bergman 1991).

On the basis of the dominant minerals present (phlogopite, olivine, diopside, sanidine, and leucite), lamproites can be divided into several groups. The existing nomenclature of the lamproitic rock clan, on the basis of mineralogy, is shown in Figure 7.1 (by Mitchell 1985).

Noticeably nearly absent from olivine and phlogopite lamproites are megacrysts that are characteristic of most kimberlites, i.e., Ti-pyrope, picroilmenite, subalkalic diopside, bronzite, and zircon. Cr-diopside and enstatite xenocrysts are rare. If present they have compositions identical to minerals in lherzolite (Mitchell 1985; Mitchell and Bergman 1991).

Some lamproites may contain dog-tooth olivine, a rounded mantle xenocryst with faceted overgrowths of magmatic olivine. The xenocrystal olivines are identical to those

TABLE 7.1 Historical names and mineralogic lamproite classifications

Historical Name	Mineralogic Description
Cancalite	Enstatite-sanidine-phlogopite lamproite
Cedricite	Diopside-leucite lamproite
Fitzroyite	Leucite-phlogopite lamproite
Fortunite	Hyalo-enstatite-phlogopite lamproite
Gaussbergite	Olivine-diopside-leucite lamproite
Jumillite	Olivine-diopside-richterite-madupitic lamproite
Madupite	Diopside-madupitic lamproite
Mamilite	Leucite-richterite lamproite
Olivine orendite	Olivine-diopside-sanidine-phlogopite lamproite
Orendite	Diopside-sanidine-phlogopite lamproite
Verite	Hyalo-olivine-diopside-phlogopite lamproite
Wolgidite	Diopside-leucite-richterite-madupitic lamproite
Wyomingite	Diopside-leucite-phlogopite lamproite

derived from lherzolites (magnesium ratio, 0.88–0.93). All mineral phases in xenoliths are Mg rich and Ti poor. Equilibration temperatures are estimated as 880° to 1,100°C (Jaques et al. 1986). Mantle xenoliths in olivine lamproites are dominantly dunites with rare lherzolites and very rare harzburgites (Mitchell 1985; Mitchell and Bergman 1991).

Garnets, although uncommon in lamproite, range in composition from pyrope to almandine-pyrope (see Chapter 15, Exploring for Diamond Deposits). According to the classification scheme of Dawson and Stephens (1975) these garnets belong to groups 1, 3, 5, and 9. Most of the pyropes are poor in TiO_2 (<0.4 wt. %) and all are Ca saturated with only 2 to 10 wt. % Cr_2O_3. The subcalcic, Cr-rich pyropes of the G10 group that are used as diamond indicators for kimberlite are absent in both the Prairie Creek and Argyle lamproites. Such garnets are also extremely rare in the Ellendale 7 and 9 diamondiferous lamproites (Lucas et al. 1989), and only one has been found as an inclusion in an Ellendale diamond (Hall and Smith 1985).

The nomenclature of lamproites has been source of confusion. Local names were provided for many lamproite rock types in the past. For instance, Cross (1897) introduced three rock names in the Leucite Hills of Wyoming (madupite, wyomingite, and orendite). A fourth type (olivine orendite) was proposed years later by Carmichael (1967) for some olivine-bearing flows at South and North Table Mountain, Wortman dike, Endlich Hill, and Black Rock in the Leucite Hills. The system of using local names resulted in mineralogically identical lamproites from various localities around the world receiving more than one name. For example, the Leucite Hills wyomingites are equivalent to fitzroyites in Western Australia. To alleviate these problems Mitchell and Bergman (1991) suggested that the lamproite nomenclature should be modernized by using descriptive mineralogical classifications, and that the type-locality nomenclature be eliminated (Table 7.1).

GEOLOGY AND VOLCANOLOGY

Some lamproites have features described as maar-diatreme volcanoes. Maars are craters with diameters up to 3 km and depths reaching 300 m that are formed by explosive subaerial volcanic activity. The morphology of lamproitic bodies is in sharp contrast to the

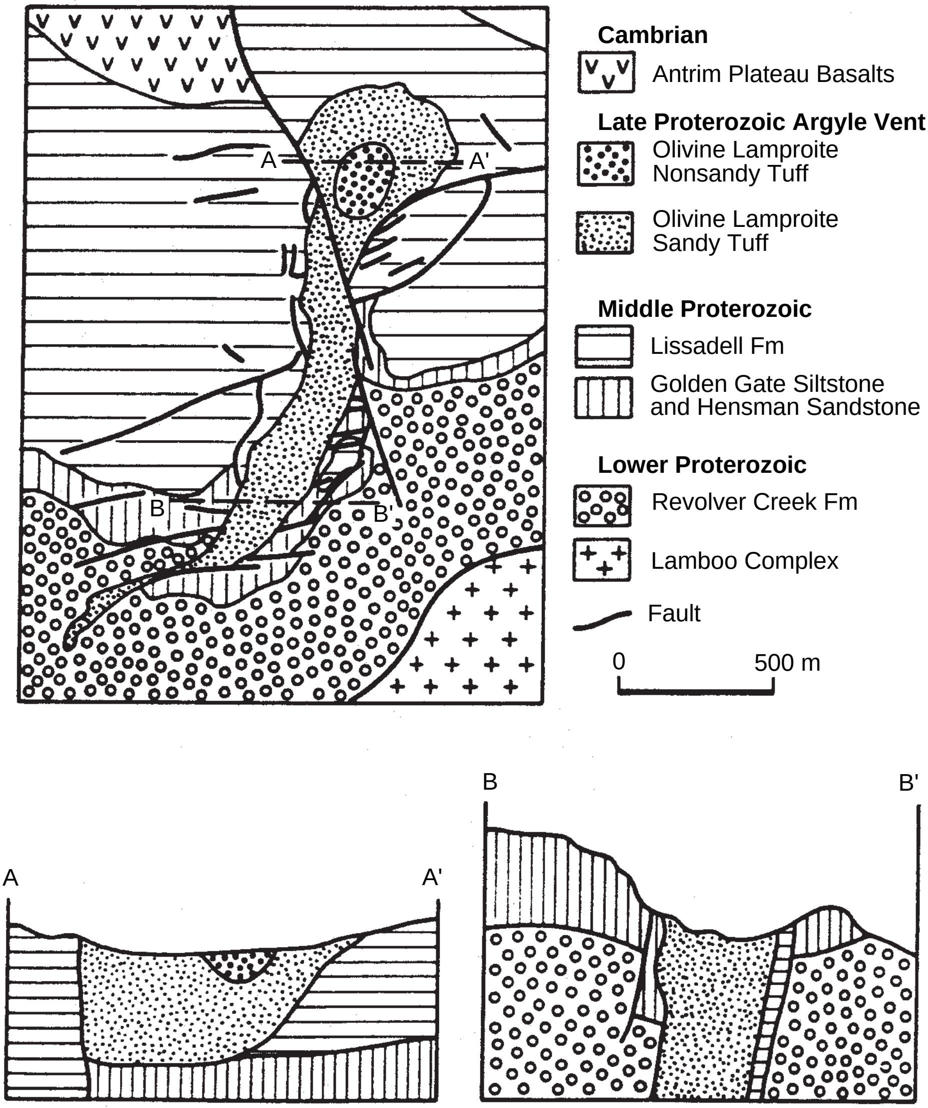

FIGURE 7.2 Geological sketch map and cross section of the Argyle pipe, Western Australia (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

typical kimberlite pipe. Typical cross sections of lamproitic bodies are presented on Figures 7.2 and 7.3 for the Argyle pipe, Western Australia, and Crater of Diamonds pipe in Arkansas. Instead of pipes with steep walls that slowly diminish in width with increasing depth, lamproites are typically characterized by "champagne-glass" vents filled by tuffaceous rocks, often with massive volcanic rocks in the core.

When discovered, the Argyle lamproite in Australia had a surface area of 50 hectares and was composed of both tuffaceous and magmatic lamproite. It was suggested

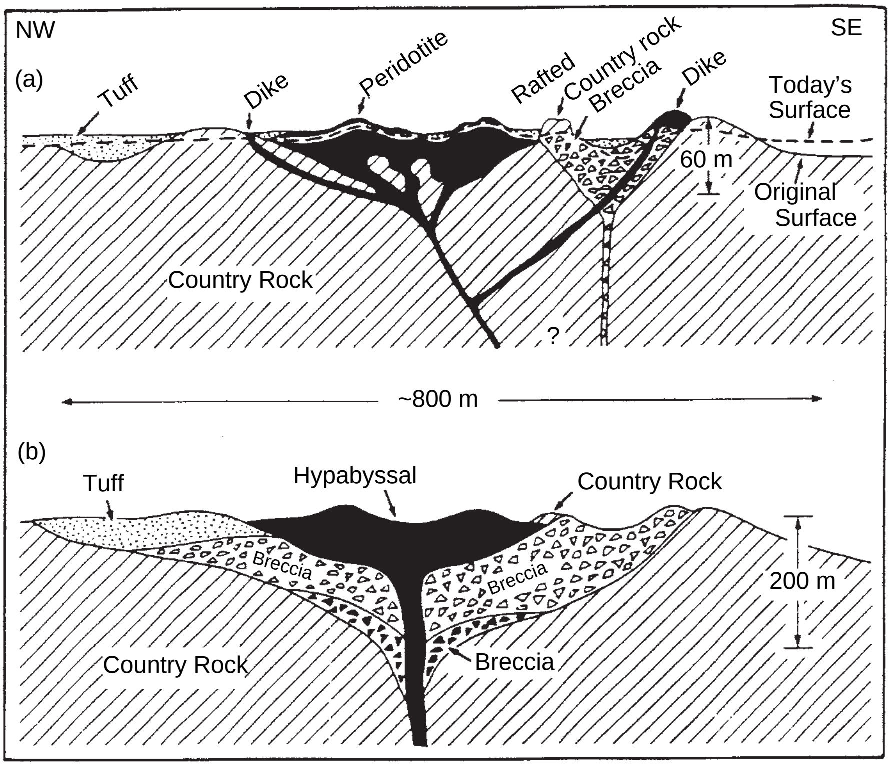

FIGURE 7.3 Schematic cross section through the center of the Crater of Diamonds vent complex, Prairie Creek, Arkansas, showing the relationships between various lamproitic facies. a. After Bolivar 1984. b. Drilling by the Arkansas State Geological Survey within Crater of Diamonds area showed a series of lava flows intercalated with tuffs (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

that lamproite magma rose along a zone of weakness until it encountered groundwater-bearing sediments that caused a series of explosions. The ejection of the magma was accompanied by a downward migration of the explosive activity producing subsidence of the volcanic rocks and surrounding sediments. This subsidence in turn led to the formation of a volcanic crater that eventually filled with ground water, ash, and sediments (Shigley et al. 2001).

Lamproitic bodies are small, reflecting small-scale magma formation at depth. The small volume of lamproitic bodies is reflected in the absence of huge shield volcanoes or stratovolcanoes. Instead, where preserved, the lamproitic volcanoes are small cinder cones, similar to those of alkaline basalts.

Lamproite magmas are volatile rich and maintain a high solubility of volatiles at all pressures. Thus explosive lamproite vents are restricted to phreatomagmatic centers where the magma contacts the water table. Three major lamproite provinces are currently recognized: the Kimberley area of northwestern Australia; the Leucite Hills, Wyoming; and Murcia–Almeria, Spain. Approximately 15 to 20 smaller provinces occur in the world.

Lamproites typically occur within groups or fields. Within each field, lamproitic dikes, sills, vents, flows, and scoria cones may be found. For example, Ogden (1979) noted that the Leucite Hills field in the Wyoming craton consisted of 22 lamproite flows, dikes, necks, plugs, cinder cones, and pumice cones lying along the northern flank of the Late Cretaceous–Paleocene Rock Springs uplift. Radiometric dates indicate the volcanic activity occurred between 3.1 Ma and 1.1 Ma (Bradley 1964; McDowell 1966, 1971).

Individual flows mapped in the Leucite Hills field range in thickness from less than 0.3 m to 37 m and include vesicular lavas, scoria, autolithic intrusive breccias, lapilli tuffs, tuff breccias, and agglomerates. Although vents were found in association with most of the flows, vent facies breccias were inconspicuous at some of the lamproites. Either these vents were removed by erosion, or they were buried by later flows. Such hidden vent facies are not uncommon in lamproites. For example, drilling the center of one lamproite flow in western Montana exposed hidden vent facies beneath the flow (W.D. Hausel, field notes, 1994).

Many flows are simple. Basal autobreccia rubble zones are overlain by dense, nonvesicular platy lavas characterized by a well-developed flow layering that is accented by abundant phlogopite phenocrysts. These platy zones grade upward into massive lava marked by an increase in vesiculation and common orthogonal cooling joints with flow layering. The layering is due to contrasting layers of vesicular and nonvesicular rock, light and dark lavas, and alternating layers of contrasting mineralogy (Ogden 1979). Alternating centimeter-scale bands of vesicle-rich sanidine-phlogopite lamproite and vesicle-poor leucite-phlogopite lamproite may be due to a difference in water contents of the magmas. A slight difference in volatile content would also explain contrasting stabilities of sanidine and leucite at atmospheric pressure (Mitchell and Bergman 1991). Flow tops are scoriaceous, and squeeze-up spines protrude above the preserved flows. Concentric flow ridges also occur at some localities.

Lamproites include both crater and hypabyssal facies. Crater facies include pyroclastic eruptions associated with cinder cones, volcanic necks, and vents. The cinder cones are composed of highly vesicular cinders (lapilli), ribbon bombs, spindle bombs, and welded blocks of lamproite scoria. Most of the cinder cones erupted after massive vesicular lava flows, although a few erupted contemporaneously with the flows. Some rootless vents at Steamboat and Black Rock in the Leucite Hills are recognized by cylindrical reentrants exposed in the walls of eroded mesa. Pyroclastic debris (cinders and ribbon bombs) line the reentrants (Ogden 1979).

Other crater facies lithologies include tuff breccias, lapilli tuffs, and agglomeratic rocks. The fragmental rocks contain a wide range of xenoliths; some are composed almost entirely of cognate, comagmatic fragmental clasts in tuffaceous matrix. At Boars Tusk, in the Leucite Hills, vesicular to dense autobreccia lavas form portions of xenolithic tuff breccias and lapilli tuffs that are virtually identical to similar lithologies from the West Hill tuff at Crater of Diamonds (Prairie Creek), Arkansas; Calwynyardah (West Kimberley), Australia; and Cerro de Monagrillo (Murcia–Almeria), Spain. However, the Leucite Hills tuffs lack the abundant quartz grains derived from disaggregated country rocks in many West Kimberley, Argyle, and Prairie Creek crater facies lithologies.

Hypabyssal facies lamproites are dominantly massive, dense, holocrystalline rocks formed by in situ crystallization of intrusive bodies of lamproite magma. The hypabyssal facies lamproites can be considered as the intrusive equivalent of crater facies. Possibly the best example of hypabyssal facies lamproite is found at Walgidee Hills, Australia, where phenocrysts of richterite, waderite, jeppeite, priderite, diopside, and apatite occur in a fine-grained groundmass (W.D. Hausel, field notes, 1986).

Dikes in the Leucite Hills are typically less than 46 to 200 m long. Ogden (1979) found that the widest dikes (8 to 11 m) tend to possess a breccia zone composed of rounded fragments of lava with sedimentary rock fragments in a volcanic matrix.

In the Smoky Butte field of Montana, 30 to 40 lava flows are associated with 11 vents. The flows are 0.2 to 40 m thick (usually 1–25 m) and cover a small area of 0.05 to 5 km^2 (average 0.7 km^2) with a volume of 0.05 to 5 km^3 (average 0.7 km^3). The volcanic cones are relatively small covering 0.7 to 1.6 km in diameter and reach heights of only 75 to 100 m (Mitchell and Bergman 1991)

The type of lava eruption can be deduced from the lava flows. For example, Gaussberg Nunatak, Antarctica, is composed of pillow-shaped tongues of vesicular glass-rich (50%–65%) lava forming a symmetrical cone 1.5 km in diameter and 373 m high. The monotonous character of pillow lavas at Gaussberg led Tingey, Macdougall, and Gleadow (1983) to conclude that it was formed by a single eruptive episode. The same conclusion can be drawn about almost all small lamproitic lava volcanoes, scoria cones, and lava flows. Their size and shape and absence of well-developed soils between lava flows indicate that most lamproites were formed during a very short period.

Circular faults mapped along the boundaries of some lamproitic craters are similar to structures associated with calderas. These caldera-forming eruptions result in a 500- to 2,000-m-diameter crater about 200 to 500 m deep. Associated with these eruptions are coarse-grained, vent-filling, tuff breccia, such as is present at the 81-Mile Vent in the Noonkobob field in Western Australia. Similar tuff breccias underlie the layers of "sandy" tuffs in the Prairie Creek, Arkansas, sandy lapilli tuff in Ellendale 9, and sandy lapilli tuff and tuff breccia in Ellendale 4 (see Figures 7.3 through 7.5).

During a series of consequent eruptions, different types of pyroclastic rocks, air-fall tuffs, pyroclastic flows, and base surge volcanics were deposited. The reworking of primary pyroclastic deposits resulted in formation of epiclasts. As a rule, later stages of centers of lamproitic activity are usually characterized by eruption of lava flows, emplacement of dikes and sills, and formation of lava lakes in craters.

The difference in morphology directly reflects the difference in volcanology of lamproite and kimberlite. In contrast to kimberlite, lamproite is characterized by air-fall tuffs, base surge deposits, lava flows, and a central core of massive lamproite. The difference in volcanic style may result from the difference in volatile content and composition. The volatiles in lamproites are dominated by highly soluble silicate melts rich in H_2O and F, and although limits of solubility have not been determined experimentally for melts of lamproitic composition, their combined solubility in a superliquid lamproitic melt is on the of order 3 to10 wt. % or more at low pressure (0.5–1 kbar). Thus, during the ascent of this volatile-rich magma, it is not until shallow depth (0.5–1 km) that significant exsolution of an aqueous fluid phase occurs.

Expansion of this fluid at these low pressures probably results in the formation of funnel-shaped lamproitic vents. In contrast, kimberlitic magmas rich in CO_2 can be expected to exsolve volatiles at much greater depths. As a result, kimberlitic magmas tend to produce much deeper explosions that form the root zone and an embryonic type of kimberlitic pipe at depths 1 to 2 km below those at which lamproitic vents were formed (Mitchell and Bergman 1991).

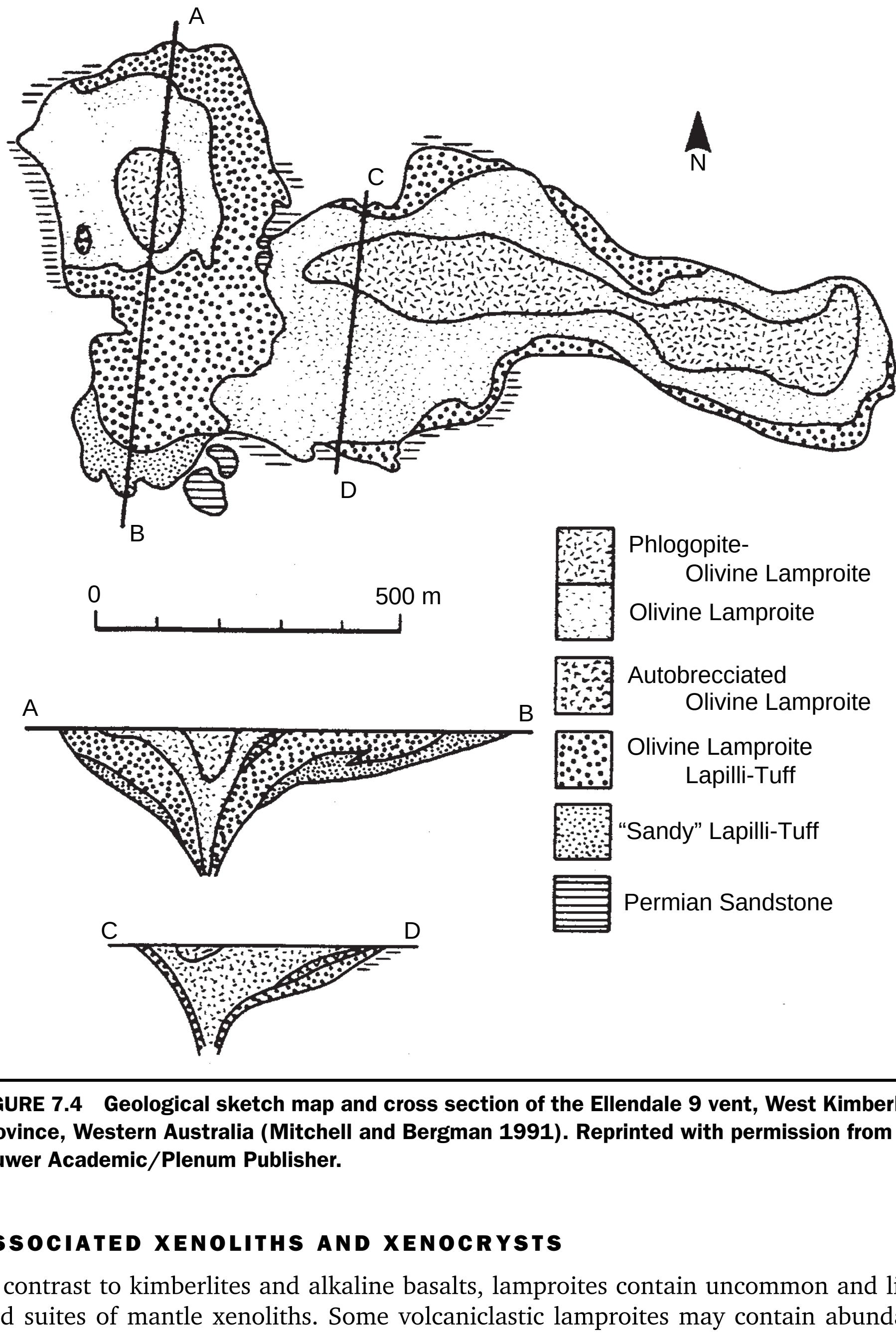

FIGURE 7.4 Geological sketch map and cross section of the Ellendale 9 vent, West Kimberley Province, Western Australia (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

ASSOCIATED XENOLITHS AND XENOCRYSTS

In contrast to kimberlites and alkaline basalts, lamproites contain uncommon and limited suites of mantle xenoliths. Some volcaniclastic lamproites may contain abundant crystal xenoliths, but in general mantle xenoliths are uncommon. Some of the crustal xenoliths may also provide evidence of relatively high temperatures in the lamproitic magma. For example, some shale xenoliths recovered from lamproites in the Leucite

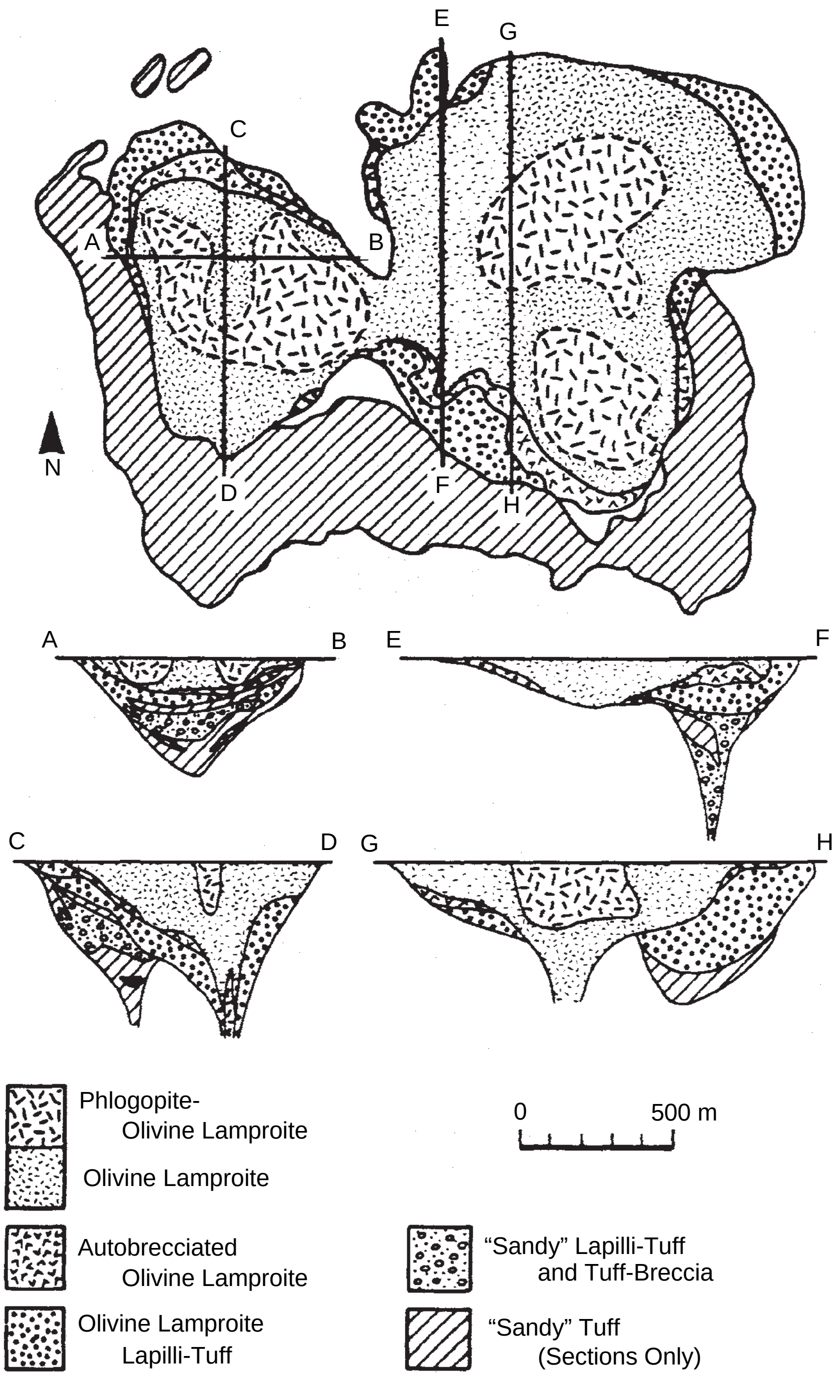

FIGURE 7.5 Geological sketch map and cross section of the Ellendale 4 vent, West Kimberley Province, Western Australia (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

Hills contain evidence of thermal baking, owing to the higher magma temperature and lack of CO_2 associated with lamproites.

Overall, xenoliths in lamproites have not been studied in as great detail as those in kimberlites, owing to the smaller number of commercial diamond mines in lamproites and the comparatively smaller percentage of xenoliths in lamproite. However, Ogden (1979) examined several xenoliths in the Leucite Hills lamproites in southwestern Wyoming and reported abundant lower and upper crustal xenoliths. Cognate phlogopite lamproite nodules are also relatively common in some of these lamproites, but mantle nodules were uncommon to rare.

The types of xenoliths and xenocrysts found in olivine lamproites are distinct from those in phlogopite- and leucite-rich lamproites. The latter are dominated by a cognate suite of phlogopite- and leucite lamproite nodules and fragments, whereas olivine lamproites are characterized by dunite xenoliths and aggregates of olivine xenocrysts known as "dogtooth" xenocrysts (Mitchell and Bergman 1991). Lherzolite and harzburgite xenoliths are rarely present in either lamproite type but are more common in the mafic varieties. Some olivine lamproites contain both varieties of xenoliths, but they are uncommon. No eclogitic xenoliths have been found in lamproites as of 2001, which seems paradoxical, because the majority of diamonds recovered from lamproite are E-type diamonds.

Deep-seated nodules found in olivine lamproites are dominated by coarse-grained dunites. The suite of nodules may also include garnet-bearing dunite, garnet harzburgite, diopside-bearing garnet harzburgite, chromite-bearing garnet harzburgite, and chromite dunite. Some diamondiferous garnet and chromite-garnet lherzolites and harzburgites have been reported from Argyle (Mitchell and Bergman 1991).

The mantle xenoliths reported in lamproite belong exclusively to the ultramafic (peridotitic) rock association. Mineral inclusions found in lamproite-derived diamonds are characterized by two distinctive suites: peridotitic and eclogitic. The peridotitic suite includes various combinations of purple Cr-pyrope, Cr-diopside, forsterite, and enstatite, whereas the eclogitic suite is composed of various combinations of pyrope-almandine, omphacite, kyanite, coesite, sanidine, and rutile.

Xenocrysts found in lamproite derived from the disaggregation of mantle xenoliths are relatively uncommon. Where xenocrysts are found, olivine is the most common. Other xenocrysts may include diamond, rare G9 (chrome-pyrope), and eclogitic (G3, G6) garnets. The megacrystal/macrocrystal Ti-Cr-pyropes (G1, G2) and magnesian ilmenites are very rare in lamproite (Mitchell and Bergman 1991). The subcalcic chromian pyropes (G10) are also very rare in lamproite.

DIAMONDIFEROUS LAMPROITE

It would appear that lamproites are developed under variable conditions with variable thermal gradients and from a depth range extending from within to well above the diamond stability field (Nixon 1995). A qualitative correlation between diamond and olivine in lamproite is confirmed in both the Ellendale and Kapamba provinces, in which diamond grades are consistently higher in olivine lamproites than in leucite lamproites. Thus any deposit of olivine lamproite is of exploration interest.

Diamondiferous lamproites have now been identified at several localities, which include the Argyle and nearby Ellendale field in Western Australia. In Zambia, the Kapamba lamproites are dikes and pipe-like intrusions with subeconomic concentrations of diamond in olivine lamproite tuffs, but some diamonds have also been recovered from leucite lamproite flows in that field. The Majhawan, India, olivine lamproite, dated at 1.14 Ga, has a reported ore grade of 10 carats/100 tonnes. The Prairie Creek, Arkansas, olivine lamproite has an average grade of about 11 carats/100 tonnes. Lamproites in the Zhenyuan field of the Yangtze craton, China, have reported ore grades of 25 carats/100 tonnes. Diamonds have also been recovered from the Bobi lamproite in the Ivory Coast, Africa.

Most olivine lamproite fields have yielded some diamonds. However, owing to the relatively slow rate of ascent, lamproites contain relatively few mantle xenoliths or xenocrysts (as well as indicator minerals). Because of this slow ascent rate, diamonds in lamproites are often resorbed and graphitized, and octahedral diamonds are relatively uncommon. Many diamonds recovered from lamproites show a variety of morphologies suggestive of resorption, and large diamonds have been uncommon in the lamproites mined to date. At the Argyle, for instance, more than 60% of the recovered diamonds were irregular in shape and included macles, polycrystalline forms, and rounded dodecahedrons (Shigley et al. 2001).

When lamproites are diamondiferous the diamonds are found primarily in pyroclastic rocks, and the magmatic lamproite phases are notoriously diamond poor owing to the relatively high temperatures that are sustained in the flows when they erupt in an oxidizing atmosphere. Thus lamproitic diamond deposits are essentially restricted to the volume of preserved pyroclastics in a given vent, where the magma temperatures cool more rapidly (Scott-Smith 1996). This restriction tends to greatly limit the available ore tonnage in most lamproites.

However, where vents flare out, a potential for substantial tonnages exists in the larger craters. At the Argyle lamproite, for instance, past reserve estimates of 94 million tons of ore at an average grade of 750 carats/100 tonnes led to its classification as a world-class ore deposit. Some of the richer portions of the lamproite yielded grades as high as 2,000 carats/100 tonnes. However, large numbers of the Argyle diamonds were graphitized and partially resorbed; the largest diamond weighed 42.6 carats and the overall size of the diamonds was quite small, averaging less than 0.1 carat. At one point, Argyle's annual production totalled about 40% of the world's production, and by the end of 2000 the mine had produced an extraordinary 558,400,000 carats (Shigley et al. 2001).

Diamonds in lamproites have many characteristics similar to diamonds in kimberlites. The size distribution of diamonds is log normal in both cases (Hall and Smith 1985), and both have similar morphology, color, and infrared-detectable nitrogen. But overall the Argyle and Ellendale stones are relatively smaller than many kimberlitic diamonds. This difference is explained as a result of strong resorption in the transporting lamproitic magma. Large (>1 mm) Ellendale stones are dominantly yellow dodecahedra, whereas stones less than 1 mm are colorless or pale-brown, frosted, unresorbed step-layered octahedra. The Argyle diamonds are mostly irregularly shaped stones, fractured, strongly resorbed dodecahedra or combinations of octahedra and dodecahedra. Deeply etched channels are also characteristic. Almost 80% of Argyle diamonds are brown, and many of remaining 20% are yellow or colorless. Very significant, however, are the rare but economically important pink diamonds that bring Argyle fame. It is important to notice that peridotitic xenoliths found in lapilli tuffs from Argyle contain small (>0.4 mm), unresorbed, strongly frosted octahedral diamonds (Keller 1995).

Many lamproite-derived diamonds are colored (or "fancy"—yellow and brown) stones, relatively small, and likely to contain eclogitic mineral inclusions. Peridotitic

(P-type) mineral inclusions in diamonds have compositions that overlap with those in kimberlite-derived diamonds but are generally richer in iron. Chrome pyropes are not subcalcic! Eclogitic E-type clinopyroxene inclusions in lamproite-derived diamonds are anomalously rich in potassium, and they are thus consistent with models suggesting that the formation of E-type diamonds may be related to potassic metasomatism of the lamproite source area. The source of the carbon for E-type diamonds is interpreted to be related to subducted sediments.

The Sm-Nd model ages of E-type Argyle diamond inclusions, which average 1.58 Ga, postdate orogenesis (1.8 Ga) but predate eruption of the lamproite (1.18 Ga). These dates may suggest that the E-type diamonds formed during postorogenic cooling of the metasomatized mantle.

Although these ages indicate that Argyle eclogitic diamonds are xenocrysts in the host lamproite, it does not preclude an original phenocrystal relationship with similar small-volume mantle magmatism about 450 million years earlier and subsequent lithosphere storage. This hypothesis is in agreement with observed $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ ratios, which clearly bear lithosphere storage signatures (Richardson 1986). In the same work, Galimov is cited as showing that carbon isotopes with ^{13}C values of Argyle eclogitic inclusion-bearing diamonds are within the range of values typical for eclogite-inclusion-bearing diamonds worldwide (2 to −25). However, they are clearly distinct from the values characteristic of diamonds derived from the kimberlitic Premier pipe.

Jaques et al. (1989a, b) noticed a distinct difference in the morphology of diamond crystals corresponding with different mineral inclusion suites. At Argyle, E-type diamonds were rounded, resorbed, and dodecahedral. However, P-type diamonds were characterized by sharp-edged octahedrons with etched and frosted surfaces. The P-type diamonds were also identical to diamonds found in the peridotitic xenoliths recovered from the Argyle lamproite.

Diamonds from the Ellendale and Prairie Creek lamproites have inclusions from both suites, whereas 76% of the diamonds from Argyle carry inclusions belonging to the eclogitic suite. However, this difference could easily be artificial and be merely a product of selective studies on the mineral inclusions from lamproitic diamonds (Pantaleo et al. 1979; Hall and Smith 1985; Jaques et al. 1989a, b). Equilibration temperatures for E-type garnet-clinopyroxene pairs are estimated at 1,085° to 1,575°C (Griffin et al. 1988). In contrast to kimberlite-related diamonds from South Africa, no low-Ca (<5.0 wt. %) pyropes have been reported from the mineral inclusion suite in lamproite-derived diamonds.

Diamondiferous lamproites possess some geochemical and mineralogic similarities to kimberlite. Principal among these is high amounts of modal olivine, Cr-rich spinels, and similar enrichments in Cr, Ni, and incompatible elements.

Some petrologists now recognize Group II kimberlites (orangeite) as chemically overlapping with olivine lamproites. Group II kimberlites have been a significant source of gem diamonds (Peterson 1966), and thus these rocks appear to be a very important source of diamonds.

Altered olivine-phlogopite lamproite dikes have also yielded diamonds from the Ivory Coast and Gabon, in western Africa. These rocks are devoid of the usual kimberlitic indicator minerals and appear as talc or phlogopite schists. The Bobi lamproite dike in the Ivory Coast is locally very diamond rich with grades up to 1,000 carats/100 tonnes (Helmstaedt 1993).

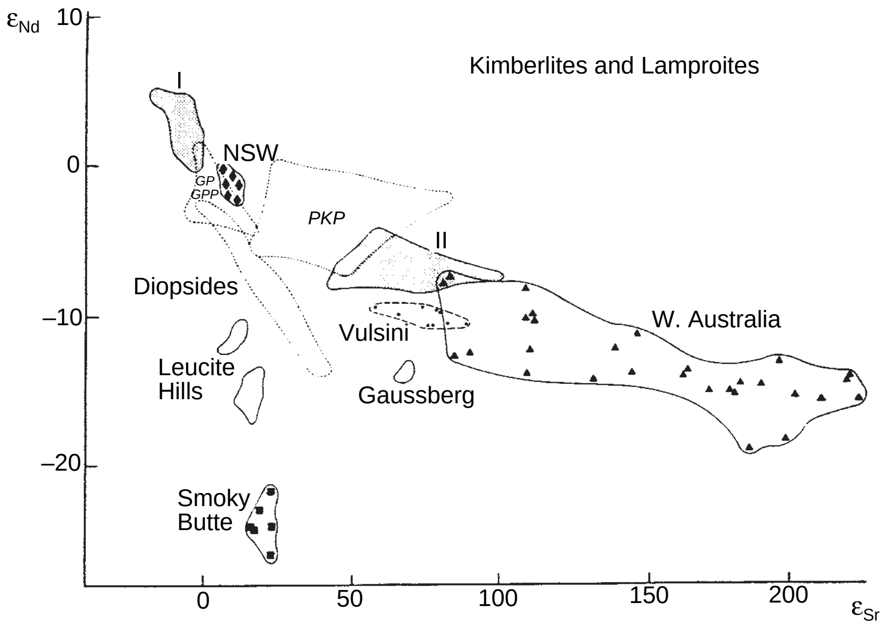

FIGURE 7.6 Nd, Sr variations in kimberlites (shaded), selected potassic and ultrapotassic volcanic rocks and mantle xenoliths. PKP (phlogopite K-richterite), GPP (garnet + phlogopite), and GP (garnet) are whole-rock peridotite xenoliths from the Kimberley area. The Group II kimberlite are mostly from Finsch Mine, South Africa, and the Group I kimberlites are of South African Cretaceous kimberlites (Hawkesworth, Fraser, and Rogers 1985). Reprinted with permission of Geological Society of South Africa.

GENESIS OF LAMPROITE IN COMPARISON WITH KIMBERLITE

Some petrologists conclude that lamproite source regions form as a result of subduction-related metasomatism, which is supported by enriched Sr and Nd isotopic signatures of the lamproites. Many lamproites occur in mobile belts where evidence of former plate subduction is present. However, the Leucite Hills lamproites and lamproites (or lamprophyres) of the Missouri Breaks region in the Wyoming craton lie within an Archon and show no apparent relationship to subduction.

As with kimberlites, lamproites are interpreted to form by the partial melting of similar, yet distinctive, peridotitic material. The magmas are interpreted to originate at depths of at least 150 km under temperatures in the range of 1,100° to 1,500°C. Kimberlites appear to have crystallized at the middle to upper part of this temperature range and lamproites at the lower end.

As will be shown later (see Chapter 14, Temporal Distribution of Diamond Deposits), both kimberlites and lamproites tend to be formed during global pulses of intense magmatic activity. In many regions—Western Australia, the Rocky Mountain region of the United States, Greenland, India, the Aldan shield, and the Siberian platform—the radiometric ages of lamproites are much younger than those of kimberlites (about

1.3 Ga in Western Australia; about 400 Ma in the Colorado–Wyoming region). Both kimberlites and lamproites tend to occur within the same regions, although many of the lamproites locations are shifted toward orogenic belts (i.e., Halls Creek and King Leopold mobile belts in Australia). The apparent difference between kimberlites and lamproites in Sr and Nd isotopic compositions reflects the evolution in mantle conditions (Hawkesworth, Fraser, and Rogers 1985; Figure 7.6). Lamproites are generally believed to form by partial melting of LILE-enriched, phlogopite-bearing harzburgitic mantle sources at substantial depths (exceeding 150 km for some olivine lamproites) under reducing, F-rich conditions. A significant role is probably played in the difference in volatile compositions in these rocks that is reflected in the prevalence of water in lamproites rather than carbon-dioxide gases as in kimberlites. The latter is also a reflection of mantle evolution in a single magma chamber and on a regional scale. This change profoundly influences the diamond content and the dissolution of mantle-derived nodules and megacryst suite minerals in lamproites.

Lamproites may originate in veins in mantle rocks composed of volatile-rich (phlogopite, amphibolite, carbonate) and incompatible-element enriched phases (such as rutile and apatite). Different degrees of partial melting, total pressure, and local variations in source rock mineralogy will result in a wide spectrum of primary lamproite melts.

Unfavorable conditions for preserving diamonds in lamproites include a quiet mode of emplacement of such as sills, rather than diatremes with associated tuffs; magmatic differentiation in the sills (which would allow for diamonds to sink and be resorbed or graphitized); and abundant water (diamond will graphitized more readily in the presence of gases such as water vapor).

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CHAPTER 8

Placers

PLEISTOCENE TO RECENT PLACERS

The chemical inertness and hardness of diamond enables it to travel great distances in streams and rivers. Many placer deposits have high ratios of gem-quality to industrial-quality diamonds, as compared with kimberlite or lamproite. This difference is partially caused by impact during stream transport that preferentially disintegrates the weaker, impure industrial stones along cleavage planes. Because cubic diamonds are about 6,000 to 8,000 times harder than any other mineral, those without defects may be scratched only by other diamonds and should not abrade during stream transport. Most alluvial diamonds thus retain pristine surface features.

Milashev (1989) described one experiment in which six gem-quality and six industrial diamonds were placed in a rotating tank with numerous boulders and 265 steel balls. After 7 hr, the industrial diamonds were completely destroyed. After 85 hr, the gem-quality diamonds had lost only about one hundredth of a percent of their original mass.

Typically, diamond size noticeably decreases with distance from the source area, primarily because larger diamonds are too heavy to be transported for any considerable distance. Diamonds have a higher specific gravity (3.52) than most rock-forming minerals (e.g., quartz, 2.65) and rock pebbles. Consequently diamonds tend to gravitate with other relatively heavy minerals to trap sites in streams, gullies, potholes, and old beach levels. In large enough concentrations, detrital diamonds form alluvial placer deposits (stream, terrace) and marine placers. They are not so different from other placer deposits such as gold and platinum-group metal placers. Such deposits will bear the usual well-known regularities of placers—in particular, concentration on bedrock with periodic rich pockets (pay streaks). Some pay streaks covering areas of only 10 m^2 to 14 m^2 have contained caches of diamonds valued at \$100,000 to \$200,000 (Milashev 1989).

Placer Resources

Despite the comparatively small reserves and erratic grades that are characteristic of placer deposits, placers have historically been and continue to be major source of diamonds, both gemstones and industrial stones. Most placer deposits are alluvial placers (recent or paleo-alluvial), although some karst-related placers have been reported. Such placers are located in karst cavities (especially at the level of ancient karst weathering in Angola and Zaire). Some of these karst pockets were reported to have diamond concentrations as high as 300 carats/m^3 (Milashev 1989). There are also reports of eolian placers formed by the concentration of heavy minerals during the eolian reworking of original diamond sources, but they appear to be rare.

Ancient sources of diamond were all alluvial placers, which included the Indian and Kalimantan placers. Until recent times, diamonds had essentially no value other than as jewelry. As a result, these placers fed all demands up to the nineteenth century, because only the very wealthy were able to acquire diamonds.

In the middle of nineteenth century, another important source of diamonds was discovered in Brazil. The Brazilian placers produced ten of world's 50 largest gemstones, including such famous diamonds as the Vargas (726.6 carats) and the Goyas (600 carats). Such large placer diamonds suggest a nearby bedrock source.

Placers were also a source of industrial diamonds. Industrial diamonds mined from alluvial and "intermediate" placers, along with the comparatively cheap Brazilian carbonado, satiated industrial demands and provided the basis for a modern metal-cutting and geological drilling industry. Since the end of the nineteenth century much of these industrial diamonds have been partly replaced by Central African "carbon" and by synthetic stones.

Even today, placers are a significant source of diamonds, especially in some underdeveloped countries where they are mined with cheap labor and little capital investment or contemporary equipment. However, actual industrial diamond production figures are difficult to obtain for a number of reasons; many are mined in underdeveloped countries that don't provide accurate figures. Diamond smuggling is also rampant in some of these regions, and by some estimates it can divert as much as 50% of official production.

All Brazilian diamonds have been produced from alluvial gravels, and in 2001 the original source of the diamonds had yet to be found. In 1980, production was reported to be 1,089,000 carats; in 1985, it was 800,000 carats.

Most Venezuelan diamonds are from placers along the Caroni and Paragua Rivers. Throughout the 1970s, the official production figures showed average diamond output to be around 250,000 carats, of which 20% to 25% were gems. The gemstones are usually small (<0.5 carats). Since the 1970s, production has declined sharply.

In Ghana, diamonds were mined from two placers, the Akwatia and Birim concessions, northwest of the capital, Accra. Annual production peaked at 2,283,000 carats in 1975 with only 850,000 in 1986. About 10% of country's output is classified as gems. The decline in production has not been due to the lack of resources, but rather to the general economic deterioration of the country. Recoverable resources are estimated to range between 20 and 50 million carats. Ghana's estimated annual production could well exceed 1.3 million carats (Miller 1987).

Most Guinean diamonds are also mined from placers. Some exceptionally rich gravels derived from the erosion of a series of small uneconomic dikes have been fairly productive. Annual production reached 200,000 carats in 1986.

Almost all Liberian diamond production (45% of which is gem quality) comes from small alluvial diggings around Gbapa and Takormah. In the nearby Central African Republic, annual production is estimated at 300,000 carats. Diamonds were first mined in this region in 1913 when two placers were found east of Brea, 64 km from the capital, Bangui.

In many regions, placer discoveries served as a precursor to the discovery of lode deposits. In India, for example, diamond mining started in prehistoric times, but kimberlites and lamproites were not found until the beginning of twentieth century. In Brazil, the original source of the diamonds was not known up to the middle of the twentieth century. In South Africa, placer diamonds were mined prior to the discovery of kimberlite in the late nineteenth century.

In West Africa diamonds were first found in 1910, and most placer deposits were found in the 1930s. Lode deposits were found much later: Sierra Leone in 1948, Guinea in 1952 to 1965, Mali in 1958, Ivory Coast in 1960. In Zaire, diamond placers have been known since 1913, but the first lode deposits (Mbuji mine and Lubi) were not found until 1946. Thus all diamond production statistics for these countries before these dates were derived from placer mining.

Marine Placers in Southwest Africa

Marine placers in southwest Africa (formerly part of German Southwest Africa) were found in 1908 at a site north of Luderitz (Figure 8.1). In 1926, marine placers were identified south of the mouth of Orange River. The initial exploration of these sites showed extremely rich placer deposits. For example, a small geological team collected 12,500 carats of diamonds including two big stones (81 and 71.5 carats) that eventually sold for about 300,000 German marks each. One site, an area about the size of a desk, yielded diamonds that exceeded £1,000,000 in value.

In September 1908, the German colonial authorities proclaimed as "Forbidden Land" the 100 km strip along the coast and stretching inland from the Orange River mouth to 70 km north of Luderitz Sperragebeit. In the next 8 years before the beginning of World War I, a total of 5.4 million carats worth £19 million were recovered from the vicinity of Luderitz by nine mining syndicates.

In 1923, as a result of so-called Halbershield Agreement, DeBeers acquired the sole right to mine and prospect for diamonds within Sperragebeit (Figure 8.2). Namibia's greatest diamond wealth was discovered 5 years later when it was realized that onshore marine terrace and beach deposits stretching 160 km north of the mouth of the Orange River contained unparalleled amounts of gem diamonds. From the Olifants River in South Africa to an area north of Hottentot Bay in Namibia, a strip of diamond-rich placers extended for about 1,000 km.

Diamond reserves were conservatively estimated at 1.5 billion carats, of which 90% to 95% are of gem quality. Thus, the marine deposits off southwest Africa apparently contain 100 times as many diamonds by weight as are presently being used in jewelry. In addition, this source contains a high percentage of rough diamonds that are suitable for cutting and very important in the jewelry industry (Miller 1993).

The diamonds in the marine placers are believed to have been derived from the erosion of kimberlite pipes located within the stable South African craton. The erosion has occurred during last 80 to 100 million years since Cretaceous time (Figure 8.2) (Gurney, Levinson, and Smith 1991).

Estimates of the volume of material eroded from the average diamond pipe and of the average diamond grade, assuming that pipe had a uniform content of diamonds throughout, indicate that about 500 million carats were eroded from one pipe alone and released into the Orange River and finally the Atlantic ocean (Gurney, Levinson, and Smith 1991). The diamond grade in the placers was increased through the continuous enrichment caused by rewashing of the primary alluvial placers, abrasion, and continuous resedimentation as sea level changed. Because surf attacked the shore at an acute angle, the diamonds had the tendency to redistribute and become dispersed along a beach zone for tens of kilometers, sometimes forming highly enriched horizons.

At the mouth of the Orange River, the average diamond size is 1.5 carats, whereas 200 km north, the average size is only about 0.2 carats (Figure 8.3). The grade generally

FIGURE 8.1 Location of marine placers in southwest Africa (Gurney et al. 1991). Reprinted with permission from J.J. Gurney.

increases in the same direction for two reasons: there are many more small diamonds than large stones, and small diamonds are more easily transported by wave and current action than large ones. As a result there are enriched placers on the beach and on terraces. Within one such horizon (the so-called "oyster horizon") the average diamond content reaches 50 to 100 carats/m^3, and the diamond quality is very high. For example, the average price for uncut gem-quality diamonds from southwestern African marine placers is estimated at $60/carat (Gurney, Levinson, and Smith 1991).

It has long been known that offshore diamond resources along the Namibian and South African coasts were much richer than those onshore. However, technological difficulties prevented any attempt to exploit them until the late 1960s. Namibia was the principal source of gem diamonds in the world, but it was overtaken in the 1980s by

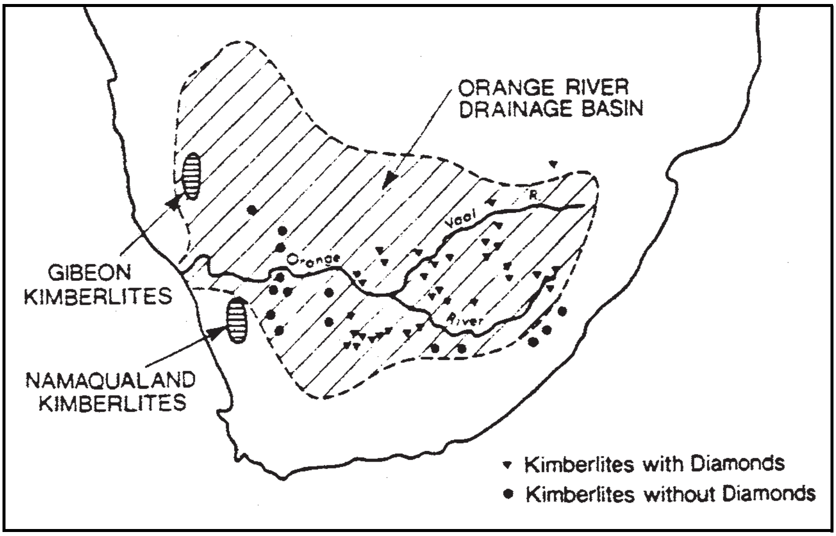

FIGURE 8.2 The drainage area of Orange River through major kimberlite fields of South Africa (Gurney et al. 1991). Reprinted with permission from J.J. Gurney.

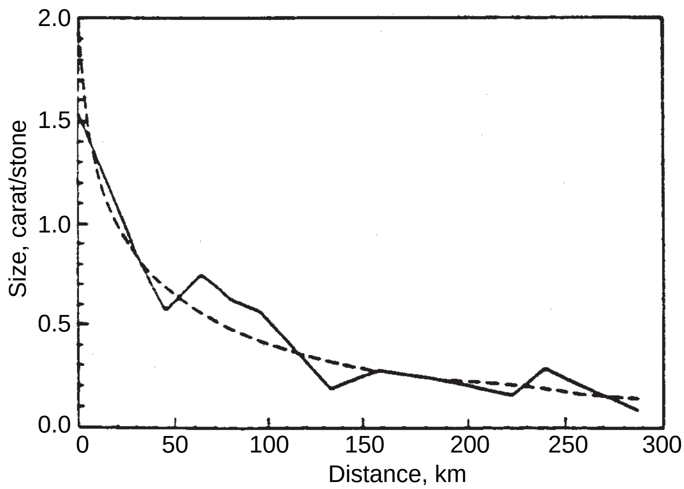

FIGURE 8.3 Size decrease of diamonds recovered from beach sands from the mouth of the Orange River northward (Gurney et al. 1991). Reprinted with permission from J.J. Gurney.

TABLE 8.1 Historic Namibian offshore diamond production (carats x 1,000)(Miller 1993)

	Diamond Source		
Year	Offshore	Tidal	Total
1965	217.2		217.3
1966	170.8		170.8
1967	134.5		134.8
1968	82.4		82.4
1969	19.1	183.8	202.9
1970	16.6	205.7	222.3
1971	40.8		40.8

TABLE 8.2 Recent CDM marine diamond production (Miller 1993)

Year	Carats
1990	29,195
1991	170,741
1992	260,298
1993 (estimate)	244,000

Botswana and Russia (Table 8.1). Miller (1993) stated that southern Africa's west coast marine deposits are "the greatest gem diamond deposit in the world."

Table 8.2 provides figures for diamond output by Consolidated Diamond Mines (CDM), a company created by Sir Ernst Oppenheimer in 1923 as a result of the Halbershield Agreement. In 1991, DeBeers was transformed from an exploration company into a mining entity.

By this time, the technology to mine the deep-water gravels became available and diamond production of the Namibian offshore deposits began. In 1992, approximately 750,000 m^2 of ocean floor was explored by the CDM/DeBeers marine ships. Exploration resulted in the identification of placers at an average diamond grade of 0.35 carats/m^3. Most of the diamondiferous gravels north of the mouth of the Orange River are 1 to 3 m thick; the average thickness is 2 m. This volume would suggest an average diamond grade of 0.175 carat/m^3 or 10 carats/100 tonnes.

Because the technology was only recently developed, exploration and mining for marine placers is in its infant stage. No exploration for marine placers has occurred off the Australian, Canadian, or Siberian coasts, although there are indications that such exploration has been initiated in Western Australia close to Kimberley (Atkinson and Smith 1995) and in the White Sea near the Arkhangel'sk diamond deposits.

The comparatively unattractive placer deposits have played a major role in every segment of the history of diamonds. During the earliest stages up to the nineteenth century, alluvial placers were the greatest source of the largest diamonds in the world. In the nineteenth century, alluvial and redeposited placers provided the greatest amount of industrial diamonds, many of which were mined in Brazil and Central Africa. Now owing to the development of mining and exploration techniques for marine-shelf diamonds (Garrett 1995), shelves could become a major source of diamonds in the twenty-first century.

PALEOPLACERS

About 20% of the world's diamond production prior to 1968 was from detrital sources. Half of the detrital production was derived from Precambrian paleoplacers, and the remaining half was from Paleozoic to Recent paleoplacers and placers. Paleoplacer diamond deposits have been found in rocks ranging in age from Archean to Tertiary. Paleoplacer diamonds tend to show etched features, frosted surfaces, and a greater degree of alpha-particle damage that produces abundant green and brown spots in the diamond crystals.

Consolidated Precambrian and Gondwana (Tertiary) paleoplacer conglomerates in India have been considered the source of diamonds in Quaternary alluvial placers, and they have been intensively mined since ancient times in the historical regions of Panno, Karnool, and Krishna. Paleoplacers have been found in the Banganapalli quartzites (Karnool district), Gallapalli sandstones, and Gondwana Rajahmundi sandstone (Mathur 1982). A resurrected interest in these placer diamond deposits recently resulted in discoveries of new placers in the Simla region (near the Himalayas) and other placer regions in the Junkel and Koel areas (see Figure 1.1).

Many of the modern placer diamonds recovered from Brazil, the Central African Republic, and Venezuela have been derived from paleoplacers, including diamonds in Witwatersrand, South Africa. This is why an intensive study and exploration effort for paleoplacers was initiated in Russia (Nemirova, Skripin, and Safyannikov 1994).

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SECTION 3

Unconventional Source Rocks

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CHAPTER 9

Overview

Presently, unusual or unconventional diamondiferous host rocks are thought of as scientific curiosities, although some may play important roles in the diamond economics in the future. We define unusual diamondiferous host rocks as those that do not have kimberlitic or lamproitic affinity.

Even though unconventional host rocks have been almost completely neglected by industry and many academicians, a variety of such deposits have now been identified that have extremely high ore grades (Wilson 1948; Kaminsky and Vaganov 1976; Dawson 1979; Kaminsky 1984; Bergman et al. 1987; Helmstaedt 1993; Hausel 1996). The number of discoveries is astonishing, particularly when one considers the overall lack of exploration effort for these types of deposits as compared with the enormous exploration expenditures placed on the search for conventional diamond deposits.

Only a few decades ago, this chapter would have included lamproites. However, today lamproites are well recognized as potential sources for commercial diamond deposits by most diamond exploration groups. The Argyle lamproite in Australia was a world-class discovery that created hundreds of mining and exploration jobs in Australia and gem cutting professions in India. It is just a matter of time until another significant diamond deposit is discovered in another unconventional host rock. Possibly lamprophyres (in particular, diamondiferous minettes) may be the next group to move to the "conventional" category.

Some recent discoveries of potential interest include the following:

1. abundant graphitic pseudomorphs after diamond in the Beni Bousera massif, Morocco, which requires serious consideration of the potential for diamonds in similar layered plutons
2. diamond deposits related to high-pressure metamorphic complexes (China, Kazakstan, Mongolia)
3. diamond in ring structures of unknown origin to which the term astrobleme has been commonly applied.

All these new types of deposits provide significant insight into the deep-seated orogenic processes and the character of development of stable cratons. Diamonds found in meteorites, which are of only scientific and mineralogical importance, provide clues to the behavior of carbon in the deepest parts of the earth.

A relatively large variety of unconventional host rocks has now been identified. For instance, diamonds have been found in some alnöites, which lack the characteristic kimberlitic indicator minerals and contain melilite and biotite mica (Helmstaedt 1993), two minerals that are lacking in kimberlite and lamproite. Diamondiferous picritic monchiquites have been reported in Western Australia (Jaques et al. 1986), and diamonds have also been reported in other lamprophyres in Quebec (Helmstaedt 1993). Lamprophyres with diamond-stability-field garnets and chromites have also been identified in Montana, although many of these remain untested (Fipke, Gurney, and Moore 1995; Doden 1997). Diamondiferous minettes, a variety of lamprophyre, are also increasingly recognized.

In eastern Australia, diamonds are associated with a variety of unusual diatremes that have been described as alkali basalt, nephelinite (an alkaline mafic rock with augite and nepheline, but lacking plagioclase), nepheline mugearite (a dark, fine-grained igneous rock of the alkaline basalt series that has a texture similar to trachyte, but which contains oligoclase-andesine feldspar with some orthoclase and olivine), and dolerite (MacNevin 1977; Sutherland, Hollis, and Raynor 1985). Diamonds have also been recovered from alkali basalt in Kamchatka and Arkhangel'sk (Barron et al. 1994; Danielson 1994), and from Late Cretaceous "kimberlitic" and "lamproitic" rocks in an island arc setting in the Valaginsky Range of the Kamchatka arc (Seliverstov, Kepezhinskas, and Defant 1994).

In the Urals Mountains of the former Soviet Union, both alluvial diamonds and paleoplacer pyrope garnets have been found downstream from peridotites. Diamonds were initially identified in this region in the 1820s, and some placers along the western slopes of the central Urals Mountains were worked for diamonds near the turn of the century and during World War II (Dawson 1967). Exploration for the source of the placers led to the discovery of diamondiferous picrites (a dark basaltic rock containing abundant olivine along with pyroxene, biotite, possibly amphibole, and <10% plagioclase) and limburgite (a mafic rock with olivine and clinopyroxene phenocrysts in an alkali-rich glassy groundmass) breccias (Luk'yanova et al. 1992).

Some high-pressure metaphyllites have also yielded small diamonds. For instance, microdiamond inclusions averaging only 12.5 microns (0.125 mm) were discovered in garnets and zircons in garnet pyroxenites, garnet-biotite gneisses, and garnet-biotite schists of metasedimentary origin in the Kokchetav massif of northern Kazakstan (Sobolev and Shatsky 1990). Diamonds have also been reported in low-grade chloritic schists from the Birrimian greenstone belt in Ghana. These schists are interpreted as metamorphosed basalts, ultrabasic igneous rocks, and graywackes (Dawson 1968).

In China, diamonds were recovered from eclogite lenses in Cambrian to Permian blueschists (Barron et al. 1994). Diamonds were also identified in garnets in coesite-bearing eclogites, garnet pyroxenites, and jadeites in the Dabie Mountains of eastern China. These diamonds average only 10 to 60 microns but include grains up to 240 microns (2.4 mm). Other diamonds associated with metamorphic rocks were discovered northwest of the Ulaan Baatar region of Mongolia. Initial bulk sampling recovered diamonds in the range of <0.1 to 1 mm in diameter. Ore grades, however, were incredible—reportedly 4,000 to 10,000 carats/100 tonnes (Helmstaedt 1993), much greater than that of any known kimberlite or lamproite. This deposit is similar to the Kokchetav deposit in Kazakstan (Edward Erlich, personal notes, 1993).

Diamonds are also reported in meteorites and impactites. Some stony and iron meteorites and impactites contain isometric and hexagonal (lonsdaleite) diamond. Lonsdaleite has also been reported from the Popigay depression in northern Siberia. The Popigay structure is interpreted as an astrobleme, although some researchers suggest that the structure may be terrestrial.

Recently, diamonds were recovered from talc schist in South America that is believed to represent altered komatiite. However, samples of the host rock show no evidence of spinifex texture, although some relict rounded olivines appear to be present that may represent cumulate texture (Carl Brink, personal communication, 2000).

Diamonds have also been found in alpine and ophiolitic peridotites and related ultramafic rocks along plate collision boundaries at several localities around the world. Rocks along portions of the Pacific Coast of the United States exhibit similarities to these types of host rocks. The Beni Bousera complex in Morocco recently was identified as having graphite pseudomorphs after diamond. This type of deposit is exciting because discovery of a similar host containing preserved diamonds could be one of the great discoveries in economic geology.

Even rock types that contain only trace amounts of diamond will bear great implications for a wide range of problems in fundamental science from general geology and petrology to cosmology. They reflect complex geological history of each region under consideration.

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CHAPTER 10

Meteorites, Ultramafic, and Mafic Igneous Rocks

METEORITES

Diamonds have been reported in several meteorites. It has been suggested that the diamonds were derived from shock during impact, or that they were originally formed by gravitational pressures in early planetoids in the solar system that were later destroyed during interstellar collisions.

Diamonds were reported in a stony meteorite from Novo-Urei, Russia, in 1886, which became known as a ureilite (an olivine-pigeonite achondrite) meteorite (Yerofeyev and Lachinov 1888). At about the same time, diamonds were also identified in the Magura, Czechoslovakia (Wenschenk 1889), and the Canyon Diablo, Arizona, iron meteorites. The Canyon Diablo diamonds were found in association with troilite (FeS), schreibersite (Fe_3P), and carbonaceous matter. Since the turn of the last century, numerous diamonds have been found and reported in both stony and iron meteorites (Urey et al. 1956; Lipschutz and Andrews 1961; Mason 1962; Vdovykin 1991).

Meteorite classification schemes are complex owing to the variety of meteorites that have been found. For our purpose, we use a simple classification: stony meteorites (chondrite and achondrite groups) versus iron meteorites (stony-iron and iron groups). The stony meteorites are dominated by silicate minerals, and the iron meteorites are dominated by nickel-iron minerals. The reader is referred to Mason (1962) for detailed information.

Diamonds have been found in both stony (carbonaceous achondrites) and iron meteorites. Some specimens have been reported to contain up to 1% diamond by weight. The majority of the diamonds are very small. Relatively larger diamonds have been found in some meteorites, such as the Dyalpur (India) meteorite, which contained a few diamonds larger than 300 angstroms. These diamonds typically exhibited octahedral form. Some meteorites also contain graphite and cliftonite. Cliftonite is assumed to be a paramorph of diamond (Mason 1962).

Urey (1956) proposed that diamonds were produced at the graphite-diamond transition at a pressure of at least 16,000 atm and a temperature of 298°C. Gravity alone will produce a similar pressure at the center of a planetoid object having a 1,010-km radius. If one requires a pressure in excess of 16,000 atm to prevail in at least half of the body, the required radius will be 1,670 km, similar to that for the Earth's moon (1,740 km). Urey therefore proposed that lunar-sized objects must have been important in at least one stage of the evolution our solar system, particularly since diamonds have been identified in many meteorites.

X-ray studies by Lipshutz (1964) support that graphite (rather than diamond) was initially present in all of the diamondiferous meteorites that he examined. Under favorable circumstances, shock-related diamonds show preferential orientation, and thus any graphite found in the meteorites would also show preferential orientation if derived from shock-produced diamond. Lipshutz's research, however, showed that the graphite did not have a preferred orientation, and he concluded that the diamonds in the Canyon Diablo meteorite (and probably in all similar meteorites) formed as a result of impact, and that the diamonds were not initially present in the samples. This conclusion is also supported by the discovery of coesite in meteorites (Chao, Shoemaker, and Madden 1960).

However, most ureilite meteorites that have been examined are too small (0.3–2.2 kg) to have hit the Earth at speeds much greater than terminal velocity. This implies that the diamonds in them could not have formed during impact. Because of this, Lipschutz (1964) suggested that these diamonds were formed by preterrestrial shock possibly related to the breakup of the ureilites' parent body or by past collisions in space.

Some evidence recently obtained from Neptune by Voyager 2 may shed new light on the possible origin of meteoritic diamonds (Rist 2000). Neptune shows great radiation imbalance: it radiates 2.6 times as much energy as it absorbs from the sun. The natural explanation is that this high level of energy output is due to an internal energy source; it has been suggested that this energy comes from methane in Neptune's atmosphere that decomposes to four hydrogen atoms and one carbon atom under the influence of the sun. Because of the tremendous pressure of the gravity field on the carbon molecules, they tend to form diamond dust. The carbon atoms tend to bind to one another and rain down onto Neptune.

The concept has been confirmed by Robin Benedetti at the University of California, Berkeley. She simulated the pressure and temperature conditions that one may find at approximately one-third the depth toward the planet's center. Such an approach provided a new idea about the source of carbon from a methane atmosphere and new mechanism for the formation of diamond.

ULTRAMAFIC INTRUSIONS

Peridotite Massifs of the World

Diamondiferous peridotite massifs have been reported at a number of localities in the world including Armenia, Australia, Canada, Indonesia, Kamchatka, Spain, and Tibet. The presence of other diamond-bearing peridotites is suggested by the presence of diamond placers in similar terranes. For example, diamonds have been recovered from placers downstream from peridotite massifs in the Appalachian Mountains and the Pacific Coast of the United States.

The source of some of the diamonds associated with peridotite massifs, ophiolites, and similar ultramafic belts may be highly magnesian chromite harzburgites. The harzburgites are similar to diamondiferous peridotite xenoliths found in kimberlites, although garnet is usually lacking. Such deposits are found in some major collision zones supposedly tectonically emplaced from the base of the lithosphere (Pearson et al. 1989). The recognition of the harzburgites is usually complicated by intense serpentinization of the peridotite.

Unusual alluvial diamonds in the Ural, Caucasus, and Koryak Mountains are believed to have been derived from alpine-type peridotites. This belief led to the successful search in 1978 for diamonds in harzburgite bedrock from the Phanerozoic Koryak Mountains of northern Kamchatka (Helmstaedt 1993). Diamond in these peridotites from the Koryak Mountains as well as from Tibet are thought to have formed during subduction and to have survived as a metastable phase as a result of rapid tectonic uplift.

Diamonds have also been identified in peridotite in the Amasiyan-Sevan-Akerinsk ophiolite complex in the Lesser Caucasus Mountains of southern Armenia. As is true in many other peridotitic diamond deposits, this deposit lacks associated high-pressure mineral indicators in the region (Bergman et al., 1987).

Diamonds have also been reported in alpine-type ultramafic intrusions and layered ultramafic plutons. Accessory diamonds were found in rocks of alpine-type ultramafic plutons in Armenia (Gevorkian et al. 1975). Later Janse (1994) made an affirmative review of these findings including the discovery of diamond in ultramafic rocks of Koryak Highland (Russian Far East, between the Kamchatka and Chukotka peninsulas).

Diamond inclusions have also been described in a native platinum nugget from Goodnews Bay in southwestern Alaska. Another microdiamond was recovered from a core sample of bottom sediments in the bay (Forbes, Kline, and Clough 1987). The source of these diamonds may be related to rocks of the Goodnews Bay layered ultramafic pluton, which is known to contain platinum minerals.

Alpine-type ultrabasites, an apparent host to some of the diamond deposits, are one of the principal members of the ophiolitic triad. It is believed that they occur at the surface as a result of obduction of oceanic plates. Before localization within the lithosphere, the ultrabasites had a long and complex history prior to uplift under midoceanic ridges accompanied by melting of basaltoid magmas from mantle pyrolite and subsequent secondary transformations. In this sense it is possible to consider ultrabasites as a kind of residual, restitic material of mantle peridotite; they have been subjected to a complex deformational and metamorphic history including an early blueschist-facies event.

Kaminsky and Vaganov (1976) discussed the general petrologic prerequisites necessary for diamond mineralization in alpine-type ultrabasites. By analyzing the Mg-Fe distribution between the mineral pairs olivine/orthopyroxene, olivine/clinopyroxene, and orthopyroxene/orthopyroxene, Kaminsky concluded that the alpine-type ultrabasites formed within a range of 50° to 1,200°C, which is within the accepted temperature range of diamond formation (1,100°–1,400°C) (Figure 10.1).

The pressure of formation was evaluated using experimentally established data on the solubility of alumina in enstatites. The results showed that the pressure varied over a wide range from 25 to 65 kbar. The data support that parts of some of the ultrabasites fell within the diamond stability field (i.e., the Chirynaiysky massif and harzburgites of Ust-Bel'sky massif in the Koryak upland; a portion of the harzburgites of the Yuzhny Kraka massif, Urals Mountains; the Webster massif, Scotland; portions of massifs in New Zealand; and dunites in New Caledonia) (Figure 10.2).

In order to preserve comparatively rare and small garnets, Kaminsky (1984) recovered the grains from diamond-bearing ultrabasites using thermochemical extraction. As a result, Kaminsky obtained garnets from 20 large samples weighing more than 100 kg each from different massifs in the Urals. The garnets were small (0.1–0.5 mm) and were associated with moissanite, graphite, corundum, and zircon.

The garnet populations included iron- and magnesium-rich varieties, and they contain a pyrope component of 30% to 47% and 61% to 73%, respectively. The Cr_2O_3

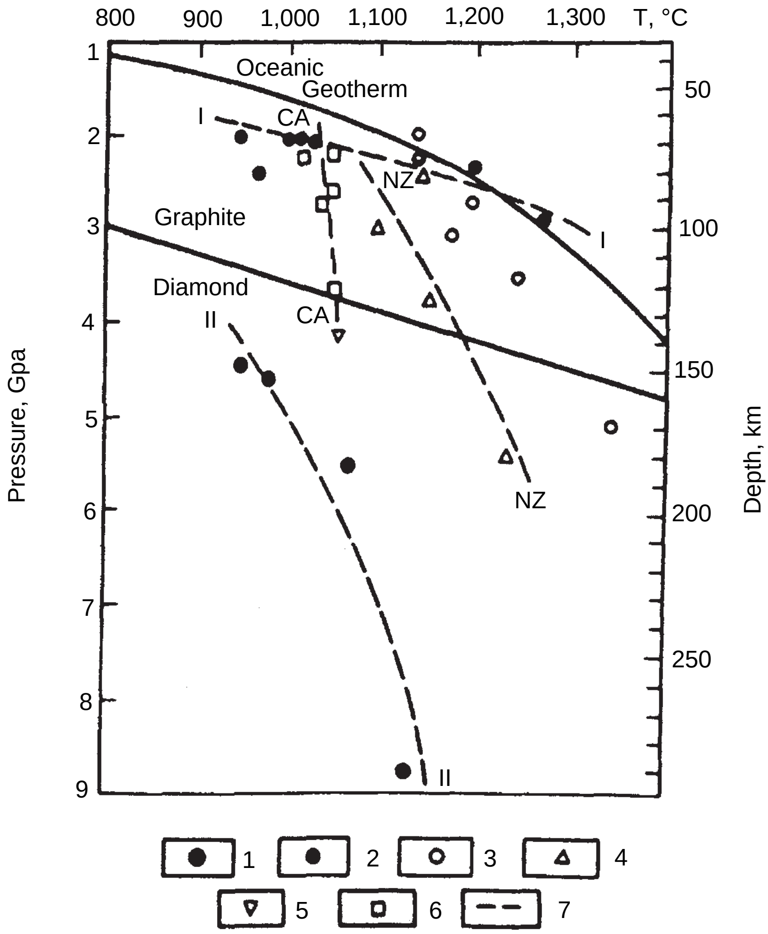

FIGURE 10.1 Thermodynamic conditions of formation of the ophiolitic series of ultramafic rocks. 1, 2, 3 = massifs of Koryak highland, northeast Russian Federation (Kaminsky didn't provide specific names of massifs apparently owing to secrecy requirements in Russia at the time of his publication), 4 = massifs of New Zealand, 5 = Adamsfield massif, Tasmania, 6 = massifs of California, 7 = evolutionary trends of massifs denoted as (I), first, and (II), the second, New Zealand (NZ), and California (CA) (Kaminsky 1984). Reprinted with permission from Publishing House Nedra.

content in the garnets, as determined by electron microprobe, was close to the threshold of analytical error (0.04%–0.16%).

In addition to garnet, another tracer mineral commonly associated with diamonds was found in the ultrabasites: high-chromium chrome spinels similar to the chrome spinels reported in kimberlites. The chromites from the ultrabasites from the Lesser Caucasus region contained 60% to 70% chromium component. Ten percent of the chrome spinels from the Kempirsaysky massif in northern Kazakstan (along the trend of the southern Urals) contain more than 62% Cr_2O_3 and 7% to 9% Al_2O_3 (Kaminsky 1984). The usual association of diamonds with platinum-group minerals and chromites in

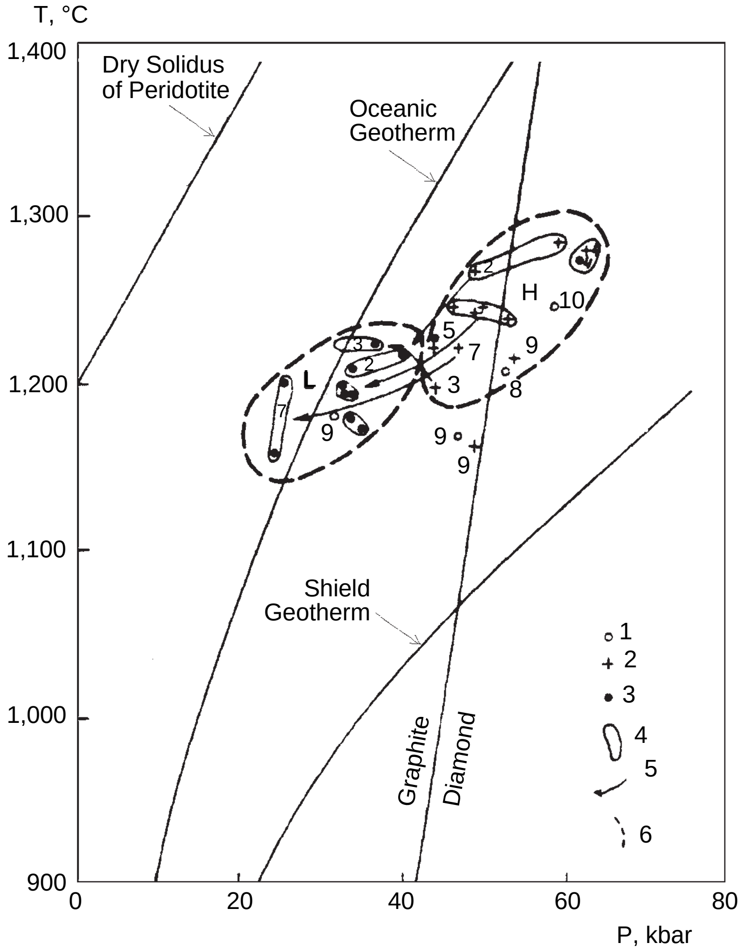

FIGURE 10.2 Thermodynamic conditions of formation of alpino-type ultrabasites from different regions. 1 = dunites, 2 = harzburgites, 3 = lherzolites, 4 = boundaries of massifs, 5 = trend of evolution of ultrabasites in the massifs shown, 6 = boundaries of composite fields of harzburgites (H) and lherzolites (L). Numbers within the field of diagram denote following massifs: 1–6 = massifs of the Koryak highland, 7 = Lizard, 8 = Webster, 9 = New Zealand, 10 = New Caledonia (Kaminsky and Vaganov 1976).

Alaska and British Columbia, in the Lesser Caucasus, and in the Kimpersay massif of northeastern Kazakstan should also be emphasized.

Elsewhere, diamonds have been identified in a metamorphosed, layered, mafic-ultramafic complex near Kaya, Burkina Faso, in western Africa. This complex consists of metamorphosed dunite overlying a sequence of layered amphibolite, metaperidotite, metapyroxenite, and biotite-plagioclase gneiss (Helmstaedt 1993).

In Kalimantan, Indonesian Borneo, diamonds have been mined from alluvial deposits since the early part of the 1700s, and diamonds were also found in Upper Cretaceous and Eocene conglomerates and sandstones (Dawson 1967). The diamonds are mainly gems, and production averaged <10,000 carats/year (MacNevin 1977). One report in the *San Francisco Chronicle* (August 1, 1872) claimed that a nearly priceless diamond was found in this region. The diamonds were traced to Lower Cretaceous peridotite breccia that intrudes the Bobaris peridotite massif in the Pamali River area of southeastern Borneo. This breccia was investigated by Anaconda Minerals and found to be a wedge of scree derived from ophiolite (Dawson 1989). The Pamali breccias lie within a geosyncline and were intruded during a orogenic cycle (Dawson 1980).

Sinkankas (1959) reported that microscopic diamonds were separated from chromite from the Montreal Chrome pit, 2.2 km southeast of the south end of Little Lake St. Francis, Coleraine Township, Canada. Microscopic diamonds were also reported from chromite from Olivine Mountain, Tulameen River region, British Columbia, and also from Scottie Creek, Bonaparte River near Ashcroft, Cariboo district, Canada.

In Australia, diamonds have been reported in Devonian peridotite at Heazelwood in northwestern Tasmania (Wilson 1948). The peridotite is part of a complex with associated gabbro and pyroxenite. North of Tasmania in the Sydney basin of New South Wales, approximately 500,000 carats have been recovered from alluvial deposits and Tertiary conglomerates as a by-product of tin mining, and diamonds have also been found in several intrusive breccias including dolerite, nephelinite, nepheline mugearite, and alkali basalt (MacNevin 1977; Jaques et al. 1986; Dawson 1989). The New South Wales coast lies near the Australian plate margin (Dawson 1989).

Beni Bousera Massif, Morocco

One of the more intriguing peridotite massifs, located in Morocco, can be considered as a Rosetta stone for describing diamond concentrations in layered ultramafic rocks (or slabs of oceanic crust). This deposit, the Beni Bousera massif, may also provide an exploration model for the source of placer diamond deposits near peridotites along some continental margins. The deposit is part of an oceanic lithosphere slab that was part of a layered massif located at Beni Bousera, Morocco, near the Strait of Gibraltar (Pearson et al. 1989; Pearson, Davies, and Nixon 1993). Pearson et al. (1989), and Pearson, Davies, and Nixon (1993) reported that the massif, which is mainly pyroxenitic, was distinguishable from normal alpine-type ultramafic intrusions, which are peridotites.

The Beni Bousera massif is located in the Rif Mountains, northern Morocco, about 60 km southeast of Tetouan on the Alboran Sea coast. It forms a part of the Betico-Rifean fold belt. A continuous gravity high between Beni Bousera and the Ronda peridotite massif in southwestern Spain suggests that the Betics and Rif are part of the same tectonic system. Both the Beni Bousera and Ronda massifs contain many mafic layers of varying composition, which form 10% of the outcrop. The Beni Bousera massif forms a 50-km-long mantle slab composed of altered spinel lherzolite accompanied by rare harzburgites and dunite pods. Four garnet peridotite layers within the massif are formed of orange pyrope-almandine garnet and omphacitic pyroxene porphyroclasts. The diamondiferous zone at Beni Bousera lies within garnet pyroxenite.

The $^{40}Ar/^{39}Ar$ ages obtained in the metamorphosed country rocks of 21.5 Ma coincide with Sm/Nd isochron ages that reflect the closure of intermineral diffusion equilibrium during cooling associated with emplacement of the peridotite massif into the crust. It is consistent with 21.5 ± 1.8 Ma Sm/Nd isochron from the adjacent Ronda massif, which is thought to have an equivalent tectonic history.

Tectonic, geophysical, and isotopic constraints indicate the pyroxenites in the mantle wedge were formed above a subducting slab during the Cretaceous. Physical and chemical evidence precludes them from simply representing ancient subducted oceanic lithosphere, thinned by diffusion. However, the petrologic and isotopic diversity of the massif supports the concept of a "marble cake" mantle capable of producing the observed geochemical diversity seen in oceanic magmas.

The depletion of light rare Earth elements that is common in many of pyroxenites, their wide variation in composition and lack of correlation between incompatible elements, and fractionation indices preclude them from representing a crystallized peridotite source (Pearson, Davies, and Nixon 1993). Instead, the massif was derived by partial fusion of an upper mantle fragment at 1,500°C and 25 kbar (80 km depth) followed by reequilibration in the spinel lherzolite field at decreasing temperatures and pressures (Kornprobst 1969).

Graphite pseudomorphs after diamond were identified in the garnet clinopyroxenite (Pearson et al. 1989). The graphite octahedra are confined to four garnet clinopyroxenite magmatic cumulate layers in the ophiolite. Two of the layers are >2 m thick. They consist of orange pyrope-almandine garnet with compositions comparable to those found in diamond-bearing eclogites (Na_2O concentrations up to 0.14%) and omphacitic pyroxene porphyroclasts including minor plagioclase, spinel, and sulfides. The graphite occurs in a garnet cumulate layer in the lower portion of the layers. The graphite-bearing garnet clinopyroxenite layers, along with wehrlites, lherzolites, and diopsidites, form an intercalated horizon up to16 m thick at the apex of the massif.

The clinopyroxenes from Beni Bousera are aluminous sodic-augites to omphacites (with as much as 11.8 wt. % Al_2O_3 and 23 mole % jadeite component) and fall in the field defined by carbonaceous eclogites. Garnets ($Py_{34–57}Al_{29–48}Gr_{8–12}$) are within the range measured in diamond-bearing eclogite xenoliths. The garnetiferous rocks from Beni Bousera have undergone more extensive reequilibration than typical eclogite xenoliths, such that a significant proportion of the garnet has been exsolved from clinopyroxene.

The possibility of diamonds in ultramafic rocks from high-pressure metamorphic belts was predicted on the basis of similarities between these rocks and ultramafic xenoliths from diamondiferous kimberlites, after it was shown that eclogite xenoliths from kimberlite and related rocks may represent fragments of subducted oceanic lithosphere. Support for a connection between oceanic lithosphere and diamond was presented by Pearson, Davies, and Nixon (1993) who showed that graphite pseudomorphs after diamond in garnet clinopyroxenite (eclogite) of the Beni Bousera peridotite massif, Morocco, have inclusion minerals and carbon isotopic compositions similar to those of eclogitic (E-type) diamond from kimberlites.

Graphite forms up to 15% by volume of the layers and does not crosscut the layer margins in the surrounding peridotites, which contain no disseminated graphite. Graphite is also found in late stage, low-temperature, Cu-Ni mineralized veins, which crosscut the emplacement fabric of the massif. The graphite in the garnet clinopyroxenite is not associated with this mineralization.

Graphite in the garnet pyroxenite consists of sharp-edged octahedra with or without rounded, fibrous, graphite coats. In addition, four other types of graphite are present:

- rhombicubo-octahedra that exhibit well-formed cubic and octahedral faces
- contact twins or macles with or without reentrant angles, most of which are coated

- irregular to rounded masses with a surface morphology resembling framesite (phanerocrystalline carbonado)
- octahedra that have been flattened and/or sheared along the {111} face.

Raman spectroscopy indicates that both rim and core graphite in the pseudomorphs are well crystallized and hence of high-temperature origin. The graphite aggregates are thus similar to coated diamonds; the latter are characterized by well-formed cores surrounded by diamonds displaying a fibrous, radial growth and giving rise to the generally rounded external morphology of thickly coated diamond.

The graphite pseudomorphs are not only identical to various forms of diamond, but they even bear the characteristic crystal face patterns:

- Scanning electron microscopy (SEM) reveals that the {111} faces of the graphite octahedral cores display pronounced steps, which are compatible with the {111} growth plates observed on natural diamonds.
- The {111} faces also preserve small pits of trigonal symmetry, which have a geometry and orientation similar to the etched trigons on the faces of diamond from volcanic pipes.
- SEM observations demonstrate that the basal {0001} planes of the graphite crystals in the pseudomorphic cores are approximately parallel to the {111} faces of octahedra. In contrast, graphite crystals in the fibrous coat (rim) are much more irregular in their orientation relative to their basal planes, and they often lie at a high to perpendicular angle to the {111} surfaces of the core.

X-ray diffraction shows that graphite produced by experimental graphitization of natural diamond at 4.8 to 10 GPa pressure will produce graphite crystallites with {0001} planes parallel to the {111} faces of the original diamond octahedra. Single-crystal x-ray diffraction patterns of the octahedral graphite show a strong preferred orientation similar to that observed on the experimentally graphitized diamonds.

Graphite with no regular symmetry or form is widespread and probably represents highly deformed cubic forms. The regular crystalline form of diamond belongs to the cubic system, whereas natural graphite crystallizes in the hexagonal system. Thus, cubic graphite can be pseudomorphic after several cubic minerals: the most geologically plausible are diamond and spinel-group minerals such as chromite and magnetite. Submillimeter-size anhedral spinel grains do occur in some mafic layers in Beni Bousera, forming less than 1% by volume. However, the relatively low (450–1,000 ppm) Cr content of the bulk rock and low Cr and Fe^{3+} content of silicate minerals are not consistent with the rock having once contained up to 15% of either chromite or magnetite. Additionally, petrographic studies show no evidence of either mineral as relicts within the graphite aggregates, and ashed graphites confirm their absence.

All usual crystalline forms of graphite are displayed by natural diamond. Since the nature of the pseudomorphs is of crucial importance in verifying the potential for diamond in these rocks, some of the following details are significant.

According to Pearson et al. (1989), inclusions of clinopyroxene and garnet were found within the graphitic octahedra in thin section. These inclusions occupied a large proportion of the interior of some pseudomorphs. Some of these inclusions possess well-formed cubo-octahedral faces identical to many inclusions in natural diamond. It has been experimentally shown that cubic diamond can impose its influence on minerals of a lower symmetry, such as monoclinic pyroxene. The physical mechanism for such influence is not clearly understood, but this phenomenon is powerful evidence for a diamond

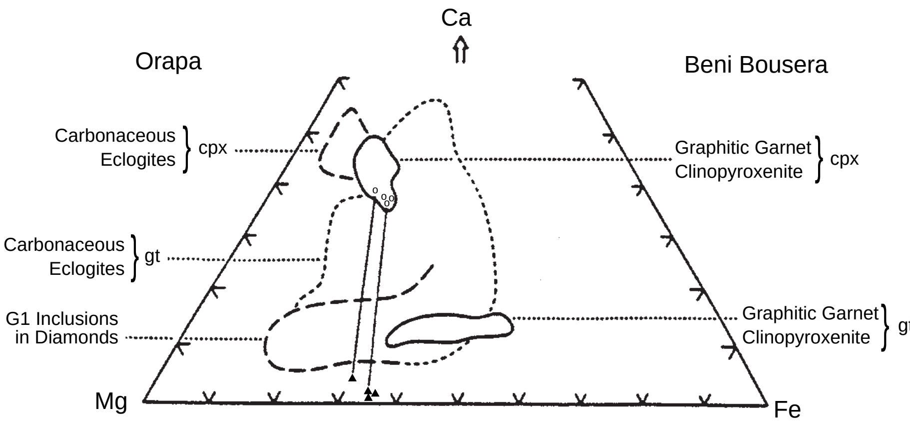

FIGURE 10.3 Ternary Ca-Mg-Fe atomic proportion plot of coexisting garnet-clinopyroxenites. For clarity the data are illustrated as fields. Circles = clinopyroxene, triangles = orthopyroxene. Fields of Orapa carbonaceous eclogites and inclusions in diamonds are shown (Pearson et al. 1989). Reprinted with permission from *Nature*.

precursor for the Beni Bousera graphite aggregates. It was also recognized that the chemical composition of pyroxene and garnet inclusions within the pseudomorphs from Beni Bousera are similar to pyroxene and garnets in eclogitic xenoliths from the Orapa kimberlite pipe, South Africa (Figure 10.3).

The host rocks at Beni Bousera are believed to represent portions of a hydrothermally altered, carbon-rich, oceanic lithosphere subducted to depths far exceeding the diamond stability limit and later subjected to decompression melting and graphitization during slow diapiric rise. Even though it appears that all of the diamonds reverted to graphite during emplacement of the Moroccan ophiolite, diamonds have been found in similar terranes elsewhere in the world. John Gurney from the University of Cape Town, for instance, notes some instances where diamonds have been reported from ophiolite sequences along plate margins. Diamonds have been preserved in an obducted ophiolite complex in Tibet as well as in alpine peridotites in Armenia and in the Koryak Mountains along the eastern coast of Russia. The source of numerous diamonds downstream from obducted ophiolite fragments in California is of interest (Hausel 1996).

Many of the typical exploration tools used to search for kimberlite would be useless in the search for diamondiferous ophiolites. However, on the basis of the Beni Bousera mineralogy, orange pyrope-almandine garnet and omphacitic pyroxene should be useful as tracer minerals to isolate the diamond deposits.

Although primary diamond was not preserved in the Beni Bousera slab, the concentration of the graphitized diamonds indicates that the slab initially contained about 15% diamond or approximately 10,000 times as many diamonds per unit mass of rock as any known high-grade kimberlite intrusive (Pearson et al. 1989). This grade is compatible with the diamond content of some small, hand specimen, diamondiferous eclogite xenoliths found in some kimberlites. It is our opinion, as in the case of the discovery of lamproite-related diamond deposit at Argyle, Australia, that the possibility of deposits with preserved diamonds similar to Beni Bousera can lead to a chain of significant new discoveries.

High-pressure peridotites have also been recognized in the Coast Range of southeastern Oregon and northern California, although no diamonds have been found in situ in these complexes (Medaris and Dott 1970). The mineral assemblages include forsterite, aluminous enstatite, chromian diopside, and $MgAl_2O_4$-rich spinel indicative of the spinel lherzolite field. Spinels from these peridotites are highly aluminous and comparable to spinels from Beni Bousera. The textural relationships indicate reequilibration at 1,200°C throughout a pressure range of 19 to 5 kbar (60–16 km depth). The peridotites are interpreted to represent oceanic upper mantle material emplaced in as a solid thrust wedge in the western margin of the Cordillera during the late Mesozoic (Medaris and Dott 1970).

In parts of western Norway, eclogites are associated with garnetiferous peridotite in a high-grade metamorphic basement complex. Carmichael, Turner, and Verhoogen (1974) suggested that these fragments might be tectonically transported mantle fragments. Mineralogically they resemble kimberlitic xenoliths. The eclogites have a high pyrope content (>55%). Other eclogites (with 30%–55% pyrope content) occur as lenses in high-grade metamorphic gneisses. They may have come from the mantle, or they may represent deep crustal gabbros that have been depressed to upper-mantle pressures and metamorphosed (Carmichael, Turner, and Verhoogen 1974).

MAFIC AND ULTRAMAFIC-ALKALINE VOLCANIC ROCKS

In Australia, diamonds have been reported in Devonian peridotite at Heazelwood in northwestern Tasmania (Wilson 1948). The peridotite is part of a 48-km-long, 3- to 8-km-wide complex with associated gabbro and pyroxenite.

North of Tasmania in the Sydney basin, New South Wales, approximately 500,000 carats of diamonds were recovered from alluvial deposits and Tertiary conglomerates as a by-product of tin mining, and diamonds have also been found in several mafic intrusive breccias (MacNevin 1977; Barron et al. 1994). The breccias include dolerite, nephelinite, nepheline mugearite, and alkali basalt (MacNevin 1977; Jaques et al. 1986; Dawson 1989). The New South Wales coast lies near the Australian plate margin and many of the intrusives lie within the Tasman fold belt, which is considered to be an atypical regime for kimberlite magmatism (Dawson 1989).

Some of the New South Wales diamonds were recovered from intrusive breccias interpreted to have sampled a diamondiferous subducted oceanic slab. Possibly, fragments of the slab were tectonically emplaced west of the diatremes, because alluvial diamonds in the Bungara field of northeastern New South Wales are thought to have originated from serpentinized peridotites in the Great Serpentinite belt. The diamonds at Bungara are found in channels near numerous faults and thrusts marked by serpentinite and other basic and ultrabasic rocks (MacNevin 1977).

Diamonds from New South Wales are unusually hard (Wilson 1948), contain mineral inclusions of coesite and grossular garnet (Sobolev 1985), and are the youngest diamonds dated. A set of five diamonds yielded an average date of 300 Ma (S.R. Lishmund, written communication, 1994). These diamonds are interpreted to have formed at relatively shallow depths (80 km) in a cool, subducted, organic-rich, oceanic slab (Barron et al. 1994).

Seventeen diamonds weighing up to 1 carat and possessing exceptional hardness were recovered from subbasal gravels and also from a dolerite (basaltic) dike in the Walcha region of New South Wales. The dike contained mantle lherzolite xenoliths that recorded pressures and temperatures of approximately 15 kbar and 1,000°C. In other words, the xenoliths record pressure–temperature (P–T) conditions outside the diamond stability field. The association of diamonds with nonkimberlitic rocks such as alkali basalts and ultrabasic intrusives is reaffirmed by examples worldwide.

Recently, diamonds were reported from a volcaniclastic komatiite from the Dachine region of the Inini greenstone belt of the Guyana shield of French Guyana (Capdevila et al. 1999). The diamondiferous komatiites are part of a Proterozoic island arc. It is suggested that the presence of diamond indicates that the komatiite originated at depths of 250 km or greater. The deposit is at least 5 km long and 350 to 1,100 m wide. Bulk samples of the rock yielded <1 to 77 diamonds per kilogram. The diamond population is dominated by microdiamonds; larger diamonds (>1 mm) are locally abundant. The largest diamond recovered measured 4.6 mm across.

The primary morphology of the diamonds is that of cubo-octahedron with low carbon isotope ratios ($^{13}C = -23$ to -27) suggestive of possible eclogite origin. Indicator minerals associated with the komatiite include a garnet population characteristic of lherzolite, with some subcalcic harzburgitic (G10) garnet and some eclogitic garnet. Other kimberlitic minerals (Mg-ilmenite, chromian diopside, and perovskite) are not reported. Chromite cores are poorer in Ti and richer in Mn that those typically associated with kimberlite and lamproite, and they are similar to spinels from greenstone belts. Most of the komatiites are highly altered and form finely foliated albite-carbonate-chlorite-talc schists, but primary volcanic textures are well preserved in some outcrops. Concentrations of immobile elements are very low, similar to those in komatiites and distinct from kimberlite and lamproite.

It is thought that the komatiite formed by melting of a hot, deep, mantle source, and the diamonds are xenocrysts derived from depths >150 km. Capdevila et al. (1999) suggest a genesis for the Al-depleted magma at depths >250 km.

Lamprophyres and Related Rocks

Diamonds have been found in mafic to ultramafic alkaline rocks (such as alnöites and minettes) that are lamprophyric in the broad sense. By definition, lamprophyres contain several generations of mafic phenocrysts (olivine, pyroxene, amphibole, biotite, and phlogopite), lack quartz or feldspar phenocrysts, and have a high volatile component. A lamprophyre is a dark porphyritic hypabyssal igneous rock characterized by panidiomorphic (euhedral) texture, a high percentage of mafic phenocrysts, and a fine-grained groundmass of mafic minerals in addition to some light-colored minerals (feldspars or feldspathoids).

Certain types of ultramafic lamprophyres, notably alnöites, contain deep-seated ultramafic xenoliths, and some have yielded minor amounts of diamond. Alnöites are lamprophyres composed of olivine, mica, augite, melilite (a feldspathoid of the composition $[Na, Ca]_2[Mg, Al][Si, Al]_2O_7$), nepheline and garnet (with or without accessory perovskite). Typically, these rocks have higher Al_2O_3 than kimberlite and lack the characteristic kimberlitic minerals such as pyrope garnet, Mg-ilmenite, and Mg-ulvospinel, as well as garnet lherzolite xenoliths. Characteristically, alnöites contain melilite and biotite mica (instead of phlogopite). Alnöites are characteristically found in rift environments. Kimberlites are found in cratonic (interrift) environments.

Apparently 10 diamonds (0.06 carats) were found in an alnöite within the Ile Bizard intrusive, which is part of a carbonatite complex about 6 km west of Montreal. The diamonds were recovered from a small bulk sample in 1968. However, there is some question as to whether the sample was contaminated during milling in Johannesburg.

Potassic Basalts

Another rock type that has been reported to contain accessory diamond is alkaline potassium-rich basalt. Kaminsky (1984) described diamonds in basalts in the Sikhote-Alin' area (Russian Far East). According to current concepts, the composition of primary alkaline basaltoid magmas is determined by depth of selective melting and the degree of melting of an original pyrolite.

Alkaline basaltoid magmas are generated within the pressure range of 1.5 to 3.5 GPa by the process of melting 3% to 25% of an original pyrolite containing 0.1% H_2O (Green 1971). In explaining of origin of such basalts, Kaminsky referred to the fact that there was a correlation between the depth of foci of earthquakes within Benioff zones and the alkalinity of basalts, and that this depth correlates with the depth of origin of diamonds.

For alkaline basalts, as for alpine-type ultrabasites, Kaminsky (1984) attempted to show that the existing data for P–T conditions for magma generation of alkaline potassium basaltoids do not contradict known constraints for diamond crystallization. The magmas are generated at an equivalent pressure of about 2.0 to 3.0 GPa. Using mineral associations in ultramafic inclusions to determine the temperature and pressure of magma generation for potassium-enriched alkaline basalts, Kaminsky determined that these formed at 3.1 GPa pressure and 1,150°C temperature for the higher-temperature inclusions in basaltoids from New Zealand and the Solomon Islands. These temperatures and pressures are within the range of P–T conditions required for diamond formation.

The success of Kaminsky's studies has been due to the treatment of large samples using thermochemical leaching, which provides a complete separation of the heavy mineral fraction. These samples are extremely costly to transport and the process is labor intensive. The application of thermochemical extraction, although it provides great advantages for preservation and extraction of small mineral grains (0.3–0.5 mm), is generally not used by exploration companies or scientific teams in the West, owing to the size of the sample (several hundred kilograms) that must be collected and transported to a lab for treatment.

Minettes

Although only a few decades have passed since lamproites were considered unusual sources of diamond, another rare rock type appears to display more and more indications that it might someday be considered as a common potential host rock for diamond. This rock type, minette, has an appearance similar to that of some lamproites.

By definition, minettes are porphyritic lamprophyres with biotite phenocrysts in a groundmass of biotite and alkali feldspar (Bates and Jackson 1980). Mitchell and Bergman (1991) describe minettes to be mildly potassic rocks (K_2O/Na_2O molar = 1.0–1.5) with biotite (or phlogopite) and clinopyroxene phenocrysts set in a groundmass of sanidine/orthoclase. They suggest that minettes can be distinguished from lamproite on the basis of their mineralogy and geochemistry. Minettes may also have minor to accessory olivine, Cr-spinel, salite, titanomagnetite, apatite, calcite, analcime, and amphibole.

Minette differs slightly from lamproite in the mineralogical and chemical composition of its parent magma. Leucite ($KAlSiO_3$) is replaced by feldspar ($MAl[Al,Si]_3O_8$), where K, Na, Ca, Ba, Rb, Sr, and Fe are assigned to M. Possibly the most significant difference between minette and lamproite is the composition of the volatiles in the magma.

Considering the similarity in compositions, it is no wonder that minettes are spatially associated with lamproites and kimberlite-like rocks and that they form similar geologic bodies (dikes and diatremes). Minettes require a different mantle source than lamproites, shallower depths for genesis, and crustal involvement.

Some minettes are reported in the Navajo volcanic field of the Four Corners region along the margin of the Colorado Plateau. Within this field is a series of volcanic bodies—Buell Park, Mule Ear, Moses Rock, and Cane Valley. These intrusive centers are composed of different types of potassium-rich rocks (including katungites, monchiquites, olivine leucitites, alnöites, minettes, and kimberlitic tuffs) and are described as being related to minettes (Roden 1981, Mitchell 1986). At Garnet Ridge within this region, lapilli tuffs are reported that are more characteristic of lamproites than kimberlites (Meyer 1976). Locally, these rocks contain both chromium diopside and pyrope garnet (Allen and Bulk 1954), although no diamonds have been found in the field.

Elsewhere, diamonds have been recovered from minettes. For instance, diamondiferous minettes are reported in Canada (MacRae et al. 1995; Kaminsky et al. 1998a). One of these, the Parker Lake minette, yielded an extremely rich sample of diamonds. Reports indicate that a large number of diamonds were recovered from a small sample (>1,500 diamonds from a 22 kg sample) (MacRae et al. 1995).

Diamondiferous minette dikes are also reported with lamproites in the Jharia field in India (Rock et al. 1992). Other deposits are reported, and we expect that others will be found as research continues in this field.

Ultramafic-Alkaline Volcanic Rocks

Common minerals in alkaline rocks include nepheline, melanite, and melilite. The compositions of alkaline rocks range from carbonatite to ultramafic to mafic free, which suggests all alkaline rocks can be derived from the same mantle source.

Most alkaline rocks are thought to result from partial melting of metasomatized mantle. Metasomatism is the process in which the chemical composition of a rock is substantially changed through metasomatic alteration of its original constituents. The process involves practically simultaneous capillary solution and deposition by which a new mineral of partly or wholly different chemical composition may grow in the place of an old mineral. The presence of interstitial chemically active pore fluids or gases within the rock is essential for the replacement process.

The diversity of alkaline rocks results from complex combinations of igneous fractionation, mixing, assimilation, and metasomatism. Mafic to ultramafic alkaline lamprophyres including kimberlites, lamproites, and minettes are components of belts of alkaline rocks where the ultramafic and nephelinite carbonatite families occur. The unusual chemical compositions of these rocks probably result from melting of variably metasomatized mantle protoliths that are highly enriched in K and other incompatible elements (Ba, Rb, Nb, rare earth elements), under a variety of P–T conditions. Any potassic (molecular K > Na) mafic to ultramafic rock should be considered potentially diamondiferous.

Mineralogically, diamond-bearing rocks are usually rich in olivine, clinopyroxene, phlogopite, and carbonate. Rocks with primary calcic plagioclase (alkali basalt family) typically do not contain diamond, although some rare alkali basalts (i.e., Mountain diatreme, Canada) are reported to contain microdiamonds. Other primary minerals not compatible with diamond include nepheline, feldspathoids, zeolites, and melilite.

High-Alumina and Calc-Alkaline Basalts

Diamonds were reported in basaltic rocks as early as in 1906, when a single diamond crystal was found in a 6-m-thick dike of amphibole diabase cutting Middle Carboniferous granite in New South Wales, Australia (David 1906). Later three other diamonds were found in the same diabase. The chemical composition of the diabase corresponds to the composition of a normal diabase having 3.7% alkali content. In the same region, diamonds were later reported in a doleritic dike (Thompson 1909). The validity of these discoveries was questioned because the samples were collected in an adit cutting the dikes, and the adit intersects a diamond placer. Follow-up sampling produced no positive results.

Rumors of diamonds in basalt were quite popular at the turn of nineteenth century in South Africa. Harger (1910) reported finding diamonds in amygdaloidal porphyrites bought from local miners. In two pebbles, diamonds weighing 0.5 carats were found. It has been suggested that the pebbles were from widespread tillites from the Dwyka Series in the region. On the basis of petrographic characteristics, the pebbles are identical to amygdaloidal diabases of the Wentersdorp Series.

Another discovery was made by a Dutch petrographer during thin-section preparation of samples collected by G. Kimmerling, the chief of the Indonesian Volcanological Service (Gisolf 1923). Gisolf noted numerous isometric and isotropic particles in the samples that were harder than corundum. The sections were of a basaltic bomb collected from the Gunung Ruang volcano north of the Island of Celebes (Sulawesi), Indonesia (Kaminsky 1984). Nowadays these findings are considered apocryphal and not reliable.

Another intriguing discovery of diamonds was apparently made in Quaternary basalts from Kamchatka. In 1969, a geologist (Fareed Kutiyev) reported finding diamonds in the heavy fraction of basalts. Felix Kaminsky confirmed the diamonds. However, doubts of this very unusual discovery were so great that it was not published until 1975 (Kutiyev and Kutiyeva 1975). The text of the article was accompanied by a comment of academician and internationally known diamond geologist, Vladimir Sobolev, that he saw the diamonds and confirmed their existence, but that he could not make any connection between diamonds and plagioclase. The presence of feldspar (in this case plagioclase) does not contradict current petrologic concepts about the physicochemical conditions of the origin of feldspar and diamond. On the contrary, it has been shown that some plagiobasalts are generated within the mantle. It should also be realized that certain magmas known for diamonds, such as minettes and lamproites, contain feldspar (although in these cases, the feldspar is sanidine).

Six elongated diamonds were recovered that measured 0.4 to 0.7 mm along their long axis. The diamonds were extracted from the heavy nonmagnetic fraction of a 40-kg sample of high-alumina basalt in the early Quaternary Maly Payalpan basaltic shield volcano, northwestern Kamchatka. In the course of this study, intergrowths of diamond and anorthitic plagioclase were found. In the same fraction, grains of pyrite, chalcopyrite, apatite, rutile, magnetite, ilmenite, and pale-pink garnet were described. The diamond and garnet were confirmed by x-ray analysis (Kutiyev and Kutiyeva 1975).

Speculating on the origin of the diamonds, Kutiyev and Kutiyeva (1975) insisted that the capture and assimilation of deep-seated diamond-bearing rocks in the basalt was unlikely, because there were no traces of contamination such as xenoliths, there were no observed lenses of ultramafic rock, and the diamond facets bore no traces of corrosion. It has been suggested that plagioclase was captured in the course of later phases of crystallization. Such capturing is possible in the process of phase transition, i.e., diopside + enstatite + spinel ⇒ anorthite + forsterite.

One constant feature of the morphology of the diamonds is elongation of the grains. The absence of characteristic rounded forms and, instead, the elongated and flattened form of the diamond grains are considered to indicate crystallization under oriented pressure (stress) in the presence of a gas rich in carbon. Kutiyev and Kutiyeva (1975) hypothesize that the diamonds crystallized in the basalt under great stress within a Benioff zone.

The regional geologic office assigned two experienced local geologists, Valery Sheimovich and Anatoly Patoka, to repeat the sampling. Their results were published in 1979 (Kaminsky et al. 1979) and essentially legitimized the Kutiyev's earlier work, because four more diamonds were found in samples of 3 to 5 kg. The Maly Payalpan volcano was renamed "Almazny" ("diamondiferous" in Russian). Speculation about the origin of the diamonds in the volcano involves the origin of the magma and its relationship to the Benioff zone.

Later, Kutiyev sampled the volcanic rocks of the Zhupanovsky volcano-tectonic structure in eastern Kamchatka but found no diamonds. In 1980, a large (500-kg) sample was taken in the valley between the Avachinsky and Kozel'sky volcanoes. The rocks were very dense and unusual olivine basalts, which contained abundant small phenocrysts of chrisolite, augite, and chrome diopside. These rocks were named "avachites" after the geographic location. An 80-kg fraction of this sample was crushed and studied for accessory minerals, but even in this comparatively small volume, three diamond crystals, about 2 mm in size, were recovered (no information on morphology is available). In 1993 a duplicate sample of avachite (150 kg in weight) contained 26 grains of carbonado (0.1–1.0 mm). One grain was 3 mm in diameter (Smishlyaev 1999).

As it turns out, the description of the "avachite" is identical to the description of the diamond-bearing basaltic bomb published by Gisolf (1923). To date, the existing theories are unable to explain why these diamondiferous basalts exist, as the physicochemical constraints on the genesis of diamond and the origin of the basalts do not appear to be compatible. More research on these deposits is needed, because the possible results of such studies can provide new insights in some important petrologic concepts. Possibly, the parent plagiobasaltic magma is influenced by unique volatile conditions in the upper mantle, as is the magma that produces minette. Here nature takes us to the threshold of the unknown when one considers the occurrence of diamonds in metamorphic rocks and in "astroblemes," as well as diamonds in meteorites as described in Chapter 12, Specific Type of Ring Structures (Astroblemes).

Metakimberlites

Another rock of interest for diamond has been described as metakimberlite. Metakimberlites occur as dikes in the Ivory Coast and northwest Gabon of Africa. The dikes are believed to be of Precambrian age (1.4 Ga and 1.9 Ga, respectively), are devoid of the usual kimberlitic indicator minerals, and have the appearance of talc schist or mica schist. In addition to talc and phlogopite, they contain chromite, anatase, apatite, zircon, and Mn-ilmenite.

The Bobi metakimberlite dike, Ivory Coast, is locally diamond rich and has grades as high as 1,000 carats/100 tonnes (some reports indicate grades as high as 3,000 carats/100 tonnes). The close association of metakimberlite with lamproite has been noted at

Bobi, and it suggested that the metakimberlite actually represents a highly altered olivine-phlogopite lamproite rather than metamorphosed kimberlite.

We also include the Norris "kimberlite" north of Knoxville, Tennessee, in the category of metakimberlite. Although no diamonds have been confirmed from the rock, two diamonds were apparently recovered from gravel downstream from the two intrusives. Meyer (1976) described these intrusives as kimberlites. The rock is foliated and contains phlogopite, vermiculite, hematite, ilmenite, serpentine, carbonate, and tiny pyrope garnets.

Phyllite Dikes

Diamond-bearing phyllitic dikes appear to be widespread in the region of Diamantina, Minas Gerais province, Brazil. These dikes appear as highly altered schistose rocks and form sharp contacts with their hosts, and locally they cut their host rocks at acute angles. But for the most part, the dike swarms parallel the strike of the host. The dikes are composed of sericite and chlorite developed along a substratum containing relicts of porphyroblasts. The dikes form clusters oriented along the general strike of dislocated host rocks. Diamonds recovered from the phyllitic dikes are typically dodecahedral.

A suite of minerals (xenotime, zircon, magnetite, and tourmaline) in phyllitic dikes that are typically associated with granite leads to the speculation that phyllites represent altered granulites. However, some researchers assume that the Brazilian phyllites are highly altered kimberlites. Similar rocks in the Ivory Coast, Africa, are widespread and interpreted as highly altered kimberlitic dikes. The degree of diamond mineralization and the form of diamond crystals of the Ivory Coast intrusives are similar to those in Brazil (Zubarev 1989). Similar intrusives may represent the original source of placer diamonds found in the Birim area, Ghana (Kaminsky et al. 1998b).

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CHAPTER 11

High-Pressure Metamorphic Complexes

Diamonds affiliated with high-pressure metamorphic terranes were discovered in Dabie Shan, China, west of Shanghai (Xu et al. 1992), Mongolia, and in the Kokchetav region of Kazakstan.

Other microdiamond deposits have been reported in metamorphic rocks in the Western Gneiss region of Norway (Dobrzhinetskaya et al. 1995), and in Saxonian Erzgebirge, Germany (Massonne 2000). On the basis of the geologic evidence, the metamorphic diamonds appeared to show close affiliation with ultrahigh-pressure collision belts. In addition, some of the deposits possess extremely high ore grades, although the diamonds are small. As interest and exploration increase, larger diamond porphyroblasts may be found.

DABIE SHAN, CHINA

Syntectonic diamonds in ultrahigh-pressure metamorphic rocks have been described in the Dabie Mountains west of Shanghai in eastern China. The Dabie Shan is one of the better-known metamorphic diamond occurrences, although most studies on these occurrences have concentrated on the tectonics and structure of the collision zone, and research on the diamonds has been of only passing interest.

The Dabie Shan site is located in the eastern Qinling-Dabie orogen, which is the site of collision of several continental plates (Xu et al. 1992). The rock types include eclogite, garnet pyroxenite, and rare jadeitic bands intercalated with, or enclosed within, leucocratic gneiss and marble. The collision orogen is about 2,000 km long, and ultrahigh-pressure metamorphic rocks occur in the three easternmost ranges: Hong'an, Dabie, and Su-Lu. Eclogitic metamorphic facies occupy an area of about 1,000 km^2; they are believed to have formed by north-directed subduction of the Yangdsu craton or a microcontinent beneath the Sino-Korean craton (Liu and Hso 1989) (see Figure 11.1).

Diamonds occur as inclusions in some garnets in eclogites and garnet peridotites. More than 20 diamond crystals were found in polished thin sections. Most of the grains are 10 to 60 microns across but some of the largest reach 240 microns. The grains are typically anhedral but a few form cubes or dodecahedra. Another 20 diamonds were extracted; most were cubes and octahedra and a few were tetrahedra. These diamonds were about 150 microns in diameter although one was 700 microns.

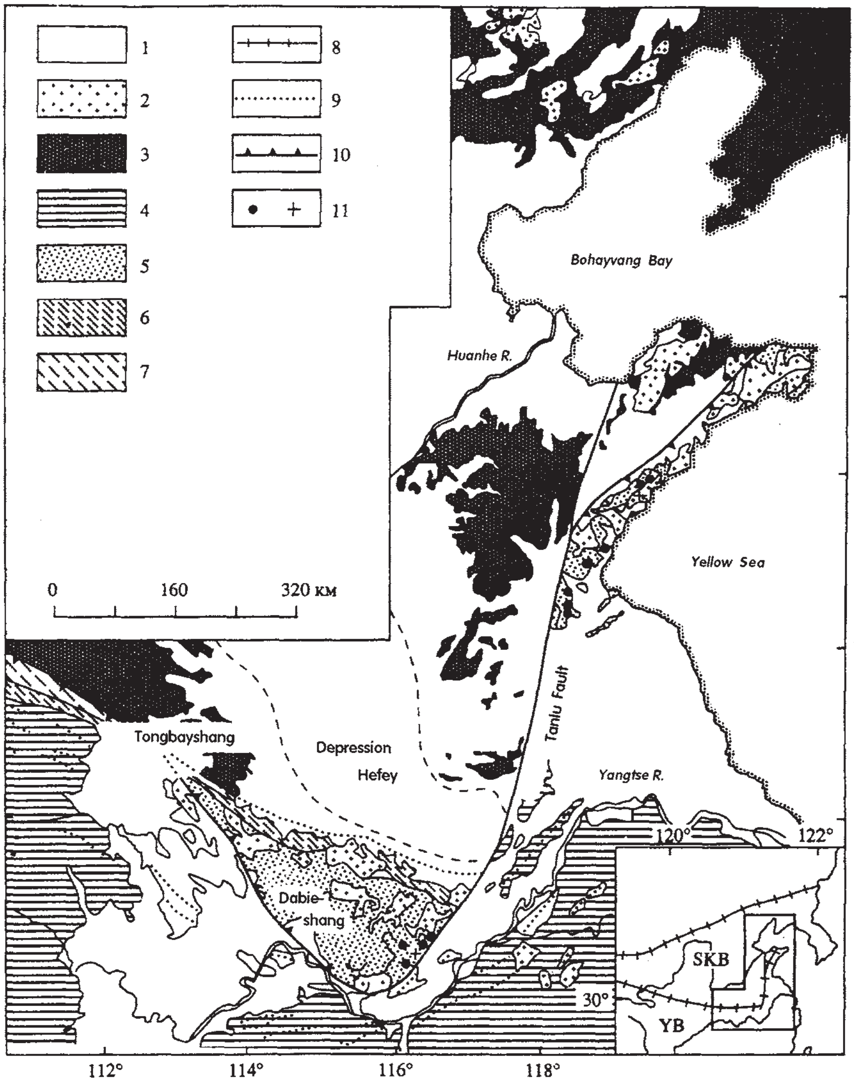

FIGURE 11.1 Geologic position of ultrahigh-pressure complexes of eastern China. 1 = Jurassic and Quaternary deposits, 2 = Late Mesozoic granitoids. Undivided Proterozoic-Triassic (?) sequences of 3 = SKB (Sino-Korean shield) and of 4 = YB (outcrops of rock sequences of the Yangdsu craton), 5 = Dabie Shang high-pressure complexes and its equivalent in Shandonte, 6 = the Rosiling complex, 7 = the Quingling complex, 8 = the zone of joining of the Sino-Korean shield and the Yangdsu craton, 9 = axes of major Mesozoic fold structures, 10 = suggested zone of late Mesozoic collision, 11 = coesite and diamond sites (Perchuk et al. 1995). Reprinted with permission from Petrologiya Journal, the Russian Academy of Science.

The radiometric age of the diamondiferous metamorphic rock was established using Sm-Nd determinations from three samples. The results yielded 240 ± 0.2 Ma, 224 ± 20 Ma, and 221 ± 142 Ma (which generally coincides with a major kimberlite episode and a flood basalt province that is widespread in Siberia (see Chapter 14, Temporal Distribution of Diamond Deposits). Along with the diamonds in garnets were other mineral inclusions that indicate high pressure, in particular coesite. The metamorphic conditions under which diamond and coesite were formed indicate a pressure of about 40 kbar and temperature of about 900°C, which is lower than that considered for diamond stability associated with kimberlite in cratonic environments (45–55 kbar and 1,050°–1,200°C). It is deduced that the original rocks were subducted to the depth of about 100 km with subsequent rapid uplift. Thus, Dabie Shan is thought to be a collision orogen between two continental plates. The area occupied by the diamondiferous rocks likely corresponds to the basement of the Yangdsu plate.

V. Vaganov (oral communication to Erlich, 1993) reported that similar microdiamond occurrences, described by Mongolian geologists, were also found in the eastern part of central Mongolia.

GEOLOGIC PROVINCE

Diamonds were initially described in the literature from thin sections of metamorphic rocks from the Kokchetav massif, in northern Kazakstan, by Sobolev and Shatsky (1987). However, the diamonds were actually found more than 10 years earlier, and placer diamonds were discovered nearly 20 years earlier (Kashkarov and Polkanov 1972). The discrepancy in the time between the discovery and publication was due to the change in the official positions of the authors. Up to the time of *perestroika* (restructuring of Soviet political system under the rule of Mikhail Gorbachev) in the former Soviet Union, any information concerning exploration for diamonds was classified and publication was forbidden. Thus publication about the discovery became possible only because of the high positions that were later obtained by the authors.

Local geologic groups recovered large (10 m^3) samples of material from the deposit (Essenov et al. 1968). The consequent treatment of these samples involved crushing and flotation, and significant amounts of diamonds were recovered and examined in the diamond laboratory of the Central Institute for Exploration for Noble Metals and Diamonds in Moscow. The diamondiferous ore and adjacent country rocks were extensively trenched and drilled to depths of several tens to hundreds of meters and were mapped in detail by geologists of the Kokchetav expedition.

Microdiamonds (average size only 12.5 microns) were found in high-pressure gneissic rocks of the Kokchetav massif. The diamonds occurred as inclusions in garnet and zircon in garnet-biotite paragneiss and paraschist in the Kokchetav massif.

A group of high-pressure minerals was found in association with the diamonds that can be considered as mineralogic and geochemical indicators. These include coesite, high-K_2O clinopyroxene, high-Al_2O_3 sphene, phengite (a silica-rich muscovite), and grossular-pyrope garnets. The diamondiferous rocks are also remarkably enriched in graphite. Studies of the various rock types have shown that different rock types host different diamond morphologies. The microdiamonds occur as cuboids, cubo-octahedra, octahedra, skeleton crystals, and reentrant crystals. Twins and aggregates of cuboids, cubo-octahedra, and dodecahedra are also found. Each rock from the massif that contained microdiamonds exhibited a distinctive range of diamond morphology, nitrogen abundance, color, inclusion content, and C isotope composition.

The diamonds from the garnet clinopyroxenites, zoisite gneiss, and dolomitic garnet clinopyroxenites typically contain water and carbonate inclusions, and in this respect they differ from nearby alluvial diamonds that do not contain these inclusions. The lack of inclusions and the particular nitrogen content and isotopic signature of alluvial diamonds suggest that they originated from a different rock type than those found in the Kokchetav massif.

The inclusions provide evidence that the diamonds grew in a fluid phase enriched in C-O-H. The carbon isotopic signatures of the microdiamonds show a restricted range of ^{13}C values between −15 and −10. This range is different from that of most peridotite and eclogite diamonds of mantle origin, which commonly yield characteristic mantle values of −10 to 0 that are commonly detected in peridotite and eclogite diamonds in xenoliths in kimberlite. Metasediments therefore appear to be the most plausible source of the carbon for the microdiamonds.

General Geology

The northern Kazakstan diamond province extends nearly 700 km; it reaches 250 km wide in the northwestern part of the province and about 100 to 150 km wide in the southeastern portion of the province. The easternmost microdiamond deposit that has been recognized in this province is located in northeastern Kazakstan near the city of Pavlodar. The westernmost sites are located around the median of the Kokchetav massif (Figure 11.2).

A characteristic feature of the geology of this province is widespread development of metamorphic complexes that form a series of median massifs within geosynclinal Paleozoic sequences. Sedimentary, magmatic, and metamorphic complexes are recognized in the province; they include eclogite-gneiss complexes of Precambrian age and sandstone-claystone rocks of Paleogene age.

The most intensely investigated site where diamond has been identified lies in the northern part of the Kokchetav massif; it is composed of pre-Vendian (uppermost Proterozoic, 560–1,100 Ma) sialic basement lying along the continental border of the Kazakstan-Siberian pre-Vendian and early Paleozoic oceanic basin. The Kokchetav massif is located at the western continental slope. The Kumdikol industrial microdiamond deposit and the lesser-known Barchinsky microdiamond site are located about 15 km northwest of Kumdikol (Shatsky et al. 1991). The diamond content of both the Kumdikol and Barchinsky deposits averages in the tens of carats per tonne.

Much of the region is underlain by rocks of the Early Proterozoic Zerendinskaya series. This series is composed of garnet-biotite gneisses and schists with variable amounts of andalusite, kyanite, sillimanite, and muscovite. Nearly 70% of the series is composed of metapelites and lesser eclogite. Two distinct units are recognized in the series: an upper unit that includes the Zerendinsky series metapelite schists and gneisses without eclogite, and a lower unit represented by a tectonic melange composed of eclogite.

The melange contains fragments of post-Vendian and pre-Vendian shallow-marine metasediments that are mixed with eclogites derived from obduction onto a passive continental margin, most probably in Early Cambrian time. The relationships between the two units are not clear owing to extensive overburden in the region. Allochthonous quartzites of the Kokchetav unit and Cambrian-Ordovician volcano-sedimentary rocks

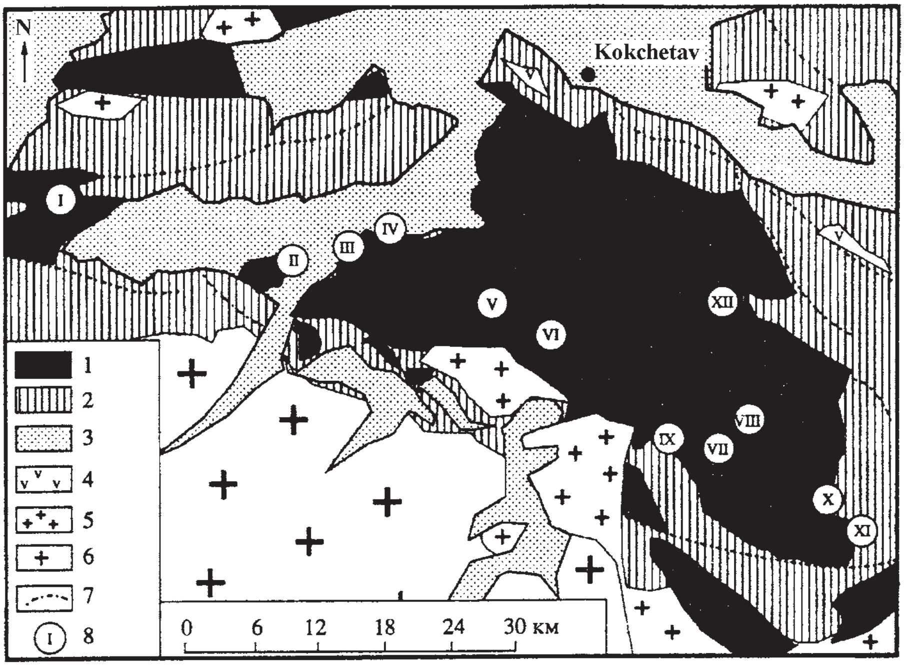

FIGURE 11.2 Simplified geological map of the northern part of Zerendinsky pluton, with host rocks of the Kokchetav massif, northern Kazakstan. 1 = Proterozoic (Pr_{1-2}) metamorphic rocks of the Zerendinsky rock series, 2 = undivided Proterozoic and Paleozoic formations, 3 = Cenozoic formations, 4 = Proterozoic mafic intrusive rocks, 5 = Paleozoic granites, 6 = granitoids of the Zerendinsky pluton, 7 = faults, 8 = areas of eclogite exposures and high-pressure micaceous schists: I = Leninsky, II = Kumdikol, III and IV = Chaglinka, V = Sulu-Tyube, VI = Yenbekbyrlik, VII–IX = Kulet, X and XI = Karabulak, XII = Uyaly (Perchuk et al. 1995). Reprinted with permission from Petrologiya Journal, the Russian Academy of Science.

both metamorphosed under greenschist facies conditions and were thrust above the protoautochthonous gneisses and melange. During the Late Ordovician and Early Silurian, multistage granites were emplaced that formed a series of granitic domal structures surrounding the Kokchetav massif. A Devonian-Carboniferous molasse lies unconformably on these units.

The Barchinsky diamond deposit, 15 km west of the Kumdikol deposit, is located in the lower part of the Zerendinsky series where the series is cut by a deep-seated fault, the Krasnomaisky fault. This area is characterized by widespread development of Early Cambrian pyroxenites and by syenites of the Krasnomaisky ultramafic-alkaline complex.

Diamondiferous Rocks and Diamonds

Some known diamond deposits in Kazakstan are located along the margin the Kokchetav massif. This massif is characterized by the development of a melange of rocks with differing degrees of metamorphic grade ranging from eclogite to amphibole facies. The different stages of metamorphism within the Kumdikol area of northern Kazakstan were reconstructed by Udovkina (1985) (Figure 11.3).

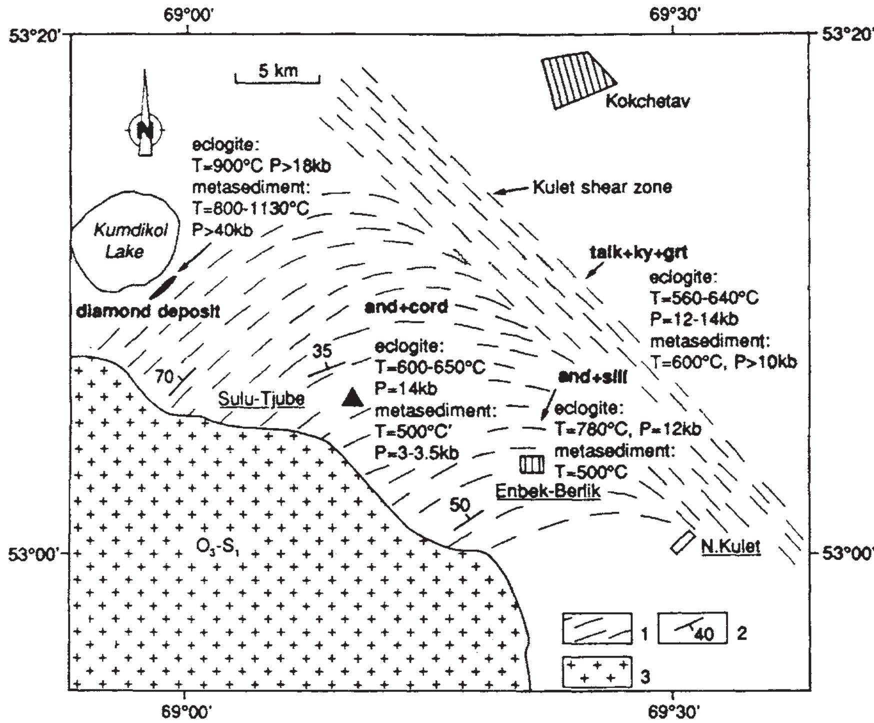

FIGURE 11.3 Schematic drawing of metamorphic conditions in the rocks of the Zerendinsky series (southern part of the Kokchetav massif). 1 = metamorphic rocks of the Zerendinsky rock series, 2 = the trend of the main planar structure and general foliation, 3 = Late Ordovician–Early Silurian granites (Dobrzhinetskaya et al. 1994). Reprinted with permission from Elsevier Science.

Characteristically, the diamonds in this deposit show no connection with the eclogites. Instead, the diamonds lie within ductile and semiductile shear zones that cut the eclogite facies rocks at acute angles (Figure 11.4).

According to Lavrova et al. (1996) microdiamonds (<1 mm) were found in and along the grain boundaries of virtually all of the rock-forming minerals: garnet, quartz, biotite, phlogopite, diopside, zoisite, and kyanite, and in accessory zircon and secondary sericite-calcite aggregates (replacing clinopyroxene). Diamond is also preserved in chlorite-sericite aggregates that replace garnet, and it is found in thin sericite pseudomorphs after plagioclase in garnet-biotite and garnet–two-mica gneisses.

Rose-like and cuboid skeletal and unshaped diamond grains frosted by many tetragonal etch pits predominate in the calc-silicate rocks. The diamondiferous lenses cut foliation in the metasedimentary rocks at a 25° to 30° angle. This pattern reflects the geometric features of the internal part of the shear zone. Maximum microdiamond and graphite content occur in the highly strained rocks. Microdiamonds occur in eclogitic facies rocks only where shear zones cut the latter. The highest concentration of microdiamonds is spatially concentrated along specific surfaces of the main shear zone.

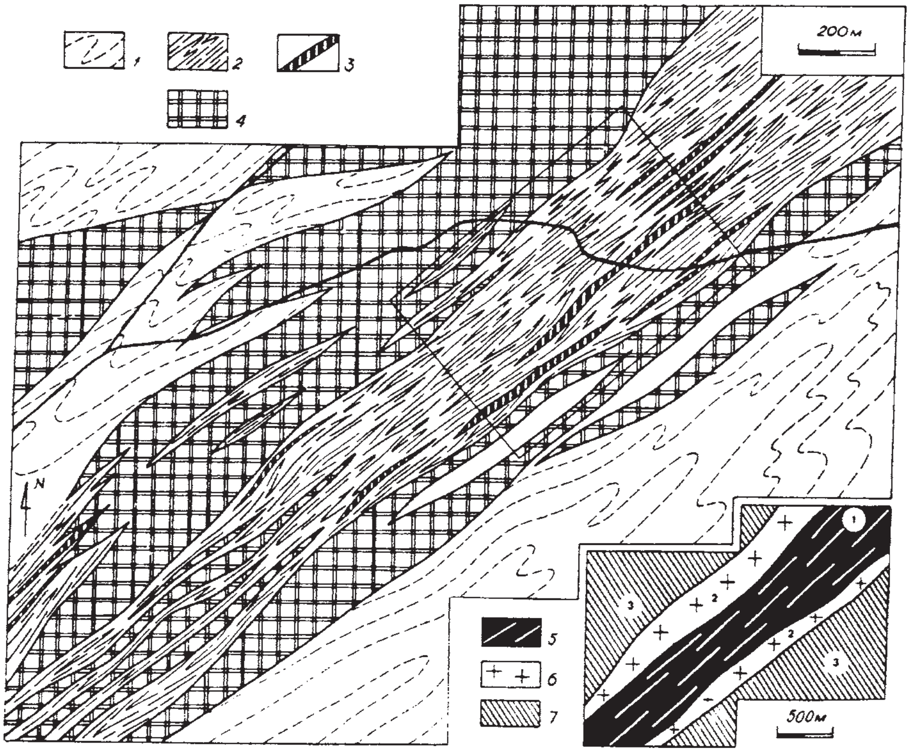

FIGURE 11.4 Tectonic map of the Kumdikol microdiamond deposit. 1 = synmetamorphic polyphase folding, 2 = strongly sheared rocks with relics of subisoclinal folds and superimposed foliation S3 and S4, 3 = tectonic lenses of marble and calc-silicic rocks, 4 = prekinematic granites and migmatites. Zones inset into the main structural zone: 5 = strike-slip shear zone within which microdiamond ore body is located (zone 1), 6 = prekinematic granite and migmatite (zone 2), 7 = areas of synmetamorphic folding (zone 3). Boxed area in the center of the figure is shown in more detail in Figure 11.6 (Dobrzhinetskaya et al. 1994). Reprinted with permission from Elsevier Science.

Microdiamonds at the Barchinsky site northwest of Kumdikol are of the same size and morphologic types as those at Kumdikol but have better crystallographic shapes. Lavrova et al. (1996) indicated that the morphology varies in accordance with the type of diamond-bearing rocks. Flat-facet diamonds are usually found not in quartz microgneisses as at the Kumdikol deposit but rather in apogneissic metasomatites and carbonate rocks. Cubes and combined crystals are prevalent in metasomatites with complex compositions and structure. Octahedrons and aggregate crystals are found in the carbonates. DeCorte et al. (2000) reported that diamonds in garnet clinopyroxenite and dolomitic garnet clinopyroxenites are predominately cuboids. Biotite gneisses are characterized by cubo-octahedral diamonds, although cuboids and growth forms transitional to cuboids also exist. Some zoisite gneisses (devoid of symplectitic zoisite) host octahedra, whereas others contain cuboids.

In thin section, many irregular diamond grains show thick graphitic rims. Thus the diamonds show the effects of changing P–T conditions.

Graphite is a typical mineral satellite of metamorphic diamond deposits. The typomorphic characteristic of graphite in these deposits is considerable imperfection of crystal structure. These crystals have a considerable amount of an amorphous phase, and they show interplane distances characteristic of the carbon phase chaoite. Another reported characteristic feature is the increased density of the graphite.

Graphite forms syntaxial intergrowths with diamond. When the crystalline structures of graphite and diamond are similar, it is a direct indication that both minerals formed under similar conditions. The considerable imperfection of graphite's crystalline structure does not permit one to accept its origin under high-pressure conditions of metamorphism (Shumilova 1995a, 1995b).

In 1992, metamorphic diamonds were also discovered northwest of the Ulaan Baatar region of northern Mongolia by the Central Institute of Exploration in Moscow. Initial bulk sampling tests recovered diamonds in the range of <0.1 to 1 mm in diameter at grades of 40 to 100 carats per tonne! The type of diamond and structural location is similar to what has been described in northern Kazakstan deposits.

In 1985, a block of garnet amphibolite was found in the alluvium of the Dzhalty River of the Russian Far East that contained two clear, colorless cuboidal diamonds. The diamonds were different from the diamonds of the Kumdikol, north Kazakstan, eclogite-gneissic complex but similar to diamonds from some kimberlites (Romashkin 1997).

Genesis

It has been suggested that the diamonds in both the Chinese and Kazakstan deposits resulted from early kimberlite eruptive rocks that were eroded, allowing many diamonds to be incorporated into adjacent sedimentary rocks that were later metamorphosed. Other theories have suggested that the diamonds formed during subduction followed by subsequent emplacement at the surface owing to exhumation of the crustal slab. Another theory suggests that they were formed in place by metastable growth.

Many researchers have been inclined to use a standard "plate tectonic" approach to explain these deposits. In this theory, the diamonds would have formed as the result of subduction of a continental crustal slab to the depth of 150 km (equivalent to a pressure >40 kbar and temperature of 900°–1,000°C) followed by rapid exhumation of the slab in order to preserve the diamonds (Shatsky et al. 1991; Dobretsov, Theunissen, and Smirnova 1998). However, the presence of diamonds with a much more restricted carbon isotope signature than eclogitic diamonds does not support this origin.

Kazakstan geologists involved in field investigations of these deposits believe that the pattern of major local geologic structures controls the spatial distribution of these diamond deposits. This pattern is a system of strike-slip faults that are transverse in relation to the main deep-seated Krashnomaysky fault zone (Figures 11.5, 11.6).

Supporters of the two different hypotheses emphasize different interpretations of the tectonic environment associated with the diamond deposits. Supporters of a subduction hypothesis interpret the presence of an active, island-arc continental margin. However, paleotectonic reconstructions suggest the presence of a passive Atlantic-type continental margin and obduction of an eclogitic slab on the continent (Mossakovsky and Dergunov 1985).

Marakushev, Pertsev, and Zotov (1995) proposed that the diamonds are of kimberlitic origin and were inherited by nearby sediments from the kimberlites prior to metamorphism. However, this concept does not explain why larger diamonds typical of

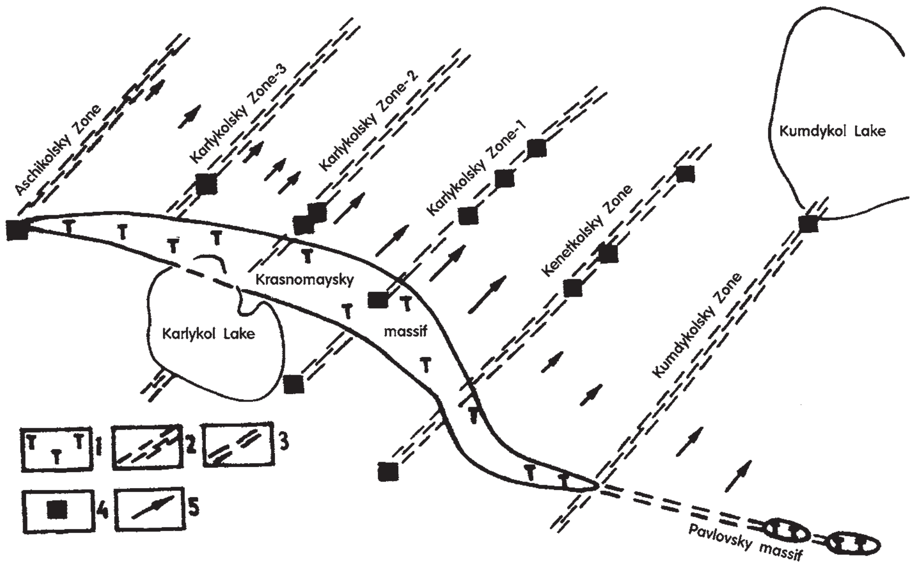

FIGURE 11.5 Model of dynamic development of the Krasnomaysky zone. 1 = pyroxenites of the Krasnomaysky complex, 2 = zones of mantle carbon degassing, 3 = Krasnomaysky deep-seated fault zone, 4 = areas of diamond mineralization traced, 5 = vectors showing direction of the block's displacement. Length of vectors indicates speed of displacement (modified from Abdulkabirova and Zayachkovsky 1996).

diamondiferous kimberlites have not been found, why eclogites with high-pressure mineral complexes rich in carbon are barren of diamonds, and why certain diamond morphologies are restricted to the various rock types. In addition, the carbon isotopic signatures of the microdiamonds show a very narrow range of signatures, unlike the characteristic broad range reported in diamonds from eclogite xenoliths in kimberlites.

The radiometric ages of zircons with diamond inclusions are on the order of 530 Ma (Claone-Long et al. 1991). Lavrova et al. (1997) reported that the age of garnet-biotite pairs in these rocks is about 500 Ma. If the garnets and zircons are relict mantle minerals as suggested by Marakushev, Pertsev, and Zotov (1995), then these relicts were formed in Paleozoic time. However, the eclogite-metamorphic complex was formed in the Precambrian. Helium isotope studies also show that $^3He/^4He$ ratios of the metamorphic rocks and diamonds are contrary to what would occur if they were of mantle origin (Lavrova 1996).

The third hypothesis, supported by most geologists involved in exploration, insists that the facts do not support a subduction genesis but instead support metastable genesis for the diamonds. This concept suggests that the diamonds crystallized metastably outside their stability field (Lavrova 1991) under stress in tectonic zones associated with slow shearing and recrystallization of the rocks associated with deep-originated fluids (Letnikov 1983). The pressures attained were probably not higher than 20 kbar (DeCorte et al. 2000). This hypothesis is supported by the typomorphic studies on the microdiamonds.

Nadezhdina and Posukhova (1990) suggested that the Kumdikol microdiamonds crystallized very rapidly under disequilibrium conditions during a series of discrete periods in a medium that was rich in carbon and other admixtures, but not necessarily under high pressure.

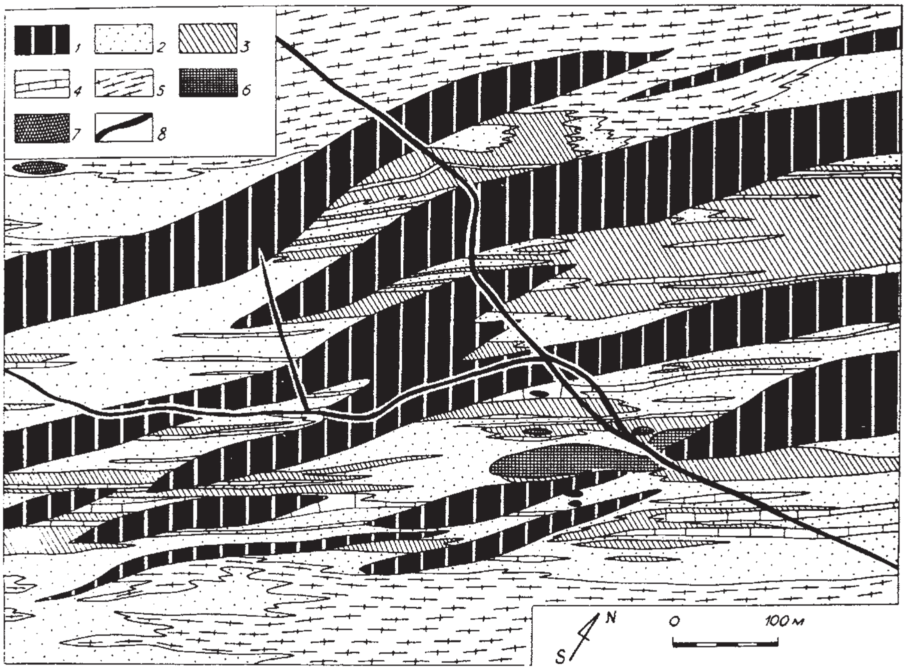

FIGURE 11.6 **Distribution of microdiamonds within strike-slip shear zone. 1 = lenses with high concentration of microdiamonds. 2, 3, and 4 = metasedimentary rocks of the Zerendinsky series: 2 = garnet-biotite and garnet–two-mica gneisses alternated with amphibole-biotite and biotite gneisses, 3 = garnet-pyroxene and tremolite-chlorite-bearing quartzites, 4 = marble and calc-silicate rocks. 5 = garnet-bearing two-mica granite, 6 = eclogite, 7 = garnet pyroxenite, 8 = dike of dioritic porphyrite (Dobrzhinetskaya et al. 1994). Reprinted with permission from Elsevier Science.**

In addition, the preservation of Ib-type diamonds in the Kokchetav massif is consistent with a short residence time and/or low temperature and supports a metamorphic genesis. Type Ib diamonds are very rare in kimberlites, and estimates of peak temperatures by use of the nitrogen data from the diamonds suggest that the diamond formed within a narrow time–temperature range.

SAXONIAN ERZGEBIRGE, GERMANY

Microdiamonds were detected in quartzofeldspathic rocks from the central portion of the Saxonian Erzgebirge gneiss-eclogite unit in 1992. The diamonds were found as inclusions in garnet, kyanite, and zircon, and they ranged from 1 to 25 microns. Partially graphitized diamonds and graphite pseudomorphs after diamond also occur in the host mineral. The diamondiferous gneisses are characterized by abundant quartz and phengite, and by peculiarly rounded, relatively large garnets of millimeter size that form 10% to 20% of the bulk rock. Graphite appears in the rock matrix.

Massonne (2000) estimated the P–T conditions of three different stages of metamorphism. The earliest stage is characterized by pressure conditions of 20 kbar at temperatures of 700°C. Temperatures of 900° to 1,000°C were later achieved when the rocks were buried to a depth of nearly 130 km, an environment postulated on evidence from significant Ti in garnet. At this stage, phengite broke down to form kyanite and a potassium-bearing phase. Exhumation of the gneisses was accompanied by cooling. During the third metamorphic phase, phengite was newly formed at temperatures around 750°C and pressures around 15 to 20 kbar (Massonne 2000).

The Erzgebirge massif is located in Saxony and the northern Czech Republic; it forms an oval antiformal structure extending nearly 80 km in a northeast-southwest direction. The central portion of the massif is dominated by orthogneiss and paragneiss with numerous lenses of eclogite.

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CHAPTER 12

Specific Type of Ring Structures (Astroblemes)

INTRODUCTION

In the 1960s, two schools of thought developed to explain ring volcano-tectonic structures. The first was developed by the American petrologist Robert Dietz (1969). Dietz used observations of underground nuclear tests to establish standards for characteristics of shock metamorphism generated by extremely large explosions, and he applied these characteristics to ring structures. In parallel, he summarized the characteristics of meteorite craters and combined them with the observed characteristics of the underground nuclear explosions. He described these features as

> *ancient erosional scars on Earth's surface produced by the impact of a cosmic body and usually characterized by a circular outline and highly disturbed rocks showing evidence of mineral formation under high pressure impact; an eroded remnant of a meteoritic or cometary impact crater.*

Dietz's work on this problem was first summarized in 1961 (Bates and Jackson 1980) and later formulated in general terms in 1968 (Dietz 1969). Dietz was attached to the University in Arizona near the Diablo meteorite crater, and the first summary on the mechanics of the formation of that crater was published about the same time (Shoemaker 1959).

The second school of thought was built on ideas formulated from research related to the early era of satellites, which turned the attention of the world's geoscientists to the possible consequence of Earth's interaction with space, and in particular, to the formation of impact structures as a result of Earth's collision with meteorites. The obsession with such an interaction was especially strong in Russia, which put the first satellite and first man into space. Russia's vast territories and technical ability to apply remote sensing methods of observation favored this tendency. Thus, as was written in a dedication of a book on impact diamonds to a distinguished Russian geochemist from Novosibirsk, Lev Firsov,

> *The discovery of the first Russian astrobleme Puchezh-Katunsky ring structure in 1965 belongs to him (Firsov 1965). Together with Academician A. L. Yanshin, he was the first who suggested an impact origin of the Popigay structure in 1964 (Vishnevsky et al. 1997).*

Many of the concepts related to impact features were developed and synthesized only after the origin of the Popigay depression, located in the northeastern part of the Siberian platform, was hypothesized. The individual credited for much of this research is a distinguished petrologist from St. Petersburg, Victor L. Masaitis. Driven by scientific curiosity, Masaitis researched the unusual structure of the Popigay depression and proposed that the structure was formed by impact.

During petrographic investigations, traces of shock metamorphism were recognized in some rocks within the depression, which led Masaitis to the idea of possible meteorite impact origin. This work led to an unusual payoff: in the volcaniclastic rocks of the Popigay depression, lonsdaleite, an hexagonal modification of diamond that is nearly three times as hard as normal diamond, was discovered. The industrial applications for an extremely hard diamond polymorph were limitless. Consequent sampling in many structures similar to the Popigay resulted in the discovery of similar diamond deposits.

The concept of impact met stiff resistance from the geologists at the Institute of Arctic Geology, which was the organization that initially described the Popigay structure as a volcano-tectonic depression. This same institute was also infamous for missing the first kimberlite pipe found in the Soviet Union in 1952 (the Leningrad pipe; see Chapter I, History of Diamond Discoveries) and all of the kimberlite pipes within the Daldyn-Alakit region (which have been actively mined to this date and became the pride of the Soviet diamond industry).

Information on the Popigay ring impact structure was published in 1971 (Masaitis, Mikhailov, and Selivanovskaya 1971), following about two to three years of research. Masaitis recognized and described the characteristic petrographic features, described indications of shock metamorphism, and discovered diamonds in rocks associated with the structure. On the basis of this research, similar structures were described in different regions of the USSR and the world. The most recent data on the Popigay structure are summarized by Masaitis, Maschak, and Raikhlin (1998).

The discovery of lonsdaleite in the rocks at Popigay led to a flurry of activity—extensive drilling and geophysical studies. The lonsdaleite-bearing rocks were tested to develop a method of extraction, and it was planned to transport the rocks hundreds of kilometers by air (!) to the nearest facility having sulfuric acid for leaching.

The discussions on the Popigay depression and its origin obtained great publicity and became internationally acclaimed. A series of similar structures was found in Russia (the Karsky and Ust' Karsky structures near the polar Urals, the Puchezh-Katunsky structure near the city of Nizhny Novgorod, the Kalyzhsky structure in the central part of Russian platform, and the Elgygitgin structure in the Chukotka peninsula). Similar structures were identified in Estonia (the Kyardla structure), Ukraine (the Ternovsky and Zapadnaya astroblemes), Belorussia (the Logovsky structure), Finland (the Lappayarvy astrobleme), and Kazakstan (the Zhenminshin astrobleme).

Although some specifics on the origin of the Popigay and related structures are still under discussion, for the readers' convenience we use the most widely accepted terminology in following description, on the basis of the presumed impact origin, and apply the term *astrobleme* to the appropriate structures. The use of this terminology is appropriate because shock metamorphism terminology more closely reflects specific features of the structure and composition of the rocks.

LOCATION

The largest astroblemes are closely associated with deep-seated fault zones that have a long geologic history. The Popigay depression itself is located at the northern tip of the Anabar shield. The depression lies along a deep-seated fault zone that is on the northeastern boundary of the shield. Along this fault, rocks of the crystalline basement are displaced to depths of about 3 to 3.5 km. This same structure controls the position of numerous kimberlitic bodies, which is a clear indication of multistage development of the deep-seated fault zone. There exists evidence that some stages of development of the fault were strike-slip in nature.

The Puchezh-Katunsky astrobleme is located at the junction of northwestern limb of the Volzhsko-Kamsky anteclise and the Moscow syneclise. A belt of northeast-striking deep-seated faults occur along which there are step-like displacements of the crystalline basement's roof throughout a distance of 1.7 to 3.0 km. The apparent structural control and position of the great astroblemes provides strong evidence for their nonmeteoritic genesis that is used in discussions by supporters of terrestrial origin (Polyakov and Trukhalev 1975; Marakushev 1993).

INTERNAL STRUCTURE

Specific types of ring structures (astroblemes) are represented by circular depressions, which sharply overlap different types of geologic structures. Like volcanic calderas, they are usually filled by volcanogenic rocks and bounded by a system of circular faults. The diameter of the depressions varies from one to tens of kilometers. For the most part, astroblemes tend to overlap the rigid structures of the ancient cratons. In contrast to calderas, one distinct characteristic feature of the internal structure of most astroblemes is the presence of a domal uplift of the basement in the central complex. In some structures, the domal uplift has been preserved; in others, it has been destroyed by explosive processes associated with the main "impact" event (Figure 12.1).

The Puchezh-Katunsky astrobleme in Russia is 80 km in diameter. The main morphostructural elements of this astrobleme are a central uplift 10 km in diameter, a ring trench 40 to 42 km in diameter surrounding central uplift, and a peripheral ring terrace.

The central uplift within the depression is formed by a granite-amphibolite-gneiss dome (the Vorotilovsky uplift). During the Permian and Triassic, continual uplift of the dome (Marakushev, Bogatirev, and Fenogenov 1993) resulted in the Paleozoic sedimentary cover being subjected to thrusts and plicate dislocations that now surround the dome as a system of folding, in which the intensity decreases radially. On the basis of drilling, the age of the central depression and a system of breccia dikes, pipes, and ring trenches that surround the Vorotilovsky uplift (Figure 12.2) were established as Jurassic.

The structure originated from high-energy explosions as a result of impact. The impact produced breccia composed of impact-transformed gneisses and amphibolites cemented by glassy and pyroclastic material. Other materials that were produced as a result of the consequent explosion included rocks resembling lava-breccia, xenohyaloclastics, hyaloclastitic rocks, and glassy basalts. As a result, numerous pipes and depressions with comparatively small diameters form part of the Puchezh-Katunsky structure (the largest has a diameter of 6 km) that cut and overlap the central part of the Vorotilovsky uplift. The central depressions in all of the pipes are filled with unconsolidated rocks of the Jurassic Coverninsky suite, the age of which signifies the time of completion of the explosive processes within the ring structure.

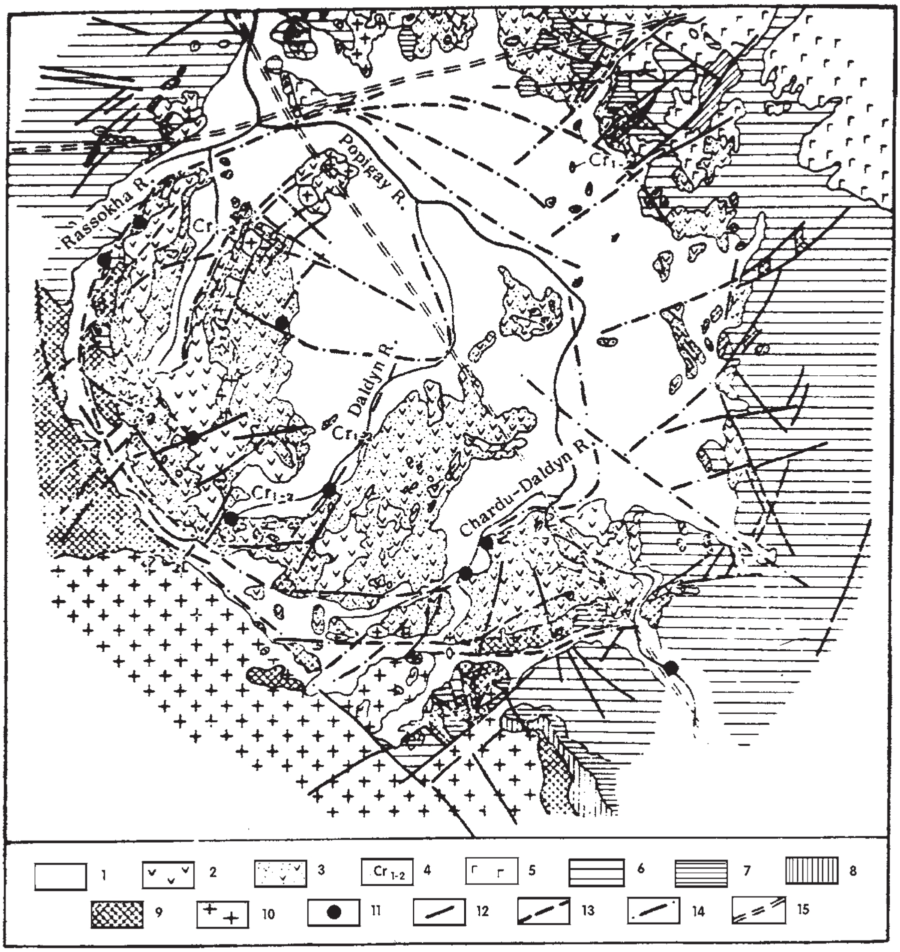

FIGURE 12.1 Schematic geologic map of the Popigay ring structure. This map provides the most complete geological description of this structure's origin. 1 = unconsolidated Pliocene-Quaternary sediments, 2 = Paleocene lavas of andesitic composition, 3 = Paleogene welded tuffs and ignimbrites of andesitic and dacitic composition, 4 = Lower and Upper Cretaceous coal-bearing volcano-sedimentary sequences, 5 = Permo-Triassic sandstones with doleritic lavas and sills, 6 = Lower and Middle Cambrian limestones and dolomites, 7 = upper Proterozoic dolomites, 8 = Middle Proterozoic sandstones and dolomites, 9 = Middle Proterozoic quartzite sandstones, 10 = Archean crystalline basement, 11 = inferred Paleogene extrusive bodies, 12 = andesitic dikes, 13 = faults, 14 = faults hidden under Quaternary sediments, 15 = deep-seated fault zones. Rocks shown by symbols 2 and 3 later described by Masaitis, Mikhailov, and Selivanovskaya (1971, 1975) as tagamites and suevites forming a coptogenic complex that fills the Popigay astrobleme (modified from Polyakov and Trukhalev 1975).

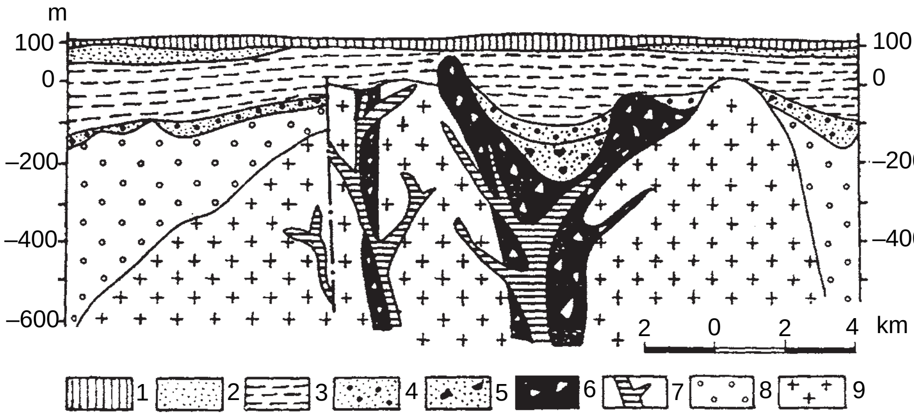

FIGURE 12.2 Cross section along a north-south line across the Puchezh-Katunsky ring structure, Russian platform. 1 = Quaternary and Neogene sediments, 2 = Upper and Middle Jurassic sediments, 3 = Middle Jurassic Koverninsky suite, 4 and 5 = the Uzol'sky suite at the basement of the Jurassic sequence (4 = redeposited explosive breccias and pyroclastic rocks, 5 = allogenic (returned) breccias of amphibolites), 6 = autochthonous gneisses and granites filling explosive pipes, 7 = hyaloclastites and basalts, 8 = rocks of platform's sedimentary cover, Vendian or early Paleozoic Vorotilovsky suite with folded structure, 9 = gneisses, amphibolites, and granites of Archean-Proterozoic age forming the Vorotilovsky dome-like structure (Marakushev 1995). Reprinted with permission of A.A. Marakushev.

The central part of the Popigay depression also has a central domal uplift that is weakly expressed by gravity and magnetic anomalies. Between the central uplift and the outer margins of the depression is a ring rampart that divides the depression into two trenches. Both the ring rampart and the trenches are well delineated by gravity studies (Figure 12.3).

The Popigay depression is displaced to the northwest in relation to the relict granite-gneiss dome. The position of the dome is schematically shown by a dotted line on the cross section in Figure 12.4. The dome represents a ring rampart, or a relic of the dome's basement, that was destroyed by the apparent impact. The dome is surrounded by a large central depression filled by glassy impactites (tagamites and suevites), which are also widespread within the deep outer trench.

The domal uplifts within the depressions were formed long before the time of the suggested explosive impact, a time relationship that contradicts an impact genesis for these features. Their formation was accompanied by the development of a complex dislocation of platform cover including formation of a fold ring. With increasing size of the pipe, domal uplift can be nearly completely destroyed and preserved only in the form of a ring rampart composed of granite gneiss overlapped by a strata of allogenic breccia. A typical example of such a structure is represented by the Kyardla structure in Estonia (Puura, Kaala, and Suuroya 1989). Details of a granite gneiss rampart, which on the basis of drill data surrounds this structure, is presented on Figure 12.5. Explosions occurred here after folding because blocks of granite gneiss basement lie on the dislocated Cambrian sandstones. The authors suggest that the destruction of the dome took

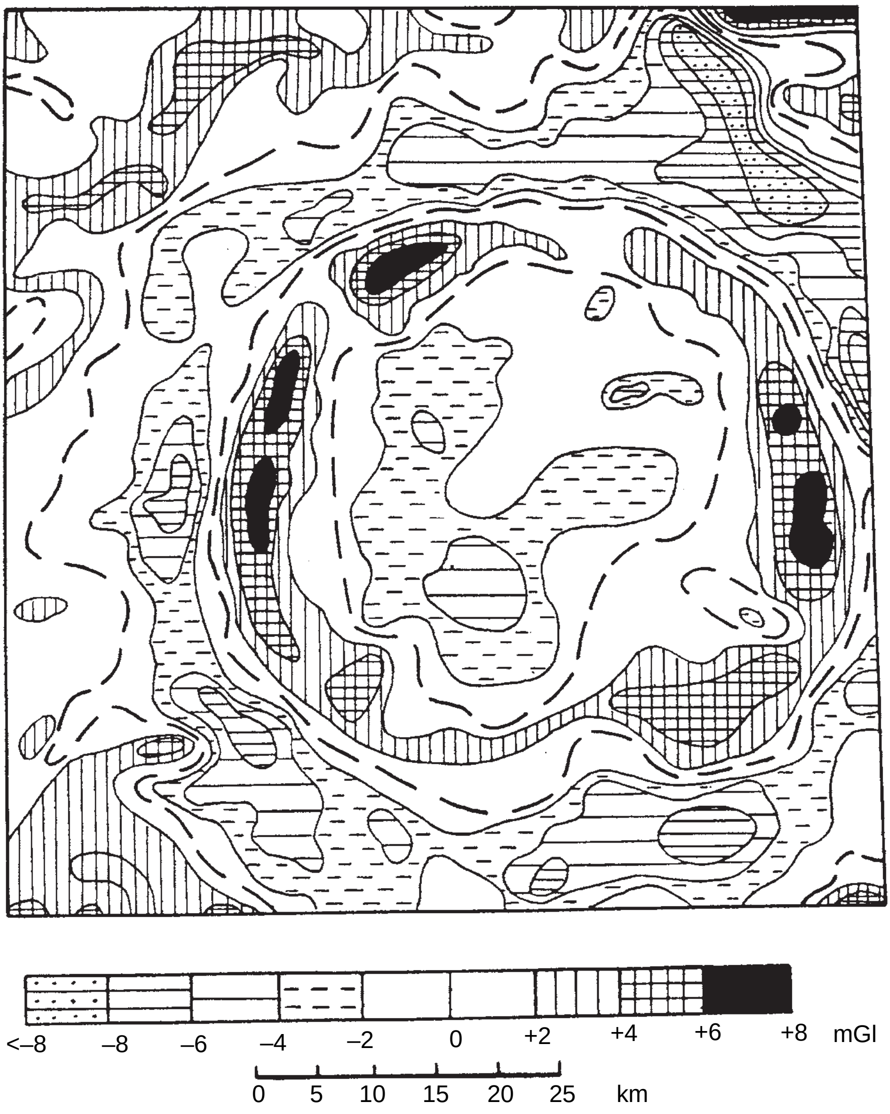

FIGURE 12.3 Local gravity anomalies within the Popigay ring structure. Ring system of positive gravity anomalies reflects the existence of ring uplift of basement rocks surrounding central depression (Masaitis, Maschak, and Raikhlin 1998). Reprinted with permission from V. Masaitis.

place after its partial denudation and surficial weathering. A similar structure is suggested for the Ries crater, Bavaria, Germany.

An extreme example of the complete destruction of the uplift can be seen within the Logoysky structure in Belorussia (Glazovskaya, Gromov, and Parfenova 1991). At this location, the explosive process affected rocks of the crystalline basement and platform cover. A breccia of blocks of crystalline Proterozoic basement is overlapped by a breccia

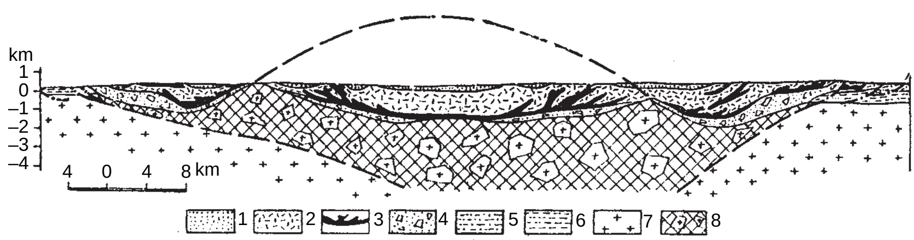

FIGURE 12.4 Geological cross section of the Popigay ring structure along a southeast-northwest line. 1 through 4 = a coptogenic complex (1 = coptoclastites, i.e., pyroclastic rocks supposedly formed by meteoritic impact, 2 = suevites, 3 = tagamites, 4 = allogenic gneiss breccia), 5 and 6 = sedimentary and volcanogenic rocks of the platform's sedimentary cover (5 = Paleozoic and Mesozoic, 6 = Late Proterozoic), 7 = Archean gneisses, amphibolites, and granites, 8 = their autogenous breccia. Dotted line indicates suggested position of granite-gneiss dome destroyed as a result of superimposed explosive structure. Only relicts of this dome are preserved in the form of granite-gneiss domal uplift (Marakushev 1995). Reprinted with permission of A.A. Marakushev.

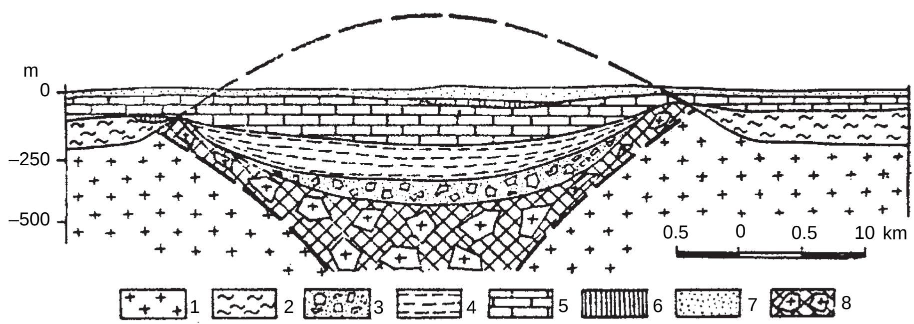

FIGURE 12.5 North-south cross section across the Kyardla ring structure, Estonia. 1 = Archean and Proterozoic crystalline basement rocks (gneisses, granites, amphibolites), 2 = Proterozoic and Cambrian aleurolites and sandstones, 3 = allogenic breccia, 4 = Middle Ordovician limestones and sandstones, 5 and 6 = limestones, dolomites, and marls (5 = Middle Ordovician, 6 = Lower Silurian), 7 = Quaternary sediments, 8 = allogenic breccia of gneisses, granites, and amphibolites (Marakushev 1995). Reprinted with permission of A.A. Marakushev.

of blocks of Devonian rocks derived from the platform cover. The main explosive event produced a huge volume of clasts as it completely destroyed a small granite-gneiss dome in the center of the structure (Figure 12.6). The rocks endured intense melting and the formation of suevites and tagamites, which were intruded as sills into the layered allogenic breccia of the platform cover.

Two other geologic features of astrobleme structure must be considered in any discussion of their origin. First is a group of comparatively small-diameter crosscutting funnels, which are considered to be feeder channels for volcanic material that erupted on the surface. Such bodies have been cut by drill holes within the Puchezh-Katunsky structure. A series of the same type of funnels was described by Polyakov and Trukhalev

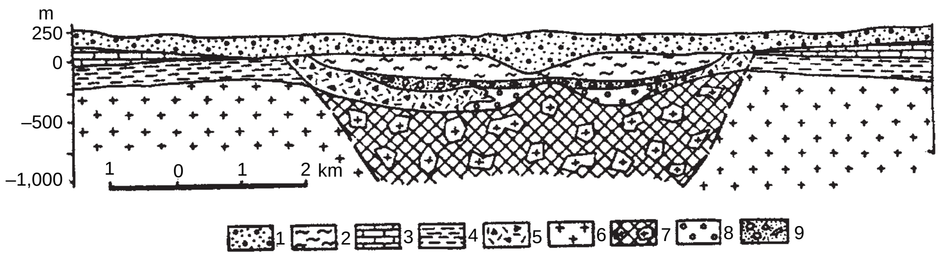

FIGURE 12.6 Southwest-northeast geologic cross section across the Logoysky ring structure, Belarus. 1 = Quaternary sediments, 2 = Paleogene–Neogene deposits, 3 and 4 = platform sedimentary cover (3 = Devonian, 4 = Proterozoic), 5 = allogenic breccia, 6 = granites and gneisses of Archean and Proterozoic age, 7 = their authigenic breccia, 8 = suevites and tagamites, 9 = allogenic breccias and boulders of gneisses and granites (Marakushev 1995). Reprinted with permission of A.A. Marakushev.

(1975) within the Popigay structure. Four similar funnel-like bodies about 20 m in diameter are located 8 to 10 km south of the Popigay depression within a field of Middle Cambrian limestones. These bodies have sharp subvertical contacts with the host Paleozoic carbonate rocks and are composed of rocks similar both petrographically and chemically to the rocks that fill the Popigay depression. These funnel-like bodies were examined by Milashev, Sokolova, and Shikhorina (1987), who confirmed that the features were independent of the main Popigay structure.

Second is data related to the formation of astroblemes that indicate that they were formed in a series of events during a comparatively long geologic period of time. The rocks of the main stage of volcanic activity within the Popigay depression are dated by fission-track and K-Ar methods at 29 to 45 Ma (Masaitis, Mikhailov, and Selivanovskaya 1975; Komarov and Raykhlin 1976). These dates clearly indicate that the duration of the volcanic activity caused by impact was about 20 million years. Moreover, Polyakov and Trukhalev (1975) cited geologic data suggesting that the structure formed during an even greater time interval. Within the Popigay depression volcanogenic material is present in Upper Jurassic–Lower Cretaceous volcaniclastic rocks exposed at the structure's northeastern boundary. Volcanogenic material, similar in composition to sequences of Paleogene volcanics that fill the main structure, is present in 100- to 120-m-thick outcrops of Upper Cretaceous sands and clays and contains bombs and lapilli interbedded with layers of volcanic ash. The petrographic and chemical characteristics of this material are similar to the main Paleogene volcanogenic sequence. The radiometric dates of rocks composing the Zapadny astrobleme are within a range of 105 to 125 Ma (Gurov and Gurova 1996) (i.e., the duration is also about 20 million years).

The endogenous nature of such ring structures considered to be astroblemes is confirmed by the presence of localized high heat flow. Such increased values were observed, for example, within the 50-million-year-old Karsky crater.

PETROGRAPHY AND PETROLOGY

Main Rock Types

Rocks within the Popigay depression include two unique petrographic types: tagamites and suevites. Tagamites are defined as impact-related rocks having massive texture and

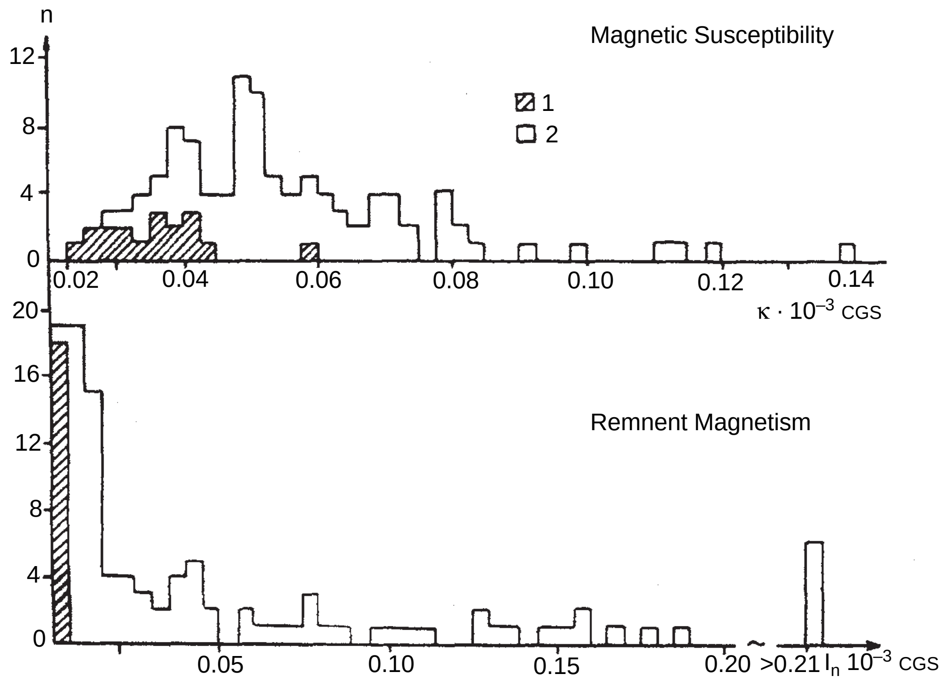

FIGURE 12.7 Histogram of distribution of magnetic susceptibility and remanent magnetization in (1) high-temperature and (2) low-temperature tagamites (Masaitis, Maschak, and Raikhlin 1998). Reprinted with permission from V. Masaitis.

a glassy to slightly crystallized matrix containing relatively large (a centimeter to several meters across) country rock clasts that form <3% to 5% of the rock's volume. These clasts are evenly distributed but may occur in lenses and lense-like bodies. Locally, blocks and other clasts may form up to 40% of the rock's volume. Tagamites form layered bodies of various thickness (from centimeters to several meters).

Two types of tagamites are recognized: high temperature (H-tagamites) and low temperature (L-tagamites). It is thought that these were formed from the initial impact melt and are characterized by different initial temperatures. The two types are distinctly different in their degree of residual magnetization and magnetic susceptibility (Figure 12.7).

L-tagamites usually form small bodies or marginal parts of large tagamitic bodies. They consist of 70% to 90% glass and 10% to 30% clasts. Of the clasts, 60% to 70% are microinclusions less than 0.5 cm in diameter and, to a lesser degree, 3 to 5 cm in diameter. Crystalloclasts in most cases bear no traces of significant shock metamorphism, although sometimes inclusions of partially crystallized diaplectic glasses are found. Holocrystalline tagamites, which form the central parts of thick-layered bodies, usually contain 5% to 10% small inclusions.

H-tagamites lie at the surface in several places in the southern part of the Popigay depression and also have been intersected by drill holes in the central part of a large layered body in the depression. These rocks contain up to 25% to 35% of unevenly distributed fragments, mainly quartz and feldspar with sharp boundaries that are rarely transformed by shock metamorphism. However, these clasts show evidence of being partially melted and are enclosed by a very thin (about 0.001 mm) rim of colorless glass.

Suevites are expressed by breccias that possess various degrees of lithification and are composed of fragments of bombs, pancake-like polymineral impact glass (vitroclasts), and rock fragments (lithoclasts) that are partially lithified as a result of welding. Any layering, rounding, and sorting of fragments is usually absent. The size of the fragments varies from a millimeter to several meters. The cement of the suevites is a clay-like material or aleuritic (silt-size) material similar in size to the fragments.

The classification of suevites is based on the composition of the fragments. Three types are recognized: sediment-rich suevites (S-suevites); vitric (glass-rich) suevites (V-suevites); and crystal-rich suevites (C-suevites). On the basis of the size of the fragments in the cement, suevites are further subdivided into three groups: ash suevites with prevailing fragment size of 0.1 to 2.0 mm, lapilli suevites with prevailing fragment size of 2.0 to 50 mm, and blocky suevites with prevailing fragment size >50 mm. Transitional subtypes may contain nearly equal amounts of fragments from more than one subtype. Impact diamonds have been found in all types of suevites and tagamites.

Petrology

Detailed studies of tagamites and impact glasses from suevites show that their compositions are homogenous within the whole structure. This indicates a high degree of homogenization of the melt. The composition of these rocks is similar to the composition of the crystalline basement underlying the impact features. The closest similarity to tagamites is shown by the composition of Archean biotite-garnet gneisses found in the basement under the impacts. Comparison of H- and L-tagamites shows that their chemical compositions are similar, although L-tagamites are a bit richer in silica. Rb and Sr content and isotopic composition of Sr in the tagamites correspond to the average composition for the gneiss. The same is characteristic of the Sm-Nd isotopes for tagamites and gneisses.

In impactites of all kinds, specific mineral associations and diaplectic glass are observed along carcass silicates (quartz and feldspars), although orthosilicates and metasilicates (such as garnet, olivine, and pyroxenes) are more stable in relation to impact melting (Feldman and Sazonova 1993).

Dense modified minerals (e.g., coesite, stishovite, diamond, lonsdaleite) are associated with low-density minerals such as quartz and feldspars in the impactites. In some cases low-pressure minerals replace high-pressure minerals. With continuously decreasing pressure (stishovite-coesite-leshatelerite), coesite in impactites is often replaced by quartz.

GEOCHEMICAL PATTERNS

Impact melting of the host metamorphic and volcanic rocks is accompanied by an increase in Au, platinum-group minerals, Cr, Ni, Cu, Zn, and other metals. These metals are present in impactites in quantities that are several times as great as in the original rocks. The Cr/Ni ratios in impactites (0.32–0.46) are much higher than in normative chondrite.

Within the Popigay volcano-tectonic depression, the average Cr and Ni content increases significantly in rocks of the lower sequence: 110 ppm Cr and 85 ppm Ni as compared with 80 ppm Cr and 27 ppm Ni in the host gneisses. The degree of enrichment increases with depth.

Rocks of the upper tagamitic sequence contain Ni (35–70 ppm) whereas rocks of the lower tagamitic sequence contain Ni (73.3 ppm). Moreover, the lower sequence lays on brecciated gneiss containing veinlets of glass rich in Ni (630–730 ppm). Veinlets of native Ni and Ni-sulfides were also found in gneissic clasts in the breccia.

A similar distribution of Ir is also noted. Rocks of the upper tagamitic sequence (depth 69–21 m) are much lower in Ir (0.0017–0.0071 ppb), as compared with the lower (474–85 m) sequence rocks that contain Ir (0.021–0.145 ppb). Additionally, the highest Ir content is associated with veinlets of impact glass found in the megabreccia, which contains impact cement at the base of the sequence.

A similar picture was observed for rocks of the Brent crater in Canada. Here, impactites of the lower sequence are enriched in Ni (370–600 ppb), compared with Ni (5–21 ppb) in the host rocks. The Ni content of the upper sequence is much lower: Ni (115 ppb).

Pohl, Stoffler, and Gall (1977) described a similar picture for rocks of the Ries impact crater in Germany. Veinlets of glass in the granites and gneisses of the Ries crater contain are enriched in metals. Native iron in some veinlets contains Cr (11%), Ni (6%), and Co (0.3%).

The abundance of Cr as compared with Ni in native iron conflicts with the traditional concept that the metal distributions in astroblemes are associated with meteoritic water and genesis. Instead, they show a distribution similar to that of native iron in the dunitic core of the Konder massif in the eastern part of the Siberian platform (Lennikov, Nikolsky, and Pakhomova 1993). In the Konder massif, native iron forms plate-like intergrowths in forsterite with Cr (0.32%), Si (2.51%–5.75%), and Mn (0.54%–0.53%). The zoned crystals of forsterite in the massif contain drop-like secretions with Cr (0.01%–0.14%), Ni (0.02%–0.19%), Cu (0.10%–0.05%), and Si (0.47%). Native silica inclusions in native iron clearly indicate that ultrabasitic magmatism is characterized by strongly reduced regimes genetically related to the ring structures that developed within platforms.

Two other concepts are stressed. Chaoite, the high-temperature modification of graphite, is uncommon in astroblemes. Instead, graphitic and diamond-lonsdaleitic aggregates are found in the tagamites of some astroblemes, particularly the Ries and Popigay, and some craters in Ukraine. However, chaoite has been reported in association with impact graphite in the Popigay crater, in some Ukrainian astroblemes (Valter, Mel'nichuk, and Rakitskaya 1985), and as a pure phase in melted impact rocks of the Ries crater (El-Goresy and Donnay 1975). There are reports of chaoite in yakutite in placers (i.e., possibly dispersed independently from the Popigay impactite rocks). Chaoite formation reflects temperatures exceeding 2,000°C.

Second, another important feature is a positive correlation between diamonds and P_2O_5 in the rocks. Masaitis, Maschak, and Raikhlin (1998) discussed possible relations between diamonds and apatite in "target" gneisses. It may also be important that lonsdaleite-diamond aggregates in iron meteorites (i.e., Canyon Diabolo, ALHA-77283) are associated with troilitic liquation nodules in kamacite that contain admixture of kohenite (F_3C) and schreiberzite (Fe_3P).

Rapid movement of the melted matter to the surface related to impact is reflected not only by the presence of native metals in the impact glass, but also by a high FeO/Fe_2O_3 ratio. For example, in rocks of the Popigay structure this ratio is 2.50 (versus 1.95 in the host rocks). The same relationship is repeated in the Yannisyarvi structure (2.60 versus 1.75), Boltishsky (3.52 versus 1.57), Vabar (2.30 versus 0.32), Ailulel (1.48 versus 0.08), and others.

Additionally, the gas inclusions in the glass of the Popigay depression are unique and contain H_2 (up to 20%) and hydrocarbons (8%–22%) (Feldman and Sazonova 1993; Marakushev 1995). The presence of hydrogen is interpreted to be due to rapid uplift of reduced, hydrogen-rich fluids from the Earth's core that were held in greatly compressed (fluid) juvenile gases.

DIAMOND DISTRIBUTION

Diamonds were initially found in rocks of the Popigay depression in the early 1970s. Since that time considerable work has been completed on the characteristics of the diamonds. That work depended on a great volume of samples of impactogenetic rock from natural outcrops and in drill cores. More than 1,800 samples were taken. Much of the work has been devoted to establishing characteristics of the diamond distribution in relation to structural elements and to petrographic and petrochemical features.

The impact diamonds recovered from nearby placers ranged from 0.05 to 2 mm across. Diamonds recovered during crushing of the diamondiferous impactite included grains up to 8 to 10 mm in diameter. Most diamonds are shades of yellow, and some are colorless, gray, and black. The impact diamonds are usually represented by polycrystal aggregates of cubic and hexagonal modifications (lonsdaleite) of carbon. They often form paramorphs with graphitic crystals.

Lonsdaleite in impactites forms small elongated or irregular grains in fine-grained aggregates with graphite and diamond, or with diamond only. It can also form a type of matrix in relation to small cubic or cubic-octahedral diamond crystals. The lonsdaleite can form as much as 60% of the fine-grained diamond aggregates (Valter et al. 1992).

In contrast, chaoite is rarely found in the graphitic and diamond-lonsdaleite aggregates in tagamites of the astroblemes in the Ries, Popigay, and some other astroblemes in the Ukraine. Where found, it reflects the fact that the temperature of impactogenesis exceeded 2,000°C.

Masaitis, Maschak, and Raikhlin (1998) reported that, overall, the impactites of the Popigay astrobleme have a comparatively low diamond content, although within some local areas the diamond content is anomalously high. In general, the diamond content in the tagamites is three to five times as high as in the suevites. Special attention was paid to sampling the host metamorphic rocks both within the area of their development and in xenoliths. It is important to stress that in most cases, the gneiss in the impactites contained essentially no diamonds, although single grains were found in some samples. The diamond content, however, was quite high (up to hundreds of carats per tonne) in some samples of biotite-garnet gneiss and plagiogneiss xenoliths that contained visible graphite.

Figure 12.8 shows the distribution of diamonds at Popigay. The maximum diamond content is found in rocks (both tagamites and suevites) within the ring depressions. The minimum diamond content is found in the central zone of the Popigay depression. A comparatively reduced diamond content within the central zone reflects a smaller rock outcrop area. Figure 12.8 shows radial zones that are distinctly enriched in diamonds.

The degree of diamond mineralization correlates with the degree of cryptocrystallinity of the tagamites, especially in moderately to weakly crystallized, high-temperature tagamites in the upper part of the section that consists of a complex layer-like body of tagamites (Masaitis, Maschak, and Raikhlin 1998). A positive correlation was also established between diamond content and P_2O_5 content of both suevites and tagamites. A meaningful positive correlation was also observed between diamond content and MgO

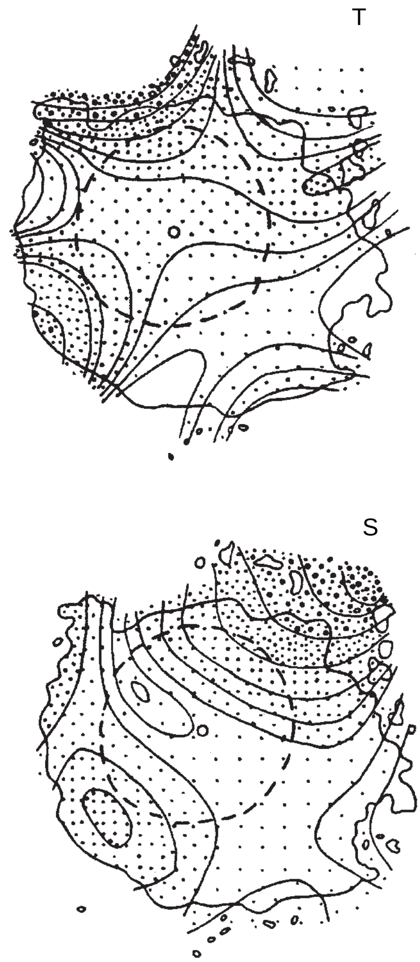

FIGURE 12.8 Relative abundance of diamond in rocks within the Popigay ring depression (T = tagamites, S = suevites). Nonpolynomial trend analysis of sampling data. Density of stipple proportional to increasing diamond content. Dashed line = axis of ring basement uplift. Solid irregularly circular line = impactites' outer boundary (Masaitis, Maschak, and Raikhlin 1998). Reprinted with permission from V. Masaitis.

content of the rocks. The diamond content is also shown to be somewhat higher in rocks with higher CaO and lesser SiO_2 content (Masaitis, Maschak, and Raikhlin 1998). The diamonds are distributed unevenly along radial zones (Figure 12.9), which is explained as a result of uneven ejection of material at the time of impact.

Greater diamond concentrations are associated with the glassy matrix of the impactites, and the diamond mineralization is proportional to the content of xenolithic blocks from rocks that are characterized by the absence of graphitic material. This is the principal reason why diamond mineralization is higher in tagamites than in suevites.

There is also a high correlation between the diamond content of weakly crystallized tagamites that form the cement of megabreccias and an increased diamond content near

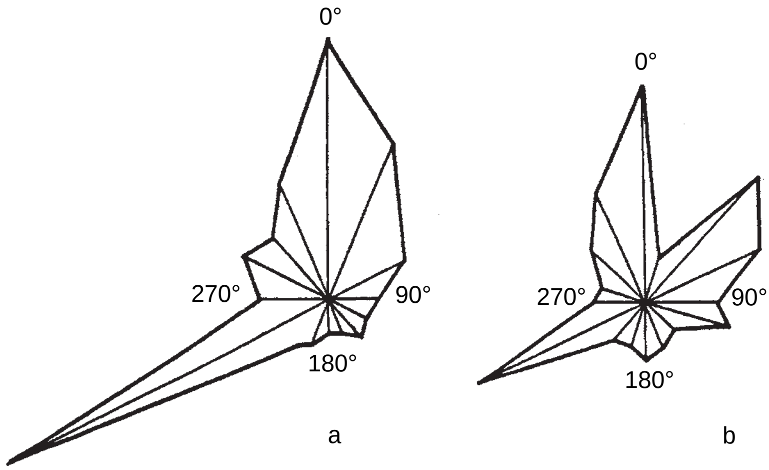

FIGURE 12.9 Azimuthal inhomogeneities in diamond distribution in (a) tagamites and (b) suevites within the Popigay volcano-tectonic structure (Masaitis, Maschak, and Raikhlin 1998). Reprinted with permission from V. Masaitis.

the contacts of large blocks of gneiss. Within these tagamites the diamond content increases and may even produce what is termed "hurricane-like" (tens to a hundred times greater than average) diamond contents. There is a clear positive correlation between the diamond content of the tagamites and the content of xenolithic fragments with diameter less than 0.5 cm. Where larger fragments form less than 10% of the rock's volume, the diamond content is not affected. This relationship is seen especially within the first hundreds of meters below the surface.

A negative correlation was observed between the diamond content and the abundance of hypersthene aggregates associated with clasts of quartz in the tagamites. The size of hypersthene microlites found as reaction rims around quartz clasts show a negative correlation with diamond content.

The geologic literature related to impactites tends to stress diamond paramorphs with graphite in the host rock (Valter et al. 1992). In addition, diamond found in the impact glass occurs as fine-grained aggregates with a morphology independent of graphite. Similar aggregates of fine diamond crystals (<1 mkm, i.e., 10^6 microns) with lonsdaleite (up to 2.2 carats) known as yakutites are described in placers. Some authors consider these yakutite-bearing placers to have formed because of dispersion caused by the impact explosion, rather than the result of normal placer accumulation in water (Vishnevsky et al. 1997) (Figure 12.10).

All of the data indicate that the diamonds of these astroblemes are not exclusively associated with carbon accumulation of the host ("target") rock, but rather with magma crystallization related to shock metamorphism.

GENESIS

Although evidence of extremely high pressure related to shock metamorphism is obvious, the origin of both the astrobleme depressions and the associated rock units remains unclear. Historically, supporters of a terrestrial origin of these structures stressed the following geologic evidence:

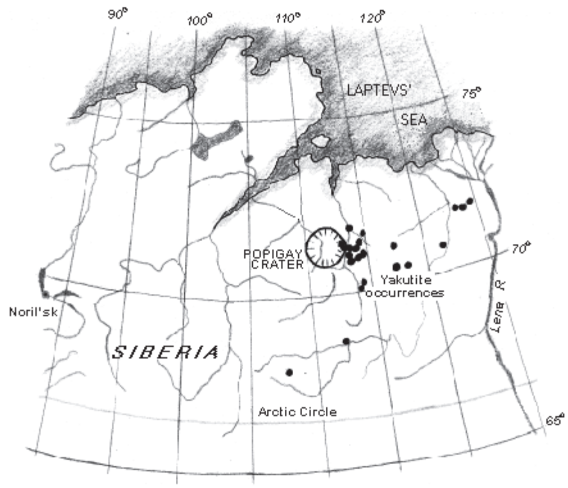

FIGURE 12.10 The strewn field of yakutites (diamond-lonsdaleite intergrowths) in the vicinity of the Popigay ring structure. Distribution may reflect the amount of pyroclastic cover erupted as the Popigay structure formed (modified from Vishnevsky et al. 1997).

- the structural position of the major astroblemes (apparently controlled by major deep-seated fault zones)
- central domal uplifts within the astroblemes that were not developed during the course of a single event, but rather exhibit a long geologic history.

Supporters of meteoritic impact consistently search for geochemical evidence and high-pressure mineral indicators. The meticulous work by the distinguished Russian petrologist Marakushev (1995) synthesized both approaches and succeeded in presenting an alternative concept on the origin of the structures and related rock formations.

From our point of view, five reasons support a terrestrial origin for these structures:

1. evidence of multistage volcanic activity within the astroblemes
2. indications that the suevites were formed as a result of both airfall and pyroclastic flows, whereas tagamites resemble lava flows and sills
3. an elongated radial distribution of diamonds that is similar to the spatial distribution of hydrothermal mineralization related to silicic calderas. A general model that explains this distribution is seen at the Uzon silicic caldera in Eastern Kamchatka (Erlich 1974, 1986) (Figure 12.11). A similar structural distribution of gold mineralization occurs in such ancient structures like Aginsky caldera in central Kamchatka and Portovelo caldera in Ecuador (van F. Thornout, personal communication, 1991).
4. numerous examples of correlation between the diamond content and petrographic features of the rocks
5. dispersion of Popigay-related diamonds throughout a large area possibly owing to the initial explosion, which ejected fine pyroclastic material.

The data do not permit a single universal model of genesis for the so-called astroblemes and associated rock formations. It may be possible to accept a meteorite-impact origin for comparatively small craters such as Arizona or even Ries craters, but the data presented for large structures such as Popigay, Puchezh-Katunsky, and others support a terrestrial origin. A model for terrestrial origin has been presented by Marakushev

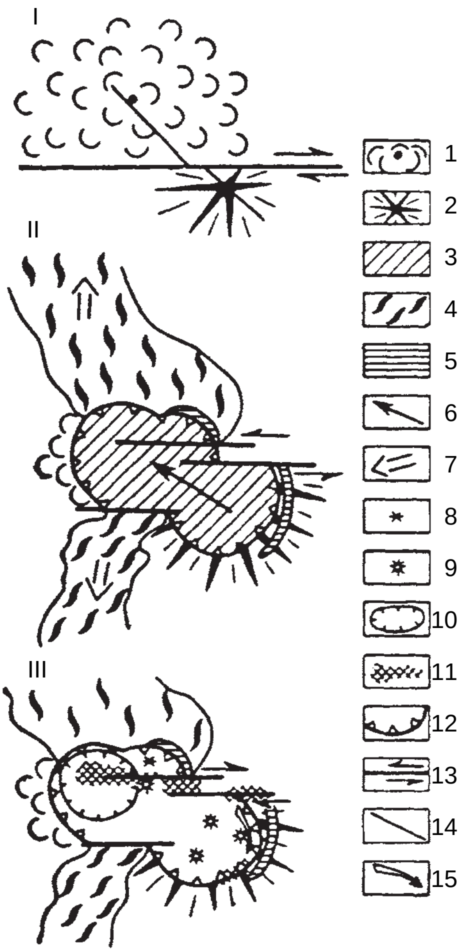

FIGURE 12.11 Formation of Uzon caldera, eastern Kamchatka, in connection with development of deep-seated strike-slip faults. I = precaldera stage of volcanic activity, II = caldera-forming stage of volcanic activity, III = postcaldera stage of volcanic activity. 1 = shield-like basaltic volcano (Q 1–2), 2 = center of silicic volcanic activity at the precaldera stage, 3 = eruptive centers of the caldera-forming stage that produced huge fields of silicic pyroclastics, 4 = ignimbrites development, 5 = arcuate fissures filled by viscous silicic magma, 6 = direction of lateral development of silicic magma chambers, 7 = suggested direction of ignimbrite flow, 8 = basaltic maar of the Dal'neye Lake, 9 = silicic volcanic extrusive domes, 10 = explosive funnel, 11 = fields of development of hydrothermally altered rocks, 12 = fault located along the boundary of volcano-tectonic depression, 13 = suggested zones of deep-seated strike-slip faults (arrows indicate relative displacement), 14 = feather fractures generated by strike-slip faults, 15 = direction of flow of thermal waters from deep-seated strike-slip fault zone along drainage fracture system. Ignimbrites characteristically distributed along radial zones (Erlich 1974).

(1995), in which he suggests that the so-called astroblemes are associated with mighty explosions related to streams of hydrogen-rich gases ascending from the earth's core. As such, diamonds were formed in the impact melts owing to input of Cr, Ni, and other metals (M) in the form of metal-organic combinations: $M(CN)_2$, MCN_2, $M(CH)_2$, and MCH_2.

The geochemical-thermodynamic regime that permitted impact diamonds to form was determined by the interaction of the host rocks with shock waves that were generated by strong explosions of hydrogenous gases containing metal-organic compounds. The melting of minerals under this regime occurred under isochoric conditions in which a temperature increase was accompanied by increasing pressure ("autoclave effect") and reached levels that were sufficient to form diamond, lonsdaleite, chaoite, coesite, and stishovite—the classic minerals in impact glass. Such a model explains the main features of impactogenesis: its development only within a rigid substratum (gneiss, granite, and extrusives) that assured isochoric melting.

Impact shock waves, which are stress related, are unfavorable for the formation of dense phases (diamond, lonsdaleite, chaoite, coesite, and stishovite). Instead, these phases are related to an increase in pressure caused by increase in temperature in a sufficiently rigid medium of isochoric melting, which is expressed by the formation of monomineralic glasses (Stofler and Hornemann 1972).

The latter is proven by the widespread development of diaplectic glasses (pseudomorphs of melts along mineral grains); the melting of rocks that accompanied diamond formation, and introduction of Cr, Ni, platinum-group metals (especially Ir), Au, Cu, and Zn, which reflect a metal-organic component of hydrogenous explosive fluids that are immediately associated with the earth's core.

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SECTION 4

Tectonic Control and Temporal Distribution of Diamond Deposits

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CHAPTER 13

Tectonic Control of Diamond Deposits

Analyses of the tectonic settings of kimberlites and lamproites show that no single model can explain all data (e.g., Dawson 1980; Mitchell 1986). Thus, we do not present a universal model of structural control of kimberlitic/ultramafic-alkalic magmatism. Instead we intend to concentrate on clear examples of structural control in order to restore the geodynamic conditions that appear to be favorable for the formation of different types of diamond deposits.

PETROGRAPHIC PROVINCES

The spatial distribution and structural control of diamond deposits result in what we term "petrographic provinces." These provinces are related to the uniformity (or comparative uniformity) of igneous rock associations. In accordance with the definition, it is natural to expect that any specific petrographic province is controlled by a specific type of tectonic structure and characterized by a uniform type of magmatic rock associations. This uniformity, in turn, has significant theoretical and practical implications; for example, which types of structural elements control the development of certain types of magmatic activity? The boundaries of specific provinces are determined by the uniformity of magma types. The designation strongly affects such practical matters as the zonal distribution of diamond deposits and the evaluation of prospective areas. It is worthwhile to refer to a discussion on the boundaries of the Yakutian (or Siberian) province by Milashev (1965, 1974) and Kaminsky (1972).

The term petrographic provinces, as well as other tectonic terms used in this book, match definitions in the *Glossary of Geology* (Bates and Jackson 1980). Some structural terms such as *anteclise* and *syneclise*, which are uncommon in the Western literature, also match definitions provide by Bates and Jackson (1980).

Mitchell (1986) recognized three types of kimberlitic provinces that are related to the timing of kimberlitic volcanic activity. He also considered timing to be one of the major factors in the definition of a province. These included Type-1 provinces, which consist of a single kimberlite field; Type-2 provinces, which consist of several fields of similar age; and Type-3 provinces, which consist of numerous fields characterized by kimberlites of different ages and petrologic character. The polychronous character of kimberlite fields and even single kimberlite intrusives is common, and the Type-1 and

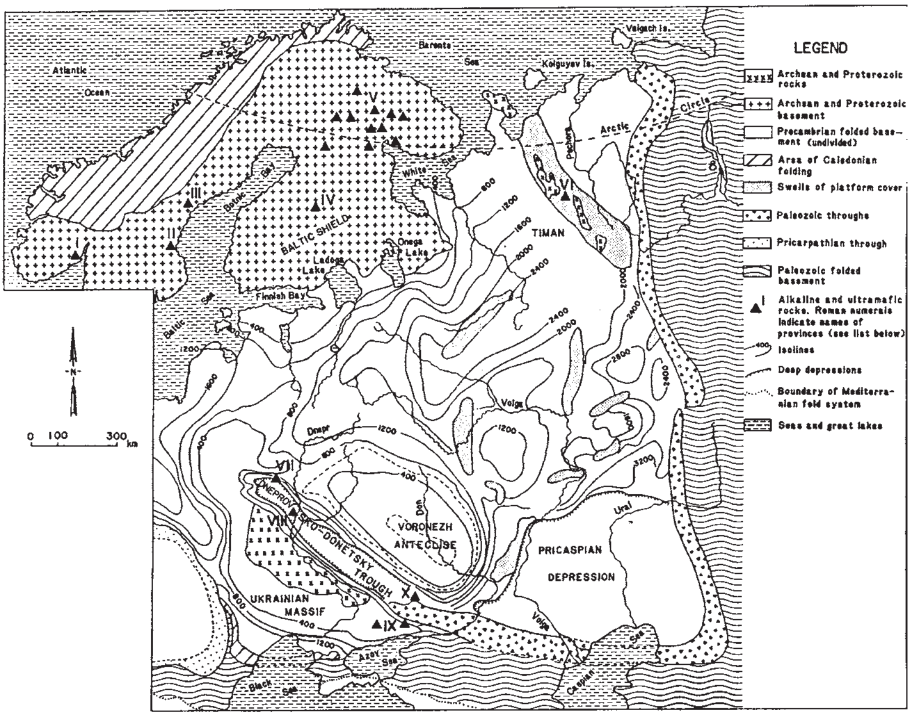

FIGURE 13.1 **Distribution of alkalic and ultramafic-alkalic complexes within the Russian platform. Main regions of magmatism: I = Oslo graben, II = Alno, III = Almunge, IV = Synnil'yarvi, V = Kola Peninsula and adjacent Finland and Karelia, VI = Timan (suggested inverted aulacogen), VII = Pripyatsky depression (aulacogen), VIII = Chernigovsky uplift, IX = Priazovsky region, X = Donetsky coal basin (aulacogen) (Erlich et al. 1989).**

Type-2 provinces as defined by Mitchell (1986) do not correspond to specific types of provinces, but rather to types of kimberlitic fields (see Chapter 14, Temporal Distribution of Diamond Deposits).

The association of the great diamond regions with platforms or with prominent structural elements (such as shields or anteclises) can be accepted as no more than a first-order connection. From both theoretical and pure exploration points of view, it is important to establish the order (and specific types) of structure with which kimberlite provinces are associated. For example, provinces of alkaline magmatism within the Oslo graben and on the Kola Peninsula both lie on the edge of the Baltic shield, but they differ in all other respects. Thus, according to our definition, they cannot be considered a single petrographic province (Figure 13.1).

The Siberian platform also cannot be considered as a single petrographic province because different parts are characterized by the development of different types of igneous rock associations. For example, alkaline rocks within the Aldan shield are quite different from alkaline rocks around the Anabar shield (i.e., the Kotuy-Maimecha, Olenek, or Udja regions). Thus these regions cannot be considered as a single petrographic province (Figure 13.2).

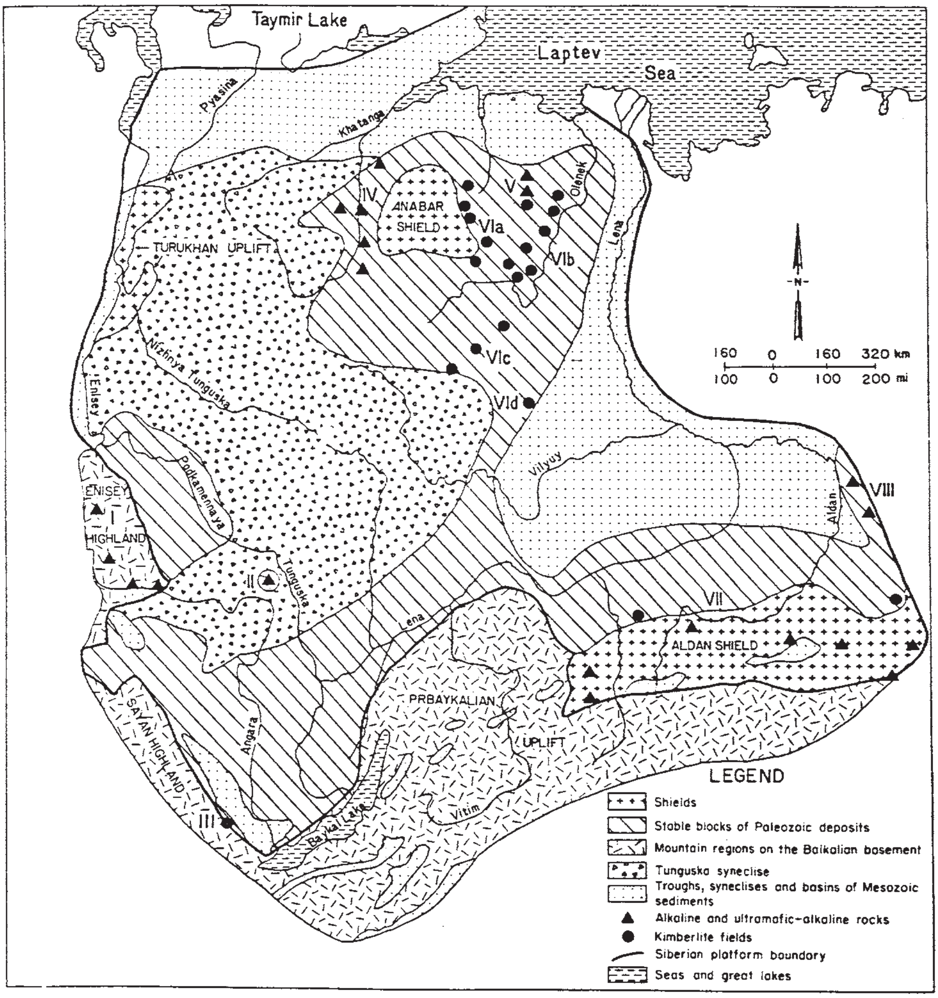

FIGURE 13.2 Spatial distribution of ultramafic-alkalic and alkalic complexes within the Siberian platform. Main provinces of magmatism: I = Enisey highland, II = Chadobetsky Uplift, III = eastern Sayan (Ziminsky), IV = Kotuy-Maimecha, V = Udzhinsky, VI = main kimberlitic fields: VIa = Anabar-Djelindinsky, VIb = Oleneksky, VIc = Daldyn-Alakitsky and Munsky, VId = Batuobinsky, VII = Aldan shield, VIII = Stanovoy (Ingili-Arbarastakh) (Erlich et al. 1989).

These examples show that the boundaries of petrographic provinces within the framework of platforms (or cratons) are determined by structures of a third order whatever their nature—aulacogens, swells, rifts, or stable blocks, rather than platforms (structures of the first order) or shields (structures of the second order).

Provinces are composed of kimberlite fields, and within each field are several to tens of kimberlitic intrusives. Kimberlite fields typically average about 50 km in diameter and are usually associated with a similar mantle source. It is assumed that the kimberlites have similar compositions that may vary depending on the mode of intrusion and differentiation.

STRUCTURES ASSOCIATED WITH ULTRAMAFIC-ALKALIC MAGMAS

The decisive stage of a platform's formation is cratonization—the consolidation of basement blocks of different ages and degrees of metamorphism into a single structure characterized by a uniform and comparatively stable regime of tectonic movement. Peneplains developed on these blocks are considered an indicator of cratonization, which is a reflection of deep transformation of the crust beneath cratonized blocks. In Chapter 14, Temporal Distribution of Diamond Deposits, we will discuss Clifford's rule in relationship to the timing of cratonization and the generation of ultramafic-alkalic magmas and kimberlite/lamproite magmas.

Structures of the first order that are related to the spatial distribution of ultramafic-alkaline and kimberlitic volcanism are also associated with the location of ultramafic-alkaline provinces in general. In addition, kimberlitic fields within large tectonic structures are characterized by long periods of protracted uplift. Such structures (anteclises) develop within platforms. This conclusion is supported by the fact that all known kimberlite fields are located within Precambrian or lower Paleozoic sequences, or along the boundaries of structures formed by these sequences, or within basins filled by upper Paleozoic and Mesozoic sequences.

The term *anteclise* was originally introduced in the Russian literature to describe co-sedimentation structures. An appropriate synonym applied to recent structures is the term *swell*. Such swells are considered first-order structures that control the generation of peralkaline and ultramafic-alkaline magmas (LeBas 1971, 1987).

The association of kimberlite with anteclises has been established for the Siberian platform (Kirillov 1961; Erlich 1963). Within the Russian platform, kimberlites are associated with the Ukrainian and Baltic shields and adjacent anteclises. Within the Siberian platform, kimberlitic fields of the Daldyno-Alakitsky, Munsky, Batuobinsky, and Anabar-Oleneksky regions lie within the Anabar anteclise, and kimberlites of the Aldansky region lie within the Aldan anteclise.

Kimberlitic fields (and ultramafic-alkaline rocks in general) can also be localized along other favorable structures, characterized by slow protracted uplift. These include old orogenic zones (for example, the Baikalian) and adjacent frontal depressions, which were related to slow uplift during middle Paleozoic time and later became anteclises of Paleozoic platforms. Examples include the Eniseysky upland and the adjacent Angarsky depression, and the eastern Sayan orogenic system with its adjacent frontal depression. Within these structures are a series of ultramafic-alkaline and kimberlitic fields and some associated diamonds.

Structures favorable for the localization of similar types of magmas are median massifs, which underwent the same deep transformation during cratonization and are characterized by lengthy, protracted periods of uplift. Examples of such structures include the Czech massif (Kopecky 1960), median massifs within the transitional zone between ocean and continent in the Russian Far East (Izosov, Konovalov, and Emel'yanova 2000), median massifs in Pamir, central Asia (Dmitriev 1974), eastern Kazakstan and Kyrgyzstan (Abdulkabirova and Zayachkovsky 1996), and a similar massif within Borneo (Sobolev 1951).

Considering the grand scale of tectonic transformations associated with cratonization, it is no wonder that kimberlitic/lamproitic magmatism can occur not only within these structures, but also within parts of orogenic systems along their boundaries. Examples include the Rocky Mountains adjacent to the Colorado Plateau and the Wyoming craton in North America. In South Africa, kimberlitic fields are located within the Transvaal craton

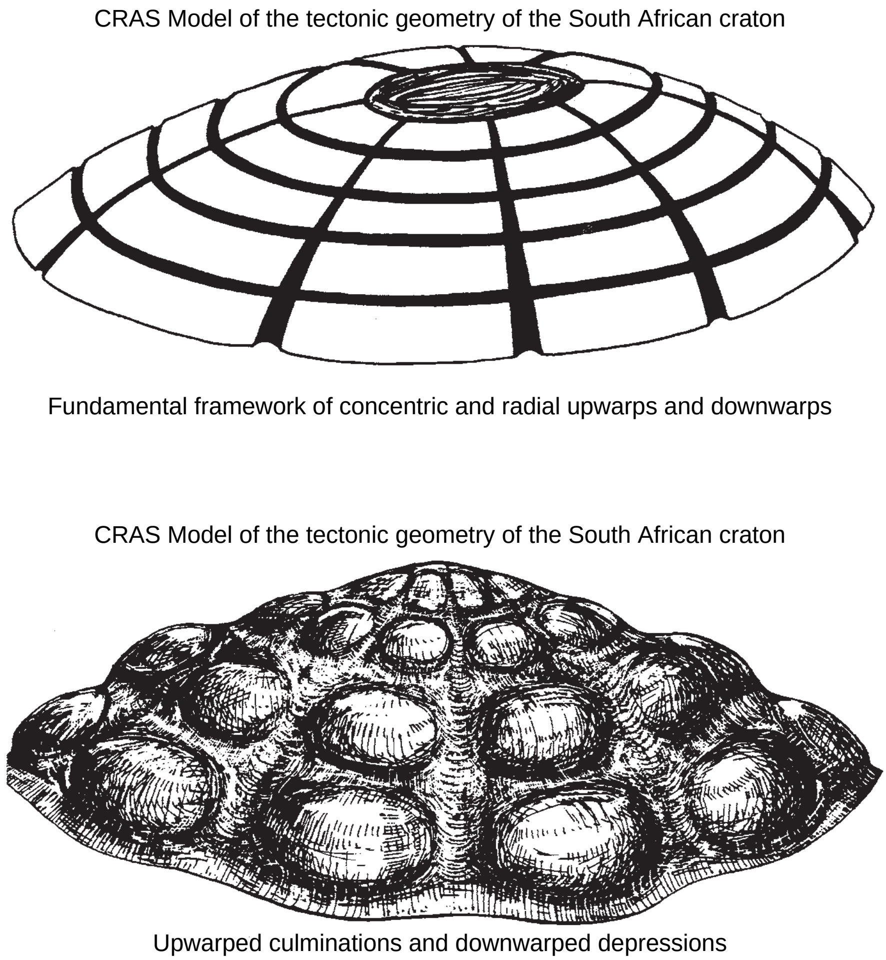

FIGURE 13.3A Structure of the Transvaal craton. CRAS (concentric radial anteclises and syneclises) model of the tectonic geometry of the South African craton. Concentric arrangement is not mandatory for all cratons, but in general the combination of anteclises and syneclises matches this system of dome-like swells within the Anabar anteclise (Anabar, Olenek, Kuoyka-Daldyn, Udja, domal uplifts and corresponding syneclises) (Pretorius 1973).

(see Figure 13.3A and 13.3B). Structures within this craton include a series of swells that shows similarities to the Anabar anteclise (Pretorius 1973). It has to be stressed that there were no indications of subduction during the time of kimberlite formation around this craton (see discussion in Sharp [1974] and in Newton and Gurney [1975]).

In Australia, alkalic complexes are distributed around stable blocks that had different ages of cratonization. The earliest phases of alkalic magmatism coincide with cratonization of the nearest stable block. This relationship has been well demonstrated in a review of timing of the earliest phases of ultramafic-alkaline magmatism and the timing of cratonization of stable blocks in different parts of Australia (Jaques et al. 1985;

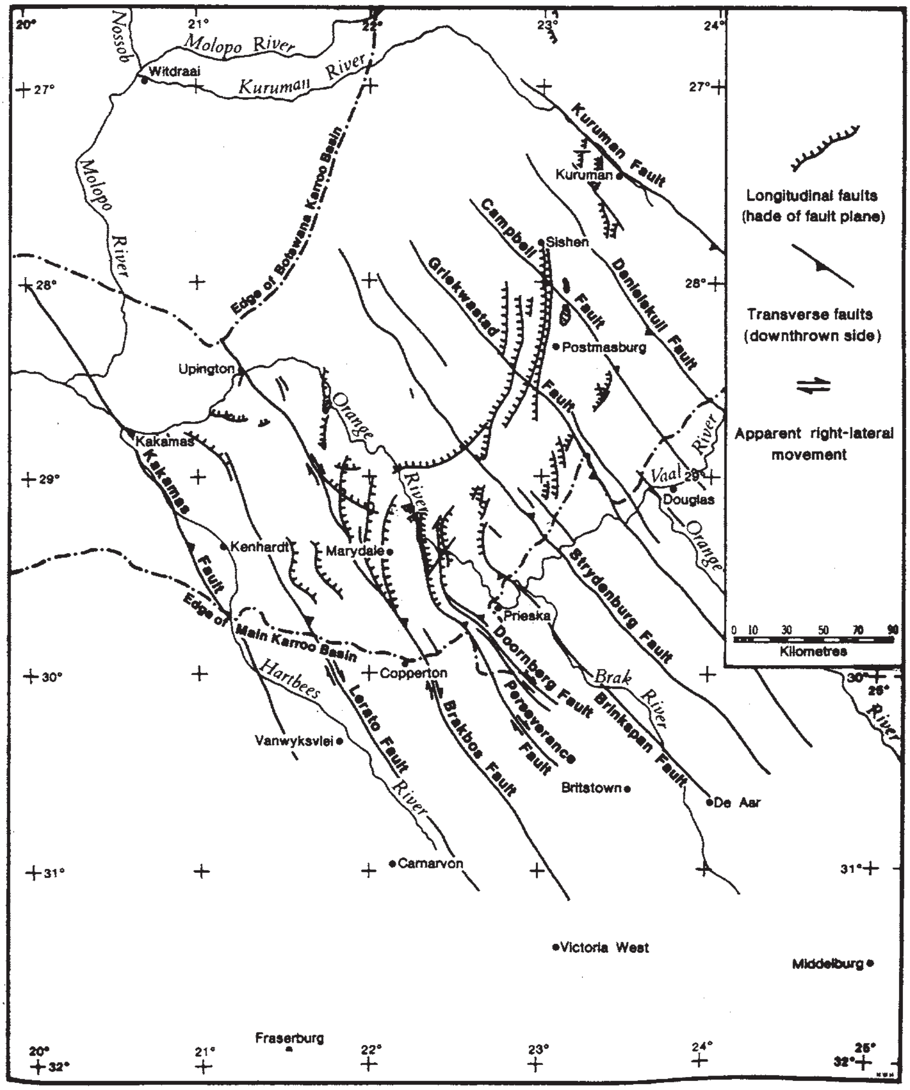

FIGURE 13.3B Major faults within the regional structure as interpreted from the gravity field over the Brakos-Doornberg lineaments in the Gordonia-Carnarvon region. Notice significant role of deep-seated strike-slip faults (Pretorius 1973).

Figure 13.4). Lamproitic deposits are located within the King Leopold and Halls Creek mobile zones near Kimberley block in Australia (see Chapter 1, History of Diamond Discoveries, Figure 1.7). It is also well known that the Roman ultrapotassic province is associated with the Tyrrhenian block.

The constant association of ultramafic-alkaline magmas with specific types of structures leads to the conclusion that both features are genetically interconnected; i.e.,

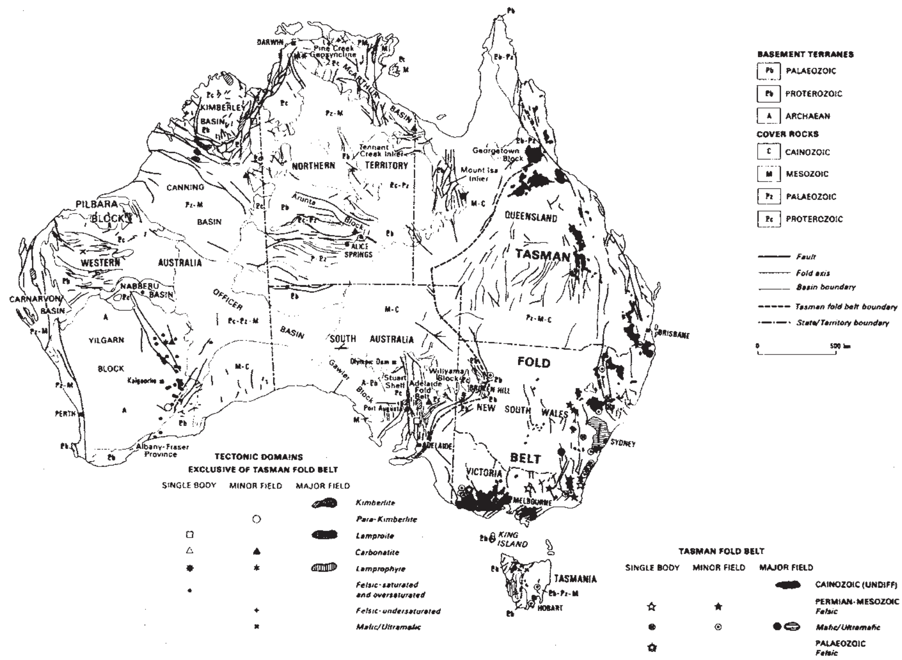

FIGURE 13.4 Distribution of alkaline rocks in Australia with respect to major crustal subdivisions (Jaques et al. 1985). Reprinted with permission of the Geological Society of South Africa.

structural development and magma generation occur as a result of same type of deep-seated activity generated by specific types of volatiles deep within the earth.

This conclusion is underscored by the almost complete absence of flood basalts within areas of ultramafic-alkalic and kimberlitic magmatism. Only rare (although lengthy) dikes and isolated sills are present, in contrast with the adjacent Karroo-Tunguska type of syneclise that is associated with flood basalts.

STRUCTURES THAT CONTROL SPATIAL DISTRIBUTION OF DIAMOND DEPOSITS AND ULTRAMAFIC-ALKALINE MAGMAS IN THE CRUST

In the sections below other types of structures that control the distribution of certain ultramafic-alkaline magmas in the crust are discussed. In contrast to the structures described in the previous section, those discussed here were formed by different types of crustal deformation processes. It is suggested that these types of processes were generated at much shallower depths—possibly at the asthenosphere/lithosphere boundary. Examples of these shallower structures are described below.

Deep-Seated Faults

Ultramafic-alkalic magmas are typically localized along deep-seated fault zones. The concept of structural zones controlling the emplacement of kimberlite was first recognized following the initial discoveries of kimberlite (DuToit 1906; Harger 1906;

Arsenyev 1962; Erlich 1963, 1985; Bardet 1964). However, the nature of the deep-seated faults remain unknown, because such faults were often drawn simply as linear zones connecting points of known kimberlites. However, most diamondiferous kimberlites lie within stable blocks with no clear trace of any type of structure. Despite this early drawing of lines connecting different kimberlitic fields, in many cases no obvious fault zones are associated with the pipes.

This absence of controlling fractures has been explained by a process in which the kimberlitic gases, which are under tremendous pressure, simply bore through the earth's crust and propagate a fracture above the rising magma (DuToit 1906). Examination of the problem in relationship to the Siberian platform led Grinson (1984) to propose that the deep-seated faults degrade upward into small faults and flexures that are assigned to the category of "hidden faults in the basement."

Extension zones in the lower part of the crust that control the emplacement of kimberlite are characterized by decreased density as a result of decompression within a deep-seated diapir formed by the magma. Such diapirs can be a source of melts in the upper horizons of the crust.

One example of the lack of direct evidence for structural control came from the most prominent (and most geologically studied) region for kimberlites in the world, South Africa. Research showed that the kimberlites are linearly distributed and associated with fracture zones. However, an underlying problem is related to the linearity that most likely reflects the deepest structural patterns, such as linear flexures along with a changing thickness of the earth's crust.

Commonly, there is no direct evidence of the postulated faults owing to poor exposure and to the specific type of manifestations of the faults at the surface. In some cases, their existence has been denied (see Milashev 1983, 1984; Milashev and Sokolova 1984, 2000). But even these extreme statements acknowledged that kimberlitic bodies are located almost exclusively within blocks characterized by intense fracturing and that fractures within these blocks are isotropic. In the Siberian and Russian platforms 190 kimberlitic intrusives are concentrated within blocks containing isotropic fracturing, and only six are found within blocks containing anisotropic fracturing (Milashev and Sokolova 2000).

Questions that remain include the following: Is fracturing isotropy in itself a sufficient and necessary condition for the localization of kimberlite? Do specific conditions apply to deep-seated fault zones that govern the spatial distribution of kimberlitic magmatism?

Deep-Seated Transform Strike-Slip Faults

Structural control is often not evident. However, the deep-seated control of ultramafic-alkaline massifs, kimberlites, and lamproites has been suggested to be associated with longitudinal and transverse strike-slip fault zones. Analysis of examples show that such zones can be equally superimposed on stable platforms or adjacent mobile orogenic belts. They may or may not be localized at the extension of transform faults cutting mid-oceanic ridges. Examples can be seen in the localization of kimberlites in Namibia and Angola (Figure 13.5). The role of transform faults controlling emplacement of kimberlitic fields in West Africa has been widely discussed (Williams and Williams 1977; Hastings and Sharp 1979).

The localization of the Murcia-Almeria lamproite province is related to a system of fractures generated by strike-slip fault tectonics (Figure 13.6). Transverse fault zones superimposed onto orogenic belts can be seen in the Cordillera province, North America,

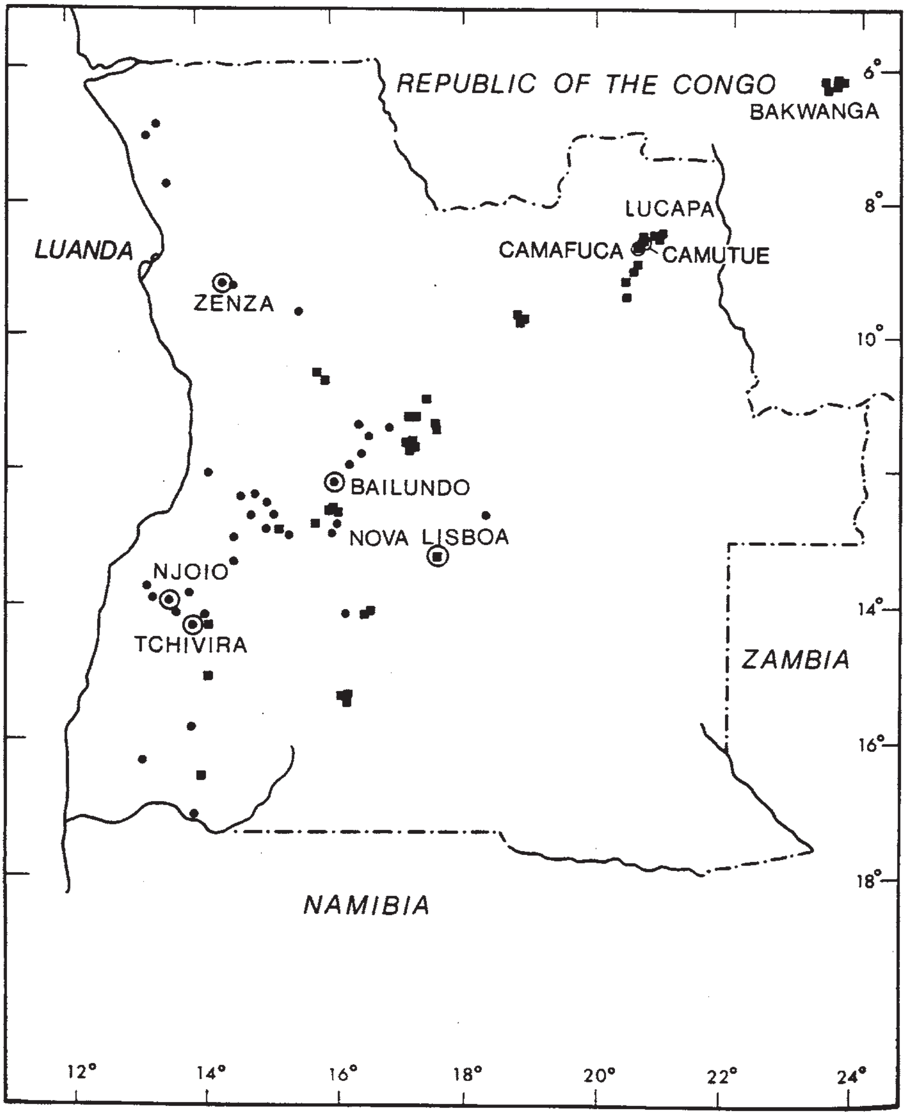

FIGURE 13.5 Location of kimberlites in Angola (+) and carbonatites/alkalic complexes (⊙) (Allsop and Hardgrave 1985). Reprinted with permission of Geological Society of South Africa.

where these types of faults determine the location of a series of kimberlite, lamproite, and lamprophyre subprovinces within the Wyoming craton. The maps and structural schemes for these fields show fissures that are quite typical of those associated with strike-slip fault tectonics.

Another type of transverse fault zone cuts mobile orogenic belts adjacent to stable blocks or cratons. Deep-seated strike-slip faults superimposed onto mobile belts adjacent to stable blocks can be seen in the Murcia-Almeria lamproitic province superimposed

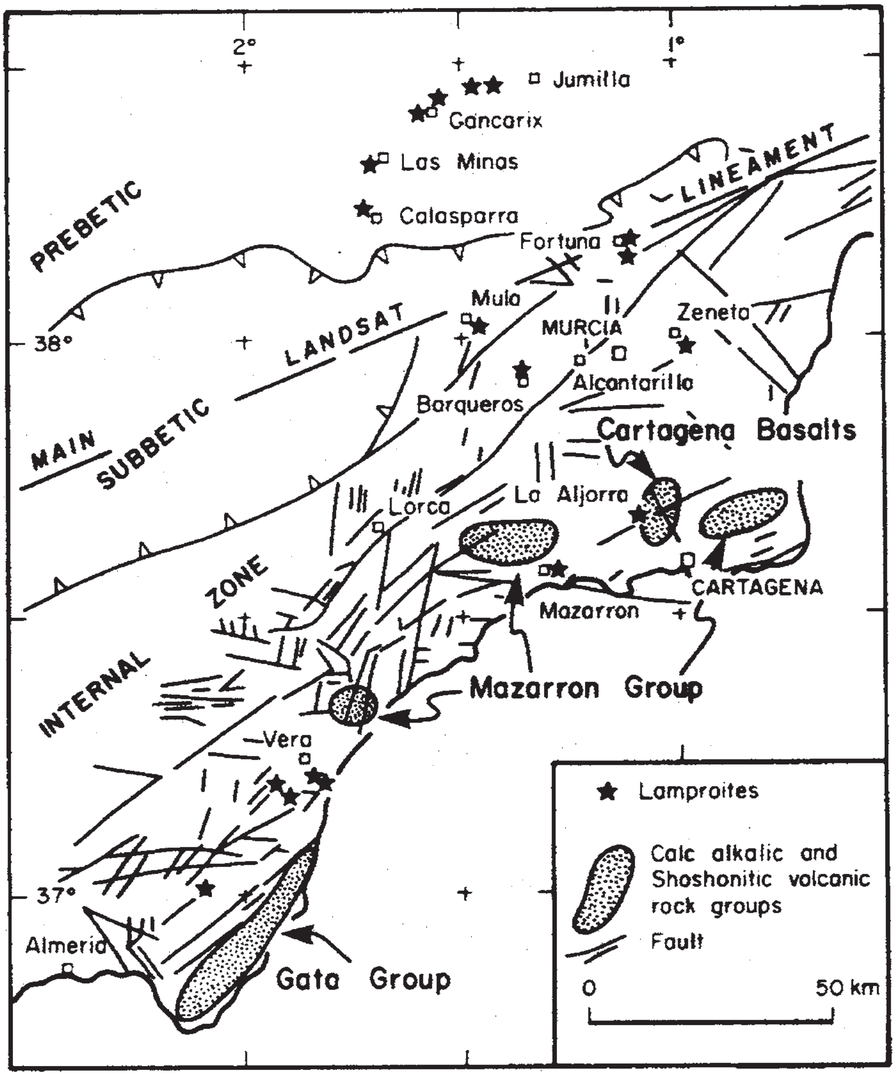

FIGURE 13.6 Spatial distribution of lamproites within the eastern Betic Cordillera including Murcia-Almeria province, Spain. General outline of structural patterns typical of strike-slip fault tectonics (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/ Plenum Publisher.

onto the Betic Cordillera in Spain (Figure 13.6). It shows structures typical of strike-slip fault tectonics. Other examples include the Argyle lamproite in Western Australia (Figure 13.4), and the Appalachian province in North America (Figure 13.7). The Smoky Butte lamproites in Montana are distributed along a strain fracture that may have formed at the time of magma emplacement (Figure 13.8).

Strike-slip fault dislocations that are superimposed onto stable massifs may give rise to kimberlitic provinces. The same type of dislocations developed over orogenic zones or

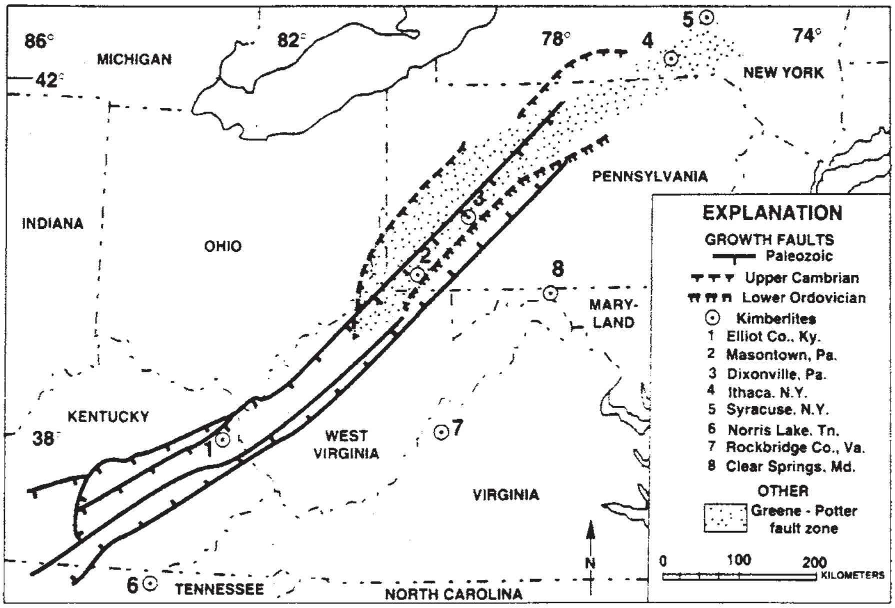

FIGURE 13.7 Structural position of kimberlites in the Appalachians (Parrish and Lavin 1982)

transverse fault zones may give rise to lamproites. Both types of rocks are characteristically associated with compressional zones.

Aulacogens and Grabens

One form of linear structures that appear to control the emplacement of ultramafic-alkalic magmatic bodies in general, and kimberlitic pipes in particular, are aulacogens—linear tectonic troughs that cut platforms. An analysis of the deep structure between the roof of crystalline basement of a platform and the overlying Paleozoic sediments shows the existence of a series of linear troughs (or aulacogens) cutting the platform.

Shatsky (1946, 1955), who first recognized these troughs as specific type of depressions, introduced the term *aulacogen*. Later it was shown that some of these structures were closely associated with the development of mobile orogenic belts adjacent to the ancient platforms, such as the Ouachita mobile belt in North America. The Wichita system in North America has also been shown to have similarities to aulacogens. Since that time extensive data on aulacogens in all ancient platforms have been accumulated, and they were summarized by Burke (1977).

Analysis of the African rift system suggests the existence of three-armed rifts (triple junctions) that are related to doming (Burke and Whiteman 1973). As in the case for "anteclise/swell," the term *aulacogen*, born in the Russian geologic tradition, reflects not only recent morphology but also the formational history of the structure. In a pure morphologic sense, the term *graben* or *rift* is used for similar structures.

Despite the fact that aulacogens and grabens are bounded by normal faults, the dynamic conditions within and around them are determined by their tendency to extend along strike. Examples of well-studied grabens include the East African, Rhine, and

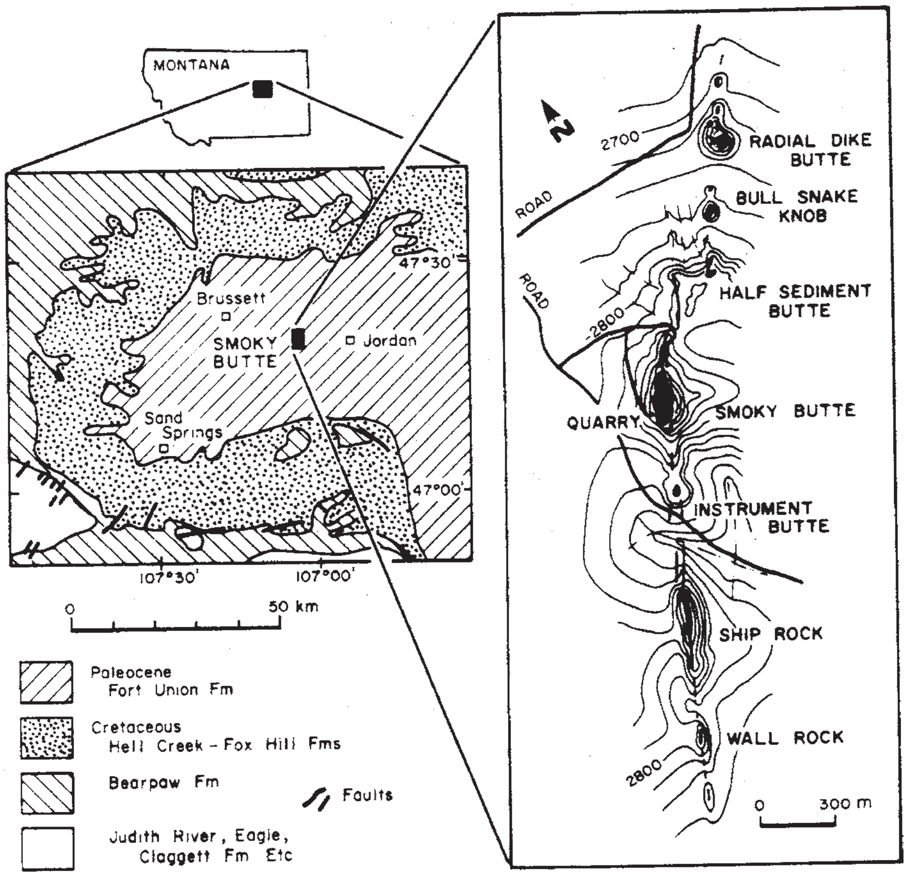

FIGURE 13.8 Geological map of the Smoky Butte lamproite field, Montana (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

Baikal grabens. Within the Ottawa-Grenville-St. Lawrence province, a Late Proterozoic (Ryphean, Late Proterozoic time) aulacogen has degenerated with time into the St. Lawrence graben.

Such troughs appear in the form of linear depressions that are several hundreds or even a few thousand kilometers long and tens of kilometers wide. These troughs are usually filled by undeformed volcano-sedimentary sequences up to several thousand meters thick. The structural position of such troughs usually coincides with sutures in the crystalline basement that divide blocks of different ages and compositions of the crystalline rocks. Many of these troughs develop as linear depressions during the early stages of the development of the platform (in Ryphean time) and tectonic reconstruction at the Cambrian-Proterozoic boundary in part by the formation of linear anticlines. Examples of such aulacogens that formed during the early stages of platform development include the Pachelmsky and Dneprovsko-Donetsky (Donbass) depressions, which cut the Russian platform, and the Midcontinent rift of the North American platform. Aulacogens developed at this stage of development in cratons are well expressed as

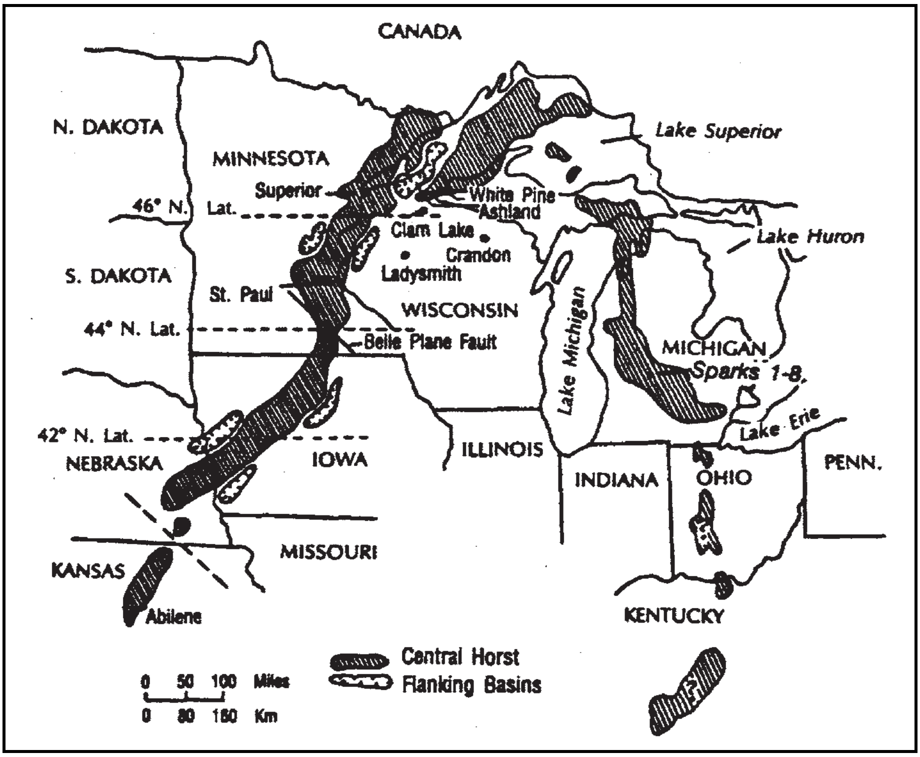

FIGURE 13.9 Midcontinent rift zone in United States, a buried aulacogen (Dickas 1984). Reprinted with permission from the Oil & Gas Journal.

linear troughs in the relief of the crystalline basement in the platform (as can be seen in the Russian platform) (Figure 13.1).

Aulacogens are expressed by positive gravity and magnetic anomalies. The Midcontinent rift in North America is synonymous with the term Midcontinent gravity high seen in positive Bouguer values for some 1,287 km (Figure 13.9). Residual values across this feature can approach 175 mgals (Dickas 1984) (Figure 13.10). Magnetic anomalies over these troughs are intricate but in general are represented by a series of weak linear positive anomalies. A series of such anomalies traces major aulacogens in the Siberian platform, which are represented by the Taymir-Baikal and Udjinsky fault zones (Savinsky 1972).

It is generally considered that graben systems are characterized by strike-slip faulting associated with tensional tectonics. Horizontal tension along the strike of the grabens produces expansion fissures along strike; radiometric dating confirms this relationship. Two reasons for tensional tectonics include plate migration over hot spots and fissure propagation.

Such a dynamic system plays a leading role not only in the distribution of magmatism within grabens, but also (and predominantly) in the creation of a system of fissures adjacent to stable blocks. For example, most kimberlitic bodies in Tanganyika are located not within grabens, but rather within adjacent stable blocks (Edwards and Howkins 1966). It is seen from the structural scheme presented by Kostyuk (1983) that

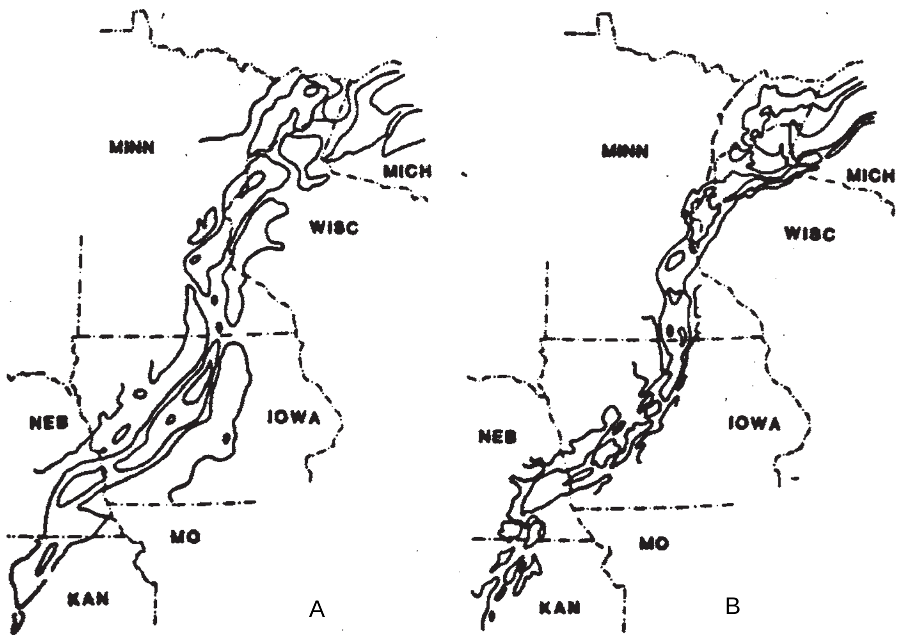

FIGURE 13.10 Geophysical characteristics of the Midcontinent rift zone. A = Positive Bouguer gravity anomalies, B = Aeromagnetic anomalies (Lee and Kerr 1984). Reprinted with permission from the Oil & Gas Journal.

the structural position of ultramafic-alkalic rocks in the Aldan province is associated with the Baikal graben tectonic system, which cuts the Aldan anteclise (see Figure 13.11).

Bodies of ultramafic-alkalic rocks typically are roughly linearly distributed along the strike of aulacogens. Ultramafic-alkalic magmatism usually begins at the early stage of their development and continues as a series of minor phases during later periods. Thus, it has been suggested that the magma persists at depth during the early period of formation of these structures, while structurally the aulacogens degenerate and cease to exist. Each later phase of magmatic activity reflects stages of compression within the platform.

Examples of associations of ultramafic-alkalic volcanism with aulacogens can be seen within all known platforms. In the most common cases, massifs of ultramafic-alkaline rocks are roughly linearly distributed along the axis of the aulacogen. An example of such locations can be seen in the form of axial dikes that follow the axis of grabens (Griffits et al. 1971). This tendency can be illustrated in most aulacogens. A few include the

- Eastern European platform—Donetsko-Dneprovsky and Pachelmsky aulacogens (Figure 13.1)
- Siberian platform (aulacogens follow deep-seated linear faults such as Taymir-Baikal and Udja faults) (Figure 13.2)
- Midcontinent rift in North America, and the Ottawa-Grenville-St. Lawrence, and Ouachita provinces
- ultramafic-alkaline rocks and lamproites within the Aldan shield (Figure 13.11)

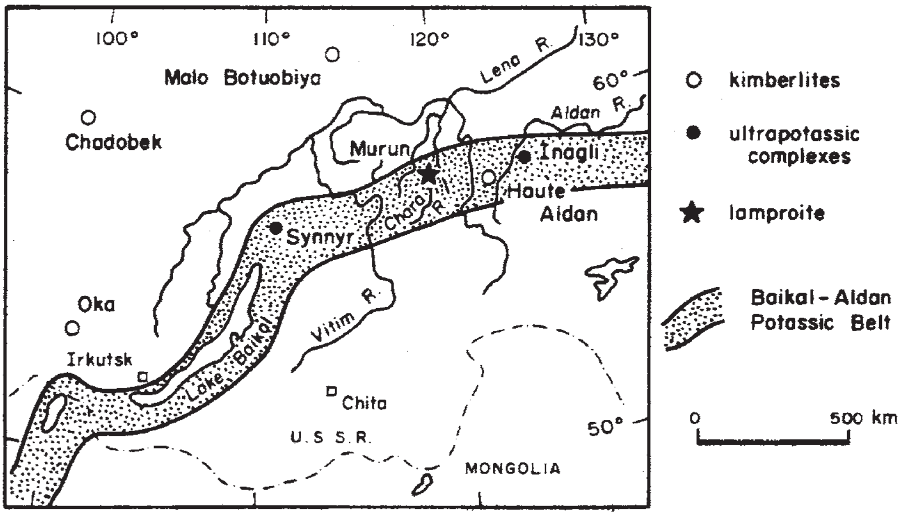

FIGURE 13.11 Lamproites of the Aldan province associated with the Baikal graben system. Potassic volcanic province corresponds with the area of the Baikal graben system (modified from Kostyuk 1983).

- linear distribution of lamproites within the Jacupiranga zone in Brazil (Figure 1.2)
- linear distribution of lamproites within the Ouachita province (Figure 13.12).

The same structural position is characteristic of alkalic volcanism related to the East African rift system, the Oslo graben, the Rhone and Rhine grabens in Europe, and the Baikal graben in Siberia. A direct association of grabens with swells has been described by LeBas (1971) (Figure 13.13).

In the Ottawa-St. Lawrence rift zone (North America) and Udja aulacogen (Siberian platform), aulacogens (or grabens) cease to exist as tectonically expressed structures, but magmas generated during the early stages of their development persist at depth. As a result, any new compressional event in the platform is followed by magmatic intrusive events, owing to the upward rise of magmas. Such relict zones reflect the existence of trough-like structure at depth. The average crustal thickness within such structures is 40 km or less, the thickness of the granitic layer is reduced by half, and the thickness of the basaltic layer is increased accordingly. In comparison with continental riftogenetic structures, they are characterized by increased density of fracturing, by increased thickness of sedimentary cover, and by anomalously low (7.3–7.4 km/s) seismic velocities at the mantle/crust boundary (Grinson 1984). The same has been observed beneath the buried North American aulacogen in the Midcontinent gravity high.

When strike-slip fault dislocations are superimposed on stable massifs, kimberlitic provinces are sometimes found. When the same types of dislocations are developed over orogenic zones or transverse fault zones, lamproites are found. Both types of rocks are characteristically associated with compressional zones.

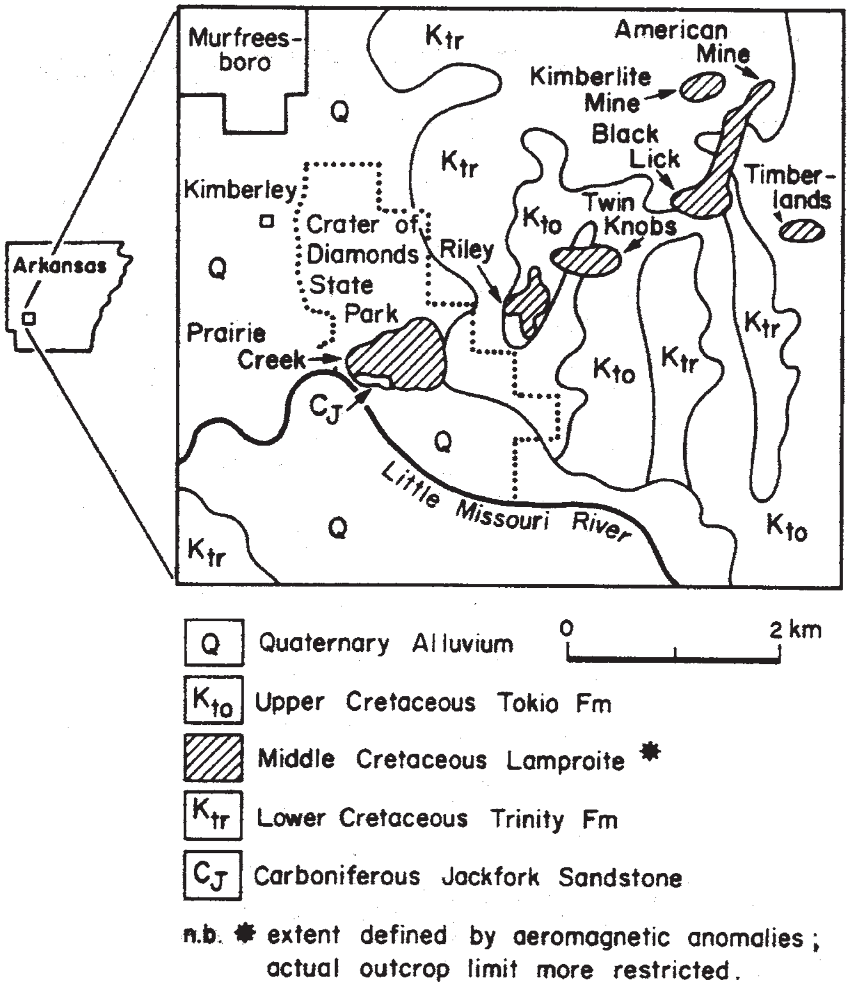

FIGURE 13.12 Linear distribution of lamproite bodies in the Ouachita province in Arkansas (Mitchell and Bergman 1991). Reprinted with permission from Kluwer Academic/Plenum Publisher.

Geodynamic Systems Within Stable Blocks of the Siberian Platform

It is quite difficult to analyze the spatial distribution of ultramafic-alkalic magmatism and kimberlite fields within stable blocks owing to the poor exposure of kimberlite. An analysis of geodynamic conditions within such blocks may provide some information. Some of the best opportunities for this are provided by a region east of Anabar shield, which can be used as a kind of "Rosetta stone," a key for decoding of the general regularities of tectonic control of kimberlite.

It is necessary to note that fault orientations and elongation of kimberlite fields do not coincide with the trend of Archean greenstone belts and supracrustal complexes within the basement of the platform, as they are traced by aeromagnetic anomalies.

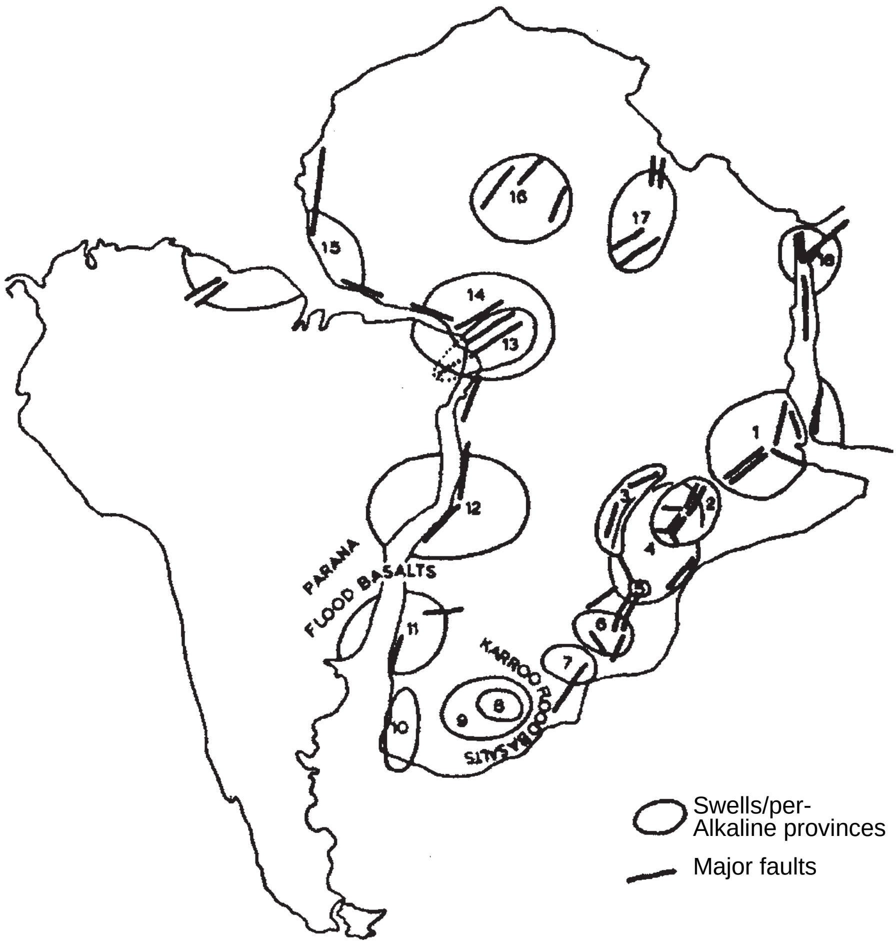

FIGURE 13.13 Swells and grabens in Africa and distribution of alkaline volcanism. Swells are defined by gravity anomalies (from LeBas 1971).

However, these complexes influence the development of structures within the cover of the platform starting from the Proterozoic through Quaternary time.

North-south–trending linear positive magnetic and gravity anomalies are notable in the platform. These can be traced for more than 300 km and are about 30 km wide. They are similar to the Midcontinent magnetic high in North America, which coincides with a buried Proterozoic Midcontinent rift. A similar north-south magnetic and gravity anomaly is located east of Anabar shield. The northernmost edge of this anomaly coincides with the Udja anticline, which was formed as a result of inversion of the Late Proterozoic aulacogen (Figure 13.14). An orthogonal (north-south and east-west) system of deep-seated faults have an appearance of being generated by a global stress field.

Although some structures during the course of geologic history have changed orientation, their boundaries remain the same. East of the Anabar shield, these faults trend northeast and northwest and exhibit a constant 30° to 40° angle with the apex. This angle is

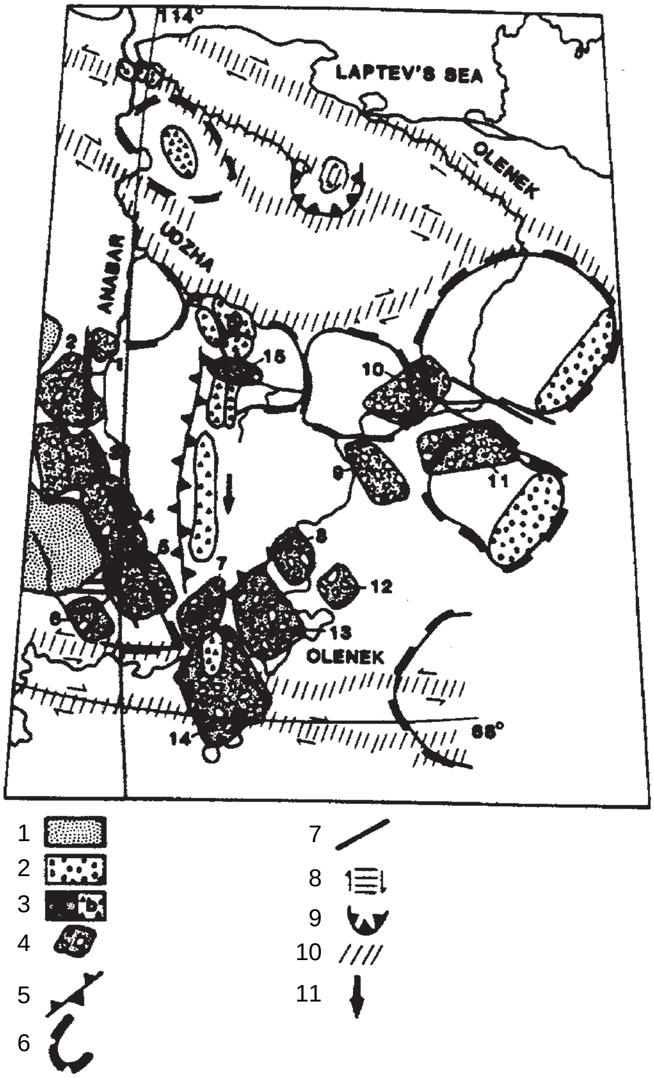

FIGURE 13.14 Kimberlitic fields within Kuonamka and Olenek regions, northeastern Siberian platform, in relation to the north-south–striking deep-seated Udja fault zone. 1 = structures within crystalline basement, 2 = anticline warp structure and suggested Proterozoic uplift, 3 = ultramafic-alkaline intrusions (a = mapped on the ground, b = inferred by geophysical data), 4 = kimberlite fields (numbers 1 through 15 indicate various fields), 5 = axes of zones of maximum Paleozoic sedimentation, 6 = Paleozoic domal uplifts (a = geologically mapped, b = inferred by geophysical data), 7 = faults mapped on the ground, 8 = faults inferred from geophysical data, 9 = inferred volcano-tectonic faults, 10 = inferred deep-seated strike-slip faults along the Anabar shield's eastern margin, 11 = inferred strain direction along north-south deep-seated fault zone in Paleozoic and Mesozoic time. Numbers on the figure designate kimberlitic fields (Erlich 1985). Reprinted with permission of the Geological Society of South Africa.

close to the theoretical angle between strike-slip fault systems under constant horizontal pressure (31° ± 2°). The north-south–trending structure of the aulacogen is reflected by a positive gravity-magnetic anomaly that forms a bisector within this system. Such regular repetitions indicate that a constant geodynamic condition persisted throughout geologic history. This condition is characteristic of the Udja anticline, where northeast-northwest faults were generated by north-south–oriented pressures. Along faults of this wedge-like system, wedges were pushed upward resulting in the formation of an anticline. Horizontal displacement along these faults may be as much as 2 to 2.5 km.

During the Cambrian, the northwest-trending eastern boundary of Anabar shield controlled intense sedimentation in the Sukhaana basin, which has a northeast-trending axis. The same northwest and northeast trends determine the boundaries of the Permian–Mesozoic Leno-Anabar linear marginal depression in the Siberian platform. Once again, the north-south–trending Udja anticline is located exactly within the southward apex of this angle.

The northwest- and northeast-trending fault systems control the location of major kimberlite-carbonatite fields in this region. Parallel to the shield's eastern boundary is a series of kimberlitic and carbonatite fields (the Djuken, Luchakan, and Kuonamka fields). The latter two fields were described by Dawson (1980) under the names of Lower and Middle Kuonamka fields. Close to these is the Kuranakh field, but it is likely that the latter is associated with east-west–trending fault zones south of the Anabar shield.

Geologic mapping clearly shows that northwest-trending strike-slip faults in the southeast corner of the Anabar shield exhibit left-lateral movement (Voronov and Erlich 1962), and the amplitude of horizontal displacement reached several kilometers. However, there is a notable difference in the orientation of elongated axes of kimberlite pipes and dikes in relation to the strike of the main deep-seated faults. The long axes of kimberlites form an acute angle to the northwest-trending fault zone and lie parallel to the Anabar shield's eastern boundary. Kimberlites located along northeast-trending zones are elongated parallel to the strike of the deep-seated faults (Figure 13.15). This difference can be explained by the tendency of the geodynamic system to exhibit counterclockwise rotation. If such a tendency exists, strain along the northeast-trending zone is characterized by extension, whereas along northwest-trending zones it is characterized by compression. The general geometry of the entire geodynamic system is analogous to the ξ-type vortex systems described by Lee (1958).

Northeast-trending fault zones control the location of the Chomurdaakh, Omonoos-Ukukit, Lower Ukukit, Motorchuna, and Omonoos-Sukhaana fields (cited by Dawson [1980] as the Middle Olenek group of fields). Considering the Paleozoic–Mesozoic ages of the kimberlite-carbonatite intrusives, such a spatial distribution of kimberlite magmatism provides direct proof that diagonal northeast- and northwest-trending faults continue to persist throughout Paleozoic–Mesozoic time. The amplitude of vertical displacement is 1,500 m, and horizontal displacement along this fault reaches many kilometers.

All kimberlites that have elongated axes parallel to the main fault zone were formed under considerable compression in the host rocks, possibly influenced by endogenous factors such as explosive processes or mechanical activity of magma. Such conditions indicate that the character of contacts with the host rocks are oriented along feather joints, which are elongated by an acute angle to the trend of a major fracture zone. Thus a significant number of kimberlites in the Kuonamka and Luchakan fields, which are primarily oriented east-west, are located along a northwest strike. The long axes of kimberlites in the Chomurdakh field are predominately oriented to the northeast and located along a north-south striking fault (Figures 13.16, 13.17).

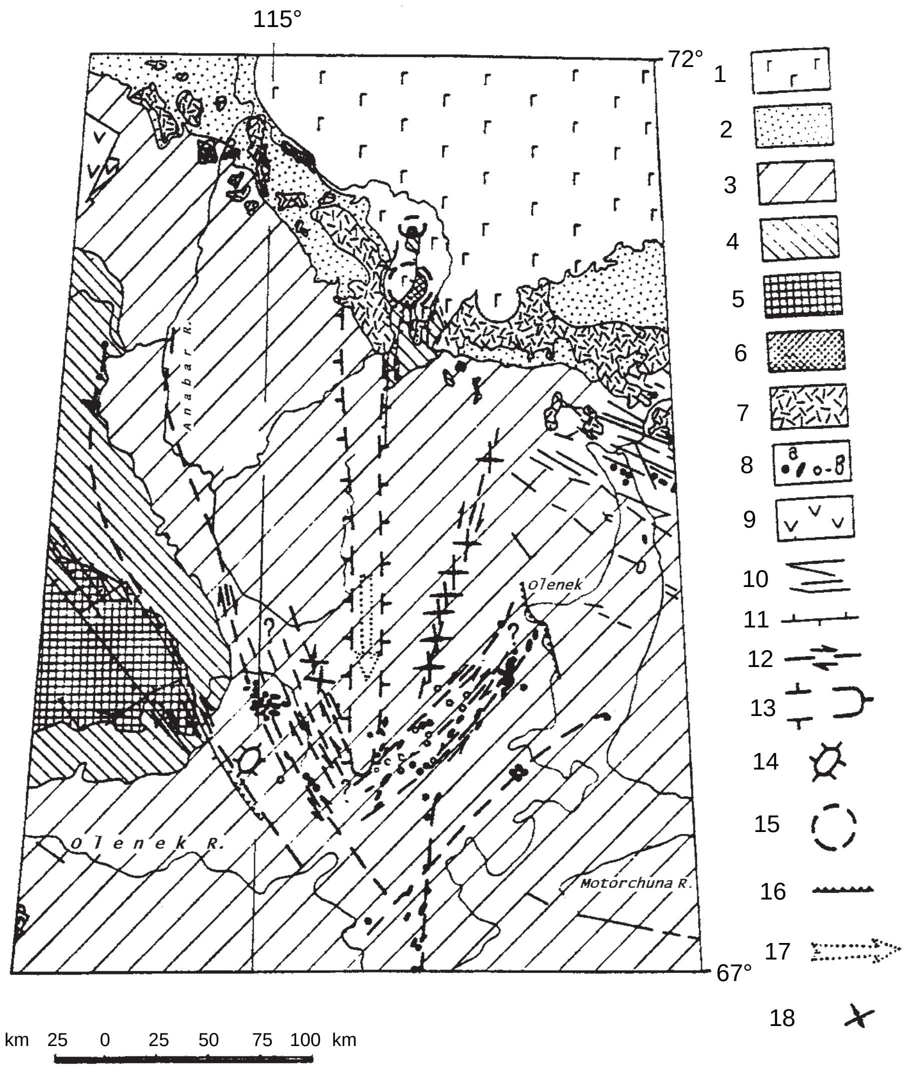

FIGURE 13.15 Geodynamics of the Anabar-Olenek region. 1 = Mesozoic deposits, 2 = Permian deposits, 3 = Cambrian deposits, 4 = Ryphean deposits, 5 = Archean crystalline basement, 6 = intrusions of ultramafic-alkaline rocks, 7 = sills and volcanic rocks associated with flood basalt formation, 8 = kimberlitic dikes and pipes (a = established, b = suggested), 9 = Cretaceous volcanic rocks associated with the Popigay ring structure, 10 = fissure-type faults, 11 = faults with observed amplitude of horizontal displacement, 12 = zone of "initial" strike-slip faults, 13 = positive magnetic anomaly along the suggested Ryphean aulacogen, 14 = domal uplift, 15 = magnetic/gravity anomalies associated with buried ultramafic-alkaline intrusives, 16 = flexures, 17 = main direction of tangential compression, 18 = folds within platform cover (modified from Voronov and Erlich 1962).

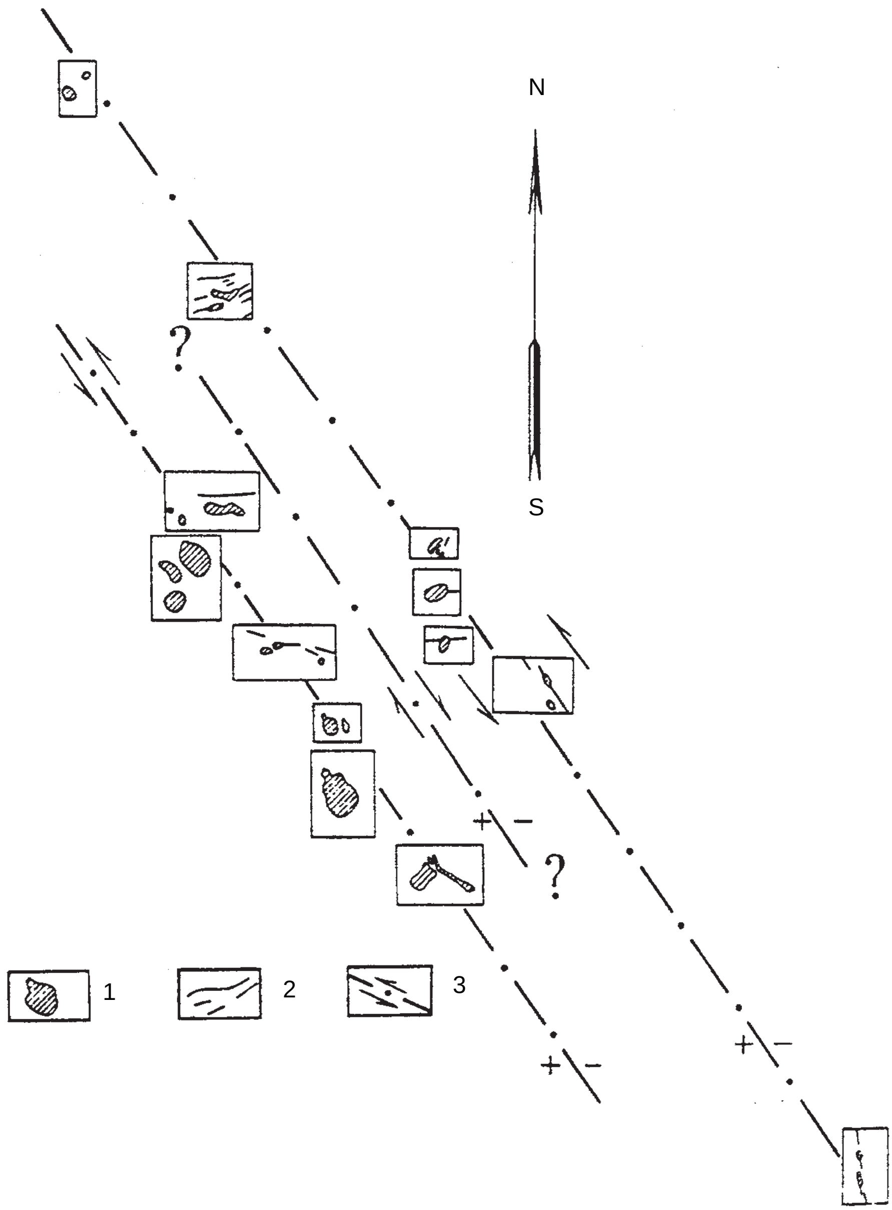

FIGURE 13.16 Suggested strike-slip fault zones within the crystalline basement, Luchakan kimberlite field. 1 = kimberlitic bodies and pipes, 2 = kimberlitic dikes, 3 = suggested strike-slip fault zone and direction of displacement at its wings (Voronov and Erlich 1962).

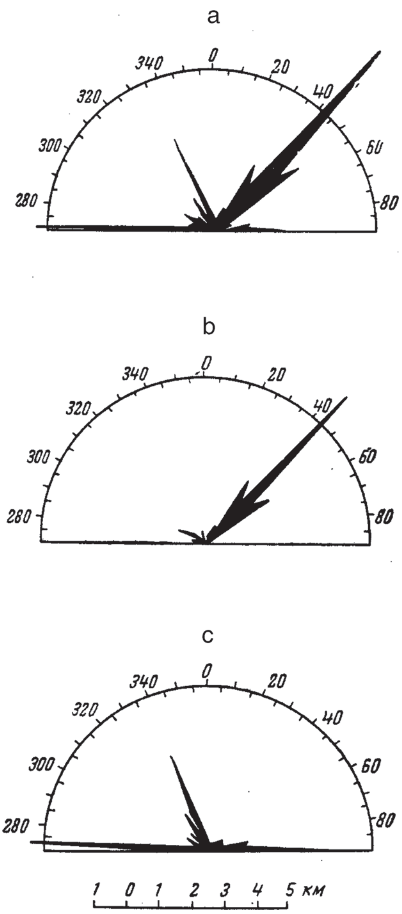

FIGURE 13.17 Orientation of long axes of kimberlitic bodies within different kimberlitic fields of the northeastern Siberian platform. a = Anabar-Olenek region (total length 44.8 km), b = Omonos-Ukukit kimberlite field (16.4 km), c = Luchakan-Kuonamsky kimberlite field (16.6 km) (Milashev et al. 1963).

It is necessary to recognize the difference in the structural position of different units of ultramafic-alkaline intrusives within this system. Whereas ultramafic-alkaline plutons of ijolite-carbonatite and agpaitic nepheline syenite compositions are controlled by the north-south–trending Udja anticline, magmatic bodies located along "diagonal" faults are characterized by the absence of alkaline and ultramafic-alkaline volcanic rocks except for kimberlites (mainly Group II kimberlite) and carbonatites. The age of these rocks is lower-middle Paleozoic to Mesozoic.

The spatial relations between ijolite-carbonatite massifs and kimberlites within the Udja anticline are not clear. But it is quite possible that the Tomtor ijolite-carbonatitic ring complex was formed at a stage of aulacogen development, and the kimberlites were emplaced during the time of the closing of the aulacogen, and a linear anticline was developed in its place.

No mafic (doleritic) intrusives are known in connection with these zones, indicating that compression conditions continued during the Permo-Triassic episode of flood-basalt eruption. Movements, or at least tension, along these fault zones has been renewed during recent stage of tectonic development, as indicated by recent stream valleys reorienting to follow the strike of these zones.

The same difference in structural position between ijolite-carbonatite massifs on one hand and kimberlite/carbonatitic formation on the other is observed within different regions. Thus in the Kotuy-Maimecha province west of the Anabar shield, ultramafic-alkalic plutons of ijolite-carbonatite affinity are distributed within a north-south–trending strip expressed by a linear positive magnetic and gravity anomaly similar to the Midcontinent magnetic high in North America and Udjinsky magnetic and gravity anomaly east of the Anabar shield. Fault zones and zones of intense fracturing in the platform's cover govern spatial distribution of the massifs within Kotuy-Maimecha province (Figure 13.18). The single kimberlite (Dalbykhsky) field in this region is located south of the Anabar shield along an east-west–trending deep-seated fault.

Similar structural locations are characteristic of Africa, where carbonatite-nephelinitic magmatic complexes and kimberlites tend to be localized along active rift zones within adjacent stable blocks (LeBas 1977), such as those observed in Tanzania (Edwards and Howkins 1966). This difference in structural localization reflect dual dynamic conditions generated in association with grabens/aulacogens: zones of general extension that are favored for the generation of carbonatite-nephelinitic and ijolite-carbonatitic magmas, whereas kimberlitic and carbonatitic magmas tend to be located within zones characterized by compression at depth associated with strike-slip faults. The spatial distribution of specific magmatic bodies is determined by zones of weakness within the sedimentary cover. An example of this is the distribution of ultramafic-alkalic massifs within the Kotuy-Maimecha province.

It has been suggested that magmas from the asthenosphere penetrate the lithosphere where extension fractures occur. In order for tensional failures to occur in the lithosphere, elastic stresses must build over a long time. As a result, the beginning and termination of cycles of volcanic activity within the Gregory (eastern) rift zone (Logachev, Belousov, and Milanovsky 1972) coincides at 23 and 16 Ma with the timing of major global episodes of kimberlite intrusion (see Chapter 14, Temporal Distribution of Diamond Deposits).

Deep-Seated Strike-Slip Faults Within the Siberian Platform

The other portion of the orthogonal deep-seated fault system, which includes north-south–trending systems of positive gravimagnetic anomalies, consists of east-west–trending fault zones. In the northeastern part of the Siberian platform, at least three major fault systems are recognized:

- deep-seated faults south of the Anabar shield that cut the Siberian platform approximately along latitude N 68° (Savinsky 1972)
- deep-seated faults along the northern boundary of the Siberian platform, which are related with the Verkhoyansky folded zone that define the strike of the Leno-Anabar linear depression
- deep-seated faults in the southern part of the Siberian platform at the edge of the Chadobetsky uplift.

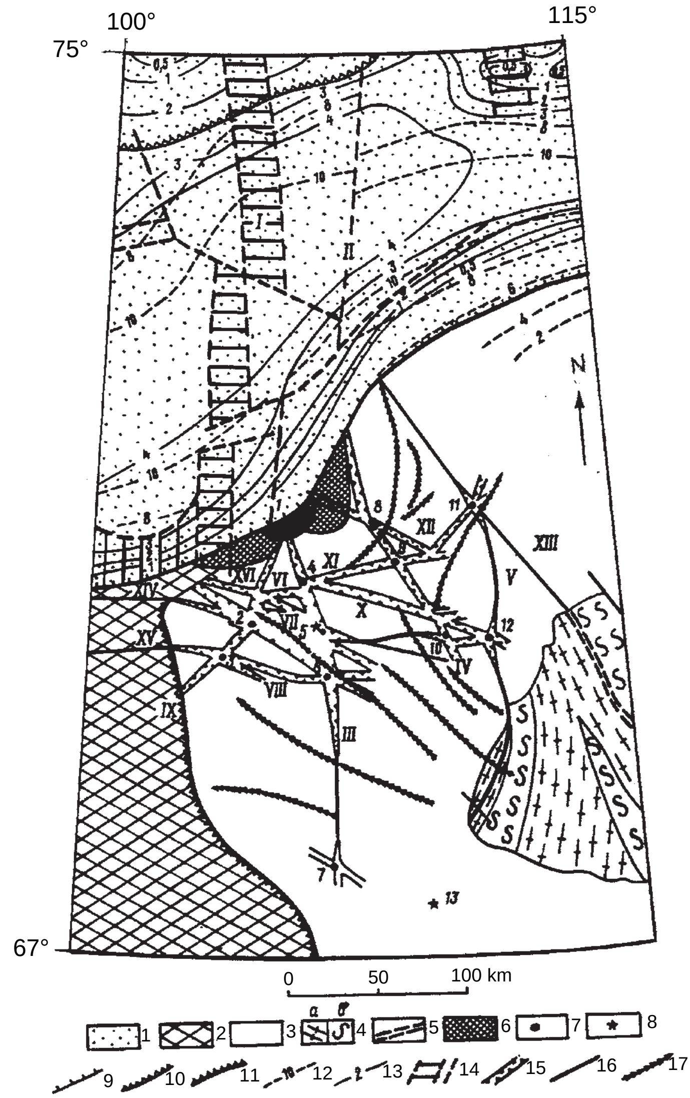

FIGURE 13.18 Distribution of ultramafic-alkaline massifs within the Kotuy-Maimecha province. 1 = Mesozoic–Cenozoic Khatanga depression, 2 = Permo-Triassic Tunguska syneclise (the field of flood basalt development), 3 = areas of intense uplift in Mesozoic–Cenozoic time (Roman numerals in the field refer to deep fault zones), 4 = Archean granulite series of the Anabar shield, 5 = Early Proterozoic shear zones within Anabar shield, 6 = depressions filled by ultramafic-alkaline volcanic rocks, 7 = Early Triassic ijoilite-carbonatitic massifs (1 = Gulinsky, 2 = Romanikha, 3 = Changit, 4 = Sedete, 5 = Dalbykha, 6 = Bor-Uryakh, 7 = Essey, 8 = Odikhincha, 9 = Kugda, 10 = Magan, 11 = Nemakit, 12 = Yiraas), 8 = kimberlitic fields (5 = Dalbykha, 13 = Kharamiysky), 9 = boundary of Mesozoic cover development, 10 = boundary of upper Paleozoic sequence development under Mesozoic cover, 11 = eastern boundary of flood basalts of the Tunguska syneclise, 12 = roof of Precambrian basement (contours in km/s), 13 = roof of pre-Jurassic volcano-sedimentary rocks, 14 = deep-seated fault zones detected by geophysical data, 15 = deep fault zones controlling ultramafic-alkaline and kimberlitic intrusives, 16 = deep fault zones in the basement of Siberian Platform by geophysical data, 17 = regional fault zones filled by dolerite dikes. (I = Central-Taymir, II = Yuka-Kotuysky) (Egorov 1991). Reprinted with permission from Publishing House Nedra.

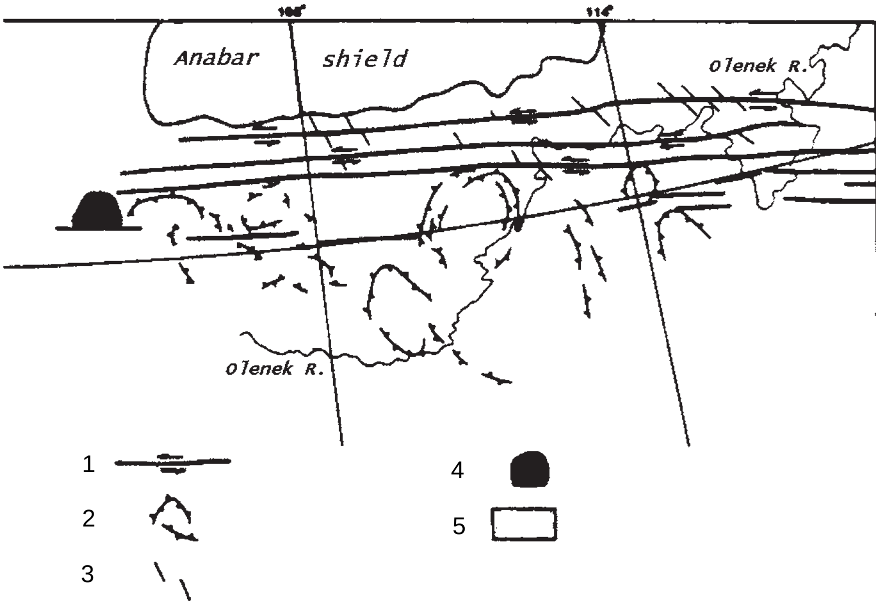

FIGURE 13.19 Horizontal displacement along east-west–striking deep-seated fault zones south of the Anabar shield. 1 = deep-seated strike-slip faults and direction of lateral displacement, 2 = linear dikes and ring complexes of mafic intrusions, 3 = axes of magnetic anomalies associated with complexes of platform's basement, 4 = magnetic anomalies associated with suggested ultramafic-alkaline intrusions, 5 = area of the Anabar shield (Erlich 1985). Reprinted with permission of the Geological Society of South Africa.

It is important to note that both north-south and east-west–trending deep-seated fault zones are "global through-structures;" i.e., they are not associated with a specific structure but instead cut a series of structures. They cut not only the platform but also cut adjacent structures, and the character of the geodynamic conditions was preserved throughout geologic time.

Both (north-south– and east-west–trending) fault systems are defined as strike-slip faults. This definition is clearly demonstrated south of the Anabar shield where axes of magnetic anomalies are interpreted to reflect igneous complexes in the crystalline basement that show displacement up to 15 to 20 km (Figure 13.19). At the same time, the surface of crystalline basement was displaced only 1.5 to 2 km.

Initial displacements along east-west–trending faults are at least late Ryphean. Regional tectonic reconstruction of Ryphean aulacogens indicates partial inversion, resulting in formation of anticlines similar to those in the Udja region. In Permo-Triassic time, faults of this system determined the distribution of lengthy doleritic dikes and ring dike complexes. These faults are partially exposed at the surface, but their extent was determined by aeromagnetic data.

Continued activity along such a fault system located south of the Anabar shield formed a series of depressions. As an example, the Aganily depression is more than 200 km long and is filled by a sequence of rapidly transported debris that is 800 m thick.

Quaternary mineralization along this fault includes some gold, galena, cinnabar, and sphalerite. This type of mineralization indicates that physicochemical conditions along east-west–trending faults are markedly different from those on the longitudinal zones, where there is an abundance of ultramafic-alkalic rocks.

Both north-south and east-west–striking elements of the orthogonal deep-seated fault system are strike-slip faults, but even a cursory comparison of the dynamics within each show that episodes of extension and compression alternated in the past.

Kimberlites of the Dalbikhsky field, located near the Kotuy-Maimecha province, are situated about 60 km south of the Essey intrusive complex at the extension of east-west–trending deep-seated faults parallel to the Anabar shield's southern boundary (Egorov 1991). Thus, it is possible to suggest that although ijolite-carbonatite massifs are associated with aulacogens and kimberlites, they are not associated with the aulacogen development but rather with the time of its closing and inversion. The data clearly show that horizontal movement is the main driving force and determines the strain within platform. The prevalence of horizontal strain at depth, in the zone of kimberlitic magma genesis, is confirmed by textures of xenoliths and inclusions in kimberlites.

ROLE OF ISOSTATIC EQUILIBRIUM

It is necessary to discuss another important point in a platform's structural development—a point that can provide a clue to the reconstruction of the stress field at different stages of geologic development and, accordingly, to the distribution of different types of igneous rock associations. At different stages of geologic development within northeastern part of the Siberian platform, pairs of structures exist that are equal in size but are characterized by opposing directions of tectonic movement (Erlich 1985). The very existence of such structural pairs fits the concept of "tectonopairs," isostatically equilibrated blocks equal in area but characterized by opposite directions of movement (Suvorov 1977). It is suggested that such tectonopairs reflect the tendency to restore isostatic equilibrium between their elements.

The first tectonopair consists of two triangular blocks. One roughly coincides with the Anabar shield, which lacks any Paleozoic sedimentary cover; the second roughly coincides with the Sukhaana basin, which is filled by the Upper Cambrian carbonates. Both blocks are divided by a system of deep-seated fault zones running parallel to the Anabar shield's eastern slope. The roof of the crystalline basement is step like and displaced several kilometers to the east, toward the Sukhaana basin. This zone shows clear indication of strike-slip fault movements, and it controls the location of series kimberlitic fields (Luchakan, Maly, and Sredne-Kuonamsky). These structures were created in the course of major tectonic transformation of the Siberian platform, which took place in late Ryphean-Vendian time. At this time, the Ryphean aulacogens closed, and a system of syneclises and anteclises characteristic of the platform stage of development were formed (Malich 1975).

The northern block, in the second tectonopair, coincides with boundaries of the first tectonopair described above. Both elements of the first tectonopair were consolidated at this stage of development into a single structure with positive relief. It is suggested that no Ordovician and Silurian sedimentation took place within its limits. The territory of the southern block is characterized by accumulation of Ordovician and Silurian sedimentary sequences. Since Devonian time and especially during the Permian and Mesozoic, the entire territory of the southern block experienced constant slow uplift. To the east and west, the southern block is bounded by fault zones that follow the boundaries of the Tunguska and Viluy syneclises. Within these syneclises, thick accumulations of Permian and Mesozoic sedimentary cover occur (Figure 13.20).

Both the northern and southern blocks of this tectonopair are separated by east-west–striking, deep-seated fault zones, which stretch to the south of the Anabar shield. This system, as it has been described above, shows clear signs of strike-slip movement and includes the Kuranakhsky and Dalbykhsky kimberlite fields along or parallel to it.

It is necessary to stress not only that the boundaries between blocks within the tectonopairs control the location of kimberlite fields, but also that the greatest Siberian diamond deposits are located in the apexes of blocks composing the second (later and larger) tectonopair. Within the apex of the southern block of the second tectonopair lies the Malo-Batuobinsky kimberlite field and the famous Mir pipe. In the apex of the northern block of the same tectonopair lies the Popigay "astrobleme" and its great concentration of lonsdaleite.

The formation of the second tectonopair was completed during the course of three stages of tectonic reconstructions of the Siberian platform (Malich 1975). These stages occurred during the Late Ordovician and Silurian, Late Devonian to Early Carboniferous, and Late Carboniferous.

The major stages of restoration of isostatic equilibrium were accompanied by emplacement of kimberlite and carbonatite. Such pulses of kimberlitic emplacement during the Ordovician and Devonian were dated by Davies, Sobolev, and Kharkiv (1982). The data indicate the importance of horizontal tension in the distribution of kimberlite fields and that the scale of horizontal displacement exceeds the scale of vertical movement.

A major consequence of the hypothesis concerning the reestablishment of isostatic equilibrium is the role of mass exchange between blocks within each tectonopair. Such a mass exchange apparently occurs within the asthenosphere. If one compares different sets of data, it is possible to conclude that localization of kimberlitic provinces is controlled by two different structural factors: zones of strike-slip fault dislocations, and the specific structures on which they are superimposed.

Where strike-slip fault dislocations are superimposed onto stable massifs, kimberlitic provinces may occur. When the same types of dislocations are developed over orogenic zones or transverse fault zones, lamproites may occur. Both types of rocks are characteristically associated with compressional zones. The association of kimberlitic volcanism with restoration of isostatic equilibrium localizes kimberlitic fields along fissure zones of increased permeability within the platform's sedimentary cover, rather than within specific linear faults. The size of tectonopairs will define the amplitude of horizontal displacement. The influence of convection currents on tension generated in the crust has also been examined in terms of plate tectonics (Liu 1980).

SUGGESTED MODEL

Two key elements may be needed to localize ultramafic-alkalic, kimberlitic, and lamproitic magmatism: structures characterized by slow constant uplift (such as cratons, anteclises/swells, and median massifs) and possibly strike-slip fault zones.

It is suggested that the first element of structural control will determine the specific composition of volatile flow from the core/mantle boundary. The second will provide a source of increased pressure and create conditions for localization of specific magmatic

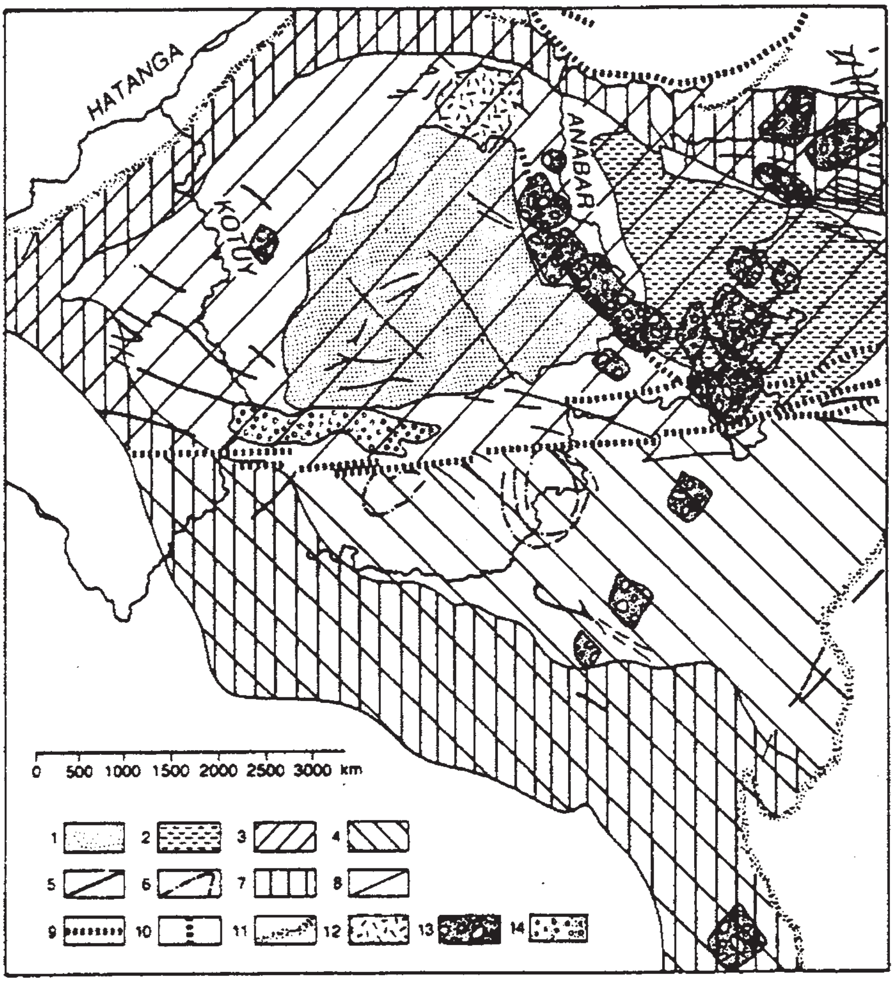

FIGURE 13.20 Tectonopairs in structures of the sedimentary cover within the northeastern part of the Siberian platform. Elements of lower Paleozoic tectonopairs: 1 = comparatively elevated blocks corresponding to the Anabar shield, 2 = comparatively subsided block corresponding to the area of Upper Cambrian sequence development, elements of upper Paleozoic tectonopairs: 3 = relatively elevated block, 4 = relatively subsided block, 5 = greatest doleritic dikes, 6 = axes of magnetic anomalies associated with doleritic intrusions, 7 = linear fractured zones saturated with doleritic intrusives, 8 = main geologically observed faults, 9 = suggested zones of deep-seated strike-slip faults, 10 = Anabar-Muna deep-seated fault zone, 11 = boundary of the area of stable Mesozoic sedimentation, 12 = Upper Cretaceous pyroclastic rocks, 13 = kimberlitic fields, 14 = superimposed depression filled by lose Quaternary deposits (Erlich 1985). Reprinted with permission of the Geological Society of South Africa.

bodies or groups of bodies. Redistribution of material in the asthenosphere will generate horizontal stress resulting in the formation of strike-slip faults that control the spatial distribution of kimberlitic fields.

Among elements of the first type, two groups of structures can be recognized: anteclises/swells within stable cratons, and zones of activation or orogenic belts adjacent to cratons or superimposed onto cratons and median massifs (microcontinents).

Two groups of structures correspond to two different types of diamondiferous magmatic rock associations. Kimberlites are preferably associated with anteclises, whereas lamproites are preferably associated with mobile zones of orogenic activation of stable cratonic blocks. The latter are developed on activated margins of cratons or median massifs.

Mitchell and Bergman (1991) noticed the tendency of kimberlites and lamproites to be associated with different types of structures. Lamproites were noted to lie within mobile belts surrounding the Kimberley block in Western Australia. However, many lamproites in the Rocky Mountains are superimposed on the Wyoming craton, and lamproites of the Chelima region are superimposed on the Indian craton. The Murcia-Almeria lamproitic province in Spain, and the recently discovered lamproites in Pamir and areas adjacent to the Urals Mountains, are associated with stable massifs within mobile belts.

Once the initial generation of alkalic magmas occurs during a "cratonization" event, continuing pulses of ultramafic-alkaline magmas occur during tectonic transformation of the ancient cratons and stable blocks. This is especially notable during the process of mass redistribution within the asthenosphere leading to the restoration of isostatic equilibrium.

In Chapter 14, Temporal Distribution of Diamond Deposits, we will show the relationships between the first occurrence of ultramafic-alkaline magmatism and "cratonization." Redistribution of material in the asthenosphere generates horizontal stress resulting in the formation of strike-slip faults, which may control the spatial distribution of kimberlitic fields. This process can explain the development of horizontal movements and horizontal strain within cratons and stable blocks, contrary to the widely accepted view that considers these structures to be dominated by vertical movement. A model of the initial phase of deep-seated fault development was presented by Voronov and Erlich (1962) (Figure 13.21).

The two different types of dynamic conditions associated with aulacogens are reflected in development of two different types of igneous rock associations: ultramafic-alkaline rocks and carbonatites, and kimberlites and carbonatites.

The dualism in geodynamic conditions in aulacogens described above explains the preferred association of lamproites with these structures in contrast to kimberlites, which are preferably localized within stable blocks (Mitchell and Bergman 1991).

The first type of igneous rock association is developed along the axis of an aulacogen and is associated with strain generated by normal faults located along the aulacogen's boundaries or within its axial zone. The second is confined to strike-slip faults associated with the development of the aulacogen.

The data presented show that the scale of horizontal movement along these faults is on the order of ten times as great as the scale of vertical movement. This difference results in considerable tension within platform's body, leading to the creation of zones of increased fracture permeability. The data suggest that horizontal movement is the main driving force and determines the strain within the platform.

The predominance of horizontal strain at depth, within the zone of kimberlitic magma genesis, is confirmed by observation of textures in xenoliths and inclusions in kimberlite. Deformation is observed in the fabric of some peridotite xenoliths from the

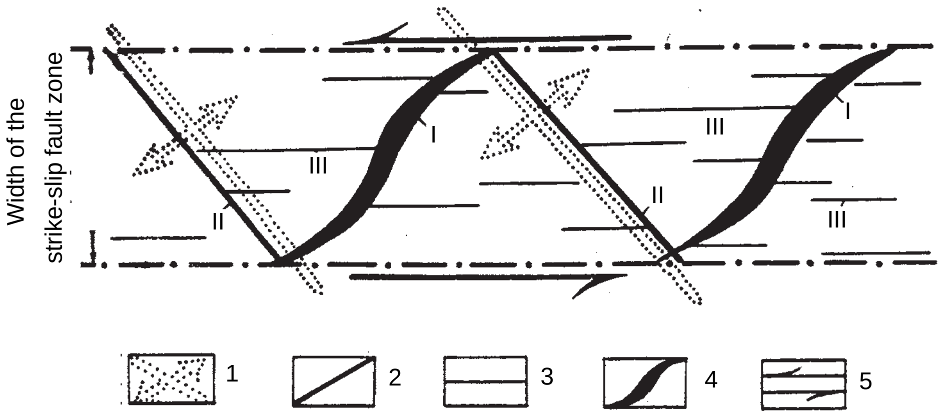

FIGURE 13.21 Schematic structure of the left-lateral strike-slip fault within a platform's cover. 1 = folds in rocks of the platform's cover over strike-slip fault zone in the platform's basement. Fractures generated by strike-slip fault: 2 = second generation, 3 = third generation, 4 = first generation, 5 = directions of displacement or tangential tension along main strike-slip fault zone in the basement. Numbers I, II, and III correspond with different generation of fissures. (Voronov and Erlich 1962, modified.)

Thaba Putsoa pipe in Lesotho, which is considered as evidence of strong horizontal movement within the mantle (Nixon and Boyd 1973). The movement has been suggested to be related to the breakup of Gondwanaland.

Dawson (1980) rejected this explanation and stressed that the range of radiometric dates (80–150 Ma) obtained for South African kimberlites exceeds the suggested time of opening of South Atlantic (130 million years). He also described other textural fabrics seen in kimberlitic nodules that support the existence of horizontal movement (or at least stress) in the upper mantle. In particular, the fabric within the deformed xenoliths is considered as proof of deformation within an anisotropic stress field, and it is interpreted as the result of superplastic flow within the upper mantle. Some of the evidence includes xenoliths with strong preferred orientation of olivine and orthopyroxene (tabular granoblastic texture), and rare coarse-grained lherzolites. These rocks contain large, deformed megacrysts that display kink-band and undulose extinction, the associated olivine having regrown from a presumed original porphyroclastic fabric.

Brittle fractures have also been observed in banded eclogite xenoliths that are oriented at a high angle to the banding within the nodule. This orientation may be attributed to a tightly spaced joint system that was imposed on the eclogite prior to its incorporation within the kimberlite (Dawson 1980).

Later phases of petrologic evolution and horizontal stress resulted in the separation and consequent injection of highly mobile fluid composed of megacrysts of olivine, garnet, pyroxene, and mica in a hot carbonatitic liquid from which olivine, spinel, perovskite, calcite, dolomite, ankerite, and quartz crystallized (Dawson and Hawthorne 1973).

The difference in the structural position of magmatic bodies is clearly illustrated by the spatial separation of the Udjinsky and Kotuy-Maimecha ultramafic-alkaline provinces (ijolite-carbonatite and agpaitic nepheline syenites), and the Oleneksky, Kuonamsky, Dalbykhsky, and Kuranakhsky kimberlitic fields.

Attempts to explain the structural position of kimberlites by the development of single types of structures (grabens, ocean openings) are insufficient, owing to the omission of different types of structural control (especially for kimberlites located within stable blocks). There have been numerous examples that attempt to connect kimberlites in South America and Africa with the opening of the Atlantic Ocean (Williams and Williams 1977; Marsh 1973; Doig 1970). As mentioned above, Dawson (1980) indicated that this concept does not fit the situation in South Africa, owing to the difference in ages of oceanic opening and kimberlite formation.

The concept that kimberlitic volcanism is preferably associated with negative tectonic structures (grabens, syneclises, and basins filled with flood basalts, such as the Karroo and Tunguska syneclises) is partially based on coincidence of timing of grabens and the formation of kimberlites (Mal'kov 1976, Milanovsky and Mal'kov 1980). These works properly emphasize one side of very complex regularities while omitting others.

Some authors attempt to establish a connection between kimberlitic volcanism and subduction of adjacent plates. In particular, Sharp (1974) developed a model connecting South African kimberlites with subduction in the Cape fold belt. Careful analysis of the tectonic and petrologic data show that this hypothesis has no basis (Newton and Gurney 1975). The objections against the role of subduction in mobile belts used by Newton and Gurney for South Africa can also be applied to the Siberian platform.

The suggested model to a certain degree is similar to the model for distribution of South African kimberlites proposed by Pretorius (1973), who acknowledged a leading role for syneclises and anteclises in the South African craton in the spatial distribution of kimberlites. In our model, we specify the role of aulacogens in this process and, in addition, we acknowledge the equally important element of structural control, i.e., horizontal strain. We also assume that volatile flow is responsible for magma generation, which is initiated at the core/mantle boundary (Bailey 1982).

This flow of fluids generated extreme metamorphic processes, which can result in the formation of magmatic melts with compositions corresponding to mica peridotite. In this process, eclogitic diamonds can be encapsulated and transported to the surface. In the meantime new P-type diamonds can be crystallized from newly formed melt.

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CHAPTER 14

Temporal Distribution of Diamond Deposits

INTRODUCTION

The timing of kimberlitic and lamproitic magmatic activity is reviewed as is alkaline and ultramafic-alkaline magmatism, with an emphasis on the timing of kimberlitic and lamproitic volcanism.

All data from 2500 Ma to the present are summarized in Tables 14.1 to 14.4. Histograms of radiometric ages for the North American and Russian platforms are divided into several time spans (see Figure 14.1 for the Siberian platform, Figure 14.2 for the Russian platform, Figure 14.3 for the North American platform, Figure 14.4 for the South African platform, and Figure 14.5 for the Brazilian platform). The tables list the age of major episodes of flood basalt formation (see Rampino and Stothers 1988), diamond deposits in metamorphic complexes, diamond-bearing ring structures, and major epochs of tectonic reconstruction. The dates in the captions of the tables are rounded within the accuracy of dating. The sources of dates listed in Tables 14.1 to 14.4 are marked by asterisks in the reference list in the end of this section.

The data clearly reflect two groups of regularities: specific features of the timing of ultramafic-alkaline magmatism within different regions, and the temporal distribution common to all regions. The tables and figures show the ages of ultramafic-alkaline magmatism and its similar timing within different platform blocks is apparent.

In different regions, the earliest phases of ultramafic-alkaline volcanism occurred at different times (Table 14.1). Some of the earliest radiometric dates are characteristic of North America, and the oldest alkalic magmatic pulses occur within the shield (Ontario, Canada) at about 2700 Ma. This pulse is reflected by a group of dates for the syenite pyroxenite intrusion near Poobah Lake, a K-Ar isochron age of 2706 ± 46 Ma (Woolley 1987), and a 2686 Ma Rb-Sr isochron age for rocks of the Kaminak Lake massif (Currie 1976).

During Australia's oldest pulse of alkaline magmatism, felsic alkaline rocks within Yilgarn block were emplaced about 2400 Ma (Jaques et al. 1985). In the Kola Peninsula, on the northern portion of the Russian platform, some of the oldest dates for alkaline rocks were obtained from the western Keiv complex: 2100 ± 50 Ma and 2400 ± 50 Ma (dates for Kola and Karelia from Kogarko et al. 1995). Similar dates characterize the Kanozersky complex (2330–2365 Ma; Pb method on zircon). In the Eltozero complex, an Sm-Nd isochron produced an age of 2080 ± 180 Ma. A similar date was obtained by

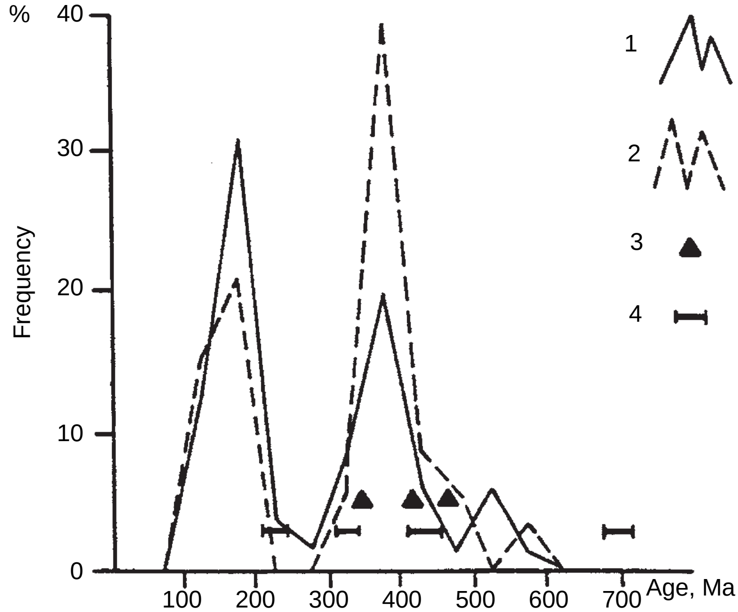


FIGURE 14.1 Line graph of radiometric dates of alkaline rocks of the Siberian platform. 1 = radiometric dates for the Lower Olenek group of fields (K-Ar, U-Pb and fission track data), 2 = radiometric data for various fields determined by fission track method, 3 = radiometric dates for the Malo-Batuobinsky and the Daldyn-Alakit regions (mainly K-Ar data), 4 = K-Ar data for minerals from Tomtor ultramafic-alkaline massif (Erlich 1985). Reprinted with permission of the Geological Society of South Africa.

the Pb method for the Ponoysky complex of ring and arcuate dikes, 2405 Ma. The age of these ancient alkalic complexes corresponds with the time of cratonization within the most ancient core blocks of the platforms. Stabilization of the entire platform followed later Proterozoic orogenies.

Within the Russian platform, a distinct peak characterizes the time interval between 1750 and 1900 Ma, and Proterozoic radiometric dates younger than 1750 Ma are rare. The same time is characteristic of phlogopite-pyroxenite calciphyres (calcite-rich metasomatic rocks) in the Aldan shield and some intrusions in North America and Australia. Radiometric dates available for alkaline rocks in northern Brazil are concentrated around 1806 ± 69, 1500 ± 38 Ma, and 1100 ± 57 Ma (Table 14.2; Ulbrich and Gomes 1981; Woolley 1987). In Greenland, India, South Africa, and North Australia the earliest dates are concentrated around 900 to 1200 Ma.

Table 14.3 shows that the youngest dates for alkalic rocks within Aldan shield yield 110 to 115 Ma. Similar dates (105–121 Ma) characterize the Barakar lamproitic field in India. In the Russian platform, the youngest dates for alkalic rocks are 162 to 163 Ma (Kogarko et al. 1995). In China, the youngest K-Ar dates for kimberlites are around 52 to 57 Ma, whereas in North America, the Leucite Hills lamproite field yielded dates of 1.1 to 3.1 Ma, and in Africa active alkalic volcanism persists to the present.

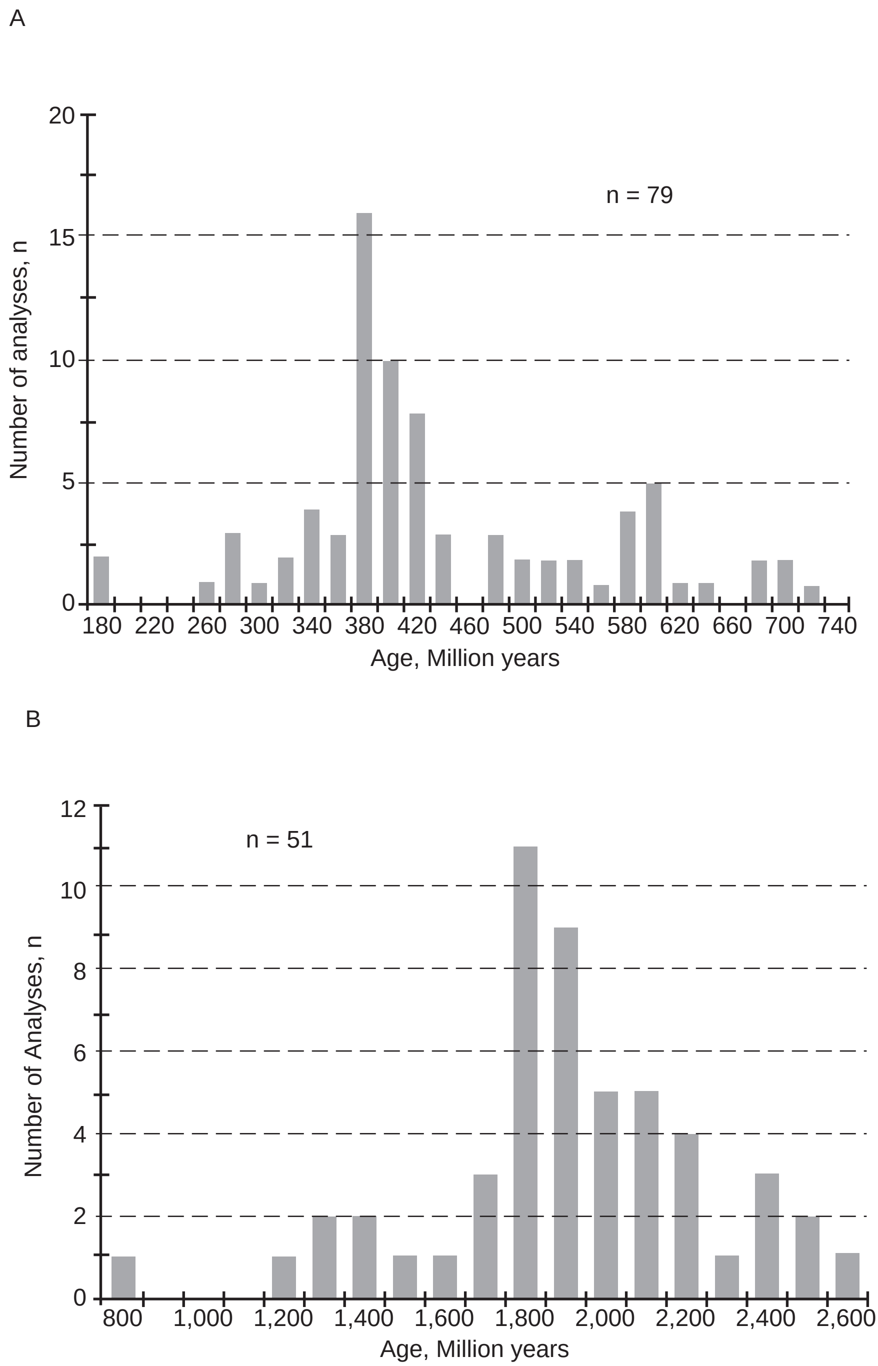
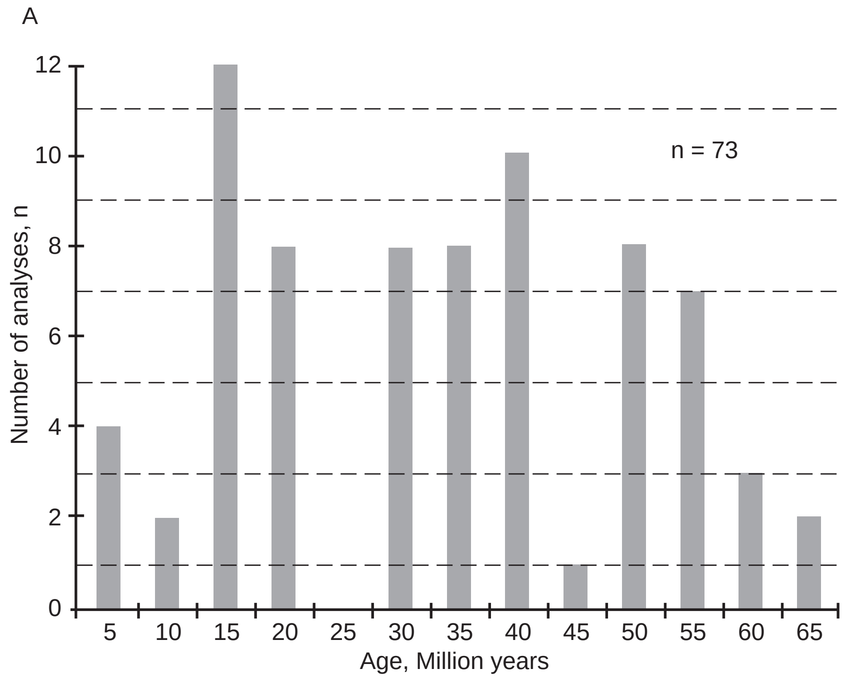


FIGURE 14.2 Histogram of radiometric ages of alkaline and ultramafic-alkaline rocks of Russian platform. A = Mesozoic and Paleozoic rocks, B = Proterozoic rocks (data from Kogarko et al. 1995).



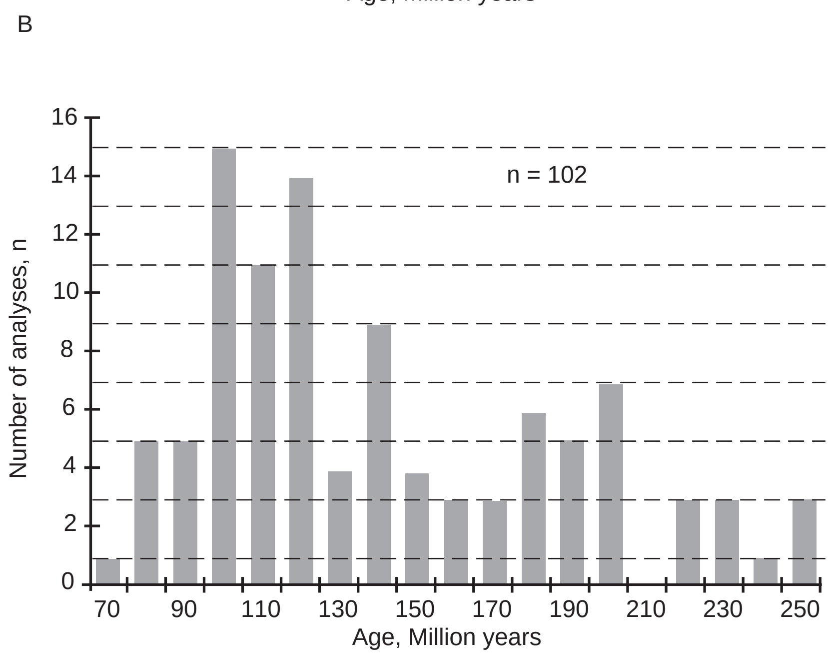
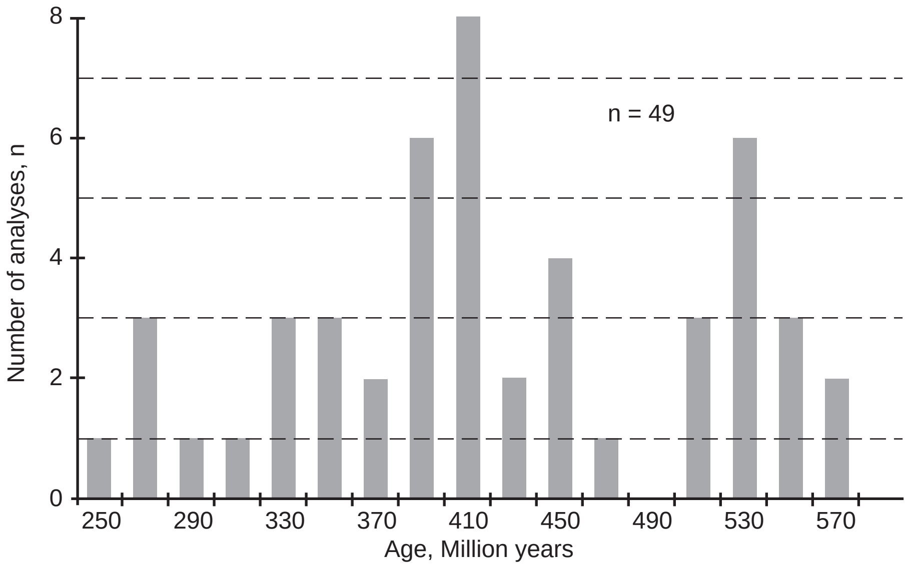


FIGURE 14.3 Histogram of radiometric ages of alkaline rocks in North America. A = post-Cretaceous, B = Mesozoic, C = Paleozoic, D = Proterozoic (data from Woolley 1987).

C



D

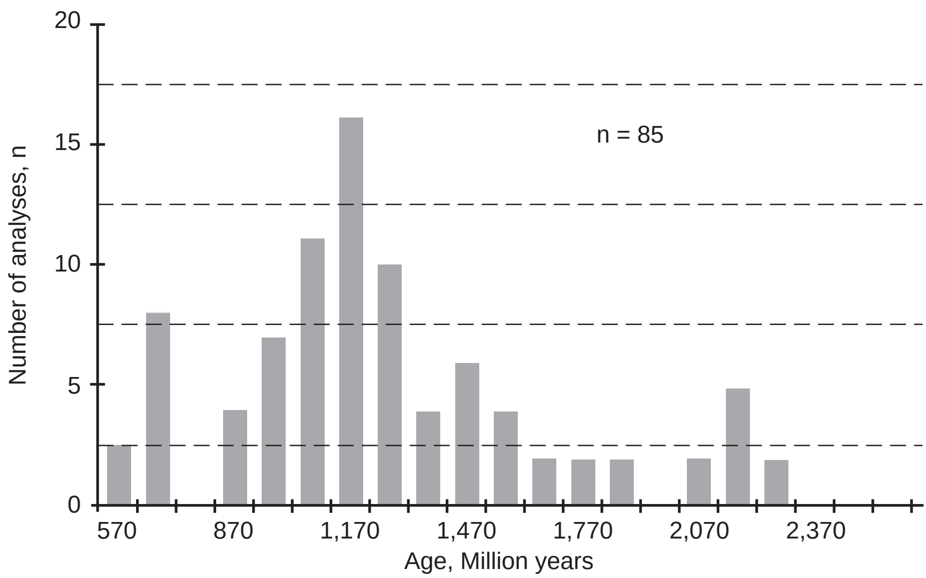


FIGURE 14.3 (Continued)

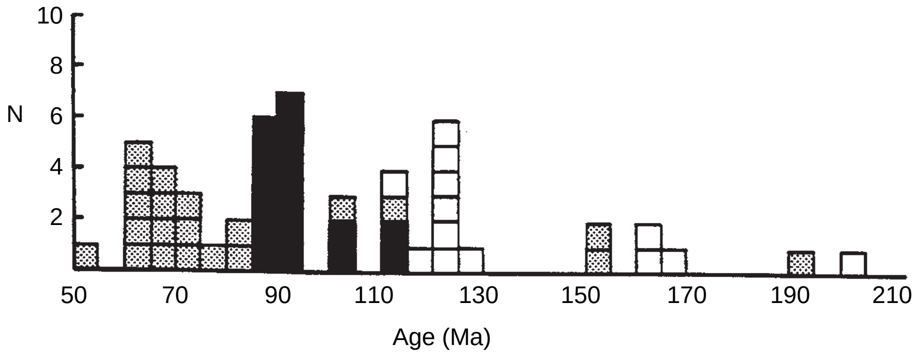


FIGURE 14.4 Histogram of radiometric dates of various types of kimberlites in southern Africa. Fifty-one samples (including one melilite basalt). Heavy stipple = Group I cratonic kimberlite, light stipple = off-craton kimberlites and melilite basalts, open symbols = group II kimberlites (Smith et al. 1985). Reprinted with permission of the Geological Society of South Africa.

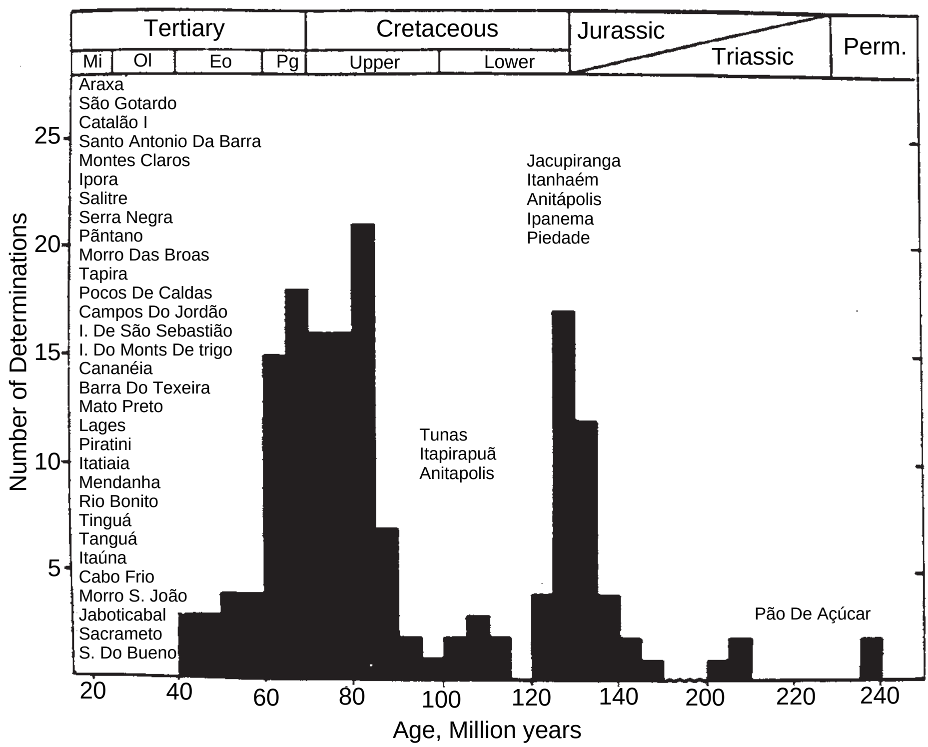


FIGURE 14.5 Histogram of radiometric dates of alkaline rocks in Brazil (modified from Ulbrich and Gomes 1981)

TABLE 14.1 Timing of ultramafic-alkaline volcanism and tectonic reconstructions (850–2500 Ma)*

Time, Ma	Regions or Provinces							
		Asia						
	Europe	Siberia	India	North America	Greenland	South America	Africa	Australia
850–1250	Kostomuksha dikes, Karelia 1120–1300 Ma	Ryphean aulacogens formed, Anabar anteclise	Majhgawan kimberlites, 840–1140 Ma and 1140±12 Ma Newania carbonatites, 959 Ma Jungel kimberlites, 974 Ma Chelima lamproites, 1140–1225 Ma Hinota kimberlites, 1170±46 Ma	*Grenville orogeny:* c. 880–1110 Ma Blue Mountain syenites, 900–1200 Ma Deloro stock, 1096±48 Ma and 1059±46 Ma Batchelor Lake, 1100 Ma NW Quebec dikes, 1100–1500 Ma	Sisimiut lamproites, 1206–1240 Ma Holsteinborg, 1227±12 Ma	Muri complex, Guyana, 1025±28 Ma. Alkaline rocks in northern Brazil, 1100±57Ma	Toubabako kimberlites, Guinea, 1145 Ma Kimberlites of National pipe, South Africa, 1180±30 Ma Premier pipe, South Africa 1200–1250 Ma	*Stabilization of Central Australia mobile belt*: 900–1000 Ma Argyle lamproites, 1104±111, and 1126±9 Ma Mordor alkaline complex, 1210±90 Ma
1250–1400		*Mongol–Okhotsly tectono–magmatic cycle* 1300±100 Ma	Chelima lamproites, 1319–1371 Ma	Blue Mtn. Ne-syenites, 1285±41 Ma				
1400–1750		*Cratonization within Anabar anteclise*		Mountain Pass formed c. 1500 Ma *Cratonization of Wyoming craton* c. 1750 Ma Argor and Goldray, central alkaline plutons, Ontario, 1655, 1695 Ma		Alkaline rocks in northern Brazil, 1500±38Ma	Bobi lamproites, 1410–1455 Ma	*Gawler province cratonized* 1470–1550 Ma *Capricorn orogen* 1500–1600 Ma Tennant Creek lamprophyres near Yilgarn block, 1660 Ma *North Australian craton stablized* 1700–1800 Ma

(*Table continues on next page*)

TABLE 14.1 Timing of ultramafic-alkaline volcanism and tectonic reconstructions (850–2500 Ma)* (continued)

Time, Ma	Regions or Provinces							
		Asia						
	Europe	Siberia	India	North America	Greenland	South America	Africa	Australia
1750–2500	*Cratonization as a result of Late Karelian folding* Kalmiussky, Yuzhno–Kalchiksky Oktyabrsky massifs, Ukrainian shield, 1680–1840, 1720–1980, 1730–1770 Ma Gremyakha–Vyrmes, Almunge, 1750–1950 Ma Central massifs in Karelia–Synnilyarvi, 1785–2030 Ma *Svecofenian regeneration*, 1950–2100 Ma El'tozero 2080±180 Ma Ponoyskii complex 2405 Ma Western Keiv 2100±50 Ma 2400±50 Ma	*Timptono–Stanovoy tectono–magmatic cycle* 1930±170 Ma 2040±90 Ma Pyroxenite–phlogopitic metasomatites within Aldan shield 1950–1800 Ma		*Elsonian stage of tectonic activity* Otto stock in Grenville–St. Lawrence province 1700–2150 Ma Castignon Lake carbonatites with kimberlitic afininty, 1873 Ma Easter Island dike, Blanchard Lake syenites, perlkaline granites 2575–2166 Ma Carbonatites of the Labrador Province 2398±72 Ma *Hudsonian orogeny Suggested cratonization in Ontario and Labrador*		Alkaline rocks in northern Brazil, 1806±89 Ma Kaituma, Guyana, 2065±100 Ma Makarapan, Guyana shield 2595±125 Ma		Phonolite dikes of the Pine Creek geosyncline, >1800–1700 Ma *Yilgarn block stabilized* 2500 Ma

*Sources of dates for Tables 14.1–14.4 are given in references marked with an asterisk at the end of this chapter.

TABLE 14.2 Timing of ultramafic-alkaline volcanism, flood basalt eruptions, and tectonic reconstructions (200–850 Ma)

Time, Ma	Regions or Provinces							
		Asia						
	Europe	Siberia	China	North America	South America	Africa	Australia	Antarctica
200–270	*Herzinian orogeny* Turiy Peninsula alkaline complex, Kola, 270 Ma	Flood basalts in Tunguska syneclise, 220–240 Ma* Alkaline central plutons, Kotuy–Maimecha Province, 220–245 Ma Last dates in Udja Province 250 Ma	High-pressure metamorphic complexes northeastern China, 212±2, 246±8 Ma 236±3.41 Ma†	Kentucky kimberlites, 257 Ma Ice River massif syenites and carbonatites, 244–280 Ma	Pao de Acucar, Brazil, Ne-syenite complex, 205±7–236±9 Ma	Kapamba lamproites, 220 Ma Mali carbonatites 262±7 Ma, 269±9 Ma	Western Victoria alkaline basalts, 201.5, 211.5 Ma Durrras alkaline basalts, 241±34 Ma Tasmania alkaline basalts, 211±3 Ma, 241±34 Ma	
270–400	Khibines and Lovozero plutons in Kola peninsula, 350–370 Ma	Kimberlites of the Sredne–Olenek region						
400–500	*Caledonian orogeny, late phase* Ultramafic–alkaline massifs in Kola Peninsula 400–500 Ma	Earliest stage of kimberlite formation, Anabar anteclise, 420 Ma Most ancent dates for Tomtor massif, Udja Province, 420 Ma		Pedernal Hill, New Mexico, carbonatites, 409 Ma				Mt. Bayliss lamproites, 413–430 Ma Priestley Peak lamproites, 482±3 Ma

(*Table continues on next page*)

TABLE 14.2 Timing of ultramafic-alkaline volcanism, flood basalt eruptions, and tectonic reconstructions (200–850 Ma) (continued)

Time, Ma	Regions or Provinces							
		Asia						
	Europe	Siberia	China	North America	South America	Africa	Australia	Antarctica
500–600	Alkaline diabases in Dneprovsko–Donetsky aulacogen, 543–610 Ma Alkaline gabbro, 534, 546, 545 Ma	High-pressure complex in Kazakstan meta-morphics, 510–530 Ma‡		Carbonatites in Cordillera: Iron Hill, 550–570 Ma McClure Mtn., 506–535 Ma Dermocrat Creek, 511–534 Ma Gem Park, 551 Ma Lobo, New Mexico syenite, 604 Ma	Alkaline rocks in Goias, eastern Brazil, 469±14, 533±21 Ma	*Panafrican orogenic event* Colossus pipe, Zimbabwe, 502±47 Ma		
600–850	Central alkaline plutons, Oslo graben, 665, 675 Ma Timan carbonatites, 680, 695, 652 Ma Kovdor, Kola Peninsula, 706 Ma	*Early phases of Caledonian orogeny* Ingili kimberlite region, Aldan 609–865 Ma		Mt. Grace carbonatite, British Columbia, 773 Ma	Southern Brazil sodalite syenites, tawaites, 600–665 Ma Alkalic rocks Goyas, Brazil, 765±40 Ma		*Stabilization of Tasmanian Fold Belt* c. 700 Ma Mud Task carbonatites, 732±5, 735±75, Ma	

* Data for flood basalt episodes in Tables 14.1–14.4 (Rampino and Stothers 1988).
† Data for high-pressure mineral complex in China (Perchuk, Yapaskurt, and Okay 1995).
‡ Data for high-pressure mineral complex in Kazakhstan (Dobrezhinetskaya et al. 1994).

TABLE 14.3 Timing of ultramafic-alkaline volcanism, flood basalt eruptions, and tectonic events (65–200 Ma)

Time, Ma	Regions or Provinces							
		Asia						
	Europe	Siberia	India and China	North America	South America	Africa	Australia	Antarctica
65–80			Kimberlites in northern China 77–80 Ma*	Laramide tectonic event	Rio de Janeiro volcanic region 48–70.5 Ma			
80–100		Central ultramafic plutons in Aldan shield, 85–115 Ma	Kimberlites in northerrn China 52.5–117 Ma	Ouachita province lamproites, 98–105 Ma	Coromandel province lamproites, 80–87 Ma	Most of South African kimberlites, 80–120 Ma	Mt. Woolome minette, 85±3 Ma Sydney basin monchiquites 101±4 Ma	
100–120		Murun lamproites, 115–143 Ma	Barakar lamproites, 105–121 Ma	Alkaline massifs of the Monteregian province, 111±6, 117±9, 120±8 Ma	Hiatus in alkaline magmatism in Brazil			
120–140		*Rifting within Aldan shield* associated with formation of central alkaline plutons, kimberlites within Anabar anteclise, ≥ 138 Ma			Serra Geral flood basalts, 130±5 Ma Alkaline rocks of Jacupiranga node, c. 130 Ma Sacuai laccolith 129–144 Ma	*South Arlantic ocean open* Karroo flood basalts, 135±5 Ma Koidu carbonatites in Sierra Leone, 92–140 Ma	Delegate breccia pipe, 138±3 Ma	

(*Table continues on next page*)

TABLE 14.3 Timing of ultramafic-alkaline volcanism, flood basalt eruptions, and tectonic events (65–200 Ma) (continued)

Time, Ma	Regions or Provinces							
		Asia						
	Europe	Siberia	India and China	North America	South America	Africa	Australia	Antarctica
140–160	Monchiquite dikes within Ukrainian shield, 162–163 Ma	Kimberlites within Anabar anteclise, ≥ 155 Ma				Pniel, Postmasburg, Swartruggens lamproites, 142–156 Ma East Griqualand kimberlites, 150–190 Ma	Wandegee picrite basalts, c.160 Ma Jingare alkaline syenites, complex, 168±2.5 Ma	
160–180							Terrowie kimberlites, 164–174 Ma Orroroo kimberlites, 170–172 Ma	Antarctic flood basalts, 170±5 Ma
180–200		Kimberlites within Anabar anteclise ≥ 183 Ma		Eastern North America flood basalts 200±5 Ma	End of formation of Pao de Acucar, Brazil c. 200 Ma	Southwest Africa flood basalts. 190±5 Ma	Western Victoria alkaline basalts, 193±10, 201.5, 211.5 Ma	Antarctic flood basalts, 190±5 Ma

* Data for kimberlites of northern China (Zhang and Liu 1983).

TABLE 14.4 Timing of ultramafic-alkaline volcanism, flood basalts, tectonic reconstructions, and formation of astroblemes (0–65 Ma)

Time,	Regions or Provinces						
Ma	Europe	Asia	North America	South America	Africa	Australia	Antarctica
0–4	Roman potassium province, 0–2 Ma Ultramafic volcanics and alkaline basalts in Germany, 0–4 Ma		Leucite Hill lamproites, 1.1–3.1 Ma Ultrapotassic and basanitic rocks, Sierra Nevada, 3–4 Ma	Alkaline volcanics on islands off Brazilian coast, 0–2.5 Ma	Active carbonatitic and alkaline volcanism, East African rift zone, 0–4 Ma		Gaussberg lamproites 0.056±0.005 Ma
4–12	Alkaline basalts, France, 4–8 Ma Murcia–Almeria lamproite province, 5.7–10.8 Ma Campus de Calatrava lamproites, 6.4–8.6 Ma					Dispersed leucitites, New South Wales, 6–16 Ma	
12–16	*Orogeny in Alps* Sisco lamproites, 13.5–15.4 Ma Ries crater formed, 14 Ma†	Alkaline diatremes in Pamir 14–16 Ma*					
16–20			Columbia plateau basalts, 17±2 Ma				
20–24	Orogeny in Alps Ultramafic volcanics and alkaline basalts in Germany. 22–24 Ma	*Orogeny in Himalaya* Lamproites in Pamir, 24 Ma†	Kimberlites around Colorado plateau, 20–22 Ma	Alkaline volcanics, island Fernando di Noroña, 22 Ma	Beni Bousera emplaced 23 Ma‡ Postcaldera volcanism in Kisingiri complex, 22.5±0.4 Ma	West Kimberley lamproites, 17–24 Ma	

(*Table continues on next page*)

TABLE 14.4 Timing of ultramafic-alkaline volcanism, flood basalts, tectonic reconstructions, and formation of astroblemes (0–65 Ma) (continued)

Time,	Regions or Provinces						
Ma	Europe	Asia	North America	South America	Africa	Australia	Antarctica
24–28							
28–32	Sessia–Lanzo lamproites, 29–33 Ma Ultramafic volcanics and alkaline basalts in Germany 22–30 Ma		Kimberlites around Colorado plateau, 28–32 Ma				
32–36	Beginning of ultramafic volcanism in Germany c. 36 Ma	Popigay astrobleme formed, 35 Ma[++]	Kimberlites around Colorado plateau, 34–38 Ma		Postcaldera carbonatites, Kisingiri, Kenya rift zone, 33±5 Ma Ethiopian plateau basalts, 35±2 Ma		
36–60				Rio de Janeiro alkaline complex, 41.7–59 Ma	Beginning of volcanic activity at Kisingiri, East Kenya rift zone, 38±5.0 Ma		
60–65	Brito-Arctic plateau basalts, 62±2 Ma	Deccan flood basalts, 66±2 Ma		Rio de Janeiro alkaline complex, Brazil 59–70.5 Ma			

* Data for Pamir (Dmitriyev 1974).

† Data for Popigay and Ries ring structures (astroblemes) (Vishnevsky et al. 1997).

‡ Data for the Beni Bousera massif (Pearson, Davies, and Nixon 1993).

PULSATIVE NATURE AND SYNCHRONEITY OF ULTRAMAFIC-ALKALINE VOLCANISM

The pulsative nature of ultramafic-alkaline volcanism is well shown by radiometric dates for different regions (Davies, Sobolev, and Kharkiv 1980; Ulbrich and Gomes 1981; Smith et al. 1985); see the histograms of radiometric ages (Figures 14.1–14.5). The existing radiometric data tend to concentrate at certain time intervals. Those intervals are in turn divided by much longer periods almost completely lacking radiometric dates. For instance, radiometric dates between the time spans such as 40 to 80 Ma are noticeably lacking and radiometric dates between 160 to 180 Ma and 270 to 400 Ma are rare.

In other words, global synchronous episodes of intense ultramafic-alkalic magmatic activity are separated by much longer hiatuses in activity. The concept of global synchroneity of major pulses of ultramafic-alkalic magmatism has been previously developed (Mal'kov 1976; Erlich et al. 1989), and the synchroneity of kimberlitic and lamproitic magmatism for the Proterozoic (around 1100–1200 Ma) has been established by Skinner et al. (1985).

Mitchell and Bergman (1991) interpreted the synchroneity in the Proterozoic in contrast to abundant kimberlite magmatism in the Paleozoic and Mesozoic, which is not contemporaneous with lamproite formation.

Data presented in Tables 14.2 to 14.4 show that such episodes existed around 700 to 720 Ma, 600 to 660 Ma, 400 to 440 Ma, 220 to 240 Ma, 100 to 120 Ma, 14 to 16 Ma, and 0 to 4 Ma. These data suggest a tendency for ultramafic-alkalic massifs to be emplaced about every 100 million years. Thus it is possible that radiometric dates of ultramafic-alkalic rocks tend to concentrate at certain time intervals.

Active ultramafic alkaline volcanism in rift zones of Africa, radiometric dates of lamproites in Antarctica (Gaussberg volcano) and North America (Leucite Hills), and the recent Roman province of ultrapotassic volcanics indicate that we currently live in a period of more intense ultramafic-alkaline volcanism.

During the Paleozoic (see Table 14.2) in Antarctica, ultramafic-alkaline volcanism occurred around 413 to 482 Ma. In Africa, activity is dated at 500 Ma. In South America such volcanism occurred around 600 Ma. In China, the earliest Paleozoic dates are concentrated around 200 Ma. Such volcanism was absent during the Paleozoic in India.

During the Mesozoic (see Table 14.3) there was almost no activity during 65 to 80 Ma, which corresponds with the age of the Laramide orogeny in North America. Exceptions, however, include alkalic volcanic activity in the Rio de Janeiro region, Brazil, and eruption of the Namaqualand, Bushmanland, and Gibeon pipes in South Africa.

Although there are many radiometric dates during 100 to 120 Ma for Africa, India, and the Aldan region of Siberia, this same time interval is characterized by the Barremian hiatus in Brazil (see radiometric data histogram, Figure 14.5). In order to establish an exact date of the Barremian hiatus we constructed another histogram of radiometric dates for alkaline rocks on the basis of data in Woolley (1987). Nevertheless, this time span includes the partial formation of the Sacuai laccolith in Brazil (110–120 Ma) and a considerable number of kimberlites in South Africa.

Radiometric dates for ultramafic-alkalic rocks are completely absent during the time frames of 16 to 20 Ma, 20 to 24 Ma, and 36 to 60 Ma, and they are scarce within 4 to 12 Ma and 28 to 32 Ma (Table 14.4). Radiometric dates for alkaline rocks in North America during the interval 21 to 24 Ma (Figure 14.3A and Table 14.4) are also absent. The same interval is characterized by an abundance of dates from ultramafic-alkaline volcanics in Germany. In Pamir, a 23 Ma age has been obtained for lamproites (Dmitriyev 1974).

Duration of Magma Storage

An important feature of ultramafic-alkalic magmas is their ability to be stored for very long periods of time at depth. Thus synchronous intensification of alkalic and ultramafic and alkalic volcanism is combined with the diachronic character of hiatuses of such types of magmatic activity within different regions of the world.

Kimberlitic and lamproitic magmatism in most provinces shows a tendency to repeat after about 1 billion years. In West Australia, Miocene lamproites erupted after an earlier pulse in the Late Proterozoic. Similar repetitions are recognized in the Aldan region, Siberian platform, where a Proterozoic pulse was followed by a Jurassic pulse. In South Africa, a Proterozoic pulse of kimberlite eruption was repeated during the last 80 to 120 million years, and the same time interval between two stages of activity has been observed in India and West Africa. Along the edge of the Wyoming craton, a Late Proterozoic kimberlite pulse was followed by a mid-Paleozoic pulse.

Confirmation of this 1-billion-year repeat interval is provided by the well-known discrepancy between the radiometric ages of host kimberlites and ultramafic nodules (cognate inclusions). For example, Rb-Sr ages of unaltered peridotite nodules in Paleozoic kimberlite (the Udachnaya pipe, Siberian platform) is 1203 Ma. The Rb-Sr age of peridotite nodules in Mesozoic kimberlite (Obnazhennaya pipe, Siberian platform) by mineral isochron data is 370 Ma. The average Rb-Sr parameters for xenoliths from different pipes within the northeastern Siberian platform form a trend with a positive correlation coefficient of 0.97. If we consider this dependence as an isochron, it is possible to approximate an average age of deep-seated material beneath the kimberlites as 1062 Ma (Zaitsev et al. 1983). Serpentinite nodules from Yakutian kimberlites have Rb-Sr ages of 741 Ma, but the Rb-Sr age of the host kimberlite is 360 Ma. U-Pb ages of two nodules in kimberlite from the Chadobetsky uplift (southwest Siberian platform) are 1600 ± 150 Ma. However, one zircon in the Bol'shaya pipe in the same region was dated 810 ± 150 Ma. The first date coincides with the time of a platform's stabilization. The second is close to the age of kimberlite in the northeastern part of the Siberian platform (1062 Ma), which (within the accuracy of dating) coincides with the time of closing of the Ryphean aulacogens. The K-Ar ages for phlogopite nodules in kimberlites located within the northeastern Siberian platform are close to these dates or coincide with them.

All of these facts suggest that after ultramafic-alkaline magmas are generated, they are stored at depth for an extremely long time. This inference is confirmed by the radiometric data obtained from diamonds (see Chapter 3, Age and Origin of Diamonds: An Overview). Because of the great depth of kimberlitic and lamproitic magma generation (more than 100 km), storage of magma for long periods is quite possible. It has been established that the Nd and Sm concentrations in inclusions in Argyle diamonds are on the average five times as great as those in their Premier counterparts, whereas Sm/Nd ratios are lower by a factor of 1.4. The carbon isotopic composition of diamonds at Argyle shows clear signs of dilution (Richardson 1986). Both isotopic features are considered to be the result of a lengthy storage time in magma within the lithosphere.

However, the data also indicate that the age of eclogitic diamond inclusions from the Premier pipe is similar to the age of emplacement of the kimberlite. Typically, diamonds that contain ecologite inclusions do not have a unique age or precursor but appear to be related in time to kimberlite or lamproite magmatism. These latter-day diamonds of eclogitic paragenesis are distinct from the dominant peridotitic diamonds formed about 2 billion years ago in the residual but highly enriched Archean lithosphere.

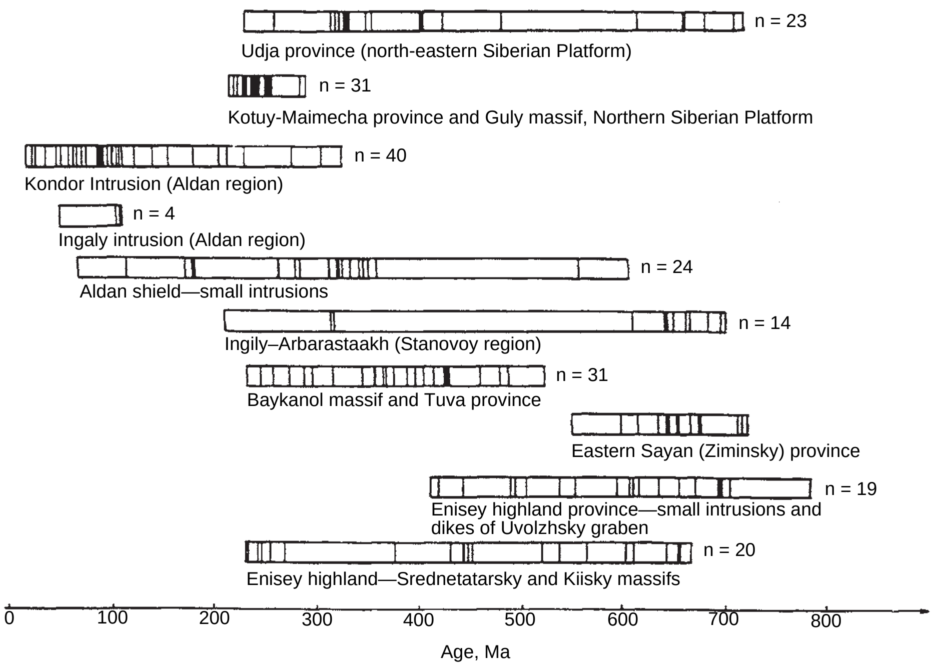


FIGURE 14.6 Range of radiometric dates for different provinces of ultramafic-alkaline and alkaline rocks within the Siberian platform (Erlich et al. 1989)

Duration of Magma Evolution

Two types of massifs exist, characterized by different ranges of radiometric dates. One type has a comparatively short duration, 5 to 20 m.y. Examples include the Khibines, Lovozero (Kola Peninsula), and Guly (Maimecha-Kotuy province, Siberian platform) massifs. Within this time interval, these largest massifs in the world evolved from ultramafic-alkalic to carbonatite-ijolitic (Guly) and from ultramafic-alkalic to peralkalic (Khibines, Lovozero). A similar massif is found in the Kruger Mountains of the Canadian Cordillera. Data for some regions of North America indicate several overlapping periods of generation of ultramafic-alkalic and alkalic magmas. The shorter time of formation for the alkalic complexes in Urals in comparison with the cratons is probably associated with the more stable tectonic conditions characteristic of cratons.

Radiometric data for the second type of massifs are characterized by dates of about 200 to 400 Ma. Examples of this type of massifs include the Kovdor massif, Kola Peninsula; Tomtor massif, Siberian platform; the syenitic massif on the Chichagoff Island, Alaska; the Quincy granites and the White Mountain rock series in the Appalachian province (Figures 14.6–14.8). In each figure, each row of points reflects different massifs or provinces that can be distinguished on the basis of two types of points: points in the beginning of each row, which reflect the time of magma generation, and all other points, which reflect the time of emplacement of next portion of magma.

It is assumed that the earliest dates in each series reflect the time of magma generation in the upper mantle beneath the ancient platform. Thereafter these magmas persist

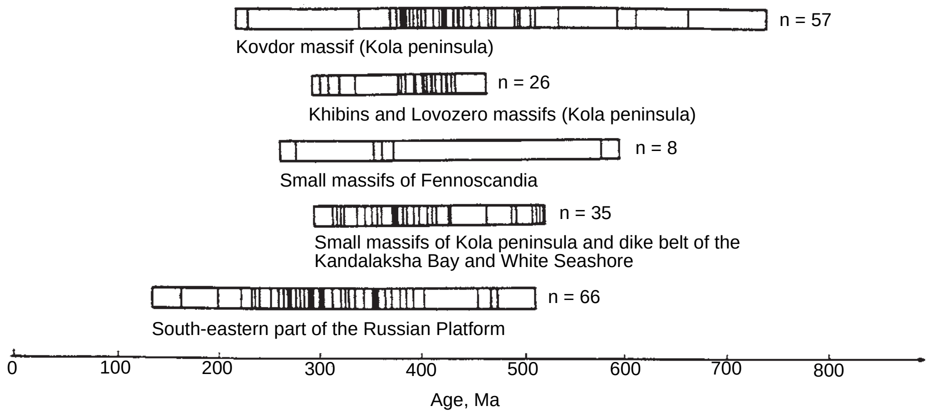


FIGURE 14.7 Range of radiometric dates for different provinces of ultramafic-alkaline and alkaline rocks within the Russian platform (Erlich et al. 1989)

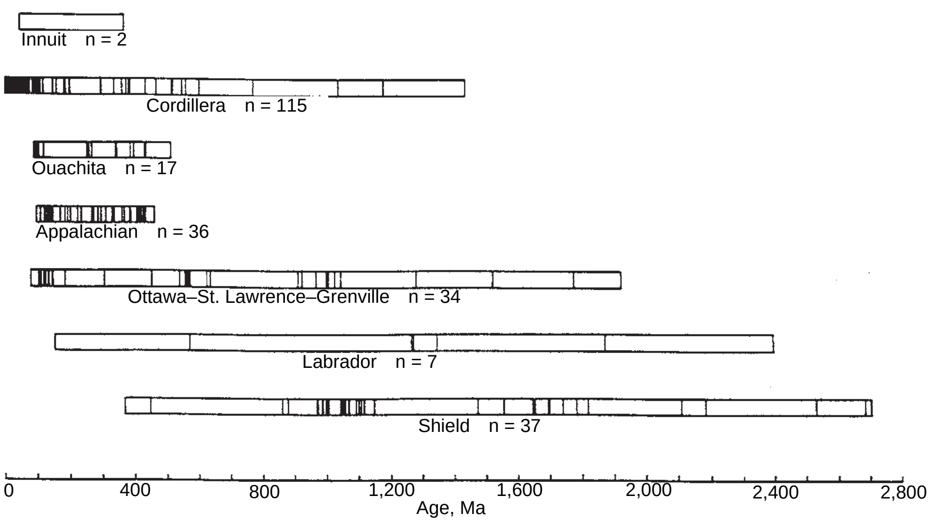


FIGURE 14.8 Range of radiometric dates for different provinces of ultramafic-alkaline and alkaline rocks within the North American platform (Erlich et al. 1989)

at depth, and each successive tectonic event is followed by a phase of magmatic activity of related composition (Kumarapeli 1970; Erlich 1985).

The size of the massifs is not related to the duration of magmatic activity. For example, the largest massifs in the world, the Khibines, Lovozero, and Guli, are relatively short lived, whereas the small Kovdor massif (Kola Peninsula) and the enormous Tomtor massif (Udja province, Siberian platform) both belong to the group with the greatest range of radiometric dates. But the range in radiometric dates for each province, as a whole, is the same in all cases: several hundred million years. Moreover, for the most part, the durations are on the order of 200 to about 400 m.y. It is important to emphasize that these durations coincide with the total range of radiometric dates that characterize different alkalic provinces within the North American, East European, and the Siberian platforms. Such consistent intervals of about 400 m.y. probably reflect the length of time required for heat exhaustion in the roots of the magmatic systems created in the asthenosphere under certain alkaline magmatic provinces and specific massifs. This regularity has been established for alkalic and ultramafic-alkalic massifs of three platforms: the Russian, the North American, and the Siberian (Erlich et al. 1989). The same can be seen in massifs within the Brazilian platform (Ulbrich and Gomes 1981).

The range of radiometric ages in hundreds of millions of years is not an indication of the time needed to produce certain massifs or intrusions, but is rather an indication of the time necessary for complete exhaustion of heat reserves within the roots of the asthenoliths within a given province.

Evaluation of the duration of volcanic pulses depends on the accuracy of radiometric dating. The dispersion of dates obtained in the analytical processes will differ depending on the episode under consideration. For example, the scatter in radiometric ages for the Upper Proterozoic can reach 100 to 200 m.y., whereas for the Upper Mesozoic may be only 20 m.y. This uncertainty strongly affects the evaluation of the duration of cycles of kimberlitic activity provided by different authors for different regions and sometimes within the same kimberlite province. For example, some authors suggest that most kimberlites in South Africa were formed within a single stage of 60 to 120 m.y. But a comparison of ages of different Mesozoic kimberlites suggests at least two episodes of kimberlite activity: one between 60 to 80 Ma and the other between 100 to 120 Ma. This suggestion agrees with radiometric dates of South African kimberlites summarized by Smith et al. (1985) and with dates of kimberlitic and lamproitic provinces of Cenozoic age which, as will be shown below, agree within 20 million years.

The clearest example of the length of time required to form kimberlite-related systems is provided by the 80 to 90 Ma Kimberley field, South Africa. This field has a characteristic linear trend, which is the result of eruption of diatremes above a system of northwest-trending dikes; examples of these dikes were found at depth in the Kimberley, Kemfersdam, and Bullfontein pipes. The Koffifontein cluster of pipes is similarly aligned and may be linked at depth by a dike (Wagner 1914). The feeder dikes cut across the east-west foliation of the basement gneisses (Dawson 1980) and lie parallel to a major system of northwest-southeast fractures (Pretorius 1973). The Koffifontein, Eberhaeser, and Klipfontein pipes are linked at depth to a northwest-southeast–trending dike. The radiometric ages cited above indicate that this field required about 10 m.y. to form.

Another way to determine the length of time needed to form a magmatic system is provided by evaluation of radiometric ages of a series of small ultramafic-alkalic plutons located within a single feeding system, such as a graben that controlled the location of small massifs. Two examples can be considered:

- In the region of the Oslo graben, recent radiometric studies have shown that alkalic rocks can be separated into two groups on the basis of age: one of about 600 Ma whose oldest dates are 665 and 675 Ma, and a second at 350 to 300 Ma. Some volcanic activity also occurred in the region about 400 to 420 Ma (Verschure et al. 1983).
- As shown, alkalic volcanism within the Monteregian region produced a series of small alkalic massifs that are distributed along the St. Lawrence graben. K-Ar ages range from 95 to 120 Ma. They lie within the range of the South African episode of kimberlitic activity and have a total duration of 25 million years.

It seems that both of these grabens control the distribution of alkalic and ultramafic-alkalic rocks and play the role of a single feeder structure for a series of intrusives. Again, the range of radiometric dates reflects the length of time required to exhaust the heat in the roots of the systems.

The same situation is represented by small plutons within the Kola Peninsula that are located along three linear zones and are characterized by two groups of radiometric ages. The first group is of Caledonian age (early Paleozoic) and contains essentially ultramafic-alkalic rocks. The second group is of Hercynian age (late Paleozoic) and contains mainly peralkalic and agpaitic intrusive complexes. Among the latter are the two largest in the world, the Khibines and Lovozero plutons. Detailed isotopic studies (Kononova 1976) show that radiometric dates even for a single massif (the Kovdor massif) embrace a time interval of about 400 m.y.

Further detailed isotopic studies show that it is impossible to separate the intrusives into two groups on the basis of their age and composition. Instead, one must discuss the general tendency of a single intrusive complex to evolve from ultramafic affinity during early stages of magmatism to peralkalic and agpaitic affinity during its later stages. This conclusion is confirmed by Rb-Sr isotopic studies for the Lovozero pluton and a series of small massifs in the Kola Peninsula, which show that the ultramafic-alkalic and alkalic rocks belong to the same episode of emplacement and have a single source (Kogarko, Krumm, and Grauert 1983; Krumm and Kogarko 1984).

The summary of Rb-Sr and Ar-Ar ages of different types of Jurassic and Cretaceous kimberlites and related rocks in South Africa provides some information about the duration of chemical evolution within the asthenosphere. Group I kimberlites tend to be younger than 114 Ma, whereas Group II kimberlites tend to be older (Smith et al. 1985) (Figure 14.4). U-Pb dates of kimberlites within different fields of the Siberian platform show the existence of several intrusive pulses within a single field (Davies, Sobolev, and Kharkiv 1980).

TIMING OF MAGMA EMPLACEMENT AND RELATED PROCESSES

Petrologic Aspects

It is not our task here to discuss in detail the petrologic problems related to kimberlitic and lamproitic volcanism and other types of volcanic rocks. Nevertheless it is necessary to emphasize some aspects of time relations between kimberlitic and lamproitic magmatism and some other rock series.

Many diamondiferous provinces are located near great flood basalt fields (e.g., South Africa, India, Brazil, and Siberia). These close spatial relationships provide a ground for forecasting diamond deposits in Siberia. A review in the previous part of this chapter showed that episodes of kimberlitic and lamproitic magmatism are closely related to (and within accuracy of dating even coincide with) the greatest episodes of flood basalts. Nevertheless, within at least three platforms—Eastern European, Australian, and North American—flood basalt eruptions are not related to diamond deposits.

It is necessary to emphasize the difference in flood basalt eruptions and kimberlitic-lamproitic volcanism. First, they differ greatly in the volume of erupted material. If flood basalt episodes erupted hundreds and thousands of cubic kilometers of material, each episode of kimberlitic and lamproitic volcanism produced no more than a single cubic kilometer of material (Mitchell and Bergman 1991). Second, although flood basalts and alkalic magmatic rocks tend to be emplaced during a single episode, alkalic magmatism occurred almost everywhere during the final stages of flood basalt eruption.

An even greater difference between these two types of volcanic events can be seen if we compare the spatial distribution of magmatic activity in the course of both types of magmatic episodes. On the basis of the volume of erupted material and the intensity of volcanism, episodes of flood basalt eruptions are considered as global volcanic events, even though each episode of flood basalt eruption occurs only within a single structure. No significant volcanic activity of the same type occurred at the same time worldwide. This localization is also true for Siberian flood basalts, Karroo basalts in South Africa, Parana basalts in Brazil, and the Columbia River basalts.

In contrast, the timing of kimberlitic and lamproitic volcanic activity shows a strong tendency to synchroneity worldwide. Kimberlitic and lamproitic magmatism can be considered as a dissipated form of internal energy, in contrast to flood basalt eruptions, which are concentrated within a single region and time interval and which represent a type of hotspot.

The timing of kimberlite/lamproite magmatism and of carbonatitic magmatism shows that they not only coincide in time, but that they are similar, and perhaps identical, in volume and character of spatial distribution.

The correlation coefficient between timing of kimberlitic and lamproitic and lamprophyric activity in Arizona, New Mexico, Colorado, and Wyoming during last 30 million years and the time of subalkalic silicic pyroclastics associated with caldera-forming events in the same region equals 0.88 and 0.90 of the age of mainly postcaldera silicic lavas (Erlich 1993). These coefficients are a numerical expression of what has been stated above: the timing of alkalic magmatism within stable blocks coincides with the timing of orogenic events in adjacent mobile belts.

If the concept of the major role of core-derived volatiles in magmatic activity is valid, it is possible to conclude that the flow of volatiles from the Earth's core endured chemical transformation within the upper horizons of the crust. Depending on tectonic regime, it resulted in two different types of magmatism: highly silica-undersaturated alkalic magmatism enriched in alkalis within stable blocks, and silicic-granitic magmatism and caldera-forming silicic pyroclastics within adjacent orogenic systems.

A comparison of timing of kimberlitic and lamproitic activities within different regions clearly shows that lamproitic intrusions were formed during the same pulses. For instance, the Proterozoic Argyle lamproitic pipe formed during the same pulse of kimberlitic activity that formed the Premier and National kimberlitic pipes in South Africa and kimberlites in Guinea, India, and Greenland (Skinner et al. 1985). The age of the Jharia lamproites in India (105–121 Ma) coincides with the age of a series of southern African kimberlites (Uintjiesberg, Finsch, Poortjie, and Bellsbank) that date from 99.5 ± 3.8 Ma to 118 to 121 Ma (Smith et al. 1985). Similar dates are reported for ultramafic-alkalic rocks from the Aldan shield. The Arkansas lamproitic province in the United States (110 Ma) also corresponds with the same pulse.

A practical consequence is obvious. Fragments of ultramafic-alkalic rocks in conglomerates and gravels within certain regions do not mean that younger ultramafic-alkalic rocks are absent. Unfortunately, the data do not permit the assessment of a dependence between the degree of diamond mineralization of specific kimberlitic and lamproitic events and the sequence of magmatic events within any region.

Magmatic Activity and Tectonic Transformations

The timing of kimberlitic and lamproitic pulses shows a strong tendency for synchroneity worldwide. The timing of kimberlitic and lamproitic volcanic activity or alkalic volcanism, in general, corresponds with the timing of major tectonic reconstructions in some regions. The forms of tectonic reconstruction vary in time and space, but such reconstructions are represented by cratonization, orogeny, breaking up of stable blocks, opening of oceans, and the formation of the great Karroo-type syneclises associated with extensive eruptive episodes of flood basalts (Karroo syneclise, South Africa; Tunguska syneclise, Siberian platform; and the syneclises associated with Deccan and Parana flood basalts, in India and Brazil, respectively).

A coincidence of pulses of alkalic magmatism with tectonic transformations became apparent as we reviewed the timing of the earliest stages of alkalic magmatism within the ancient platforms prior to cratonization. On this basis, "Clifford's rule" states that kimberlitic activity is specifically limited to stable Precambrian cratons (Clifford 1966).

In Australia, compliance with Clifford's rule was considered to be a major precondition for any diamond exploration program (Atkinson and Smith 1995). However, Australia, with its mosaic of blocks that stabilized at different times, became the area where this rule was reevaluated. The new concept considers that the earliest stages of alkalic volcanism coincide with the timing of cratonization, no matter if the pulse is Precambrian or not (Jaques et al. 1985). A major exception to the inferred first order correlation of cratonization with alkalic magmatism occurred in the West Kimberley region straddling the Proterozoic King Leopold mobile zone and the Phanerozoic Canning basin, both of which are inferred to have been cratonized around 1700 to 1800 Ma.

It required a considerable number of geologic studies to show that Clifford's rule needed to be corrected to incorporate the concept that the earliest stages of alkalic magmatism coincides with the time of cratonization. In an earlier section, we proposed that the earliest phases of ultramafic-alkalic magmatism within different cratons are associated with the timing of their cratonization.

As an example, the earliest phases of ultramafic-alkalic volcanism in different parts of Australia, can be compared with the time of stabilization of various blocks:

- The felsic alkalic rocks of the Yilgarn block are dated about 2400 Ma, which agrees well with the 2500 Ma cratonization age inferred from the last metamorphic event.
- The 1660 Ma Tennant Creek lamprophyres are slightly younger than the inferred 1700 to 1900 Ma cratonization age of the North Australia craton.
- The ages of the Mud Tusk carbonatite and Morder complex (730 and 1210 Ma, respectively) are similar to the 900 to 1000 Ma cratonization age inferred for the Central Australia mobile belts (age determinations were made using K-Ar and Rb-Sr methods on whole rock and mica).
- The South Australian kimberlites within the Adelaide fold belt are Jurassic. The Orroroo kimberlites and carbonatite are dated by K-Ar, Rb-Sr, and U-Pb methods at 170 to 172 Ma. The Terrowie kimberlites and lamproites are 164 to 174 Ma (Rb-Sr method). Within the Wandegee alkalic volcanics, 22 intrusives originally described as kimberlite are now considered as picrite basalts (160 Ma, dated by U-Pb method on zircon).

- Within the Tasman fold belt, alkalic basalts associated with alkalic magmatic activity, range from Permian to Cenozoic and yield K-Ar ages of 241 ± 34 Ma, 193 ± 10 Ma, 204 Ma, 211 ± 3 Ma, 198 ± 8 Ma, and 185 ± 8 Ma. Olivine nephelinite and alkalic basalt breccia in this region yield K-Ar ages 163 ± 8, 187 ± 2 and 178 Ma. Ninety-five known and 60 inferred diatremes of Lower Jurassic age yield dates from 138 ± 2 Ma to 180 Ma with a series of dates around 163 to 167 Ma. Mesozoic lamprophyres are also widely distributed within the Tasman belt. The McBride and Nulla provinces measure 80 to 100 km across and contains numerous bodies of alkalic olivine basalt, hawaiite, and basanite. In northern Queensland four periods of volcanic activity are recognized that range from >53 Ma to 3 Ma and that include fields with volcanism as recent as 10,000 years. In New South Wales similar rocks yield ages 70 to 10 Ma. These rocks include leucitites of the ultrapotassic series that are dated from 16 to 6 Ma (Jaques et al. 1985).

Synchronous magmatic activity typically coincides with tectonic transformation. These transformation events have been determined by age dates of granitic intrusives within adjacent linear orogenic belts surrounding stable blocks, by caldera-forming eruptions that produced huge amounts of silicic pyroclastic material, and by other types of activities that accompanied mountain building.

The oldest Paleozoic ultramafic-alkalic magmatic activities are associated with the formation of the Oslo graben, Scandinavia (665 and 675 Ma). Simultaneously, 1.6 km of ultramafic-alkalic volcanics were deposited within the Dneprovsko-Donetsky depression (aulacogen) within the Ukranian shield of the Russian platform (543–610 Ma by the K-Ar method). Within the Timan aulacogen of the northern part of the Russian platform, carbonatites were emplaced at 680, 695, and 652 Ma (K-Ar method). Alkalic gabbros from the same region yield K-Ar ages of 534, 546, and 545 Ma (Polevaya 1974). These rocks cut Precambrian sequences and are overlapped by Middle Devonian sediments.

Ultramafic-alkalic rocks associated with carbonatites are located in the eastern and southeastern part of the Aldan shield, Siberian platform, and were emplaced coincident with graben formation. This complex includes the Konder, Arbarastaakh, Ingily, Chad, and Inagly massifs. The oldest K-Ar dates obtained from different minerals within the massifs are in the range of 725 to 610 Ma (Figure 14.7). The dates (609–865 Ma) were obtained by K-Ar from the off-craton Ingili kimberlites.

It is interesting to note that the radiometric age characteristic of the rocks of the Beni Bousera ultramafic massif in Morocco, at the opposite, easternmost edge of the Alpian-Himalayan mobile belt, corresponds with orogenic events in the Alps and Himalayas and the time of eruption of Western Australian lamproites (Allsop, Bristow, and Skinner 1985). Naturally, orogenic processes are not expressed worldwide. In regions where a specific orogeny did not occur, tectonic transformations may take other forms. In particular, within stable blocks and ancient cratons as a whole, tectonic transformations may be expressed by a change in structural outline, creation of new structures (such as aulacogens), the closing and inversion of previously existing aulacogens, and the transformation of older ones. Examples of orogenic episodes are discussed below.

A Late Proterozoic kimberlitic event corresponds with the formation of a series of orogenic belts within or adjacent to most regions of Precambrian stabilization. A few of these include the King Leopold and Halls Creek mobile belts in Western Australia; the Eduran orogeny in West Africa (1.6–1.8 Ga) (Mitchell and Bergman 1991); the Orange River or Namaqua mobile belt in South Africa; and a metamorphic event (1360 Ma) near the Late Proterozoic kimberlite fields of Chelima and Majhgawan in India (Mitchell and Bergman 1991).

Dates of 900 Ma in North America are in many cases considered to reflect a metamorphic event, as confirmed by radiometric ages for metamorphism on nepheline gneisses of York River, Ontario. A similar radiometric age is also characteristic of magmatic rocks such as the Meach Lake carbonatites. Within the Cordilleran province, the most ancient radiometric dates were obtained from the alkalic complex at Mountain Pass, California, which includes alkalic granites, shonkinites, and syenites associated with carbonatites.

The Pan-African orogenic event (500–600 Ma) coincided with the formation of Caledonides in Europe and the Avalonian event that is well expressed along the southeastern edge of the Appalachian orogenic belt and in rocks beneath the sediments of the Atlantic coastal plains as far southwest as Florida. During this time span kimberlites in northern Brazil, the Colossus kimberlite pipe in Zimbabwe, South Africa, and a series of ultramafic-alkalic plutons in the Kola Peninsula were formed. Within the same time interval are some of the earliest radiometric dates for ultramafic-alkalic rocks of the Oslo graben and a series of ultramafic-alkalic plutons of the Aldan shield of the Siberian platform.

During structural development of the platform stage, radiometric dates of ultramafic-alkalic massifs are typically concentrated along time intervals coinciding with major episodes of structural reconstruction, which match orogenic events within the adjacent mobile belts. The earliest phase of Paleozoic alkalic magmatism is associated with the closing and partial inversion of Ryphean aulacogens (about 600–800 Ma).

In Africa the Pan-African orogeny embraced the entire continent. During this time, a high-pressure mineral complex was developed in metamorphic terranes in Kazakstan (510–530 Ma). This coincidence possibly reflects a Caledonian (early Paleozoic) orogenic event.

Magmatic activity within this interval is typical of ancient platforms. The earliest Paleozoic alkalic rocks in southern Brazil (nephelinite, sodalite syenites, and tawaite from Itaju do Colonia and Patiragua) yield K-Ar ages of 660 to 665 Ma. In North America the earliest phase of alkalic volcanism during the early stages of platform development took place during the Avalonian episode when carbonatites of Mt. Grace, British Columbia, were emplaced (773 Ma). Additionally, the Lobo syenite, New Mexico, intruded Precambrian schists about 604 Ma.

Within the Cordilleran province of North America, an intensive carbonatitic volcanism event occurred around 506 to 570 Ma. This included Iron Hill (550–570 Ma), McClure Mountain (506–535 Ma) Democrat Creek (511–534 Ma), and Gem Park, Colorado (551 Ma). The youngest Paleozoic date is from Rb-Sr isochron dating of a syenite dike from Pedernal Hill.

The Late Caledonian orogeny (early Paleozoic) is reflected in synchronous magmatic activity within Kola Peninsula and the Siberian platform. In the Siberian platform, alkalic rocks of Tomtor massif are dated at 400 Ma. A series of similar dates was obtained using $^{206}Pb/^{238}U$ method on zircon for kimberlite from different regions (Malo-Batuobinsky, Verkhne-Munsky; see Davies, Sobolev, and Kharkiv 1980). Ages obtained by Pb-U method were repeated by Rb-Sr and K-Ar methods on mica for the Mir and Udachnaya pipes.

At 400 Ma, ultramafic-alkalic massifs within the Kola Peninsula continued to erupt for about 400 million years along with a series consequent magma pulses (Figure 14.7).

Within the Siberian platform, the formation of the Tunguska syneclise and the major stage of flood basalt eruption (220–240 Ma) coincide with the Herzinian stage of tectonic reconstruction. Simultaneously, the largest ultramafic-alkalic massif in the world (Gulinsky) and a series of comparatively small intrusive bodies within the Maimecha-Kotuisky province west of the Anabar shield was emplaced within anteclises. K-Ar dates for rocks of the Gulinsky massif range from 248 to 264 Ma (Prokhorova, Evzikova, and Mikhailova 1966; Bagdasarov, Voronsky, and Arakelyants 1983). A similar date is reported for the Tomtor massif (K-Ar date of 250 Ma) (Erlich and Zagruzina 1981).

Occasional radiometric dates indicating the existence of the same episode within other platforms have been recognized.

- A Rb-Sr age of 257 Ma has been obtained for the Kentucky kimberlites that coincides with the age of molasse formation in the Appalachians.
- Small circular alkalic plutons and carbonatitic bodies of Lower Permian age were dated by Rb-Sr technique in Mali (262 ± 7 and 269 ± 9; Liegolis et al. 1983). The Kapamba lamproites (about 220 Ma on the basis of the modal age of a phlogopite concentrate or <250 Ma) belong to the same episode.
- In the Tasman belt of Australia, alkalic basalts are dated at 241 ± 34 Ma, and 211 ± 3 Ma (K-Ar) (Jaques et al. 1985).
- During this episode, nepheline syenites of Pao de Acucar were formed in Brazil (240 Ma) (Ulbrich and Gomes 1981).
- The Doklowayo pipe, Zimbabwe, is overlain by Middle to Upper Triassic sedimentary rocks indicating a pre-Middle Triassic age for the kimberlite.

These data show that the ultramafic-alkalic magmatic activity is synchronous with major stages of flood basalt eruptions (compare appropriate data in Tables 14.2 and 14.3). Cited radiometric dates of ultramafic-alkalic intrusives within the northeastern Siberian platform nearly match the timing of the main flood basalt event and formation of the Tunguska syneclise (250 ± 5 Ma). Contemporary kimberlites belong to the Permo-Triassic episode. The Kimmerian tectonic transformation also occurred during the Middle Jurassic time.

A series of dates that corresponds with this episode was obtained for South African kimberlites, including the East Griqualand (150–190 Ma) and Swartruggens (150 Ma), and for the Pniel and Postmasburg lamproites. Approximately synchronous with this activity was the formation of Karroo basalts in South Africa and flood basalts in Antarctica (190 ± 5 and 170 ± 5 Ma, respectively). In Australia, the Orroroo and Terrowie kimberlites erupted (170 to 172 Ma and 164 to 174 Ma, respectively). During the same time, around 160 Ma, Wandegee picrite basalts formed.

Within the Siberian platform, dates in the range of 159 to 146 Ma were obtained from kimberlites of the Prilensky and Nizhne-Oleneksky region. It was concluded from determinations in nearby pipes that they were Late Jurassic (143–135 Ma; Davies, Sobolev, and Kharkiv 1980). The age of a belemnite xenolith in the Obnazhennaya pipe put the age of emplacement as Late Jurassic or possibly Cretaceous (Milashev and Shul'gina 1959).

In the Eastern European platform, monchicite dikes were emplaced (K-Ar ages 162 and 163 Ma). In Brazil, a significant group of dates cluster around 130 Ma. During this period alkalic rocks of the Jacupiranga node were formed. Within the accuracy of radiometric dating, the Serra Geral episode of flood basalts simultaneously erupted in Brazil forming the Parana basin (130 ± 5 Ma).

In India, lamprophyric dikes erupted (105–121 Ma). In Sierra Leone, the Koidu complex formed (92–141 Ma). Within the Aldan shield of the Siberian platform, the Ingily ultramafic-alkalic massif was formed during this time. Rocks of this massif yield K-Ar dates of 85 to 115 Ma. Lamproites of the Murun massif yield dates of 115 to 143 Ma (cited in Mitchell and Bergman 1991). Unpublished K-Ar dates for the Tommot massif are about 110 Ma (I. Zagruzina, personal communication, 1983).

The opening of Atlantic Ocean occurred around 127 million years ago (Moore 1976). During this time, the Sierra Geral flood basalts erupted and major structural transformations occurred along with some ultramafic-alkalic magmatic activity.

Most South African kimberlites yield radiometric ages between 83 and 120 Ma (Figure 14.4) (Smith et al. 1985). During the same time period kimberlites of North China (52–117 Ma, Zhang and Liu 1983) and numerous massifs of the Monteregian province (28 determinations from seven units yielded a K-Ar isochron of 117 Ma for the Oka massif) erupted. Rb-Sr dates for the St. Johnson massif are 111 ± 6 Ma with K-Ar ages of 117 ± 9 and 120 ± 8 Ma. Within the Cascade subprovince, the Crownest trachytes are of similar age. Similar ages were reported for the Tombstone syenitic batholith and Simmilkananeen pluton. In the Alaskan subprovince, the Hunt massif formed.

At the same time, lamproites formed that are associated with the development of the Oklahoma rift system. Numerous lamproitic dikes and sills are confidently dated at 105 to 121 Ma. (Rock et al. 1992). In West Africa, the Koidu complex in Sierra Leone formed at the same time (Dalrymple, Gromme, and White 1975). The Rajmahal flood basalts in India erupted (110 ± 5 Ma) during this time, and a similar age (105–121 Ma) is reported for lamprophyric dikes.

A new intensification of ultramafic-alkalic magmatism took place at the end of Cretaceous time. In South Africa, a series of kimberlitic bodies and melilitic intrusives located at the periphery of the Transvaal craton was formed. Radiometric ages in the range 60 to 70 Ma characterize the Gibeon field, the Pofadder-Riftfontein (70 Ma) field, and the melilite-“kimberlite”-carbonatite suite of Namaqualand and Bushmanland (Moore and Verwoerd 1985).

In Brazil, following a short period of decreased magmatism, eruptions vigorously resumed during the Late Cretaceous and formed alkalic rocks of the Minas Gerais province and Sao Paolo–Rio de Janeiro coastal belt. Flood basalt eruptions in the Deccan (India) and Brito-Arctic provinces correspond with this period (66 ± 2 and 62 ± 2 Ma, respectively).

One series of radiometric dates from ultramafic-alkalic rocks is concentrated during the Tertiary (35 Ma, 21–23 Ma, 13–16 Ma, and 0–4 Ma). The Sessia-Lanzo lamproites (29–33 Ma) and flood basalts in Ethiopia (35 ± 2 Ma) erupted during the first series of this episode.

A second series is characterized by the formation of lamproites in Pamir (Dmitriyev 1974) and the emplacement of the West Australian lamproites (21–23 Ma). At the same time, the formation of the Kisingiri complex in Kenya lies near the time boundary between first and second stages; lavas of the first cycle are older than 20 Ma and sedimentation at the margin of the postcaldera extrusive dome is estimated at 20 Ma (LeBas 1977).

The time of emplacement of the Sisco lamproites in Corsica (13.5–15.4 Ma) coincides or immediately follows the 15 Ma stage of rejuvenation of metamorphic rocks reflecting the uplift of the Alps. The Columbia Plateau basalts in North America erupted during this period (7 ± 1 Ma), as did alkalic pipes in Pamir (14–15 Ma) (Kogarko et al. 1995). The formation of lamproites of the Murcia-Almeria province (Spain) occurred within the time frame of 5.7 to 10.8 Ma. Simultaneously alkalic rocks of Campas de Calatrava formed (6.4–8.6 Ma) (Mitchell and Bergman 1991).

The Roman ultrapotassic province in Italy had a major peak of eruption at 2 to 6 Ma that shifted to an interval at 0 to 2 Ma with minor eruptions. These dates are matched by radiometric dates of lamproites from Leucite Hills, Wyoming (1.1–3.1 Ma).

Several attempts have been made to correlate the timing of specific episodes of alkalic magmatism on both sides of the Atlantic Ocean and to explain their coincidence with the opening of the ocean (Doig 1970; Williams and Williams 1977; Marsh 1973). There is no ocean between the African, Siberian, or East European cratons; nevertheless, processes within them are synchronous. Moreover, as it has been mentioned, the formation of kimberlites in South Africa is synchronous with the emplacement of lamproites in India and North America and with the formation of kimberlites in northern China.

The Laramide tectonic uplift in North America occurred in Late Cretaceous time, probably within the time span from 50 to 85 Ma. During this time intense alkalic magmatism occurred in Brazil along with the emplacement of the Paranaiba kimberlite (Svisero et al. 1984). The Deccan flood basalt eruptions in India also occurred during this period.

Alpine tectonic events in Eurasia and North America created the Alps and Himalaya Mountains. At the same time the circum-Pacific orogenic zones were formed, which included the Rocky Mountains and Andes at the American side of the ocean and a series of island arcs and related orogenic systems in Kamchatka on the Asian side.

Many attempts have been made to explore subducted plates for kimberlitic volcanism, because these plates are considered to be the major source of alkalis and volatiles needed to generate kimberlite and lamproite. Because kimberlite and lamproite occur in places unrelated to plate boundaries, it is thought that kimberlitic activity is associated with sources that are much deeper than the lower boundary of the lithosphere and possibly as deep as the outer boundary of the Earth's core. This conclusion imposes certain constraints on ideas related to the genesis of kimberlitic and lamproitic magmas and, at the same time, provides opportunities to explain major magma-generation processes.

Realizing that the Earth's core is a major source of volatiles provides us with an universal explanation for three things: the source of major components needed to generate magma, the synchroneity of mountain building, and the association of granitic magmatism and alkalic magmatism within adjacent stable blocks. Such synchroneity leads us to acknowledge that the driving force for tectonics and volcanism is located in the deep mantle (or, more properly, at the core/mantle boundary) and is fed by mantle-derived metasomatic fluids. These emissions are part of the Earth's continued degassing, and there is a persistent or renewable source of volatiles within the mantle (Bailly 1982).

As we have mentioned earlier, the intensification of synchronous alkalic and ultramafic-alkalic magmatic activity is combined with diachronous lulls of activity in different regions. This difference reflects different underlying reasons: synchroneity reflects pulses of volatile flow generated at core/mantle boundary. These pulses are common for the globe as a whole, whereas lulls in magmatic activity reflect strain in the crust that differs from region to region and prevents magma from ascending to upper levels.

It is important to stress that the timing of the different types of diamond deposits (such as those in kimberlites, lamproites, ultramafic plutons, high-pressure metamorphic terranes, and ring structures) is synchronous. This synchroneity is a clear indication that the deposits reflect the same deep-seated process—pulses of intensification of heat and volatile flow from the core/mantle boundary. The differences in the types of deposits reflect their different tectonic settings.

SUGGESTED MODEL

Synchronous pulses disregard any plate boundaries and apparently have no association with subduction zones. Some pulses coincide with the opening of the Atlantic Ocean, but this coincidence isn't compelling and doesn't set any precondition for a pulse. Thus we conclude that the type of magmatism under consideration is associated with a very deep-seated source probably located beneath plates at the mantle/core boundary. Pulses of magmatic activity also coincide with major orogenic stages. Thus, it is no wonder that within the same pulses different types of diamond deposits were formed that probably reflect root zones of orogenic belts. Appropriate examples can be seen in high-pressure metamorphic terranes in Kazakstan and China, stratified ultramafic complexes like in Beni Bousera, or specific types of ring structures (astroblemes) such as the Popigay ring structure.

The initial stage of magma generation is associated with deep transformation of the crust and mantle, and it is reflected in the stable blocks during the process of their cratonization. The later stages coincide with major reconstruction of stable blocks. These cases require restoration of isostatic equilibrium accompanied by mass exchange in the lower parts of asthenosphere.

The synchronous, polychronous character of ultramafic-alkalic magmatism through time is a result of several different and often opposite and controversial causes.

- Pulses of intensified terrestrial heat flow result in magma generation.
- Time spans with a complete absence or scarcity of radiometric dates apparently reflect general compression in the crust that prevented magma from rising to surficial levels.
- Once generated at great depth, a magma remains stored for a very long time (up to 1 billion years) and can be squeezed to shallow crustal levels during episodes of tectonic reconstruction (folding or orogeny).
- The timing of major stages of ultramafic-alkalic magma generation coincides with the timing of cratonization, resulting in deep transformation of the crust and upper parts of the mantle.
- Repetitive pulses of activity (both magmatic and hydrothermal/metasomatic processes) occur that utilize the same pipes or magmatic plutons, which play a role as chimneys that conduct volatile flow from the deepest levels in the Earth to the surface as the earth degasses. Consistent radiometric dates reflect the timing of heat exhaustion in the root of magmatic systems—apparently within a kind of asthenolith-feeding magmatic system.

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SECTION 5

Exploration, Mining, Economics, and Extraction

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CHAPTER 15

Exploring for Diamond Deposits

INTRODUCTION

Hausel, McCallum, and Woodzick (1979), Hausel, Glahn, and Woodzick (1981), Macnae (1979, 1995), and Nixon (1981) provide overviews of exploration methods and techniques used in the search for kimberlite, and Atkinson (1989) summarized the exploration techniques used to search for lamproite. Some recent advancements in the attempt to isolate probable diamondiferous deposits have compared the geochemistry of the tracer minerals with that of diamond-inclusion minerals. Diamond exploration methods are also being developed for other diamondiferous host rocks, although the technology for the exploration of most of these unusual diamondiferous host rocks is considered to be in the infant stages.

Since the early 1980s, following the discovery of diamondiferous lamproite in Western Australia, exploration models and techniques used to search for these rock types have been evolving. Mitchell and Bergman (1991) suggested that diamondiferous lamproites are restricted to craton margins in crustal domains that experienced Proterozoic to Archean accretional and or other orogenic events.

Since the discovery of the world's largest known primary source of diamonds at Argyle, Australia, lamproites have become an important exploration target. Many diamondiferous lamproites possess geochemical and mineralogical similarities to kimberlite. However, there are also differences that require the use of other exploration methods to search for lamproite.

Many kimberlites are known for their dispersion of unique kimberlitic indicator minerals, which are used as pathfinder minerals that may lead to the source rock. However, the classic kimberlitic indicator minerals are rare to absent in most lamproites and lamprophyres. As a result, other minerals are commonly used in addition to the kimberlitic indicator minerals to search for lamproite and lamprophyre.

Exploration for diamonds requires an understanding of the favorable geology and geochemistry that may produce diamond deposits at the surface of the Earth. The many nuances of diamond deposits require the exploration geologist or prospector to have considerable experience in searching for diamonds. An understanding of the geology and geomorphology of these deposits, as well as a good grasp of the unique mineralogy and petrology of the potential host rocks, is essential.

Without a good understanding and considerable experience, fairly competent exploration geologists can easily overlook potential host rocks. As an example, in the spring of 2000 a large group of geologists and prospectors were purposely led across a few hundred meters of kimberlite on a diamond field trip in northern Colorado, and not one person realized that he or she had walked across kimberlite. The purpose of the exercise was to demonstrate the difficulty in recognizing kimberlite in the field.

During the middle 1990s, some exploration geologists and prospectors filed claims for diamonds on the basis of a geologic map. After laying out the claim blocks, they requested help in finding the kimberlites in the field. Even with a map in hand, most of the kimberlites remained invisible to the untrained eye. These are just two of many examples of the difficulty in finding kimberlite. Such difficulties are primarily due to the indistinct nature of weathered kimberlite.

Most diamond source rocks (e.g., garnet harzburgite, chromite harzburgite, lherzolite, and eclogite) appear to be concentrated in the roots of Archean cratons and along craton margins where they are not accessible except by the sampling of rare magmas such as kimberlite and lamproite. In some circumstances, diamondiferous source rocks outside of cratonic environments appear to reach the surface by tectonic (rather than magmatic) processes (e.g., Beni Bousera).

Kimberlitic, lamproitic, and lamprophyric melts may develop at depth, once the geothermal gradients have been reduced sufficiently to promote partial fusion at about 150 km depth. A low seismic velocity zone may be coincident with this zone of melting; however, individual zones of melting associated with kimberlite magmatism are too small to identify. Adjacent cratonized Early to Middle Proterozoic mobile belts may also have similar low-velocity zones at depth.

Cratons provide relatively stable geologic environments with low geothermal gradients. The low geothermal gradients are necessary for generation of deep-seated magmas within or near the diamond stability field. In cratonic environments, many post-Precambrian magmas typically have ultrabasic, ultrapotassic, or alkalic affinity indicative of a deep-seated origin. In such areas of high geothermal gradient, one would not anticipate eruption of primary diamond deposits.

Models of the formation of Archean mantle roots view the depleted host rocks of diamonds either as a residue of komatiite magmatism or as products of tectonic underplating by imbricate slabs of subducted oceanic lithosphere. The primary lithospheric repositories for diamonds are suggested to be the cool keels of the cratons that developed more than 2 billion years ago (Nixon 1995; Spetsius 1995; Helmstaedt and Gurney 1995). This repository location accounts for the close association of diamondiferous kimberlites with cratons as pointed out by Kennedy (1964). Diamondiferous lamproites appear to be primarily localized in marginal cratonic terranes that are of sufficient age to be regarded as stabilized (cratonized).

According to Clifford's rule, diamondiferous kimberlites are restricted to cratonic regimes that have been relatively stable for about 1.5 billion years. Janse (1984, 1994) suggested that these cratons be separated into areas of favorability to facilitate initial evaluation for diamond potential. He termed these regions archons, protons, and tectons. Because the term proton has been extensively used in nuclear physics, we elect to use the term Proterozoic accretionary terranes.

The archons (Archean basement terranes that stabilized more than 2.5 billion years ago) are considered to have high potential for commercial diamond deposits hosted by kimberlite and possibly by lamproite and lamprophyre. The Proterozoic accretionary terranes (Early to Middle Proterozoic [2.5–1.6 Ga] basement terranes) have moderate potential for commercial diamond deposits in kimberlite and high potential for commercial diamonds deposits in lamproite. Tectons (Late Proterozoic [1.6 Ga–600 Ma] basement terranes) are considered to have low potential for commercial diamond deposits hosted by kimberlite, lamproite, and lamprophyre. Kimberlites and lamproites are rare in areas underlain by younger or rejuvenated basement. This concept applies to kimberlite, lamproite, and lamprophyre; other types of tectonically derived diamond deposits (such as high-pressure metamorphic complexes, astroblemes, and subduction-related complexes) are found in other geologic environments.

On the basis of this concept, the cores of cratonic terranes (Archons) have the highest potential for economic diamond deposits hosted by kimberlite. Nearly every commercial kimberlite-hosted diamond deposit lies within terranes underlain by Archean basement (Levinson, Gurney, and Kirkley 1992). A notable exception is the Kelsey Lake diamond mine in Colorado, which is underlain by Proterozoic basement. Early to Middle Proterozoic basement terranes, although not considered to have as great a potential as Archons, should still be evaluated for diamondiferous kimberlite and for diamondiferous lamproite. This conclusion is based on the fact that the Argyle lamproite in Western Australia lies within a Proterozoic fold belt cratonized about 1.8 Ga, and it is a very high-grade diamond deposit (Figure 15.1). Argyle is not an exception. The George Creek kimberlites in the Colorado–Wyoming State Line district yielded some relatively high grade bulk sample tests (18–135 carats/100 tonnes). However, diamondiferous deposits found within the Proterozoic accretionary terranes so far have yielded smaller diamonds on the average.

The Argyle lamproite contains predominantly Proterozoic E-type diamonds. That these diamonds occur in lamproite rather than in kimberlite is immaterial, because the diamonds are older than the lamproite. This deposit illustrates that large accumulations of post-Archean magmatic eclogitic diamonds lie outside Archean cratons (Kirkley, Gurney, and Levinson 1991). Thus, any mantle-derived magma (lamproite, kimberlite, and lamprophyre) could potentially sample similar eclogitic source terranes. The lesson from Argyle is that Proterozoic orogenic belts, as well as other geologic environments, should not be written off in diamond exploration unless all of the facts indicate that diamond mineralization and preservation is unlikely.

Diamond-bearing kimberlites and lamproites have intruded the earth's crust for a very long time, as is demonstrated by diamonds found in the 2.6 Ga Witwatersrand conglomerates of South Africa. The oldest known preserved kimberlites are 1.6 Ga intrusions (nondiamondiferous) near Kuruman in the northern Cape province of South Africa, situated on the Kaapvaal craton. Information on lamproite is less available. The known diamondiferous lamproites range in age from the Argyle intrusive (1.2 Ga) to some Ellendale lamproites in the Kimberley field of Western Australia (20 Ma). Much younger lamproites are found in the Leucite Hills of Wyoming (3–1 Ma) and in Antarctica (<1 Ma), although no diamonds have been found in them (Figure 15.2). But neither field has been searched in detail for diamonds. Lamproites and kimberlites typically show more than one episode of emplacement in most districts (Nixon 1995; Hausel 1996).

One school of thought suggests that kimberlite and related magmas begin their journey by incipient melting in the mantle. As the melt gains upward momentum, fragments of the mantle become entrapped in the melt as xenoliths, which are also transported in the magma to the earth's surface. Many diamonds found in kimberlite and lamproite are thought to have originated from the disaggregation of some of these mantle nodules.

a

b

FIGURE 15.1 View of the Argyle (a) mine and (b) mill, Western Australia (photos by W.D. Hausel, 1986)

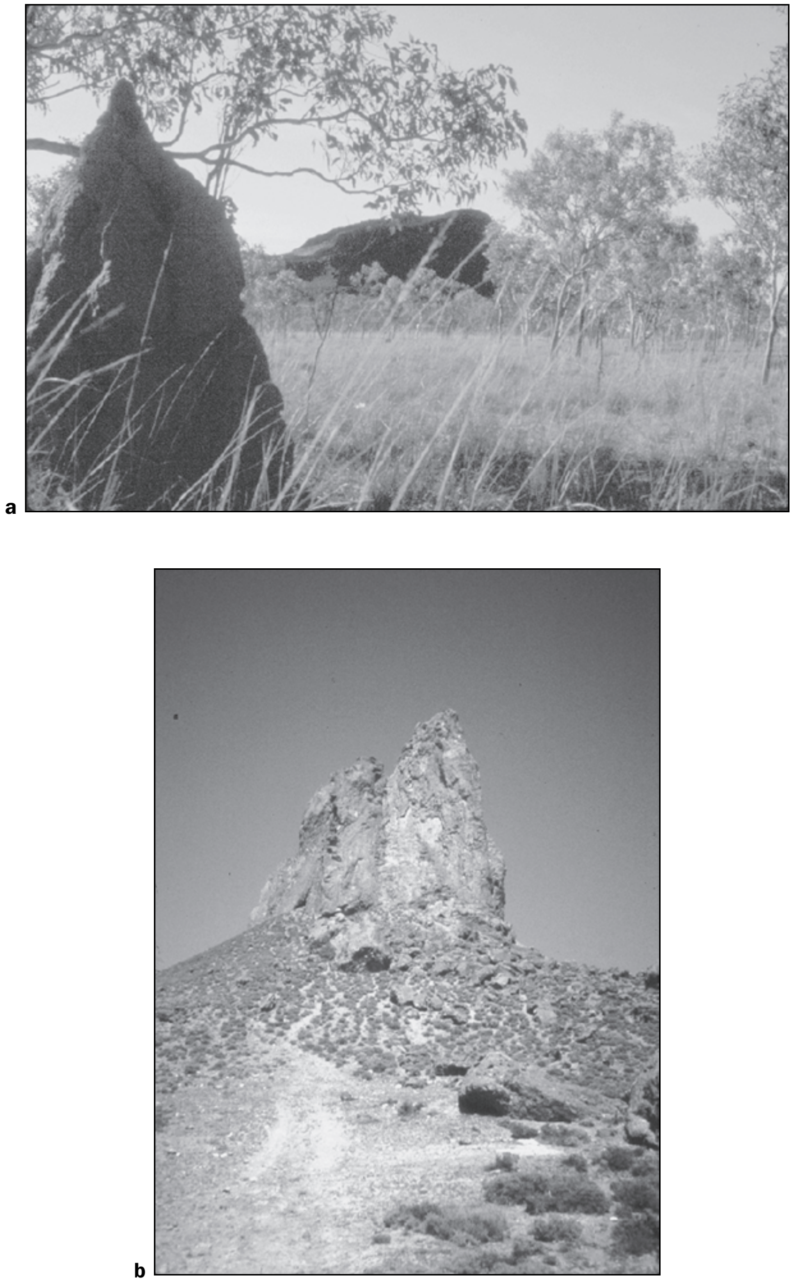

FIGURE 15.2 Phlogopite-leucite lamproites typically are resistant to erosion and form positive features. (a) Mt. North, behind a termite mound in Western Australia. (b) Boars Tusk, a prominent volcanic neck in the Leucite Hills, Wyoming (photos by W.D. Hausel, 1986).

If the kimberlitic or lamproitic magma originates from too shallow a depth, then graphite, not diamond, will be preserved, and the mineral components that form the cognate xenoliths trapped within the magma will reflect the unique chemistry of that source region. For example, diamondiferous peridotites originate from a high-magnesian, high-chromian, subcalcic geochemical environment that is imprinted on the mineralogy of the peridotite. A similar peridotite derived from the graphite stability field is typically a high-magnesian, high-chromian, calcic peridotite. Thus the minerals derived from the disaggregation of these peridotites (and of eclogites) are many orders of magnitude more common than diamond within the kimberlite host. Because of their abundance, these minerals are commonly used as indicator (or tracer) minerals to find the hidden kimberlite, and they are termed "kimberlitic indicator minerals." These indicator minerals are much less abundant (to nonexistent) in most lamproite and lamprophyre.

Once kimberlitic indicator minerals are identified, their chemistry is determined to ascertain the probability that they originated from a diamond source rock. It is important to test many indicator minerals, because the indicator minerals from a single diamond-bearing intrusive will be derived from different depths within the earth's mantle and crust (i.e., some from the diamond stability field and some from the graphite stability field).

On the basis of geobarometry, the silicate phases in peridotites associated with diamond are estimated to have formed at depths of 150–200 km and at temperatures generally not exceeding 1,200°C (Kennedy 1964). In order for diamond to survive transport to the surface in the magma from these depths, the magma emplacement must be rapid, because diamonds are unstable at shallower depths. Slow transport at elevated temperature will expose diamond to retrogressive metamorphic effects, thereby allowing diamond to resorb and convert to graphite. The environment near the earth's surface is also detrimental, because diamond will tend to oxidize (burn) in a hot magma to produce CO and CO_2 and/or graphite. Information on the oxidizing or reducing environment during emplacement is thought to be related to the chemistry of picroilmenite from the host rock. In addition, Zn and Ni in peridotitic pyropes are also thought to provide an indication of diamond preservation, and the morphology of diamonds contained in the rock may also provide information on the degree of diamond resorption.

Methods used to prospect for diamondiferous host rocks depend on a variety of factors. Different climate regions require different exploration methods, and regions that have been uplifted since the emplacement of the diamondiferous host rock require a different exploration philosophy than regions that have experienced few or no orogenic events since the emplacement of the mantle-derived rock. Other significant factors that affect exploration include political regimes and land ownership. These factors were emphasized by the policies of the Clinton administration in the United States in the late 1990s, which greatly restricted access to public land. These policies eliminated much of the incentive to explore within the United States, while at the same time exploration increased dramatically in Canada.

Lamproite and kimberlite may exhibit trends and lie along mappable fractures. Such fractures are typically restricted and difficult to recognize, and they may be indicated only by alignment of a cluster of intrusives or by the elongation of a diatreme. When fractures and trends are recognized, they may provide orientation information useful for locating additional intrusives. However, many orientations may be recognized in a single district.

Some kimberlites and lamproites may be seen on high- or low-altitude remote-sensing imagery. Such imagery is used to search favorable regions for geomorphic features, vegetation anomalies, or the characteristic reflectance of kimberlitic clays. Remote-sensing methods are also useful in mapping structures and fractures that may control groups of kimberlites or lamproites. Once a controlling structure is identified, field reconnaissance will be necessary to search for poorly exposed kimberlites or lamproites or for related hidden intrusives.

Both kimberlite and olivine lamproite typically contain abundant serpentinized olivine, which renders these rocks susceptible to erosion. In some lamproite fields, barren phlogopite-leucite lamproites may stand out as positive features, giving the prospector and geologist the erroneous impression that olivine lamproites are absent. However, nearby diamondiferous olivine-rich lamproites may be buried by a thin veneer of soil, owing to serpentinization and erosion. This shallow burial is especially notable in the Ellendale field in Western Australia, where the diamondiferous lamproites lie hidden within a field of well-exposed leucite lamproites. The Argyle diamondiferous lamproite and diamondiferous lamproites in the Murfreesboro field of Arkansas were also hidden by a thin soil cover. Kimberlites also tend to be highly serpentinized, making kimberlite outcrops the exception rather than the rule. Kimberlites and lamproites also form relatively small intrusives, ranging from a few meters and typically less than a thousand meters across, thus providing small exploration targets.

When lamproites weather, they produce an exotic suite of secondary minerals such as nontronite, smectite, and barite. Lamproites also possess anomalous amounts of Ti, K, Ba, Zr, and Nb compared with most other rock types. These trace metals may favor the growth of different flora than is common in the surrounding terrain, and they may stress the vegetation. As an example, the Big Spring vent, West Kimberley, Australia, is characterized by anomalous faint pink tones reflecting growth patterns of grass over the vent (Jaques, Lewis, and Smith 1986). Weathering of kimberlite produces a carbonated montmorillonite soil enriched in Mg, Nb, and Cr. This clay will also stress flora and produce vegetation anomalies over the intrusives.

GEOLOGIC CLUES

Structural Trends

Some researchers suggest that kimberlites are related to extensive regional fractures or structures, although a clear association with structures on a district scale is not always evident. Diamondiferous intrusives usually appear to be controlled by discrete fractures of limited surface extent.

However, many of the intrusives do occur in swarms, and the identification, mapping, and sampling of district-scale orientations is important and can sometimes lead to the discovery of associated intrusives and hidden pipes. Additional discoveries are due to the fact that kimberlite pipes ascend from a feeder-dike complex at depth (possibly 1–2 km beneath the surface). The feeder-dike complex can be envisioned as a long, linear dike (or dikes) with periodic eruptive centers that produce pipes that emanate upwards from the dike. This configuration has been mapped in a number of districts around the world. Such swarms may host barren intrusives among a few subeconomic pipes and one or two commercial pipes.

In the Iron Mountain district, Wyoming, erosion is deep enough to have removed the former kimberlite pipes and to have exposed the feeder-dike complex. Here the feeder-dike complex is roughly linear, forming a 7.5-km-long, east-northeast trend

(Hausel et al. 2000). Several smaller dikes and a few sills branch from the main dike. Along the trend several enlargements ("blows") contain frothy kimberlite or carbonatized kimberlite breccia (as opposed to the massive kimberlite porphyry) that forms much of the dike. These enlargements are interpreted as eruptive centers or roots of former pipes (Smith 1977).

A similar complex was identified in the State Line district, Colorado–Wyoming, although the erosional level is not as deep. A group of pipes containing diatreme facies kimberlite (the Schaffer pipes) were mapped as roughly elliptical pipes that are distinctly elongated along a northwest trend (McCallum and Mabarak 1976; Hausel, McCallum, and Woodzick 1979; Hausel, Glahn, and Woodzick 1981). Electrical resistivity surveys showed that most of the pipes were connected by a feeder dike at depth (Cominco American 1982).

Even though this group of kimberlites showed a prominent northwest trend, the Sloan 1 kimberlite in the southern portion of the district was emplaced along an east-west fracture, the Copper King fault, and the adjacent Sloan 2 was emplaced on a northwest-trending fault (McCallum and Mabarak 1976). Thus more than one prominent orientation can be found within the same district.

It is important that the geologist pay close attention to apparent structural trends during mapping. Simply by following trends, Hausel, Glahn, and Woodzick (1981) and Hausel et al. (2000) discovered several kimberlites in districts that had already been thoroughly mapped. Most structural trends appear to be relatively short and are for the most part district-scale fractures. Regional controlling lineaments may or may not be recognizable (see Chapter 13, Tectonic Control of Diamond Deposits).

For example, near the southeastern edge of the Wyoming craton, a group of kimberlite fields is scattered throughout a north-south region extending about 170 km within the accreted Proterozoic terrane. One major lineament in this region is the northeast-trending Cheyenne belt, which is a major Precambrian suture separating the Archean basement to the north from the Proterozoic basement to the south. The association of kimberlites with this suture is not clear. Kimberlite in the Middle Sybille Creek and Iron Mountain kimberlite fields lies near the suture. Kimberlites in the Iron Mountain district a few kilometers south of the suture lie essentially parallel to this structure; however, the one known kimberlite in the Middle Sybille Creek field trends nearly perpendicular (N30°W) to the suture. Kimberlites farther south in the State Line district, Estes Park, and the Boulder area show no apparent relationship to this structure. Many kimberlites in the State Line district about 48 km to the south have several trends, but for the most part they are dominated by N35°W trends that are roughly perpendicular to the Cheyenne belt.

The tendency of kimberlite pipes to occur in groups or clusters was also recognized in Lesotho. In Lesotho, South Africa, Dempster and Richard (1973) reported a close association of kimberlites with lineaments. The orientations of the majority of linear structures mapped in the region showed west-northwest trends. Ninety-six percent of all kimberlites were found along west-northwest trends. Many kimberlite pipes were also located where these west-northwest trends intersected west-southwest fractures.

Gurney (1972) suggested that this clustering may relate to favorable tensional crustal fractures within deep-seated zones of weakness. Thus it was suggested that the location of the kimberlite swarms may have been controlled by unseen deep fractures with limited surface expression.

Taylor (1984) suggested that kimberlites in the eastern United States were structurally controlled. According to Taylor, kimberlites and kimberlitic rocks in the Appalachian Mountains showed a close relationship with wrench faults associated with Mesozoic rifting. Mitchell and Bergman (1991) report that several lamproite fields are located along continental extensions of oceanic transform faults. However, on a regional scale, the controlling structures may not always be readily apparent.

In the West Kimberley province of Western Australia, some lamproites were shown to be spatially associated with the Sandy Creek shear zone, a Proterozoic fault. In the Ellendale field within this province, several of the lamproites were found near cross faults perpendicular to the Oscar Range trend. But the intrusions, in general, do not appear to be directly related to any known faulting. The diamondiferous olivine lamproite to the east at Argyle has an elongated morphology suggestive of fault control, and it intrudes a splay of the Glenhill fault (Jaques, Lewis, and Smith 1986).

Lamproites of the Leucite Hills field in western Wyoming lie near the Farson and Rock Springs lineaments (Gordon Marlatt, personal communication, 1996). On a district scale, the lamproites lie along the flank of the Rock Springs uplift (a Laramide anticline exposed in Cretaceous sediments), and most occur on distinct east-west–trending fractures perpendicular to the axis of the Rock Springs uplift and parallel to the Rock Springs lineament (W.D. Hausel, unpublished data, 1997; Hausel, Gregory, and Sutherland 1995).

One of the arguments concerning kimberlite and lamproite emplacement is whether these magmas propagate fractures as they begin to rise from the mantle, or whether deep fracture systems reaching to mantle depths initiate magma rise following a pressure drop caused by the fracture. If the latter is the controlling mechanism, then it is important to identify such megastructures during exploration. But such megastructures may not always be evident in the field (if they exist at all), and they are almost never observed on the surface. For instance, many oceanic transform faults mapped along the sea bottom are projected hundreds of kilometers into cratons and are assumed to exert some control on the emplacement of potentially diamondiferous magmas. One problem with this concept is that the location of the transform faults within the craton is not known with any kind of accuracy—particularly on the district scale. Thus, for the exploration geologist, such structures more often than not offer little assistance in the search for diamonds on a district scale.

If a major tectonic fracture system is present, kimberlite magma may accept a conduit within this system. Vertical fractures are assumed to be favored for kimberlite/lamproite emplacement, owing to the rapid rise of the magmas and the rarity of kimberlite sills. Thrust faults are considered unfavorable owing to their orientation and their rarity in most cratons.

The dynamics of some of these fracture systems are quite impressive. The surface expression of the fracture may be only about 1.2 km long, but the fracture may be as much as 120 km deep. Such structures suggest magmatic boring to the surface.

In many districts, evidence of multiple emplacements of mantle-derived magmas suggests that the same conduit is reused. This reuse is also evident in provinces where mantle-derived magmas have erupted during several epochs. For example, Hausel (1996) reported several mantle-related magmatic events in the Wyoming craton. Erlich et al. (1989) also summarized the repetition of worldwide events.

Geomorphology

Serpentinization of kimberlite and olivine lamproite favors relatively rapid weathering and erosion compared with the country rock. Depending on the intensity of weathering, silicification, and erosional history, different geomorphic patterns may develop over kimberlites and lamproites. Many stable cratonic regions are relatively flat with mature topography, and the basement rock is generally hard and resistant gneiss or granite. In these terranes, kimberlite and olivine lamproite may hide by blending into the surrounding topography.

Most kimberlites in semiarid environments erode at a faster rate than the country rock; thus the intrusive will blend into the surrounding topography. A few kimberlites have been expressed as slight mounds, such as the Kimberley pipe in South Africa. Nearby pipes may be expressed by slight depressions, such as the Wesselton pipe. Others may produce a subtle modification of the drainage pattern over the intrusive (Mannard 1968). But in most cases, the subtle geomorphic expressions produced by the intrusives are difficult to recognize.

In subarctic environments where glaciation has scoured the landscape, some kimberlites form noticeable depressions that may be filled with water and produce circular to elliptical lakes. Some of the diamond deposits discovered during the 1990s and in the twenty-first century in the Northwest Territories of Canada are associated with shallow lakes.

Many kimberlites in mature topography, such as those in northeastern Kansas, lie buried under a thin soil layer. A few are expressed as depressions (Figure 15.3), and small reservoirs or stock ponds may be constructed over kimberlite to take advantage of the natural depression. In mature topography where such depressions are evident, or a roughly circular to elliptical xenolithic block of rock is out of place with no apparent explanation, these may be described as cryptovolcanic structures, outliers, circular grabens, and even impact structures. The Winkler kimberlite in eastern Kansas, formed a distinct circular depression, which was initially misclassified as an impact crater (the Winkler crater)(Brookins 1970).

The first two kimberlites found in Wyoming in the early 1960s were emplaced in 1.4 Ga granitic country rock. The two elliptical features contained small blocks of Ordovician and Silurian sandstone and limestone that were clearly out of place and scattered throughout a small area of a few hundred square meters. The first explanation for these features was that they were "piston grabens," and they were termed "lower Paleozoic outliers" (see McCallum and Mabarak 1976).

In the Wyoming craton, where portions of the craton have been modified by the Laramide orogeny (a brittle mountain-forming event), many kimberlites in the mountain terrain are buried by alluvium shed from the surrounding country rock. Others form indistinct outcrops of kimberlite. And a few kimberlite dikes in the Iron Mountain district lie within saddles between distinct high walls of granite as a result of weak silicification of the country rock.

The weathering effects on most kimberlite in semiarid environments are noticeable. A cap of intensely oxidized kimberlitic "yellow ground," similar in appearance to limonite-rich gossan, may blanket the kimberlite. Lying beneath the yellow ground may be a layer of "blue ground." The blue ground formed of a carbonated montmorillonite clay-rich soil is produced from the weathering of the serpentine-rich kimberlite. Quite commonly, blue ground occurs at the surface with no noticeable expression of yellow ground.

a

b

FIGURE 15.3 Aerial views, kimberlite in Kansas producing distinct depressions. (a) The Winkler kimberlite forms a prominent circular depression that was originally misinterpreted as an impact feature. (b) The Fancy Creek kimberlite also forms a distinct depression and was a favorable site for construction of a small stock pond.

Many of the Colorado–Wyoming State Line kimberlites have deeply eroded profiles. Resistivity and seismic sounding over many of these intrusives produced profiles suggestive of a weathering depth of 25 to 38 m (Hausel, McCallum, and Woodzick 1979). In other words, some of these kimberlites are covered with a thick blanket of blue ground (Figure 15.4). The blue ground provides favorable soil for burrowing animals (including rattlesnakes) and is also favorable for mining. For example, mining costs at the Kelsey Lake open pit diamond mine in Colorado are relatively low, as the ore is easily scraped

FIGURE 15.4 Exposed DK pipe; Richard Kucera examining the intrusive breccia (photo by W.D. Hausel, 1995)

and no blasting has been necessary as of 1998 (Howard Coopersmith, personal communication, 1998).

In the exploration for the source of a major kimberlitic indicator mineral anomaly in the greater Green River basin of southwestern Wyoming (McCandless, Nash, and Hausel 1995), a group of small pipes known as the DK pipes were found near Cedar Mountain. The pipe breccias were buried, and the diatremes were outlined by abundant rounded boulder xenoliths of quartizite, granite, and gneiss (Figure 15.5). These xenoliths give an impression of small, isolated remnants of a paleoplacer, because the boulders are rounded as if transported by streams and are concentrated over the pipes. The adjacent surrounding terrane is formed primarily of Tertiary siltstones and claystones.

Digging into the "pseudo-paleoplacers" revealed blue ground breccia similar to kimberlite (Richard Kucera, personal communication, 1997). The early "dry diggings" over the kimberlites along the Orange River in southern Africa had a similar appearance (see Bruton 1978).

Unlike kimberlite, most lamproites (unless olivine rich) produce topographic highs. Many phlogopite-leucite lamproites are resistant to erosion and stand out as positive features. This topographic expression results in many lamproites being preserved as small volcanic cones, plugs, and/or flows that rise above the surrounding topography. Examples are seen in the Leucite Hills, Wyoming, the Missouri Breaks lamproites and lamprophyres in eastern Montana in the Wyoming craton, and the Noonkabob and Ellendale fields in the Kimberley craton of Western Australia.

However, most diamondiferous lamproites are olivine rich. Like kimberlite they are serpentinized, and they tend to erode rapidly and blend into the surrounding topography (Figure 15.6). Thus, any exploration program in a lamproite field should use phlogopite leucite lamproites as a guide that may lead to nearby, hidden olivine lamproites.

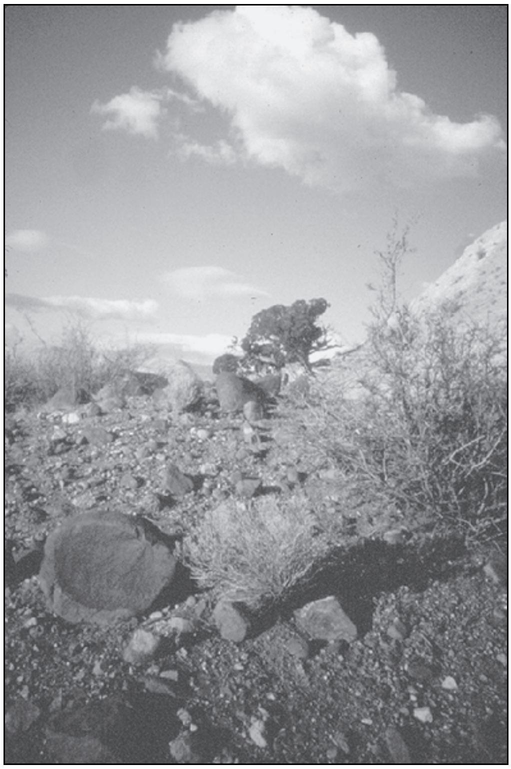

FIGURE 15.5 The DK pipe in southwestern Wyoming, expressed as a boulder field that might be misinterpreted as a paleoplacer (photo by W.D. Hausel, 1995)

FIGURE 15.6 The hidden Ellendale pipe, a tuffaceous olivine-phlogopite lamproite exposed in a trench. In the background behind the buses is a well-preserved phlogopite-leucite lamproite that forms a positive feature (photo by W.D. Hausel, 1986)

Stream Sediments

The rarity of diamonds, their ability to resist abrasion during stream transport, and the apparently enormous distances of stream transport of some gem-quality diamonds generally make diamond a poor tracer mineral for pinpointing in situ diamond deposits. In some cases, a proximal source may be indicated where diamonds are found in restricted drainage basins, or where large, unbroken, fragile, industrial diamonds with no impact marks are found in high-energy streams, such as the 32.99 carat diamond (and others) reported by Kopf, Hurlbut, and Koivula (1990) in Hayfork Creek in northern California. In other cases, relatively restricted concentrations of diamonds in alluvial deposits may indicate a nearby diamond source. For instance, the Argyle diamondiferous lamproite was discovered by following a 20 km trail of diamonds in Smoke Creek west of Lake Argyle, and some diamonds have been found in the Ord River as far as 160 km downstream from the Argyle (Gregory and White 1989). The diamond concentrations in Smoke Creek near the Argyle pipe were rich enough for commercial mining. Samples collected in the drainage also contained minor chromite, but the diamonds alone provided a clear trail to this deposit (Jaques, Lewis, and Smith 1986).

In most cases diamonds are poor tracer minerals. Diamond-bearing alluvium of the Sewa deposits of Sierra Leone lie 160 km downstream from the Yengema and Tongo kimberlites (Gregory and White 1989). Many diamonds found in beach placers along the coasts of South Africa, Nambia, and Angola are interpreted to have originated in the Orange River and its tributaries from the Kimberley region of South Africa, 800 to 1,600 km to the east. It would have been an enormous task to search so large an area for the source of these coastal diamonds; luckily the location of the source rocks was already known.

Thus tracer minerals used to search for in situ diamond deposits need to have some important qualities. They need to be for practical purposes unique to diamondiferous host rocks; they need to be relatively easy to concentrate and identify, and they need to be destroyed by stream abrasion over relatively short stream-transport distances. Some of the unique, mantle-derived minerals that have relatively high specific gravity, such as pyrope garnet, chromian diopside, chromian enstatite, picroilmenite, and chromite, are readily abraded during stream transport and make good tracer minerals for diamond deposits.

Stream-sediment sampling for such tracer minerals has been used successfully to locate diamondiferous kimberlite and lamproite in a number of regions around the world. As an exploration tool, this method has proven to be successful, although it is labor intensive and time consuming. Additionally, the extent of dispersion of the unique indicator minerals usually depends on climate and geomorphic conditions.

In the subarctic to glacial regions of Siberia, Canada, and the Great Lakes region of the United States, indicator-mineral dispersion patterns are commonly modified by glacial events. According to Mannard (1968), tracer minerals from kimberlite may have large transport distances in these environments. Maximum transport distances for three of the classic kimberlitic indicator minerals in the subglacial environment of Siberia were suggested to be 48 km for chromian diopside, 150 to 200 km for pyrope garnet, and more than >200 km for picroilmenite. Even with such wide dispersal, these indicator minerals were successfully used to find kimberlites in Siberia and Canada.

Satterly (1971) reported that pyrope garnet and picroilmenite showed little decrease in quantity approximately 1.6 km downstream from the Zarnitsa pipe in Siberia. Even so, the grain size of the minerals decreased to a fraction of a millimeter in this short distance owing to abrasion. Another mineral that is sometimes used, olivine, seldom persisted more than 5 to 6 km from the pipe, and chromian diopside barely persisted beyond a few hundred yards. Afanasev, Varlamov, and Garanin (1984) report that olivine appeared to have been transported as much as 100 km downstream from the Upper Muna kimberlite field.

Laboratory tests designed to simulate the flow of the average Yakut streams showed that pyrope and picroilmenite would be reduced by 90% of their original mass within a distance of 155 km. Fipke, Gurney, and Moore (1995) reported Cr-diopsides were present in alluvium for distances up to 50 km downstream from the Blackfoot diatreme in southeast British Columbia. In the Northwest Territory, Canada, indicator minerals appear to have been transported by glaciers and preserved for distances of 600 km from the Lac de Gras area.

In semiarid regions of the western United States, disaggregation of tracer minerals during stream transport appears to be much more efficient. Leighton and McCallum (1979) report that kimberlitic indicator minerals dispersed downstream from known kimberlites in the Colorado–Wyoming State Line district survived only 0.3 km (chromian diopside), 2.4 km (pyrope), and 4 km (picroilmenite) before complete disaggregation. These observations were similar to those reported by Mannard (1968) in Africa.

While mapping in the Iron Mountain district, Wyoming, Hausel (personal field notes, 1999) found competent kimberlite cobbles as much as 0.9 km downstream from known kimberlite outcrops. Any indicator minerals dispersed from the cobbles would increase the maximum transport distance of the minerals from the source kimberlite. This implies that the maximum transport distance of kimberlitic indicator minerals in semiarid environments, such as those in the Rocky Mountains, is probably on the order of 3.6 to 6 km. Even so, the presence of the classic kimberlitic indicator minerals in stream-sediment sample concentrates in this region would indicate a nearby source.

INDICATOR MINERALS

Exploration for kimberlite, lamproite, and lamprophyre is facilitated by stream-sediment samples collected to search for unique indicator (tracer) minerals derived from these rocks (Table 15.1). Similar tracer minerals will probably be developed for other types of diamond deposits. For example, tracer minerals used to search for Beni Bousara-type diamondiferous layered intrusives includes some typical kimberlitic indicator minerals, as the graphitized diamonds found in this deposit are closely associated with garnet pyroxenites with compositions similar to eclogite. The geologic environments and tectonic settings of kimberlite and Beni Bousara-type complexes are different, and thus where kimberlitic indicator minerals are found along coastal margins, such as northern Africa or southern Spain, a search for the source of the indicator minerals should consider the geologic environment.

Exploration for high-pressure (also referred to as “ultrahigh-pressure”) metamorphic diamond deposits would require evaluation of the proper geologic environments and tectonic setting, i.e., blueschist facies. The set of indicator minerals used to search for these types of deposits may include coesite, K-clinopyroxene, Al-sphene, phengite, and grossular-pyrope garnet. Owing to the low specific gravity of coesite and phengite, these two minerals would rarely be of value as indicator minerals. However, the search for grossular-pyrope garnet and K-clinopyroxene may provide useful guides in blueschist environments. The search for other unusual types of diamond deposits, such as Popigay-type deposits, would lead one to look for ring-dike complexes with anomalous nickel and chromium and lonsdaleite.

TABLE 15.1 Indicator minerals for diamond deposits

Type of Deposit	Geologic Environment	Indicator Minerals
Ultrahigh pressure metamorphics	Blueschist facies	Coesite, K-clinopyroxene, Al-sphene, phengite, grossular-pyrope
Popigay	Structurally controlled ring structures	Lonsdaleite, Chaoite
Kimberlite	Archean cratons and craton margins	Pyrope, Cr-diopside, Cr-enstatite, picroilmenite, chromite
Lamproites	Cratons and craton margins Proterozoic mobile belts	Chromite

For decades, geologists and prospectors have successfully used a suite of tracer minerals to search for kimberlite. Although these minerals are not restricted to kimberlite, for the most part they are derived from mantle xenoliths trapped in kimberlite and related rocks (a few of these minerals, such as some picroilmenites, are thought to crystallize from the kimberlitic magma). Thus a few nonkimberlitic intrusives originating in the upper mantle may contain these kimberlitic indicator minerals. The chemistry of many indicator minerals derived from diamondiferous kimberlite and lamproite will sometimes contain a unique enrichment of trace metals derived from the extremely high-pressure conditions related to their formation.

When recovered from the heavy mineral concentrates in stream-sediment samples, the kimberlitic minerals are identified in a lab, plotted on a map, and then later traced upstream to the source area. Ideally, the tracer mineral trail is followed upstream by collecting additional stream-sediment samples until samples are found that contain a noticeable increase in indicator minerals, followed by samples with no indicator minerals. At this point, the region between the two sets of samples is searched, both in the drainage and along the slopes above the drainage.

Although the process seems relatively straightforward, in practice it can be complex owing to terrain effects and past erosional and geologic history, which may have greatly modified the terrain since the intrusion of the pipes or dikes. Additionally, laboratory reports on the geochemistry and mineralogy of the follow-up samples require time, because the samples must be concentrated and examined mineralogically and then geochemically, which may require weeks or months. More often than not the source of the indicator minerals is not found, and even when it is, months and sometimes years of diligent prospecting may be expended.

Many of the classic kimberlitic indicator minerals are rare to absent in lamproite. Thus lamproite exploration requires the use of particular tracer minerals that are unique to that rock suite.

In any exploration program, stream-sediment sample concentrates should be examined for indicator minerals, and an attempt should be made to differentiate diamond-stability-field indicators from graphite-stability-field indicators. The stream-sediment sample concentrates should also be examined for other gemstones and precious metals unrelated to diamond deposits, as they can lead to the discovery of other valuable mineral resources. For example, several stream-sediment samples collected in the Wyoming craton in the search for kimberlite in the 1980s contained gold, ruby, sapphire, and other gemstones (Hausel, Sutherland, and Gregory 1988). Near some of the ruby-bearing samples, a corundum-cordierite gneiss was found that contained gem-quality ruby, sapphire, and iolite (Hausel and Sutherland 2000). One of the more unexpected ancillary discoveries in the search for diamond was a world-class nickel deposit in Labrador in the late 1980s (Richard Garnet, personal communication, 1999).

Kimberlite Indicator Minerals

Common tracer minerals include Mg-ilmenite (picroilmenite), pyrope garnet, Cr-diopside, and chromite. These minerals are part of a unique suite of minerals commonly associated with kimberlite that includes Cr-pyrope garnet (G9), subcalcic Cr-pyrope (G10), Mg-almandine (G5), G3 and G6 garnet, Cr-rich spinel, megacrystal/macrocrystal Ti-Cr pyrope (G1 and G2), Mg-Cr-ilmenite, Cr-diopside, and Cr-enstatite.

Picroilmenite. Kimberlitic ilmenite is an important constituent of kimberlite worldwide. In some kimberlites, ilmenite may be relatively common and form 1 to 10 wt. % of the kimberlite.

Ilmenite ($FeTiO_3$) is found in several geologic environments and associated with several rock types. However, unlike ilmenite in most other rock types, many kimberlitic ilmenites contain considerable geikielite ($MgTiO_3$) in solid solution. The ilmenites also have anomalous chromium such that ilmenites associated with kimberlite form a solid solution between geikielite and ferroilmenite that is referred to as picroilmenite. Other ilmenites recovered from kimberlite may contain considerable manganese; and are termed pyrophanite ($MnTiO_3$).

Because of its high specific gravity (4.2), ilmenite is recovered with other heavy minerals in black sand concentrates during stream-sediment sampling. Ilmenite forms black to gray-brown, submetallic to metallic, weakly magnetic to nonmagnetic, rounded to subrounded grains with conchoidal fracture. The ilmenite will produce a black streak when scratched on a streak plate. Picroilmenite's unique Cr-Mg-rich chemistry allows identification of suspect picroilmenites relatively quickly by use of an electron microprobe.

Mitchell (1986) reported that three types of ilmenite were found in kimberlite.

- Megacrysts—large, rounded, single polycrystalline megacrysts and macrocrysts
- primary groundmass—euhedral to anhedral primary groundmass ilmenite
- intergrowths—rounded to subhedral inclusions of ilmenite in olivine and/or phlogopite macrocrysts, lamellar intergrowths with clinopyroxene and less common orthopyroxene, ilmenite-spinel intergrowths, ilmenite-ilmenite intergrowths, ilmenite-rutile intergrowths, and ilmenite-perovskite intergrowths.

The amount of ilmenite varies considerably from one kimberlite to another. Some kimberlites in the Iron Mountain district, Wyoming, and in Chicken Park, Colorado, contain as much as 5% to 10% ilmenite. The Monastery, Morkoka, and Mir kimberlite pipes are also enriched in ilmenite. However, other kimberlites such as the Peuyuk, Finsch, and Koffiefontein appear to be devoid of ilmenite.

Picroilmenite compositions are reported in the following approximate ranges: TiO_2 (35%–56%), Al_2O_3 (0%–1.0%), Cr_2O_3 (0.5%–11%), Fe_2O_3 (3%–32%), FeO (11%–30%), MnO (0%–0.5%), and MgO (3%–23%). Some manganese-rich kimberlitic ilmenites are reported to contain TiO_2 (48%–54%), Al_2O_3 (0%–0.8%), Cr_2O_3 (0%–1.3%), Fe_2O_3 (0.7%–11.7%), FeO (21.5%–41%), MnO (0.26%–21%), and MgO (0.18%–12.7%) (Mitchell 1986).

Large, megacrystal picroilmenites found in kimberlite form solid solutions between ilmenite, geikielite, and hematite. Chemically, these picroilmenites contain significant amounts of Cr, high Mg, relatively low Fe, minor Mn, and anomalous amounts of Ni (180–

1,360 ppm). Some rare megacrysts may also have significant Mn enrichment. In addition to Ni, ilmenites from kimberlite are characterized by enrichment in Nb, Ta, Zr, Hf, and V, compared with Mg- and Cr-poor ilmenites from mafic igneous rocks (Mitchell 1986).

Groundmass ilmenites show geochemical trends similar to those of the megacrysts. However, groundmass ilmenites may be slightly enriched in MgO (12%–24.3%) compared with the megacrysts. The remaining paragenetic varieties of ilmenite have similar chemical ranges that fall within the ranges of the megacrystal and groundmass compositions.

Proximal ilmenite grains recovered from stream sediments tend to have abundant fresh fractured and rounded surfaces that may be pitted. The mineral may also possess a thin, white to tan leucoxene coating. Some megacrystal ilmenites may be as much as 10 cm across, making them relatively easy to pick out of the kimberlitic blue ground. Distal ilmenites lack leucoxene coatings owing to abrasion, and the grain size of the mineral will rapidly decrease with greater distance from the source.

The durability of ilmenite during transport and weathering is exceeded only by that of the indicator minerals chrome spinel and diamond. Ilmenite transport distances of greater than 300 km have been reported in glacial terrains (Mannard 1968). Transport distances of tens of kilometers are typical in some alluvial systems, especially in temperate and humid climates. In arid climates, maximum transport distances may be only a few kilometers.

For a first approximation, kimberlitic ilmenite may be identified by its lack of magnetism, unlike the magnetic ferroilmenite commonly found in granites or anorthosites (Leighton and McCallum 1979). Many ilmenites containing <4% MgO are generally considered to be derived from the surrounding country rock (Fipke, Gurney, and Moore 1995), although kimberlitic ilmenites enriched in manganese usually also have <4% MgO. In terranes such as the Laramie anorthosite complex, where considerable titaniferous magnetite is found, verification of picroilmenite is complicated by abundant titaniferous magnetite in the sample concentrates. For this reason the magnetic portion of the concentrates is removed during sample processing. The titanomagnetites from the anorthosite typically have much lower MgO (0.01%–2.4%) and higher Fe (73%–89% FeO) than the kimberlitic ilmenites, in addition to being highly magnetic.

Ilmenites may also provide information on diamond preservation. Diamonds are susceptible to burning in the high-temperature, low-pressure, oxidizing environment that may be encountered by the intrusive near the earth's surface. In such environments, diamonds may partially or completely resorb producing graphite that is stable at the earth's surface, or they may burn producing CO and CO_2. The conditions required to burn diamonds are not extreme. During one test by W.D. Hausel (personal notes, 1985), a 0.1-carat industrial diamond was burned in the oxidizing flame of a Bunsen burner in less than 2 min. Green (1981) notes that diamond will burn between the temperatures of 700° and 900°C.

Gurney and Moore (1991) suggested that ilmenites provide data on the oxidizing conditions in the host kimberlite. Ilmenites with low Fe^{3+}/Fe^{2+} ratios are associated with increased diamond preservation; higher ratios correspond to diamond resorption. It has been suggested that MgO depletion in ilmenite can provide a measure of the oxidizing conditions during kimberlite emplacement. Thus highly oxidizing conditions are thought to be associated with low MgO content in the ilmenites (see Chapter 16, Predicting Diamond Content), although exceptions occur.

TABLE 15.2 Classification of kimberlitic garnets (Dawson and Stephens 1975)

Group	Name	TiO_2	Cr_2O_3	FeO	MgO	CaO
1	Ti-pyrope	0.6	1.3	9.3	20.0	
2	High-Ti pyrope	1.1	0.9	9.8	20.3	
3	Ca-pyrope-almandine	0.3	0.3	16.5	13.4	6.5
4	Ti-Ca-Mg Almandine	0.9	0.08	18.0	10.0	9.4
5	Mg-almandine	0.05	0.03	28.0	8.0	2.5
6	Pyrope-grossular-almandine	0.25	0.25	10.8	10.0	15.0
7	Fe-Mg-uvarovite-grossular	0.3	11.5	5.0	8.5	21.5
8	Fe-Mg-grossular	0.25	0.04	7.0	5.0	25.0
9	Cr-pyrope	0.2	3.5	8.0	20.0	5.0
10	Sub-Ca, Cr-pyrope	0.04	8.0	6.0	23.0	2.0
11	Ti-uvarovite pyrope	0.5	9.5	7.5	16.0	10.0
12	Knorringitic uvarovite-pyrope	0.18	16.0	7.5	15.5	9.5

Garnet. Garnet is one of the more important tracer minerals used in the search for kimberlite, because it is relatively common in many kimberlites and some varieties are recognizable in the field. Garnets are unfortunately rare in lamproite and lamprophyre (See Table 15.2).

Garnets are found in kimberlite as megacrysts and xenocrysts, and as components of xenoliths and cognate nodules trapped in the kimberlite magma. Five groups of xenoliths, cognate nodules, and megacrysts appear to be the dominant source of the garnet xenocrysts found in kimberlite. They are (1) peridotites (wehrlites, lherzolites, harzburgites, dunites), (2) garnet pyroxenites (garnet xenocrysts from pyroxenites generally cannot be distinguished from garnets from peridotites), (3) eclogites, (4) some crustal xenoliths, and (5) Cr-poor garnet megacrysts.

Garnet xenocrysts recovered from kimberlite include yellow-orange Na-Ti eclogitic pyrope-almandine garnet; slight reddish, purple-red, gray-lavender, and purple to lavender peridotitic pyrope; green knorringite-uvarovite pyropes; and pink, reddish-brown and orange-red almandine garnet.

Locally, some garnets may be transparent and of gem quality. Other minerals of possible facet grade include chromian diopside and enstatite.

Well-formed garnet crystals are notably absent in the kimberlitic pyrope garnet suite because the garnets tend to round during transport to the earth's surface. Pyrope has no cleavage, has conchoidal fracture, and has a hardness of 7.25. According to Sinkankas (1964), pyrope has a specific gravity ranging from 3.65 to 3.84 and an index of refraction of 1.730 to 1.760. Almandine, a metamorphic garnet commonly found in abundance in amphibolite-grade metamorphic terranes in cratons, has an index of refraction greater than 1.760. Thus index of refraction can be used as a first approximation to differentiate these two garnets.

Many Cr-poor (<2% Cr_2O_3) garnets are assigned to an eclogitic (E-type) or a lower crustal source and are therefore xenocrysts. Cr-rich pyropes (>2% Cr_2O_3) are assigned to peridotites. Some of the more useful peridotitic garnets used in exploration are the G9 and

G10 (P-type) pyropes. They are the calcic-chrome pyropes (G9) derived from lherzolitic peridotites, and the subcalcic chrome pyrope (G10) derived from harzburgitic peridotites.

Garnets derived from harzburgite have geochemical signatures similar to diamond-inclusion garnets that are designated as G10. Thus, the presence of G10 garnets generally indicates that the host intrusive originated in the diamond stability region within the upper mantle. The stronger the G10 signature, the greater probability of finding diamonds as well as ore grade material. These garnets are slightly purplish-red, purple to lavender, to distinctly gray-lavender in color. Unfortunately, the garnets do not provide information on diamond preservation, quality, or size.

Peridotitic (P-type) garnet inclusions in diamond have higher Cr_2O_3 and lower CaO than garnets that crystallized in the graphite stability field. The majority of the P-type diamond-stability-field garnets (G10) are from harzburgites, although some G9 garnets derived from the disaggregation of lherzolites are also formed within the diamond stability field, but to a lesser extent. The P-type diamond-stability-field garnets typically have MgO ≥14 wt. %, and Cr_2O_3 ≥4 wt. %. Low CaO is also significant, and most diamond-stability-field garnets have CaO <5 wt. %. Some diamond-stability-field garnets of lherzolitic paragenesis contain higher CaO contents, but statistically they are less common. Diamond-stability-field garnets of eclogitic paragenesis generally contain Na_2O >0.7 wt. % (Gurney 1985).

Mg-almandine (G5) garnet has been used as a tracer mineral in the exploration for lamproite along with G9 garnets, whereas G10 garnets are rare to absent in lamproite. The Mg-almandines recovered from diamondiferous lamproite are pink to purple, or pale orange to brownish and tend to have a broken or anhedral form. Their composition ranges as follows: FeO (24.96%–29.94%), TiO_2 (<0.35%), Cr_2O_3 (<0.13%), MgO (5.26%–10.9%), and CaO (1.07%–5.66%). Although G5 garnets could be used in some regional heavy mineral prospecting programs to locate lamproite and kimberlite, G5 garnets do not discriminate between barren and diamondiferous sources. They are also easily mistaken for regional metamorphic garnets (Fipke 1994).

The Cr-poor megacrystal garnets form single, large, rounded, highly fractured, reddish-brown megacrysts. They tend to appear orange, similar to the eclogitic garnets, when they are found as small crystal fragments. Some megacrystal garnet recovered from the Schaffer kimberlites in Wyoming have been as large as 13 cm across. Similar megacrysts have also been found as intergrowths with ilmenite megacrysts in some kimberlites (Dawson and Stephens 1975). The megacrysts are distinct from the red to purple to grayish lavender P-type pyropes, and they have higher FeO and TiO_2 and lower Cr_2O_3 than lherzolite or harzburgitic garnets.

Some researchers interpret the megacrysts to represent phenocrysts in kimberlite formed at depths of 150 to 200 km. They are thought to have formed in a high-temperature magma by fractional crystallization at a depth where the kimberlite magma initially acquired xenoliths. Most researchers believe that they have no genetic link to diamond (Schulze 1995).

Kostrovitsky et al. (1997) reported that garnet megacrysts recovered from barren kimberlite contain relatively high mean FeO and low TiO_2 and Cr_2O_3 compared with diamondiferous kimberlites. On Mg-Fe-Ca diagrams, the compositions of megacrysts from diamondiferous kimberlites trend parallel to the Mg-Fe axis with relatively constant CaO, whereas those from the single barren kimberlite examined during their study trend parallel to the Mg-Ca axis at relatively constant Fe. The megacryst garnets with the highest MgO and Cr_2O_3 contents were compatible in composition with garnets from diamond-garnet aggregates.

Some pyrope-almandine garnets may also be derived from high-pressure schists or from tectonically transported mantle material. For example, almandine garnet (with <30% pyrope component) and acmitic omphacite are commonly associated with glaucophane schists and serpentinized peridotites in orogenic belts of low-temperature, high-pressure metamorphism (Carmichael, Turner, and Verhoogen 1974). Although such high-pressure metamorphic environments are not common in cratonic environments, they may occur in collisional zones.

Pyroxene. Many pyroxenes found in kimberlite are xenocrysts and represent the disaggregated products of xenoliths or cognate nodules. Others may represent phenocrysts.

Fipke, Gurney, and Moore (1995) concluded that emerald-green to bright-green Cr-diopside was a useful visual indicator mineral for the location of kimberlite and/or lamproite diatremes and that elevated K_2O in clinopyroxene provided encouragement for diamond potential. McCandless and Gurney (1989) report that clinopyroxenes from eclogite associated with diamond commonly are enriched in K_2O (>0.07 wt. %). The trace amounts of K_2O are imprinted on the mineral owing to the high-pressure conditions during genesis. Such high-pressure K_2O-enriched pyroxenes are unstable at shallow depths and tend to decompose and lose potassium. Thus when one selects samples for microprobe analysis, only those grains that appear fresh with no evidence of alteration should be selected.

Along with distinctive color, many pyroxenes in kimberlite have distinct cleavage and parting and produce prisms with square cross sections (orthopyroxene), or parallelograms with one inclined cleavage (clinopyroxene). Many megacrysts recovered from kimberlite occur as large rounded grains with a distinct white carbonate reaction rim. Rounding of the grains reflects partial assimilation of the nodule or grain by the kimberlitic magma. These megacrysts, when broken, will produce distinct prismatic crystals with cleavage parallel to the m{110} prism and parting parallel to the c{001} face. Their specific gravity averages 3.2, and thus pyroxenes are typically recovered with the black-sand suite in stream-sediment sample concentrates.

Chromian diopside, one of the more prominent pyroxenes, is an important tracer mineral used to search for kimberlite and some lamprophyres. It forms a distinct emerald-green crystal that has excellent cleavage and parting resulting in well-formed monoclinic cleavage fragments both in the kimberlite and in stream-sediment samples. Mitchell (1986) reports that megacrystal pyroxenes in kimberlite contain chrome contents of 0.08% to 2.9%. Chrome diopsides that correspond to possible diamondiferous kimberlite typically contain Na_2O (>0.5%) and Cr_2O_3 (>0.5%).

Chromian diopside forms under high pressure and temperature. The mineral is typically transported from the mantle in kimberlite, but other sources are possible. For instance, chromian diopside was encountered in serpentinized peridotites in the Hayfork region of northern California. Chromian diopside has also been found in mantle xenoliths in basalts in the Russian Far East.

Chromian diopside occurs in both megacrysts and groundmass in kimberlites, whereas orthopyroxenes are reported only as megacrysts or as intergrowths with clinopyroxene. Clinopyroxene is much less common in kimberlite than orthopyroxene.

Other pyroxenes include dark-green chromian enstatite, bottle-green diopside, and dark-brown orthopyroxene. Some of these pyroxenes will occur as monomineralic

TABLE 15.3 Classification of kimberlitic orthopyroxenes, showing mean chemical values (Stephens and Dawson 1977)

Group	Name	TiO_2	Cr_2O_3	FeO	MgO	CaO	Al_2O_3	Na_2O
1	Enstatite	0.02	0.29	4.5	36.2	4.8	0.8	0.08
2	Cr-Al enstatite	0.02	0.65	4.9	35.2	0.6	2.8	0.08
3	Na-Ca enstatite	0.14	0.32	5.31	33.8	1.42	1.45	0.43
4	Ti enstatite	0.14	0.25	5.98	34.5	0.71	0.93	0.17
5	High-Ti enstatite	0.19	0.10	8.64	32.2	1.07	1.06	0.22

rounded nodules with microcrystalline reaction rims as large as 15 cm in diameter. Others may occur as distinct orthorhombic to monoclinic crystals. Fipke, Gurney, and Moore (1995) report that enstatite associated with diamond is higher in MgO and lower in FeO, Al_2O_3, CaO, and Na_2O compared with other orthopyroxenes.

Stephens and Dawson (1977) recognized the following compositions and 15 groups of clinopyroxene and orthopyroxene in kimberlite. Recent research suggests that the compositional ranges for some of the pyroxene groups may need to be revised and extended.

Orthopyroxenes. The orthopyroxene group includes the following (Table 15.3).

Group 1 orthopyroxenes have very low TiO_2 (0%–0.08%) and moderate Al_2O_3 (0.09%–1.26%), Cr_2O_3 (0.11%–0.43%), and CaO (0.16%–0.83%). They also contain FeO (3.6%–5.9%), MgO (34.7%–37.9%), and Na_2O (0%–0.23%). The group is dominated by enstatite derived from garnet lherzolite.

Group 2 orthopyroxenes, Cr-Al enstatites, contain relatively large amounts of Al_2O_3 (1.63%–5.24%) and Cr_2O_3 (0.27%–0.95%), and they are characterized by low Na_2O (0%–0.18%) and TiO_2 (0%–0.01%). There is a positive mutual association between Al_2O_3, Cr_2O_3, and Na_2O. FeO (4.0%–6.7%), MgO (32.4%–36.5%), and CaO (0.11%–1.6%) are not diagnostic. The group is dominated by lherzolitic and harzburgitic enstatite devoid of aluminous phases (garnet and/or aluminous spinel).

Group 3 orthopyroxenes are Na-Ca enstatites with considerably more TiO_2 than enstatites of group 1 or 2. Typically, they have the following compositional ranges: TiO_2 (0%–0.22%), Al_2O_3 (1.0%–2.52%), Cr_2O_3 (0.19%–0.48%), FeO (4.18%–6.6%), CaO (1.32%–1.67%), MgO (31.7%–34.9%), and Na_2O (0.29%–0.47%). Many pyroxenes from this group are derived from undeformed and sheared garnet lherzolites.

Group 4 orthopyroxenes, Ti-enstatites, apart from having relatively high average TiO_2 (0.14%) are not chemically distinctive. They have been found in garnet lherzolite, garnet-olivine websterite, garnet pyroxenite, dunite, and chromite lherzolite xenoliths, and in megacrysts in kimberlite. Compositions range as follows: TiO_2 (0.03%–0.27%), Al_2O_3 (0.54%–2.0%), Cr_2O_3 (0.06%–0.56%), FeO (3.2%–7.55%), MgO (32.9%–36.6%), CaO (0.28%–1.18%), and Na_2O (0.03%–0.27%).

Group 5 orthopyroxenes, high-Ti bronzites, include orthopyroxenes from deformed and undeformed garnet lherzolite, garnet olivine pyroxenite, garnet pyroxenite, megacrysts from kimberlite, and inclusions in ilmenite and diamond. Compositions range as follows: TiO_2 (0.04%–0.35%), Al_2O_3 (0.7%–2.26%), Cr_2O_3 (0.0%–0.27%), FeO (6.7%–10.9%), MgO (30.4%–33.9%), CaO (0.63%–1.60%), and Na_2O (0.03%–0.34%).

TABLE 15.4 Classification of kimberlitic clinopyroxenes, showing mean chemical values (Stephens and Dawson 1977)

Group	Name	TiO_2	Cr_2O_3	FeO	MgO	CaO	Al_2O_3	Na_2O
1	Sub-Ca diopside	0.31	0.43	5.17	20.71	13.8	2.51	1.58
2	Diopside	0.26	0.71	4.16	16.91	18.44	2.69	1.78
3	Ti-Cr diopside	0.80	1.02	2.61	15.99	19.51	3.86	1.94
4	Low-Cr diopside	0.50	0.09	5.86	16.88	17.55	3.19	1.85
5	Cr diopside	0.09	1.45	2.02	16.8	20.66	2.50	1.68
6	Ureyitic diopside	0.27	2.99	2.37	15.19	17.94	3.14	3.11
7	High-ureyitic diopside	0.19	11.8	1.68	19.27	0.6	3.14	7.07
8	Jadeitic diopside	0.44	0.10	6.10	11.54	14.52	7.61	4.5
9	Omphacite	0.27	0.15	3.29	10.35	14.62	11.34	5.09
10	Diopsidic jadeite	0.22	0.02	2.42	6.36	10.19	16.87	7.64

Clinopyroxenes. Clinopyroxenes from kimberlitic sources have been subdivided into 10 groups by Stephens and Dawson (1977). They include the following (Table 15.4).

Group 1 clinopyroxenes are dull-green subcalcic diopsides with high MgO/CaO ratios. They are found in sheared garnet lherzolites and have been reported as inclusions in diamond. Compositions range as follows: TiO_2 (0%–0.46%), Al_2O_3 (0.7%–3.77%), Cr_2O_3 (0.13%–0.99%), FeO (1.6%–8.16%), MgO (18.9%–27%), CaO (9.3%–16.2%), and Na_2O (0.4%–2.13%).

Group 2 clinopyroxenes, diopsides, apart from their MgO/CaO ratios are similar to Group 1 clinopyroxenes. They have the following oxide variations: TiO_2 (0%–0.48%), Al_2O_3 (0.93%–6.06%), Cr_2O_3 (0.01%–1.46%), FeO (2.6%–5.9%), MgO (14.3%–20.8%), CaO (0.12%–3.15%) and Na_2O (0%–0.18%). They are found in garnet lherzolite, kimberlite, garnet-olivine pyroxenite, garnet pyroxenite, eclogite and diamondiferous eclogite, and as inclusions in diamond. Their bright-green color is similar to that of chromian diopside (see Group 5 below).

Group 3 clinopyroxenes are Ti-Cr diopsides with characteristically high TiO_2 (0.8%) and moderate Cr_2O_3 (1.02%). Their composition ranges as follows: TiO_2 (0.62%–1.07%), Al_2O_3 (0.36%–6.93%), Cr_2O_3 (0.13%–1.82%), FeO (1.6%–3.8%), MgO (14.6%–17%), CaO (16.1%–23.8%), and Na_2O (0%–3.74%). They are derived from garnet lherzolite and garnet pyroxenite, and they occur as groundmass grains in kimberlite.

Group 4 clinopyroxenes are similar to the Group 2 diopsides, although they contain much lower Cr_2O_3 and high TiO_2. The low Cr_2O_3 is notable because it is common in kimberlitic omphacite rather than diopside. The group is dominated by large, rounded, bottle-green diopside megacrysts from kimberlite. Many form intergrowths with ilmenite. This group of low-Cr diopsides contains TiO_2 (0.26%–0.82%), Al_2O_3 (0.86%–7.6%), Cr_2O_3 (0.0%–0.45%), FeO (4.8%–7.0%), MgO (11.6%–18.8%), CaO (15.75%–21.8%), and Na_2O (1.24%–2.49%).

The Group 5 clinopyroxene, emerald-green chromian diopside, is one of the more useful kimberlitic tracer minerals because of its distinctive color, excellent cleavage, and short stream-transport distance before complete disaggregation. Compositions for this mineral are reported to have the following ranges: TiO_2 (0%–0.45%), Al_2O_3 (0.92%–6.7%), Cr_2O_3 (0.2%–2.8%), FeO (1.1%–3.3%), MgO (13.3%–20.9%), CaO (16.1%–23.5%), and Na_2O (0.5%–3.83%). This group is dominated by pyroxenes from garnet lherzolites. Some are associated with eclogites containing chrome pyrope, and some are reported as diamond inclusions.

Compared with chromian diopside (Group 5), Group 6 clinopyroxenes, ureyitic diopsides, contain more Cr_2O_3 and Na_2O. They are found in garnet lherzolite, garnet-olivine pyroxenite, and chromite lherzolite, and as inclusions in diamond. Compositions range as follows: TiO_2 (0.02%–0.45%), Al_2O_3 (0.99%–8.0%), Cr_2O_3 (1.7%–6.15%), FeO (1.6%–3.2%), MgO (8.45%–17.2%), CaO (11.0%–23.86%), and Na_2O (2.13%–6.6%).

Group 7 clinopyroxenes, bright-green, high-ureyite diopsides, are also found as inclusions in some diamonds. Stephens and Dawson (1977) report the following compositions: TiO_2 (0.19%), Al_2O_3 (3.14%), Cr_2O_3 (11.8%), FeO (1.68%), MgO (19.27%), CaO (10.6%), and Na_2O (7.07%).

Group 8 clinopyroxenes, jadeitic diopsides, are found in garnet-olivine pyroxenites and eclogites, and in diamond as inclusions. Compositional ranges are reported as follows: TiO_2 (0.04%–0.87%), Al_2O_3 (4.4%–13.1%), Cr_2O_3 (0.0%–0.34%), FeO (4.7%–7.9%), MgO (6.56%–15.6%), CaO (9.13%–18.5%), and Na_2O (1.98%–8.8%).

Group 9 clinopyroxenes, omphacites, have higher Na_2O and Al_2O_3 but lower MgO and FeO than the Group 8 jadeitic pyroxenes. They are found in a variety of eclogites, in grospydites, and as diamond inclusions. Compositions range as follows: TiO_2 (0.04%–0.48%), Al_2O_3 (8.0%–17.8%), Cr_2O_3 (0.0%–0.92%), FeO (1.4%–5.55%), MgO (8.2%–12.1%), CaO (12.7%–18.23%), and Na_2O (3.3%–6.8%).

Group 10 clinopyroxene, diopsidic jadeites, are found in a variety of eclogites and in grospydites. Compositions range as follows: TiO_2 (0.03%–0.42%), Al_2O_3 (12.8%–19.26%), Cr_2O_3 (0.0%–0.1%), FeO (1.35%–5.4%), MgO (2.5%–8.84%), CaO (2.9%–12.94%), and Na_2O (6.0%–12.8%). They have high Na_2O and Al_2O_3 and a relatively low FeO and CaO.

Zircon. Zircon is an accessory mineral in many kimberlites. However, its abundance in granitic terranes limits its usefulness in diamond exploration. Owing to zircon's high specific gravity (4.6–4.7), resistance to abrasion, and fluorescence, zircon is recoverable by gravity methods and by fluorescence. Most granitic zircons form distinct euhedral yellowish-brown prisms capped by bipyramids with poorly developed cleavage. Kimberlitic zircons lack euhedral morphologies, are typically rounded to subrounded with a slightly pitted or frosted surface, and range from colorless to gray, yellowish to honey colored, and reddish brown to reddish-violet (Kresten, Fels, and Berggren 1975; Krasnobayev 1979). Fipke, Gurney, and Moore (1995) also report tiny (<0.5 mm), rounded purple to brown, nonfluorescent zircons and pink to honey colored, orange fluorescent (in ultraviolet light) zircons in kimberlite and lamproite. In some, concentric zones with light-colored cores are enclosed by dark-purple to pink overgrowths. They typically have glassy to frosted textures similar to E- and P-type garnets. Kimberlitic zircons also show one or more directions of perfect cleavage, unlike their granitic counterparts.

Picroilmenite, phlogopite, chromite, chrome diopside, and diamond have been reported as inclusions in zircon, suggesting that kimberlitic zircons are accessory to rocks of peridotitic paragenesis (Butkunov et al. 1980; Kresten, Fels, and Berggren 1975). Zircons have also been found as inclusions in diamond in association with olivine, ilmenite, phlogopite, and clinopyroxene. Meyer and Svisero (1973) reported one zircon inclusion in diamond that had the following composition: ZrO_2 (69.7%), SiO_2 (31.1%), TiO_2 (0.03%), CaO (0.01%), and MnO (0.02%).

More than one generation of zircon derived from mantle xenoliths is found in kimberlite and lamproite. Near the margins of cratons, xenolithic zircons can be used to determine the age of consolidation of the basement sampled by the host kimberlite or lamproite and determine if the intrusive originated from an Archon or Proterozoic accretionary terrane.

In addition to microzircon, some megacrystal zircon has been reported in Group I kimberlites. Where found, the megacrysts may be several centimeters across and are honey colored to colorless. Some zircons might have limited use as tracer minerals, but more research is necessary to positively identify zircons of kimberlitic affinity.

Tourmaline. Tourmalines are found in both kimberlite and lamproite. Minor amounts of spheroidal, light translucent to opaque brown, dravitic tourmalines have been recovered from kimberlite. Rounded tourmalines with elevated K_2O and TiO_2 were proposed by Fipke (1994) to be the result of alteration of diamondiferous eclogites. A large percentage of tourmalines recovered from four diamondiferous lamproites and one diamondiferous kimberlite showed elevated K_2O and TiO_2 (Fipke 1994).

Chromite. Chrome spinels recovered during stream-sediment sampling programs were analyzed by electron microprobe in an attempt to differentiate kimberlitic and lamproitic chromites from other potential source rocks found in cratons, such as komatiites, ultramafic massifs, alpine-type ultrabasites, basalts, and ophiolites. Chrome spinel compositions were categorized in the Russian Far East in an attempt to identify potential host rocks. The geochemistry of the chrome spinels were plotted on Cr_2O_3-TiO_2 and Cr_2O_3-Al_2O_3 diagrams and compared with known kimberlitic and lamproitic chromites. Chromites of probable diamond association contained high Cr_2O_3 (53%–68%) but low Al_2O_3 (5%–15%) and TiO_2 (0%–0.3%). Those of probable kimberlite genesis were described as having high Cr_2O_3 (20%–50%), a wide range in Al_2O_3 (18%–50%), and relatively low TiO_2 (0%–0.8%). The chromites of probable lamproitic origin typically contained high Cr_2O_3 (40%–50%), low to moderate Al_2O_3 (6%–21%), and low to high TiO_2 (1%–4%) (Romashkin 1997, figure 3 of that publication).

The chemistry of some spinels are used to assess the probability of diamond derived from the disaggregation of chromite harzburgites (see Chapter 16, Predicting Diamond Content).

For example, the MgO and Cr_2O_3 content of kimberlitic and lamproitic spinels (chromite) has been examined. Chromites with compositions similar to those of diamond-inclusion chromites show a distinct range of compositions with high Cr_2O_3 and high MgO content. Those of the diamond association typically have unique compositions of Cr_2O_3 (57%–69%; the majority have >61%) and MgO (10%–19%).

Olivine. Olivine is generally not considered as an indicator mineral in the search for kimberlite, because most kimberlitic olivines have altered to serpentine and have been completely destroyed. However, Atkinson (1989) reports that diamond-inclusion olivine contains anomalous Cr_2O_3, and olivine has been used as a tracer mineral in prospecting for kimberlite in Yakutia.

Lamproite Indicator Minerals

Kimberlitic indicator minerals are uncommon to rare in olivine and phlogopite-leucite lamproites. Megacrysts characteristic of some kimberlites (Ti-pyrope, picroilmenite, subcalcic diopside, and bronzite) have not been reported in lamproite. Cr-diopside and enstatite xenocrysts are rare, and when present their compositions are identical to minerals found in lherzolites. Because of the rarity of kimberlitic indicator minerals in lamproite, larger sediment samples are required in the search for lamproite, and a suite of lamproitic minerals is used in conjunction with kimberlitic minerals.

Stream-sediment sampling was successfully used to locate lamproite in West Kimberley, Australia (Jaques, Lewis, and Smith 1986; Atkinson 1989). Chromite was shown to be the most useful indicator mineral for olivine lamproite in the Ellendale field, Australia, because of its relatively high abundance and distinct composition (Mitchell and Bergman 1991). Diamond-inclusion chromites contain TiO_2 (<0.8%), whereas Cr-Ti chromites contain 0.8% or greater TiO_2 and plot within the high Cr-Ti field.

Cognate spinels with high Cr/(Cr + Al) and Mg are regarded as one of the most useful diamond indicator minerals in lamproite. The diamond-facies chrome spinel is rich in Cr and Mg (Cr_2O_3 ≥61 wt. %, MgO 11 wt. %) and poor in TiO_2 (<0.5 wt. %). Other potential lamproitic indicator minerals include Ti-rich phlogopite, K-Ti richterite, low-Al diopside, olivine, perovskite, armalcolite, priderite, wadeite, and shcherbakovite. The latter four minerals and K-Ti richterite are very rare but highly diagnostic of lamproite.

According to Fipke (1994), chromite, tourmaline, zircon, Mg-almandine (G5), and ilmenites were consistently abundant in the heavy-mineral concentrates in bulk rock samples taken from the olivine lamproite deposits of Prairie Creek, Arkansas; Argyle and Ellendale 4, Australia; and Jack, British Columbia. The classic kimberlitic indicator minerals were found only in trace amounts in concentrates recovered from the Argyle and Jack lamproites. Thus these diamond-bearing lamproites would probably not have been detected had only traditional sampling methods for kimberlitic indicator minerals been used.

Garnet. Garnets are uncommon to rare in lamproite. Those found typically have pyrope to almandine-pyrope compositions. Using the classification of Dawson and Stephens (1975), these garnets belong to groups G1, G3, G5, and G9. Most of the pyropes are poor in TiO_2 (<0.4 wt. %), and all are CaO saturated, and may contain Cr_2O_3 (2–10 wt. %).

Mg-almandines (G5) are morphologically indistinguishable from most metamorphic almandines and therefore have limited use in the search for diamond deposits in amphibolite-grade metamorphic terranes. These Mg-almandine garnets, as well as Mn-ilmenites and picroilmenites, may occur in barren lamproite or basic alkaline diatremes as well as in diamondiferous lamproite (Fipke 1994).

The subcalcic, chrome-rich G10 pyropes, which are considered as indicators of diamondiferous harzburgite, have not been found in the Prairie Creek or Argyle diamondiferous lamproites. Such garnets are also extremely rare in the Ellendale 7 and 9 lamproites in Western Australia (Lucas et al. 1989), and only one has been found as an inclusion in an Ellendale diamond (Hall and Smith 1985). Many of the mineral phases found in lamproitic xenoliths are Mg rich and Ti poor, and they yield equilibration temperatures estimated at 880° to 1,100°C (Jaques, Lewis, and Smith 1986).

Olivine. Iron-poor (Fo_{84}–Fo_{92}), low-calcium (CaO < 0.5%) chromium olivine is a potential indicator mineral for lamproite and kimberlite. This mineral may also characterized by high nickel (0.2–0.5 wt. %) content (Romashkin 1997). Jaques et al. (1989) described a small sample of olivines from Argyle and Ellendale 4 that had compositions within the range reported for diamond-inclusion olivines.

Although olivine has not been used extensively in the search for lamproite, it could be an important indicator mineral for diamondiferous lamproite. For example, diamond-inclusion olivines from kimberlite and lamproite plot within a distinct field with high forsterite content (MgO [90.2%–96.3%] and trace chrome (Cr_2O_3 [0%–0.2%]) (Fipke, Gurney, and Moore 1995). Many researchers tend to overlook olivine because it is difficult to find in highly serpentinized kimberlite and olivine lamproite. In particular, diamondiferous olivine lamproites are generally highly serpentinized and little olivine remains. However, lamproites such as Black Rock and Table Mountain in the Leucite Hills, Wyoming, contain relatively fresh olivine. More than 13,000 carats of fresh olivine (including gem-quality peridot) were recovered from just two anthills near the Black Rock lamproite (Hausel and Sutherland 2000). Olivine megacrysts as large as 1 cm are found in the lamproite.

Rutile. Fipke, Gurney, and Moore (1995) noted that rounded, brown and gray to black, translucent to opaque, Nb-rich rutile occurred as a minor to abundant accessory mineral in lamproite. Both rutile and Nb-bearing rutile have also been documented as relatively common accessory minerals in kimberlite (Fipke 1994; Haggarty 1991). Diamond inclusion rutiles in lamproitic diamonds are reported to contain 0.13% to 1.77% Nb_2O_3.

In most circumstances, rutile is not sufficiently abundant to be used as a tracer mineral. However, its relatively high specific gravity (4.2) suggests that Nb-rutile could offer potential as a tracer mineral for lamproite.

Ilmenite. Common ilmenite (MgO < 3%, MnO < 1%), high Mn-ilmenite (MnO >1%) and picroilmenite were recovered in trace to minor amounts from all diamond-bearing lamproites tested by Fipke (1994). Moderate amounts were also recovered from the barren Smoky Butte lamproitic plug in Montana. Fipke, Gurney, and Moore (1995) also report that trace to minor amounts of high-Mn ilmenite and high-Cr ilmenite were recovered from diamondiferous and barren lamproites. The unusual saucer-shaped or broken morphology of these ilmenites was used to distinguish those of lamproitic paragenesis from those of a regional origin. However, the relatively low abundance of ilmenite in lamproite suggests that it would not be a good indicator mineral in exploration.

Mitchell and Bergman (1991) report that ilmenites are not a characteristic mineral of most lamproites. Representative ilmenite compositions show groundmass ilmenites have significant MgO (1–1.7 wt. %) and MnO (1–2 wt. %). However, these compositions are similar to those of ilmenites in a wide variety of rocks including leucitites, and hence they are not diagnostic of lamproite. The ilmenites are unlike those in kimberlite and are on the average both Mg and Cr poor.

Zircon. Pink and mauve rounded zircons are present in most diatremes (i.e., kimberlite, lamproite, and alkali basalts). Such zircons recovered from lamproite yield ages as old as 3.3 Ga, suggesting that they are derived from cratons and/or overlying clastic units intruded by the lamproite (Fipke, Gurney, and Moore 1995).

Tourmaline. Minor to abundant spheroidal, light translucent to opaque brown, dravitic tourmalines were recovered from all diamondiferous lamproites tested by Fipke (1994) and were absent from barren lamproite. Minor amounts of similar tourmalines have also been recovered from kimberlite as well as from a lamprophyre upstream from an alluvial diamond placer in Brazil. The tourmalines are unzoned and have glassy to frosted textures typical of kimberlitic P and E garnets, and they generally lack abraded surface textures (Fipke, Gurney, and Moore 1995).

Rounded tourmalines with elevated K_2O and TiO_2 were proposed by Fipke (1994) to be the result of alteration of diamondiferous eclogite. A large percentage of tourmalines recovered from four diamondiferous lamproites and one diamondiferous kimberlite showed elevated K_2O and TiO_2. Thus, these tourmalines may be considered as tracer minerals for both kimberlite and lamproite, and they potentially may be indicators of a source area within diamond-stability-field eclogite. Tourmalines derived from lamproite, as well as titaniferous and diamond-inclusion chromites, can be distinguished from regional tourmalines and chromites on the basis of color, morphology, and chemical composition.

The detection of anomalous amounts of rounded brown tourmalines led to the discovery of the diamond-bearing Jack lamproite in British Columbia, Canada. High Cr-Ti chromites and rounded tourmalines have been recovered from a diatreme near Presidente Olegario, Brazil. Such Cr-Ti chromites that plot within an elevated TiO_2-Cr_2O_3 field have not been identified in source rocks other than lamproite or kimberlite. A high proportion of diamond-inclusion composition chromites and elevated K-Ti tourmalines are thought to possibly indicate high diamond grades in lamproite (Fipke 1994).

Other lamproite indicator minerals. Minerals usually associated with lamproites include phlogopite, Ti-phlogopite, diopside, diopsidic augite, alkali amphibole, leucite, leucite pseudomorphs, sanidine, Sr-apatite, sphene, Sr-rich barite, perovskite, and ankeritic and dolomitic carbonate. However, these minerals have low specific gravity and are typically not recovered in the heavy-mineral concentrates during stream-sediment sampling. As a result, they have limited application as tracer minerals.

Armalcolite and some other rare minerals found in lamproite, such as K-Ti richterite, K-riebeckite, K-arfvedsonite, priderite, wadeite, and shcherbakovite, are commonly present in hypabyssal and late stage lamproites but have not been reported in early crater-infill or diatreme facies samples (Fipke, Gurney, and Moore 1995). Armalcolite is uncommon in most lamproites, but it appears to be relatively abundant in the Smoky Butte intrusive in Montana. It has been found primarily as tiny euhedral groundmass prisms (Mitchell and Bergman 1991).

EXPLORATION PROGRAMS

Exploration programs for diamonds should be designed to first ensure that the target area's geologic environment and tectonic setting is suitable for the type of diamond deposit of interest. For example, when searching for diamondiferous kimberlite, one should concentrate on cratonic regions and cratonized Proterozoic margins. The search for diamondiferous kimberlites in provinces with high heat flow, such as the Basin and Range of the western United States, would be considered counterproductive.

At the beginning of an exploration program, the region under consideration is examined on topographic maps, aerial and satellite photography, geologic maps, and any aerial geophysical data maps. Any unusual circular depressions or anomalous circular drainage patterns on the topographic maps are noted, and noteworthy structural trends, vegetation anomalies, and circular features are marked on the aerial photographs (Figure 15.7). With the use of geologic maps, the geology is noted, and any unusual geologic features are identified such as cryptovolcanic structures, ring dike complexes, and circular outliers of rock. The geophysical data are used to search for any localized conductors and magnetic anomalies. If geochemical data are available, they are combined with other data sets (see Chapter 16, Predicting Diamond Content, Figure 16.1). Any location where Cr, Ni, and Nb anomalies line up with any other anomaly is marked.

The search for kimberlitic indicator minerals in stream-sediment concentrates has been one of the more effective means of identifying new target areas. In outlining a stream-sediment sampling program, sample sites are initially marked on all large drainages. Sample spacing is designed to take advantage of the region under consideration. For example, in arid regions, sample spacing is typically designed to take advantage of the relatively short transport distance of the indicator minerals. However, in many subarctic to arctic areas of Canada, Sweden, and Russia, sample density may be considerably lower owing to the greater transport distance and the logistical difficulties of collecting samples.

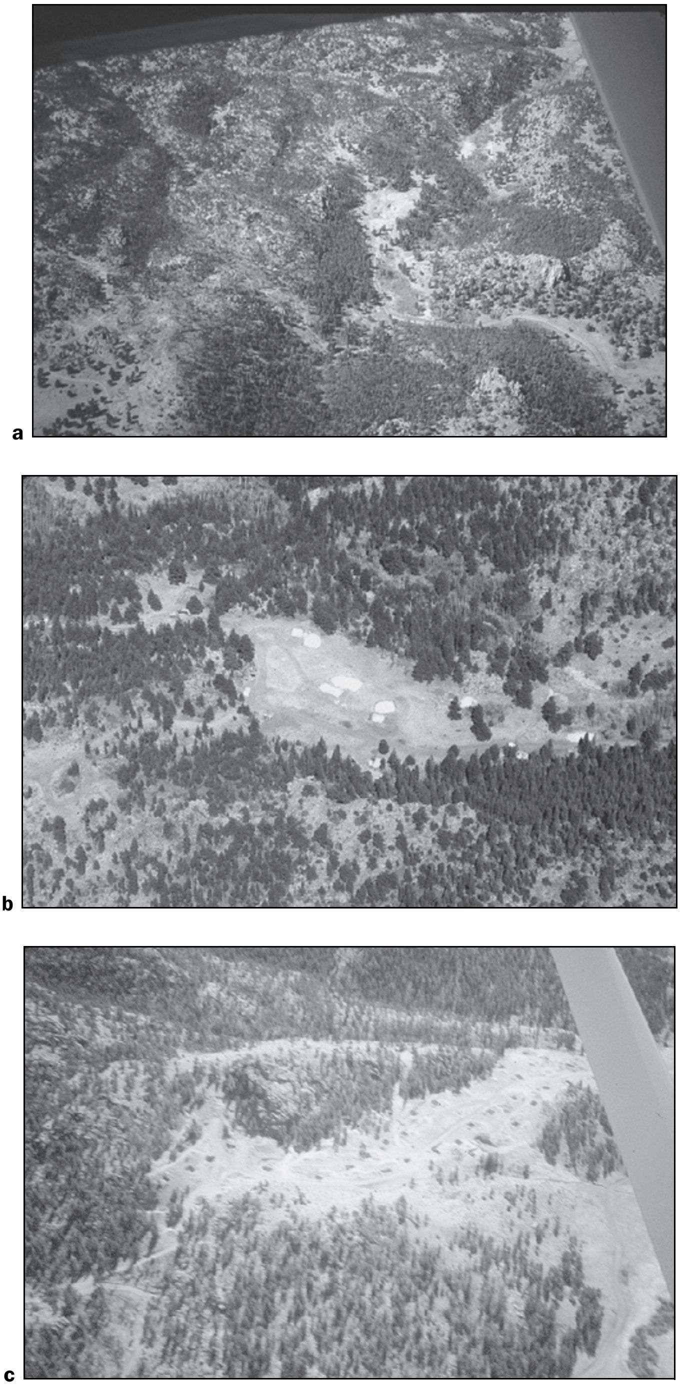

FIGURE 15.7 Aerial views. (a) Sloan 5 kimberlite, Colorado, showing distinct circular depression. (b) Closer view of Sloan 5 outlined by a distinct treeless vegetation anomaly. (c) Sloan 1 kimberlite with bulk sampling pits (photos by W.D. Hausel, 1982).

In Sweden, some orientation surveys were designed to sample drainages and eskers (discarding material >1.2 mm). In one survey, 157 samples were collected on a spacing of one per 100 km^2. Only one anomalous sample site was identified at this spacing. The sample spacing apparently was influenced by airborne geophysical data that identified 295 geophysical targets. In follow-up sampling, material was collected down ice and from till, and positive results were further investigated. A further 516 samples were collected, and reconnaissance till sampling was set at a nominal spacing of 400 m perpendicular to glacial transport directions (Phelps 2001).

In the semi-arid environment of the Wyoming craton, samples are ideally collected on a spacing of about one per 2.5 km along all significant drainages (discarding material >2 mm). The sites are initially plotted on topographic maps and adjusted in the field to take advantage of factors such as access, ideal heavy mineral traps, and vegetation.

In one project, 19% of all samples collected yielded kimberlitic indicator mineral anomalies (approximately 300 anomalies were detected in 1,600 samples) (Hausel, Sutherland, and Gregory 1988). This rate of detection is probably much higher than in most grassroots surveys, but it provides an indication of the potential, widespread, undiscovered kimberlite swarms in the Wyoming craton.

After the initial reconnaissance samples were collected, highly anomalous areas were reevaluated and sampled in greater detail. In some highly anomalous areas detected during this survey, samples were collected every 0.3 km.

Once the site was selected in the Wyoming craton, a one- or two-person crew occupied the site for about 2 or 3 hr, digging in the same hole in the drainage to recover progressively deeper material from the site. This step ensures that material is sampled throughout much of the drainage history, including past flash-flood events. For example, in other areas of this craton, gold recovered in the South Pass greenstone belt was found to concentrate at specific horizons in various drainages, possibly owing to flash-flood events. Thus it is important to sample as much of the stratigraphic section in drainages as possible in regions susceptible to flash floods.

The successful recovery of indicator minerals requires effective methods of recovering black sands; i.e., minerals with specific gravity >3. Sample recovery from the various sample sites in wet drainages was completed with a traditional gold pan. To expedite the process, samples are initially sieved in a grizzly pan with 6 mm holes. The coarse material is quickly discarded following cursory examination of the >6-mm fraction in the field for possible indicator minerals. Of several hundred stream-sediment samples collected, none contained indicator minerals >6 mm. However, during geologic mapping in the Colorado–Wyoming State Line district, some large 7.5 cm pyrope-almandine megacrysts were found in a drainage with no exposed kimberlite. The samples led to the discovery of a buried kimberlite by use of a geophysical survey during follow-up geophysical reconnaissance (Hausel, Glahn, and Woodzick 1981).

The next step was to sieve the <6 mm material with a fly screen and to pan the material that filtered through the screen (<2 mm^2). The medium-grained material (<6 mm and >2 mm^2) is discarded. Of 1,600 samples collected, only three retained indicator minerals on the fly screen, thus retention of the +2 mm^2 material was abandoned (−2 mm^2 indicator minerals were also recovered from these samples). The purpose of the screens is primarily to decrease panning time as well as sorting the material. The −2 mm^2 fraction was retained for further processing the laboratory.

The −2 mm^2 fraction is panned in the field to reduce the amount of material of specific gravity <3.2. At some localities, as much as 90% to 95% of the lighter material (e.g., quartz, feldspar, and mica) was removed by panning. Some care is taken in the panning

TABLE 15.5 Average specific gravity of indicator minerals

Specific Gravity	Mineral
3.2	Chromian diopside Chromian enstatite Forsterite
3.5	Diamond
3.8	Pyrope garnet
4.2	Picroilmenite
4.7	Chromite Zircon

process to retain black sands and a minor amount of light material. The panner should not attempt to pan as hard as one would if prospecting for gold, because gold has a very high specific gravity (>15) and the black sands of interest for diamond exploration range from 3.2 to 4.7 (Table 15.5).

To be sure that panners are correctly panning, it is recommended that samples be periodically salted with "density beads." Density beads consist of nonwettable Teflon impregnated with barium to achieve the desired specific gravity. Density beads with different densities are color coded. A bright, fluorescent orange density bead with a specific gravity of 3.53 is typically used to train samplers; the fluorescent orange composite is distinctly visible, and the specific gravity of the bead mimics the density of diamond. Density beads are manufactured in South Africa by Van Eck and Lurie (Carl Brink, personal communication, 2000).

Collecting samples from dry drainages presents a different problem. Relatively large amounts of material should either be dug from the dry placer and transported to water for processing, or water must be transported to the dry placer. Typically, samples collected from dry drainages are sieved on site and the fines transferred to a nearby wet drainage for panning.

In the exploration for kimberlite, the heavy minerals sought include pyrope garnet, chromian diopside, picroilmenite, chrome spinel, and diamond. Because these minerals may be rare or lacking in lamproite, they need to be supplemented by other minerals, some which may be less heavy, less resistant, and difficult to identify, such as zircon, phlogopite, K-richterite, armalcolite, and priderite. In Australia, the best indicators for lamproite are diamond and magnesiochromite (which appear as phenocrysts in most lamproites). Garnet, chrome diopside, and picroilmenites are rare in lamproite.

The lack of usable indicator minerals in lamproite required prospecting in a two-tier system in Australia. The search for kimberlite required samples in the range of 22.5 to 189 kg. For lamproite, where diamond is the targeted indicator mineral, samples can range from 15 to 200 tonnes!

Important clues to proximity to source are determined by which indicator minerals are present and how fresh they appear. The highly abrasion-resistant nature of some garnets, spinels, zircons, diamonds, and ilmenites allow for widespread dispersion. These minerals will generally show abrasion features indicating that they are from a distal source. Features that suggest a nearby source include kelyphitic rims on garnet; carbonate on garnet, chromian diopside, and picroilmenite; leucoxene coating on picroilmenite; a pitted, orange-peel surface on garnet; aggregate grains (e.g., garnet, pyroxene, and diamond), and bort diamonds.

Several assumptions are made in the interpretation of heavy mineral results. Errors in assumptions about direction of transport and distance from source can be fatal. Another pitfall is to assume there is only one source rock in the area under investigation.

After sample collection, processing will reduce the sample volume to a workable size. The initial sample may be 20 to 45 kg or more. Looking for one 0.5 mm pyrope in these samples can be a formidable task.

Sample Processing

The panned concentrates are transported to a laboratory for further concentration. If panning has been efficient, little additional processing will be needed. However, further processing is still necessary in order to reduce the amount of material to be examined microscopically.

In the laboratory, the sample concentrates are dried in a drying oven and then classified in a magnetic separator. The extracted magnetic fraction is discarded, and the remaining concentrates are further concentrated. Concentration can be completed by a number of processes, such as a side-shaking Wilfley table. Whatever recovery method is used, it needs to be adjusted to recover material of specific gravity of 3.2 and greater. In place of a Wilfley table, other laboratories may use jigs, heavy liquids, or other techniques to concentrate the minerals of interest.

The adjustments on the concentrating (Wilfley) table are checked by using a referee sample with common, visible, chromian diopside (specific gravity 3.2), bright-orange carbon composite chips (density beads of specific gravity 3.5), and quartz (specific gravity 2.87). The referee sample is run across the table to test its ability to separate these minerals by specific gravity; the diopside and composite chips should be separated from the quartz. After the test is completed and the table adjusted, the table and recovery buckets must be thoroughly cleaned to ensure that none of the chromian diopside is retained on the table or in the recovery system. Sample contamination should be a major concern in any laboratory. Even after precautions are taken, some laboratories still contaminate samples. In 1997, a major mining company sampled kimberlite from the Iron Mountain district in Wyoming and recovered synthetic and natural diamonds from their samples. The synthetic diamonds were apparently contaminants from crushers normally used for crushing rock core. The source of the natural diamonds was never determined. Such carelessness is inexcusable in the laboratory, because contaminated samples can lead to additional exploration of barren areas.

Following processing, the final concentrates are examined with a binocular microscope, and possible kimberlitic indicator minerals are separated from the concentrates for petrographic tests and microprobe analyses. Petrographic analysis of garnets includes checking for isotropism and index of refraction; pyrope garnets should display an index of refraction <1.760, whereas most amphibolite grade almandine garnets from the country rock have an index of refraction >1.760.

During microscope studies, the mineralogist should evaluate the general characteristics of the kimberlitic indicator minerals. In addition, value-added commodities should be noted. During sampling the in the Laramie Mountains of Wyoming, some sample concentrates contained other potentially valuable minerals such as gold, ruby, sapphire, amethyst, and aquamarine.

Both anomalous and barren sample sites are labeled on the map to assist in evaluation of targets. Typically, the anomalous samples are labeled to represent the number of indicator-mineral grains found. This information is used to determine priority of targets, because sample sites containing more than a dozen indicator minerals typically signify a proximal target. Sites containing more than 100 indicator minerals are of high priority, especially if the site has yielded diamond-stability-field minerals. Anomalies are further evaluated by more closely spaced samples in order to help pinpoint the source area.

Remote Sensing

Lamproite, kimberlite, and lamprophyre may be emplaced along deep-seated fractures. When such fractures are recognized, they may provide a unique orientation that is useful in the search for other similar intrusives. Some lamproites and kimberlites are detectable by remote-sensing exploration methods, such as high- or low-altitude imagery.

The imagery is used to search regions for geomorphic features that are characteristic of kimberlite and of lamproite flows and plugs, identify controlling structures for hidden lamproite, or detect spectral anomalies that indicate the presence of kimberlite or lamproite. Once controlling structures are identified, field reconnaissance is necessary to search for poorly exposed kimberlites and lamproites and for hidden intrusives along the structures.

Of particular interest are kimberlites and olivine lamproites. Most olivine lamproites contain abundant serpentinized olivine and tend to be buried by thin veneers of soil or regolith, making these targets much more difficult to find than the phlogopite lamproites.

Remote sensing is designed to search for anomalous structures, vegetation, and/or clay using high-altitude or satellite images. Kingston (1984) reported that remote-sensing techniques have been widely used in the search for kimberlite and have included conventional and false color aerial photography, LANDSAT multispectral scanner satellite data, and airborne multispectral scanning.

With multispectral scanning data, it is possible to identify spectral anomalies related to Mg-rich clays (i.e., montmorillonite), carbonate, and material with silica deficits—in other words, those types of altered rocks and soils that would be expected to blanket weathered kimberlite or olivine lamproite. Image enhancement techniques (contrast enhancements, ratios, principal components, and clustering) can be used to produce an image that is optimum for discrimination of kimberlitic and olivine lamproitic soils. These and other photo images can also be used to search for vegetation and structural anomalies.

Nixon (1980) reported that LANDSAT provides large-scale spectral data with pixels representing 80×80 m on the ground. Such data are generally too coarse to detect individual kimberlites and lamproites, many of which are less than 150 m across. However, some larger kimberlites may be detected, and Longman (1980) reported that LANDSAT data were used to characterize lamproite in northern Australia. Airborne multispectral scanning provides higher resolution data than LANDSAT, and this method can be used to measure the reflectance qualities of clay in soil.

Several kimberlites in the State Line district, Colorado–Wyoming, were identified with various color-band composites owing to the unique reflectance of the weathered kimberlite (Marrs et al. 1984; Marks 1985). Principal component analyses proved to be the most effective method for detecting blue ground associated with weathered kimberlite, and the second and fourth component contributions were the most distinctive for identification of kimberlite. Color composite renditions that include principal components 2 and 4 were selected as the most interpretable of the enhanced products. From these enhanced products, the target areas were selected according to their similarity with known kimberlite (Marrs et al. 1984). In South Africa, Kingston (1984) reported that reflectance spectra in the range of 0.4 to 2.5 micron wavelength were used to detect surficial occurrences of kimberlite.

Aerial photography has been used extensively in diamond exploration. Many pipes and dikes possess distinct structural qualities or vegetation anomalies that allow detection on photographs. Mannard (1968) reported kimberlites in southern and central Africa were identified on aerial photographs on the basis of vegetation anomalies, circular depressions or mounds, and/or tonal differences. Low-level aerial photographs were used successfully to located kimberlite in the USSR (Barygin 1962) and in the United States (Hausel, McCallum, and Woodzick 1979; Hausel et al. 2000). Hausel, McCallum, and Woodzick (1979) demonstrated that some kimberlites in Wyoming were readily detectable on conventional and false color infrared aerial photography. Kimberlite dikes and several circular depressions were also identified on aerial photographs in the Iron Mountain district of Wyoming.

The search for lamproite by remote-sensing exploration may simply search for distinct volcanic structures and associated prominent fractures. Some lamproites occur along visible fracture systems that are mappable for 1.5 to 3 km along strike length (W.D. Hausel, unpublished data, 1996). When such structures are identified, similar structures with similar orientations are considered high priority; however, more than one orientation may control the emplacement of these intrusives in a given district.

Many structures and lineaments are detectable on aerial photography. For instance, Hausel, McCallum, and Woodzick (1979) identified a distinct trend for several kimberlite groups in the Colorado–Wyoming State Line district. Follow-up surface mapping led to the discovery of a few hidden kimberlites along the identified trends. In the Leucite Hills lamproite field, distinct east-west–trending fractures were identified on high-altitude aerial photographs. These structures may also be useful in the search for hidden olivine lamproite. Aerial photography was also used to identify the controlling structure of the Cedar Mountain breccia pipes in the southern Green River basin of Wyoming (Richard Kucera, personal communication, 1996).

Geophysical Surveys

Geophysical exploration methods have been used in the search for hidden kimberlite and lamproite at a number of localities in the world (Litinskii 1963a, b; Gerryts 1967; Burley and Greenwood 1972; Hausel, McCallum, and Woodzick 1979; Hausel, Glahn, and Woodzick 1981; Patterson and MacFadyen 1984). The contrasting physical properties of the diamondiferous host rock (in particular kimberlite, lamproite, and minette) and the country rock commonly allow these rocks to be distinguished by surface and airborne exploration geophysics.

The diamondiferous host rocks are high-magnesian, ultrabasic, ultramafic, or mafic rocks that have relatively high density. Weathering of kimberlites and olivine lamproite commonly forms clay-rich layers at the surface that are less dense and conductive. Magnetite produced during serpentinization of the kimberlite or olivine lamproite is commonly destroyed in the clay layer. These characteristics produce different geophysical signatures over the various intrusives.

Airborne. Airborne geophysical surveys have proven useful in locating hidden kimberlite and lamproite within known diamond fields. INPUT™ is a time-domain electromagnetic-survey system in which measurements are made during the off-periods between source pulses (Sheriff 1973). INPUT™ airborne surveys have been proven to be effective in districts that contain both serpentinized hard rock exposures and weathered kimberlites, owing to the combination of conductivity and magnetics used by the survey. In particular, kimberlite rock exposures typically yield magnetic signatures but are poorly conductive, while deeply weathered kimberlites are conductive but poorly magnetic.

Because of the relatively small size of the targets, close flight-line spacing is necessary. In one airborne INPUT™ survey over the State Line district along the edge of the Wyoming craton, a flight-line spacing of 200 m effectively detected several kimberlites (including the Kelsey Lake kimberlites) and several distinct magnetic anomalies interpreted as blind diatremes (Patterson and MacFayden 1984). None of these magnetic anomalies had been drilled by August 2001, even though they are surrounded by 40 diamondiferous kimberlites and lie within 3.2 to 4.8 km of the Kelsey Lake diamond mine.

An aeromagnetic (200–400 m line spacing) survey flown over parts of northeastern Kansas by Cominco American identified several anomalies, and later drilling confirmed previously unknown kimberlites (the Baldwin Creek and Tuttle kimberlites in Riley county and the Antioch kimberlite in Marshall county). Thirteen kimberlites (and at least one olivine lamproite) are now known in eastern Kansas (Berendsen and Weis 2001).

Aeromagnetic surveys have been very successful in Australia in locating lamproite. Flight line spacings of 50 to 100 m are commonly used. INPUT™ surveys were successfully used in the Ellendale lamproite field (Atkinson 1989), and airborne magnetics and radiometrics effectively identified diamondiferous lamproite in the Ellendale field (Janke 1983; Jaques, Lewis, and Smith 1986). The olivine lamproites, in particular, produced distinct dipolar magnetic anomalies.

Detecting lamproites devoid of olivine requires the surrounding country rock to have a simple magnetic signature. If airborne surveys detect a lamproite, then ground magnetic surveys can be used to help delineate its extent. In some cases, magnetics can be used to search for hidden lamproites along fractures and under anomalous geomorphic features. Flight line spacings and elevations of regional aerial magnetic surveys are typically 235 m and 50 to 100 m, respectively (Mitchell and Bergman 1991).

However, the Argyle olivine lamproite intrudes Precambrian rocks with complex magnetic signatures, and the lamproite produced an indistinct magnetic signature (Drew 1986). Rock samples collected by Hausel from the Argyle olivine lamproite are not magnetic, although Atkinson (1989) reports that the tuff phases of the pipe are weakly magnetic.

Gravity. The high relative density of kimberlite and lamproite should make these rocks detectable by gravity and seismic surveys. However, most diamondiferous intrusives are small and weathered, and gravity and seismics are generally not sensitive or practical enough to use in the search for kimberlite or lamproite. For example, Hausel, McCallum, Woodzick (1979) noted that diamondiferous kimberlite intruded in granite in the Wyoming craton showed no detectable density differences with the host granite. However, Fipke, Gurney, and Moore (1995) suggested that where kimberlite and lamproite have a specific gravity that strongly contrasts with the country rock, ground base gravity surveys might be used to search for the intrusives.

The major drawback in using gravity is the low-amplitude responses over the pipes (Atkinson 1989). Gravity surveys over pipes may show negative responses (amplitudes commonly less than 1 mgal), owing to the presence of weathered kimberlite or a thick sequence of crater sediments. If the host rock (for example, undeformed clastic sediments) is of lower density than the pipe, then a positive anomaly should result. However, the overall lack of response over kimberlite can be attributed to a density similar to that of the basement country rock, depth of burial, and the small cross sections of the pipes and dikes.

Seismic. Few seismic surveys have been used to search for diamond deposits, because kimberlite, lamproite, lamprophyre, and basement country rocks all have similar velocities. In some cases where diamondiferous pipes have intruded clastic sediments in the platform, the velocity differences between the intrusive and country rock should be great enough to differentiate the pipes from country rock on detailed seismic surveys.

Hausel, McCallum, and Woodzick (1979) used seismics to distinguish three layers in weathered kimberlite in the Aultman 1 pipe in Wyoming. The upper layer of soil mixed with blue ground yielded an average velocity of 670 m/s. The intermediate layer, interpreted as blue ground, yielded an average velocity of 1,590 m/s. Beneath these two layers, a third layer, interpreted as dense kimberlite, yielded a velocity of 3,529 m/s. These results were similar to Puckett et al. (1972), who detected four layers over the Sloan 1 kimberlite in Colorado.

Hausel, McCallum, and Woodzick (1979) noted that the seismic velocities in the host granite were on the order of 2,136 m/s to 4,877 m/s, suggesting that seismic surveys would not uniquely differentiate kimberlite from granite in this region.

Magnetic. Both surface and airborne magnetic surveys have had limited success in detecting kimberlite and lamproite. Bolivar and Brookins (1979), Macnae (1979, 1995), Nixon (1981), Jaques, Lewis, and Smith (1986), Waldman, McCandless, and Dummet (1987), Coopersmith and Mitchell (1989), Atkinson (1989), and Sarma, Verma, and Satyanarayana (1999) report some success with magnetics. Hausel, McCallum, and Woodzick (1979), and Hausel, Glahn, and Woodzick (1981) reported only limited success.

The magnetic susceptibility of kimberlite and lamproite is controlled by various iron and titanium oxides in the intrusives. The magnetic susceptibility of kimberlite depends heavily on the abundance of magnetite produced by serpentinization. In lamproite, the magnetic susceptibility is controlled by the abundance of titaniferous magnetite, other spinels, and other Fe and Ti oxides, as well as by serpentinization of olivine.

Fresh diamondiferous host rocks may contain as much as 5% iron oxides. These oxides consist predominantly of nonmagnetic picroilmenite, but the intrusives may contain minor to trace magnetite and mixed magnetite-ilmenite. Both unweathered kimberlite and lamproite may produce weak to moderate magnetic signatures.

Lamproites possess a wide range of magnetic susceptibilities depending on the abundance of the iron-oxide phases and the amount of weathering and serpentinization. Madupitic lamproites contain magnetite. Thus these lavas should yield detectable magnetic anomalies, whereas other lamproites may only yield weak magnetic anomalies.

Macnae (1995) indicated that lamproite is less magnetic than kimberlite by an order of magnitude, and that magnetic anomalies over pipes usually are highly complex, indicating inhomogeneities. These complexities commonly reflect the multiple intrusions and facies in individual pipes, as well as differential weathering. For example, Sarma, Verma, and Satyanarayana (1999) detected distinct magnetic complexities in the Majhgawan olivine lamproite in India.

It is possible that magnetic mapping could be used to identify regions of higher ore-grade potential in some lamproites. For example, the Majhgawan pipe contains a central core of massive medium-grained olivine lamproite as opposed to the overall fine-grained texture in the surrounding rock. The diamond grade of the central core is higher than the outer portion of the pipe. This difference suggests that the medium-grained rock might be differentiated from the fine-grained rock by magnetic mapping in conjunction with detailed geological mapping.

In the Yakutia province in Russia, magnetic surveys were used successfully in areas where differences between the magnetic susceptibility of kimberlite and the intruded carbonate sedimentary rocks was high. Anomalies as great as 5,000 gammas were successfully detected from airborne surveys (Litinskii 1963b). Equally good results were reported from Mali, West Africa, where the magnetic contrast between kimberlite and schist and sandstone country rock produced 2,400-gamma anomalies over kimberlite (Gerryts 1967). In Lesotho, anomalies over kimberlite were reported to be comparable with those in the Yakutia province (Burley and Greenwood 1972).

Magnetic surveys in the United States have had limited success. In Arkansas, a 2,000 gamma magnetic anomaly was recorded by Stern (1932) over the Prairie Creek olivine lamproite. However, later surveys reported no distinct anomalies. Fipke, Gurney, and Moore (1995) indicated that the barren peridotite phases in the area yielded magnetic highs, but that the diamondiferous phases were not detected by magnetics.

In northeastern Kansas, Brookins (1970) reported large positive (550 to 5,000 gamma) and negative (0 to −2,800 gamma) anomalies over some kimberlites. These kimberlites were emplaced in sedimentary rocks with low magnetic susceptibility.

Most kimberlites in the Colorado–Wyoming State Line district, however, show only small dipolar and complex anomalies in the range of about 25 to 150 gamma. A few exceptions yielded isolated anomalies of 250 and 1,000 gamma (Hausel, McCallum, and Woodzick 1979; Puckett et al. 1972). Blue ground in the district tended to mask the magnetic anomalies. In the Iron Mountain district, Wyoming, where much of the kimberlite is relatively homogeneous, the massive hypabyssal-facies kimberlite yielded only weak to indistinct anomalies (Hausel et al. 2000).

At the Radichal kimberlite in the Middle Sybille Creek district northwest of Iron Mountain, an outcrop of highly serpentinized hypabyssal-facies kimberlite yielded a magnetic high of 700 gamma (Hausel, Glahn, and Woodzick 1981). But for the most part, the surface magnetic surveys in the Wyoming–Colorado region were of little value, because the kimberlite outcrops that produced distinct magnetic anomalies had been previously found during geologic mapping. A problem recognized in the State Line district is that disseminated magnetite and ilmenite in the granite and anorthosite country rock can produce readings similar to kimberlite.

Resistivity. Electrical resistivity depends on rock porosity and the salinity of a solution within pore spaces. Rocks that are more porous or more highly altered with respect to their surroundings will yield different resistivities. Rock replaced by clay will usually have resistivities one to three orders of magnitude less than those of igneous and metamorphic rocks.

This geophysical method requires placing four electrodes into the ground along a survey line. Current is passed through two of the electrodes, and the potential voltage is measured across the other two. A Wenner array is generally used, where all four electrodes are inserted in the ground in a line.

Resistivity can detect both lateral and layered variations in the subsurface. Typically, weathered kimberlite will yield a prominent resistivity low over the pipe and an increase in resistivity at depth. The increase at depth is essentially due to the lack of weathering in the deeper kimberlite.

The uppermost portion of a kimberlite pipe will weather into a disc-shaped, lower-density, highly conductive clay-rich horizon depleted in iron, calcium, and magnesium.

Magnetite in the weathered zone will oxidize to hematite. A modest but still detectable conductivity anomaly in some fresh kimberlites may be due to serpentinization of olivine during initial diatreme emplacement. In plan view, a diatreme may show a circular to elliptical conductivity response coincident with a strong magnetic anomaly of slightly smaller diameter (owing to the convergent shape of the pipe and depth of weathering).

Clay produced from the weathering of kimberlite and olivine lamproite is conductive and is susceptible to detection by electrical methods. Typically, kimberlites will erode more rapidly than surrounding country rocks. Depending on the local environment, some kimberlites may be capped by a layer of yellow ground with an appearance similar to gossan. The yellow ground is altered kimberlite that has been converted to dry mud clay. In some places in Africa, the yellow ground is as much as 30 to 60 m thick. In the Colorado–Wyoming State Line district, yellow ground is rare.

Yellow ground is commonly underlain by a thick layer of blue ground. The blue ground is composed of bluish-gray to light-gray weathered kimberlite that has been replaced by montmorillonite clay with mixed calcium carbonate. The blue ground is soft, and where it is clay rich it may have a greasy touch. Blue ground in the State Line district, Wyoming, was interpreted to continue to depths of 24 to 36 m on the basis of electrical-resistivity sounding methods. At depth the blue ground grades into hard-rock kimberlite.

Macnae (1979) reported that the apparent resistivity of yellow ground may range from 2 to 5 ohm-m, the less altered blue ground 50 to 100 ohm-m, and the fresh hard rock kimberlite around 500 ohm-m. Resistivity surveys in the Colorado–Wyoming State Line district detected apparent resistivities of 25 to 75 ohm-m over weathered kimberlite, compared with 150 to 2,250 ohm-m in the country rock granite (Hausel, McCallum, and Woodzick 1979).

The electrical resistivity of weathered lamproite is generally thought to be less than that of most country rocks, owing to the conductive nature of smectitic clay relative to illite, kaolinte, and other clay minerals (Gerryts 1967; Janke 1983). However, olivine lamproite vents such as Argyle possess moderate to strong resistivity anomalies of 40 to 100 ohm/m compared with the surrounding country rock (200 ohm/m) (Drew 1986).

Conductivity. Surficial weathering of serpentine-rich magmatic rocks results in a clay layer in the zone of weathering that is rich in montmorillonite. During this process, magnetite in the clay is replaced by hematite, masking the near-surface magnetic affinity of the host rock. The clay however, promotes water retention. Thus blue ground over kimberlites may produce vegetation anomalies, and because the blue ground is conductive it will be susceptible to detection by electrical methods. Thus, hard rock exposures of kimberlite are typically anomalously magnetic, but weathered kimberlite will lack magnetism and will be anomalously conductive.

A portable Geonics™ EM31 has been used successfully to search for hidden kimberlite in the Wyoming craton. Unlike the EM31, resistivity equipment is not as portable and resistivity surveys are more labor intensive. However, resistivity surveys can detect materials at greater depths than the EM31.

The EM31 electromagnetic unit induces circular eddy current loops that generate magnetic fields. A part of the magnetic field from each loop is intercepted by the receiver coil and produces an output voltage that is linearly related to the terrane's conductivity. Clay soils are conductive and produce conductivity highs. One drawback of the instrument is that it detects conductivity to depths typically less than about 6.25 m. Thus the search for blind pipes requires either airborne surveys or a geophysical sounding method such as resistivity or seismics.

In the State Line district, weathered kimberlite was successfully detected by electromagnetic (EM) surveys. One anomaly (38 mmhos-m) was detected in a drainage with considerable numbers of kimberlitic indicator minerals. The anomaly was trenched, exposing blue ground. An EM survey was also run across an exposure of hypabyssal facies hardebank kimberlite in the central Laramie Range. The fresh kimberlite (6 mmhos-m) was for all practical purposes not discernable from the surrounding anorthosite host (Hausel, Glahn, and Woodzick 1981).

Lamproite in Kansas also was successfully detected by EM surveys. Coopersmith and Mitchell (1989) indicated that the Hills Pond and Rose olivine lamproites in eastern Kansas were delineated using ground-based EM surveys.

Very low frequency. Very low frequency (VLF) surveys use the worldwide network of individual low-frequency Naval radio-communication transmitters as a source of electrical currents. Electric currents induced in conductive bodies produce local secondary electromagnetic fields that can be detected at the surface as deviations in the normal field. Data collection is rapid and inexpensive, and the data are easily interpreted. However, owing to target size and prominent ground effects, airborne VLF is generally not used in the direct detection of diatreme targets.

Biogeochemical and Geochemical Surveys

Kimberlites and lamproites are potassic alkaline ultrabasic igneous rocks with high Ba, Cr, Cs, K, MgO, Nb, Ni, P, Pb, Rb, Sr, Ta, Th, U, and light rare earth elements (LREE). The high Cr, Nb, Ni, and Ta levels may show up in geochemical tests of nearby soils and stream sediments (Jaques 1998). However, the dispersion of these metals in soils does not appear to be great.

Lamproites, which are ultrapotassic mafic igneous rocks, posses elevated compatible (Co, Cr, Ni, V, and Sc) and incompatible (Ba, Hr, Nb, P, Rb, Sr, Th, Ti, U, Y, Zr, and REE) trace elements. Stream-sediment geochemistry does not appear to be useful in delineating lamproites, because dispersion of these metals in stream sediments is too great.

Gregory and Tooms (1969) investigated the Prairie Creek lamproite, Arkansas, and found that Mg, Ni, and Nb anomalies did not extend farther than two-thirds of a kilometer from the vent. In addition, Haebid and Jackson (1986) noted that soil geochemical anomalies (Co, Cr, Nb, and Ni) were developed in sand and soil overburden immediately above lamproite vents in the West Kimberley province, Australia. This association may prove useful in the search for olivine lamproites within some lamproite fields.

Lithochemistry may be used in rock identification where lamproitic rocks are weathered or drill cuttings are involved. For example, Gregory (1984) used lithochemistry to map lamproite intrusions and to distinguish olivine lamproite from leucite lamproite on the basis of Mg, Ni, Cr, and Co ratios.

Bergman (1987) suggested that olivine lamproites are generally enriched in compatible elements relative to leucite lamproites, primarily as a result of the abundance of xenocrystal olivine in the former. Mitchell and Bergman (1991) also emphasized that "recognition of olivine lamproite is the only guide to the possible presence of potentially diamondiferous rocks." Barren lamproites (i.e., phlogopite-leucite lamproites) contain elevated alkali and lithophile content (K, Na, Th, U, Y, and Zr) relative to typical diamondiferous varieties. Despite wide ranges, diamondiferous lamproites possess twice the Co, Cr, Mg, Nb, and Ni, and half the Al, K, Na and as barren lamproites.

Many diamondiferous lamproites possess geochemical and mineralogic similarities to kimberlite. Principal among them are high modal olivine, Cr-rich spinels, and similar enrichments in Cr, Ni, and incompatible elements. The typical kimberlitic indicator minerals are notably absent in lamproite. In particular, subcalcic, chrome-rich pyropes (G10) typical of diamondiferous kimberlites are absent in the Prairie Creek and Argyle diamondiferous lamproites. Such garnets are also extremely rare in the Ellendale 7 and 9 diamondiferous olivine lamproites in the Kimberley block of Western Australia (Mitchell and Bergman 1991).

When lamproites weather they produce an exotic suite of secondary minerals such as nontronite, smectite, and barite. Lamproites also possess anomalous amounts of Ti, K, Ba, Zr, and Nb as compared with most other rocks; these anomalous components may favor the growth of different flora than is found in the surrounding terrane, and they may also stress the local vegetation. In addition, lamproites are enriched in Hf (Jaques 1998). The Big Spring vent, West Kimberley, Australia, is characterized by anomalous faint pink tones that reflect the growth pattern of grasses on the vent (Jaques, Lewis, and Smith 1986).

Many kimberlites around the world produce some sort of biogeochemical and geochemical anomaly owing to greater quantities of rare earth elements, Cr, K, Mg, Ni, and P, montmorillonite clays, and water in the kimberlitic soils than in the adjacent country rock. In the Colorado–Wyoming State Line district, Cominco American outlined the known kimberlite intrusives and dike extensions identified on the basis of Cr, Nb, and Ni soil geochemical anomalies. However, because of localized dispersion patterns, geochemical surveys have limited practical use, particularly when one considers that these extensions could have rapidly been determined with EM surveys and detailed geological mapping.

The geochemistry of many diamondiferous host rocks produces unique vegetative cover. Biogeochemical anomalies are noticeable over many kimberlite diatremes. In the Colorado–Wyoming State Line district, weathered kimberlite will not support woody vegetation, which results in open parks over kimberlite diatremes in otherwise forested areas (Figure 15.6). Additionally, these same kimberlites support a lush stand of grass, which commonly outlines the limits of the kimberlite, whereas the grass in the adjacent granitic soils is not as dense or as high. Distinct grassy vegetation anomalies over kimberlites in the Iron Mountain district were used successfully to map many intrusives. The vegetation anomalies are especially distinct following the short rainy season in the late Wyoming spring. The anomalies tend become less distinct in the late summer and early fall following several weeks of drought.

Siberian kimberlites have been noted to support denser stands of larch (*Larix dahurica*) and a more abundant undergrowth of shrub willow (*Salix*) and alder (*Alnus*) than the surrounding Cambrian carbonate country rocks. Phosphorus released from apatite (in the ultrapotassic kimberlite) was considered to be the element that produced the healthy vegetation. In central India, trees over the Hinota pipe were healthier, taller, and denser than those in the surrounding quartz arenite. The difference in the vegetation over the kimberlites and the adjacent country rock soils is attributed to the greater availability of K, P, micronutrients, and water in the kimberlite.

Vegetation over the Sturgeon Lake diamondiferous kimberlite in Saskatchewan was tested for 48 elements; the kimberlite showed a consistent spatial relationship with Ni, Sr, Rb, Cr, Mn, and Nb, and to a lesser extent with Mg, P, and Ba. In addition, relatively high Ni concentrations occurred in dogwood twigs. In hazelnut twigs, Cr levels were greater than 15 ppm near the kimberlite but only 5 to 8 ppm elsewhere, and the Nb content was higher in hazelnut twigs. Sr and particularly Rb were relatively enriched in some plant species on the kimberlite. The Sr was probably derived from the carbonates associated with the kimberlite, whereas the Rb was derived from phlogopite. The Ni, Rb, and Sr distribution and the Cr enrichment associated with Mn depletion in the twigs were used to identify nearby kimberlite.

Geologic Mapping

Detailed geologic mapping has proven to be invaluable in the search for diamond deposits. Following the discovery of a new kimberlite, lamproite, or lamprophyre, detailed geologic mapping around the new discovery is imperative. Geological mapping commonly reveals controlling structures that can lead to the discovery of other intrusives, and it forces the explorationist to pay attention to details. Such deep-seated intrusives as diamond pipes typically occur in clusters that commonly number in the dozens to a hundred or more. Usually, more than one favorable orientation is identified during mapping.

Geologic mapping is necessary to find many of the mantle-derived intrusives, because most are not detectable with remote-sensing, geophysical, or geochemical surveys. For example, in the State Line and Iron Mountain districts, the great majority of kimberlites were found through geologic mapping. Later stream-sediment sampling surveys led to the discovery of a few additional kimberlites, and a very small number were found through geophysical and remote-sensing methods.

Following the discovery of new kimberlite or lamproite pipes, it is important to map the different facies and phases of intrusive rocks in the pipes. For example, McCallum (1991) mapped six different kimberlite phases within the Sloan 1 and 2 pipes in Colorado. Differentiating the facies is important because each facies may have a different average diamond grade, and thus geological mapping may be used to outline potential commercial tonnages. Geological mapping of lamproite is also often used to differentiate tuffaceous and brecciated facies with near-commercial or commercial diamond grades as compared with lava phases that contain few or no diamonds.

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CHAPTER 16

Predicting Diamond Content

INTRODUCTION

Estimates suggest that only about one in 300 kimberlite pipes is sufficiently mineralized to be considered economic. Even in the richest pipes, diamonds are sparsely disseminated and typically occur in concentrations much less than 1 ppm. Thus to test the ore grade of a pipe requires extensive bulk sampling, and as much as 5,000 to 100,000 tonnes may be required to appraise and determine the average ore grade. In addition, the different facies of kimberlite must be noted during testing because grades may vary from facies to facies. For example, two different hypabyssal facies kimberlite of the Sloan 2 pipe in Colorado revealed different ore grades (Bernie Free, personal communication, 1994; Shaver 1994).

The geochemistry of mineral suites may be evaluated to provide an estimate of P–T, geobarometry, and geothermometry conditions during the genesis of some ultramafic nodules in kimberlite or mineral inclusions in diamond. In some cases, analysis of just one chemical element from a single mineral inclusion may provide some helpful information. In other cases (particularly in geothermometry) coexisting mineral phases need to be present. In geobarometry, minerals and elements often used include alumina substitution in orthopyroxene (enstatite) coexisting with garnet, potassium substitution in clinopyroxene, and sodium substitution in garnet. In all of these cases, elevated Al, K, and Na indicate high pressure.

Examples of geothermometry include methods based on the partitioning (relative proportions) of Ca and Mg contents of coexisting orthopyroxene (enstatite) and clinopyroxene (diopside), the relative abundance of Fe and Mg in these same pyroxenes, and the relationship between Fe and Mg in coexisting garnet and orthopyroxene (Kirkley, Gurney, and Levinson 1991).

Besides the P–T estimates, the chemistry of various "kimberlitic" rocks is determined and compared with diamond-inclusion chemistry to aid in predicting diamond content and estimating diamond grade. These data are typically used to make an initial appraisal of intrusives within a district to determine the higher-potential targets.

The geochemistry of kimberlitic indicator minerals is often tested to aid in the selection of targets. Depending on the type of deposit, the size of samples may range from 45 to 70 kg for initial examination of kimberlite. The size of the initial sample for lamproite will need to be considerably larger owing to the rarity of indicator minerals and diamonds. Gregory and White (1989) report that their targeted indicator mineral in lamproite is diamond, and initial test samples will range from 15 to 200 tonnes. Sampling of lamproites and lamprophyres will also consider chromite and pyrope compositions. Indicator minerals collected from stream-sediment samples during reconnaissance sampling should also be tested to aid in selection of targets.

The samples are processed to recover the heavy minerals (minerals with specific gravity ≥3.2). Sample processing can be done by any method that is effective in extracting heavy minerals and may include the Wilfley table, rotary pans, spirals, and/or heavy liquids. The final concentrates will need to be examined microscopically and the indicator minerals picked by hand.

The indicator minerals are extracted for microprobe analysis, and the geochemistry of the minerals is used to predict the diamond potential of the intrusive. However, the data can be misleading because there are exceptions to the rules. Thus it is important to collect as much data as possible to assist in the assessment of diamond potential. For example, information on the chemistry of the indicator minerals is important, but other data such as the rock type, bulk chemistry, diamond morphology, and the geologic environment will also aid in decision making.

In the search for kimberlite, indicator minerals collected for analysis include pyrope garnet, clinopyroxene, picroilmenite, and chromite. The garnets in kimberlite can have several different sources.

DIAMOND CONTRIBUTION OF PERIDOTITE

Information about the geochemistry of garnet, chromite, and clinopyroxene will provide valuable information for assessing the diamond contribution of a peridotite. Garnets from peridotites are initially selected for microprobe analysis based on color. The peridotitic garnets are reddish to mauve. Specific types of peridotite such as harzburgite have garnets that are generally purplish-red or purple to lavender to distinctly gray-lavender. Almandine-rich garnets from lherzolitic peridotites may be deep red or reddish brown to reddish pink.

Cr-rich garnets ($Cr_2O_3 > 2\%$) are assigned to a peridotite source. These garnets are then subdivided to G9 or G10 based on the CaO/Cr_2O_3 ratios (Sobolev 1977; Gurney 1984). Garnets of harzburgitic affinity have G10 geochemical signatures similar to those of diamond-inclusion garnets. The G10 garnets contain relatively high Cr_2O_3 and low CaO compared with garnet of lherzolitic genesis (G9).

Following analysis, the CaO/Cr_2O_3 ratios of the garnets are plotted to differentiate G10 from G9 garnets (Figure 16.1a). The presence of G10 garnets indicates that the host intrusive originated in the diamond stability field, and the stronger the G10 signature (the greater the number of garnets and the greater the subcalcic signature), the greater probability of finding diamonds as well as ore-grade material.

In order to differentiate garnets of diamondiferous lherzolite and barren lherzolite paragenesis, a Ni thermometer (see below) may be evaluated, and clinopyroxene derived from lherzolite may be tested for trace K_2O. Trace K_2O (≥0.7%) indicates high pressure, and clinopyroxene with relatively high K_2O is thought to have originated within the diamond cogenetic field (McCandless and Gurney 1989). Calcium saturation levels of the lherzolite pyropes may also provide important data. For example, G9 pyropes with the lowest calcium contents that plot near the 85% line (Figure 16.1a) indicate the highest equilibration pressures (Fipke, Gurney, and Moore 1995).

MgO and Cr_2O_3 content of kimberlitic and lamproitic spinels are also used to assess the probability that the diamond is derived from the disaggregation of chromite harzburgites. Chromites with compositions similar to diamond-inclusion chromites show a distinct range of compositions with high Cr_2O_3 and high MgO content. Those of the diamond association typically have compositions of Cr_2O_3 (57%–69%), and the majority have compositions of Cr_2O_3 (>61%) and MgO (10%–19%) (Figure 16.1c).

DIAMOND CONTRIBUTION OF ECLOGITE

Garnets from eclogite are used to appraise diamond potential from an eclogitic source (McCandless and Gurney 1989). These garnets are distinctly yellow orange and chromium poor (Cr_2O_3 < 2%). Most garnets derived from a diamondiferous eclogite contain elevated TiO_2 and Na_2O (Na_2O ≥ 0.08%), but exceptions occur; only about half of the diamondiferous garnets from the Star mine in South Africa have Na_2O (≥ 0.07%), and all are unusually low in TiO_2 (Gurney and Hatton 1989).

Those that plot within the diamond-inclusion field (see Figure 16.1b) are interpreted to have been derived from the diamond stability field. In addition, Group I kimberlitic clinopyroxenes contain K_2O contents in clinopyroxene of ≥ 0.08%. Plots of Na_2O in garnet (Na_2O ≥ 0.07%) versus K_2O in clinopyroxene (K_2O ≥ 0.08%) from eclogite are also used to assess diamond contributions from an eclogitic source (see Schulze 1993).

McCallum and Waldman (1991) provide an excellent review of how to use indicator mineral geochemistry to predict diamond grades. For example, an intrusive with a large percentage of diamond-stability-field (G10) subcalcic peridotitic garnets compared with graphite stability (G9) peridotitic garnets suggests a higher potential ore grade. The greater the subcalcic and Cr_2O_3 enrichment signature of the garnets, the higher the probability of ore grade material derived from a garnet harzburgitic mantle.

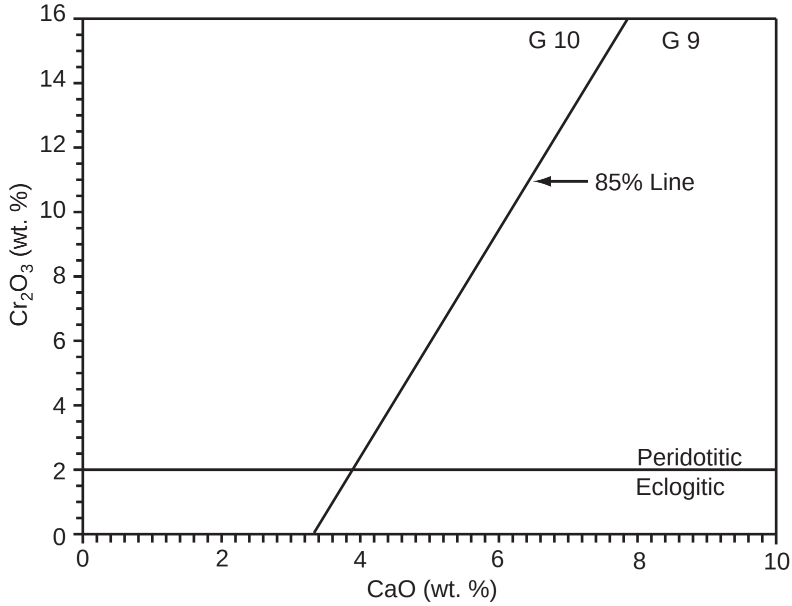
But exceptions occur. For example, The Argyle lamproite, which has produced the highest diamond ore grades in the world (average 680 carats/100 tonnes), has yet to produce a G10 garnet. This exception is very significant and needs to be considered in any exploration program. The fact that the diamonds are found in lamproite rather than kimberlite is immaterial, because the lamproite is simply a means of transporting diamonds from the mantle to the earth's surface. The Argyle example provides evidence of significant lherzolitic mantle diamond sources that could be overlooked using conventional diamond-content predictions! Although fewer than 100 diamondiferous peridotite nodules have been recovered from kimberlite and lamproite, a significant percentage of them have been lherzolites (the source of many G9 garnets).

The sodium signature of eclogitic garnet and the Cr-Mg signature of chromite are also evaluated (Figure 16.1b, d). Overall, the stronger the G10 signature, the stronger the Group I eclogitic garnet signature, and the stronger the Cr-rich chromite signature, the higher the potential diamond content of the host rock.

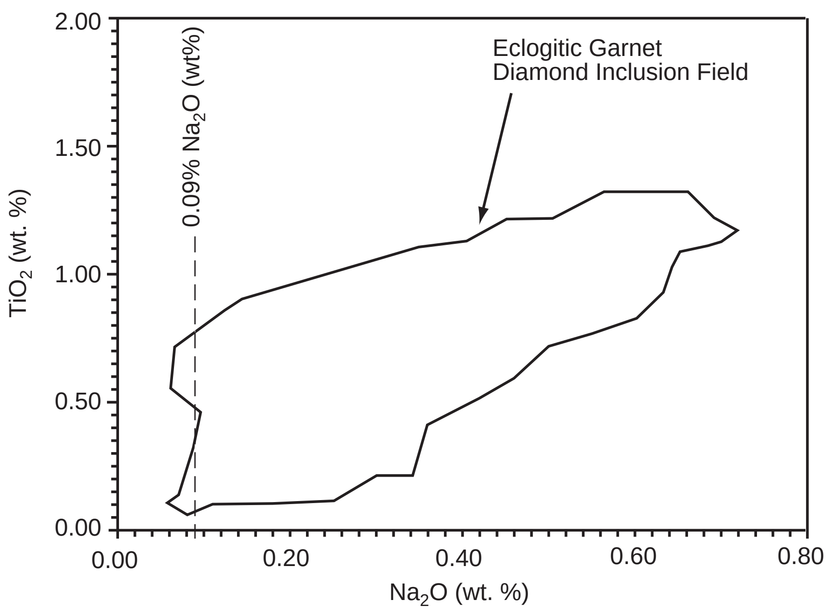
DIAMOND PRESERVATION

The above data are used to predict the diamond potential of the intrusive. But in addition to the original diamond budget sampled by a kimberlite or lamproite, diamond preservation can affect how many diamonds survive transport to the surface. Many exploration groups rely on data generated from ilmenite to evaluate the redox conditions of the magma and thus diamond preservation.

Ilmenites in diamond-rich kimberlites are suggested to possess low ferric iron and relatively high Cr and Mg content. Most picroilmenites are not related to diamond

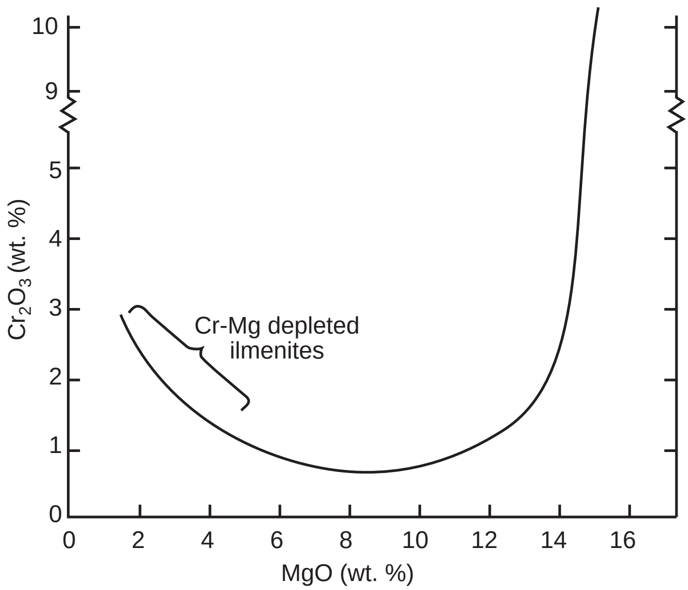


a



b

FIGURE 16.1 Plots of geochemistry used in the evaluation of indicator minerals. (a) CaO-Cr_2O_3 graph used to differentiate subcalcic G10 from calcic G9 garnets of peridotite paragenesis. (b) Na_2O-TiO_2 plot for eclogitic garnets showing diamond-inclusion field. (c) Cr_2O_3-MgO plot for chromites showing diamond-inclusion field. (d) MgO-Cr_2O_3 plot for picroilmenites showing composition parabola. Ilmenites that show Cr_2O_3 and MgO depletion are interpreted to be associated with high oxygen fugacity and resorption of diamonds (McCallum and Waldman 1991; Fipke, Gurney, and Moore 1995).



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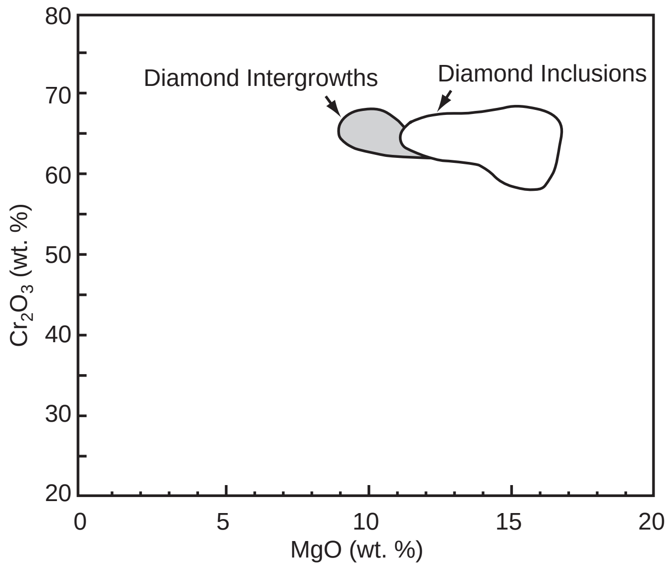


d

FIGURE 16.1 (continued)

crystallization, but instead are thought to form during later kimberlitic magmatic processes and metasomatic activity. As such, the importance of ilmenite composition during the evaluation of a pipe for diamond content may be related to diamond preservation (McCallum and Waldman 1991). Even though kimberlite and lamproite magma may originate within the diamond stability field, the magma may be subjected to later near-surface oxidizing environments. Such oxidation may show up as high Fe^{3+}/Fe^{2+} ratios (high hematite component) in ilmenite. In such cases, it has been suggested that the metastable diamonds in the host magma may be substantially resorbed to produce graphite, CO_2, or CO.

As the kimberlitic magma ascends through the uppermost mantle, the diamond xenocrysts are subjected to rapidly changing conditions that could lead to resorption and/or graphitization. Survival of diamond at elevated temperatures outside the diamond stability field is linked to low oxygen fugacity; elevated oxygen levels favor resorption. Ferrimagnetic ilmenite high in Cr_2O_3 is found in some diamond-poor kimberlites, and these ilmenites characteristically show exsolution texture. It is speculated that exsolution textures develop during slow ascent of the kimberlite, during which time diamonds are resorbed.

In contrast, homogeneous ilmenites are found in kimberlites that are interpreted to have risen comparatively rapidly. Within a fractionating magma, early-formed ilmenite may be Mg and Cr rich. Subsequent coprecipitation of ilmenites, silicates, and possibly carbonates from the megacryst parent magma typically results in later ilmenites that have lower MgO and Cr_2O_3 contents.

It has been reported that ilmenite in equilibrium with diamond contains almost no Fe^{3+}, indicating a highly reduced environment. Thus maximum diamond preservation would be expected in kimberlites where low Fe^{3+} picroilmenites prevail.

High Cr_2O_3 and MgO components in ilmenite relate to low oxygen fugacity. This association has led to the use of Cr_2O_3/MgO plots to evaluate ilmenite trends for diamond preservation (Figure 16.1c). MgO depletion along with Cr_2O_3 enrichment is thought to relate to coupled substitution accompanying the addition of Fe^{3+} in Fe^{2+} lattice sites (i.e., increased oxidation state). One means of increasing the Mg content, but not the Cr content, of the magma hosting the ilmenite megacrysts is the decomposition of accompanying magnesite megacrysts.

Picroilmenite in weakly diamondiferous kimberlite may have either a low or a high average hematite component. Barren pipes generally contain high hematite in the ilmenite suite, which is expressed as lower Mg and low to high Cr content, representing the low-Mg side of the parabolic Mg-Cr distribution (Figure 16.1c). The ilmenites from most of the significantly diamondiferous kimberlites contain only 10% hematite component on average, and they tend to be dominated by high-Mg, high-Cr ilmenites (McCallum and Waldman 1991).

Gurney (1989) and Gurney, Helmstadt, and Moore (1993) report that "ilmenites with low Fe^{3+}/Fe^{2+} ratios are associated with higher diamond content than those with more Fe^{3+}, whereas diamonds are not associated with ilmenites of high Fe^{3+} content at all." Griffin et al. (1997) reported that ilmenite megacrysts from several significantly diamondiferous kimberlites (Orapa, Koffiefontein, Kimberley/Wesselton, Liqhobong, and Schuller), several weakly diamondiferous kimberlites (Frank Smith, Monastery, Kao, Klipfontein, Kamfersdam), and several barren kimberlites support Gurney's observations.

However, this association is not supported by all observations. As pointed out by Schulze et al. (1995) and Coopersmith and Schulze (1996), on the basis of ilmenite geochemistry, an exploration geologist would be forced to conclude that finding diamonds in the Mir, Frank Smith, DeBeers, Monastery, and Kelsey Lake mines would be unlikely because these kimberlites all have ilmenites with high hematite component. Yet, unresorbed diamonds and relatively high ore grades are found in kimberlites at Mir (200 carats/100 tonnes), Frank Smith (known for its sharp-edged octahedrons), DeBeers (90 carats/100 tonnes), and Monastery (50 carats/100 tonnes). Low diamond grades are reported at the Kelsey Lake mine, but the diamonds are excellent and include many spectacular gem-quality octahedrons with little evidence of resorption. The ilmenite geochemistry of Kelsey Lake shows as much as 38% hematite component (Schulze et al. 1995; Coopersmith and Schulze 1996) which would lead to a prediction, based on ilmenite geochemistry, that these kimberlites would be devoid of diamond. However, diamond production at the mine includes a large percentage of high-quality gemstones with octahedral habit indicating that diamond preservation was favorable.

Another group of nearby kimberlites in the Iron Mountain district of Wyoming is a frequently cited example of a diamond-free group of kimberlites with oxidized ilmenite compositions (Gurney 1989; McCallum and Waldman 1991; McCallum and Vos 1993). Because of the ilmenite geochemistry, the majority of these kimberlites unfortunately have not been systematically sampled for diamond even though the indicator mineral geochemistry is essentially identical to that in the Kelsey Lake kimberlites. Limited sampling at Iron Mountain in the early 1980s recovered one macrodiamond (H.G. Coopersmith, personal communication, 1997).

In all probability, many picroilmenite nodules did not coexist with the magma at the time they were incorporated into the kimberlite. Therefore, unless the ilmenites have completely reequilibrated with their host kimberlite, their oxidation state would have little bearing on the diamond resorption potential (Schulze et al. 1995; Coopersmith and Schulze 1996). Schulze (1995) suggests that the cores of picroilmenites will not reveal the oxidation state of the host magma, and therefore microprobe analyses should concentrate on gathering data from the rims of picroilmenite if a link between ilmenite compositions and diamond preservation can be verified.

Ni and Zn Thermometers

Cr-pyrope garnet and high-Cr chromite are widely used as indicator minerals in exploration for diamond. The use of the garnets in evaluation of prospects requires a search for G10 harzburgitic garnets and high-Cr chromites that have chemistry similar to diamond inclusions. But according to Griffin, Gurney, and Ryan (1992), this criterion is ambiguous in several respects. Some diamondiferous pipes, such as the Argyle, contain few (if any) G10 garnets, whereas some barren pipes such as Zero and Buljah, Western Australia, contain abundant G10 garnets.

The use of the nickel and zinc thermometer may account for some of these shortcomings. The following summary is based on work by Griffin, Gurney, and Ryan (1992). According to their work, the nickel thermometer, T_{Ni}, is based on strongly temperature-dependent partitioning of Ni between Cr-pyrope garnet and olivine in mantle-derived xenoliths. The variation of Ni distribution coefficient D^{Ni}, defined as (ppm Ni in garnet)/(ppm Ni in olivine) is almost entirely due to variation in the Ni content of the garnet. The Ni content of mantle olivine is assumed to be essentially constant at 3,000 ± 300 ppm. This relationship allows the construction of a thermometer based on the assumption that each garnet has equilibrated with olivine that had a uniform Ni content of 3,000 ppm.

Thus the temperature of equilibration of a single grain of Cr-pyrope to a typical accuracy of ±50°C is possible using a proton microprobe analysis of its Ni content. Only grains with Cr >1% (Cr_2O_3 1.5%) are analyzed in order to eliminate most websteritic garnets, which did not equilibrate with olivine. The thermometer is useful on both G10 and G9 garnets.

Garnets with T_{Ni} in the range of 950° to 1,250°C are thought to reflect derivation from within the diamond stability field. Diamond-rich kimberlites appear to have abundant Cr-pyrope Ni temperature estimates within this "diamond window." Pipes containing a high proportion of garnets with T_{Ni} <900°C are suggested to contain low diamond grades. Many poorly diamondiferous and barren pipes contain abundant high T_{Ni} (>1,250°C) garnets with high Zr, Ti, and Y contents. Garnets rich in Ti and Zr with equilibration temperatures above 1,250°C (i.e., derived from high-temperature deformed peridotite xenoliths) are negatively correlated with diamond content.

Geotherm parameters are estimated directly from the garnet concentrates. The Cr content of chromites coexisting with garnet increases as a function of pressure and is insensitive to temperature. The Cr content of the coexisting garnet, however, is a function of temperature as well as pressure. Thus, the maximum Cr content of garnet at any temperature will be found in garnet coexisting with chromite, and the Cr content will depend on the P–T relationship (i.e., on the geotherm).

Semiquantitative estimates of diamond potential are possible. Ω = (percentage of garnets in the "diamond window") − (percentage of garnets with T_{Ni} >1,250°C). This parameter takes into account the degree of sampling of potential diamond-bearing mantle and the detrimental effects of magma interaction on diamond preservation. Barren pipes have values of Ω <10; no pipe with Ω <30 has a grade above 15 carats/100 tonnes, and all pipes with grades >30 carats/100 tonnes sampled by Griffin, Gurney, and Ryan (1992) have values of Ω ≥ 50.

Other compositional parameters in concentrate garnets allow a refinement of the grade estimation provided by T_{Ni}, by identifying diamond-poor mantle. For example, the Zr content of concentrate garnets differs from 0.250 ppm. The median Zr content of the garnets in the 950° to 1,250°C range of each concentrate and the grade of the pipe in carats/100 tonnes is plotted against Ω. It is apparent that the pipes with moderate Ω (20–50) but low grades have higher median Zr contents. The variation in Zr may be related either to the degree of original depletion of the mantle or to later episodes of metasomatism. In either case, it appears that high median Zr in median-temperature garnets is not compatible with high diamond grade. The data indicate that the grade of a pipe is determined primarily by the depth range it sampled and secondarily by processes such as metasomatism (by magmas or fluids) that have affected that mantle volume.

Griffin, Gurney, and Ryan (1992) also indicate that the Zn content of chromite coexisting with olivine is inversely proportional to temperature of equilibration, whereas Ni contents show a weaker positive correlation with temperature. The Zn relations, in particular, provide a thermometer that can be used to establish the relative temperature (and thus pressure) of individual chromite grains and to identify grains that have been derived from the diamond stability field.

In a study of the Prairie Creek and Twin Knobs #1 diamondiferous olivine lamproites near Murfreesboro, Arkansas, Griffin et al. (1994) showed that the higher-grade lamproite at Prairie Creek (13 carats/100 tonnes) had incorporated a larger proportion of mantle-derived material from the diamond-stability field than the Twin Knobs #1 lamproite (0.17 carats/100 tonnes), on the basis of an evaluation of nickel thermometry of pyrope garnet and Zn thermometry of chromite. The garnets in the lamproites show some mildly subcalcic G10 pyropes but no strongly subcalcic garnets. Griffin et al. (1994) noted that this association appears to be characteristic of lamproites worldwide. On the basis of the Ni thermometer, a greater proportion of the Prairie Creek lamproite garnets showed characteristics consistent with derivation from the diamond window.

Fipke, Gurney, and Moore (1995) criticized some of the assumptions related to the thermometers. For instance, they report that the Ni thermometer

> *ignores the vagaries of the kimberlite sampling processes in the mantle and also fails to consider the paragenesis of the garnets. Furthermore, the "diamond window," initially selected with hindsight provided by available information on the Kaapvall Craton is not universally applicable.*

Additionally, variable geothermal gradients introduce further uncertainty in using the Zn thermometer. In other words, the geothermal gradient must be assumed prior to establishing an equlibration depth from the calculated temperature; thus incorrect assumptions can lead to errors in pressure estimates (Kjarsgarrd 1992; Schulze 1995).

Diamond Facial Factors

Attempts to evaluate diamond potential using indirect indicators are designed to avoid costly bulk sampling tests. These indirect methods are based on assumptions about geologic factors that are assumed to reflect the conditions of crystal growth and preservation of diamonds. Another attempt at determining the diamond potential relies on the recovery of diamond in bulk samples. This method requires large bulk samples to recover diamonds to test, which makes it less practical than using small samples to test the indicator mineral geochemistry.

In one bulk-sample exploration test method used by the Russians, the following factors were considered: chemistry of kimberlitic magma, degree of diamond crystal preservation, and the physical properties of diamond, in particular luminescence.

The following equations can be found in a series of publications by Milashev (1965, 1972, 1974) that were later summarized by Milashev and Sokolova (2000). Refer to these works for all appropriate details of the derivation of equations.

This method evaluates the chemistry of kimberlite to assist in the determination of diamond content. This concept assumes that the diamond content in kimberlite is a function of chemical and facial factors.

$$D = f\ (CD, FD) \qquad \textbf{(EQ 16.1)}$$

where:

D = diamond content of kimberlite
CD = chemical factor of diamond
FD = facial factor of diamond

The diamond quality is affected by the chemical composition of the kimberlite. The quality is interpreted to be a quantitative ratio between different crystalline forms (octahedral, dodecahedral, and transitional) and polycrystalline aggregates.

Equations expressing the chemical potential for diamond (CPD) were described in Chapter 6, Kimberlite. In the evaluation of diamond deposits, the quality of diamonds is often more important than quantity. Thus another formula was introduced to express the morphology of diamond, and it is known as the degree of diamond preservation of the crystal (DPC) (see Chapter 6, Kimberlite, Equation 6.3).

$$D = f(\text{CPD}) \times \psi(\text{DPC}) \tag{EQ 16.2}$$

An empirical justification of equation 16.2 is made in order to obtain results in mg/m^3. Considering this, Equation 16.3 is developed:

$$D = \left(\text{CPD} + \frac{\text{CPD}^3}{120} - 1.75\sqrt{\text{CPD}}\right) \times \frac{\text{DPC}^2}{72} \tag{EQ 16.3}$$

This equation can be used to calculate the diamond content for an average weight of diamond crystals equal to 7mg.

In general CPD, DPC, and the average weight of diamonds in kimberlite are necessary and sufficient for the calculation of the diamond weight content in kimberlite using the following equation:

$$D_k = \left(\text{CPD} + \frac{\text{CPD}^3}{120} - 1.75\sqrt{\text{CPD}}\right) \times \frac{\text{DPC}^2}{137} \times \sqrt[3]{p} \tag{EQ 16.4}$$

where:

D_k = diamond content in kimberlites, mg/m^3
p = average weight of diamond crystals in the studied pipe, mg

Calculation of D_k is simplified by using a nomogram presented in Figure 16.2, which permits one to determine CPD and DPC graphically and to determine the average diamond content of the kimberlite with average weight of crystals equal to 7 mg. In pipes in which the average weight of crystals is different, D_k needs to be multiplied by coefficient $\sqrt[3]{p}/\sqrt[3]{7}$.

These equations have been checked on samples taken from several Siberian and South African kimberlites. The results show a high degree of correlation.

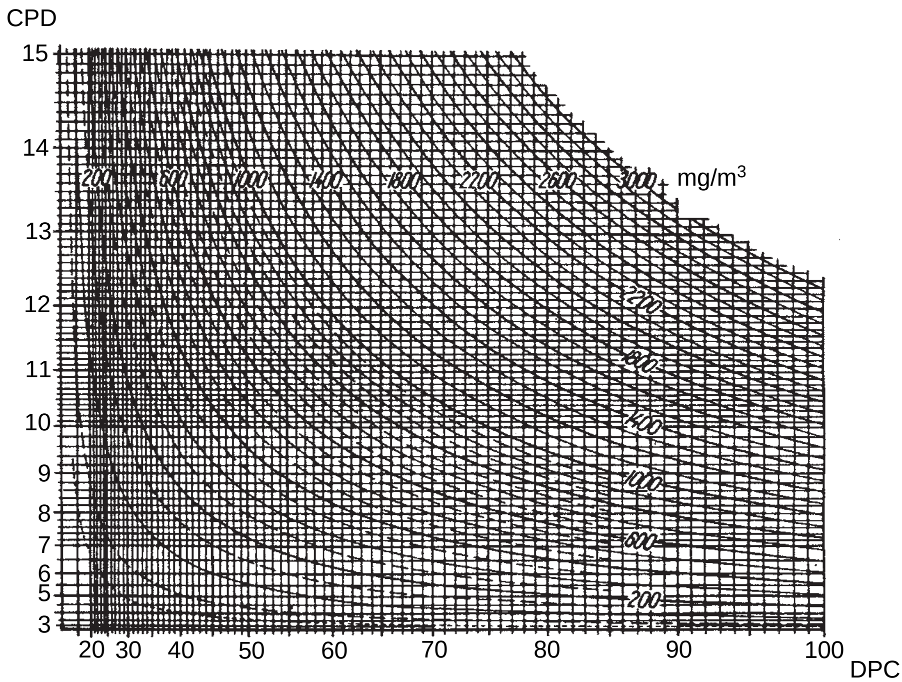
Two stages influence the formation of diamond. The first corresponds to the time of diamond crystallization. During the second, diamonds were dissolved, graphitized, and possibly mechanically deformed. Thus crystal morphology (*CM*) must reflect the degree of thermodynamic favorability for the creation of high-quality diamond crystals during both stages.

As we discussed in Chapter 6, Kimberlite, the dependence of diamond-content is expressed as CPD (Equation 6.2). Because of the leading role of titanium, it is referred to as the titanium indicator, and it is considered as an indicator of the degree of differentiation of the kimberlitic magma (DD_{Ti}).

The crystal morphology (*CM*) is considered as a function of thermodynamic conditions, and accordingly it can be used to evaluate of the facial factor of the quality of diamond crystals (*FQ*):

$$FQ = f(CM) \tag{EQ 16.5}$$

Another important factor of diamond potential is a degree of graphitization. This factor is expressed as



CPD = Coefficient of Potential Diamond Content

DPC = Degree of Preservation of Crystals

FIGURE 16.2 A nomogram for determination of diamond content in kimberlites by morphology of their crystals. Numbers near curves show diamond content of pipes with an average weight of diamond crystals equal to 7 mg (modified from Milashev and Sokolova 2000).

$$DG \approx \frac{1}{\mathrm{DPC}} \qquad \textbf{(EQ 16.6)}$$

where:

DG = degree of graphitization
DPC = degree of preservation of crystals

The rate of drop in pressure in a magmatic system is determined using the amount (or percentage) of polycrystalline diamond aggregates P, which is used as a measure of the rate of change of thermodynamic parameters.

As pressure declines, eventually a point is reached where diamond becomes metastable and is replaced by graphite. If the transition to graphite occurs spontaneously, all diamonds are replaced by graphite. The absence of the spontaneous transition of diamond to graphite in kimberlite indicates that the polymorph transformation is of a catalytic character. The role of the catalyst is played by anions of metatitanic acid $[Ti^{4+}O_3]^{2-}$.

If the graphite coating or the surface of monocrystals and polycrystalline aggregates have been eliminated during magma ascent (with consequent pressure drop), newly exposed portions of the crystals are graphitized. In the course of this process, the apex of the diamond crystals will resorb first, followed by the edges and facets. As a result, flat-faceted octahedral crystals will be replaced by curved forms with dodecahedral habit. In this process, the diamond may lose up to 75% of its original weight. Owing to the morphologic similarity of natural curved-facet diamonds with diamonds that have partially dissolved (or more precisely, burned) in melts of alkalis and some other components, the rounding of diamonds is often called "dissolution."

The relative quantity of polycrystalline diamond aggregates can be used as a measure of the favorability of the thermodynamic regime during the period of diamond crystallization.

Thus, an indication of the facial factor of the quality of diamonds can be expressed by the following equation:

$$FQ = f\left[DD_{\mathrm{Ti}}, \sqrt[3]{QP}, \frac{1}{\mathrm{DPC}}\right] \quad \textbf{(EQ 16.7)}$$

where:

FQ = diamond quality factor
DD = degree of differentiation expressed by titanium indicator
QP = quantity of polycrystalline diamond aggregates used as a measure of rate of change of thermodynamic parameters
DPC = degree of diamond preservation

Most diamond crystals are characterized by the ability to luminescence under ultraviolet light or x-rays. The main colors of luminescence are blue, yellow, and green. More rarely orange and red-pink colors are detected; they are referred to as "others." Sometimes a significant percentage of diamonds is characterized by indefinite (white) colors of luminescence or have no detectable luminescence.

An approximate expression of the relative quantity of differently luminescent diamonds and the iron-titanium ratios in kimberlite is as follows:

$$\frac{y + \sqrt{others}}{\sqrt{b + 0.5g}} \quad \textbf{(EQ 16.8)}$$

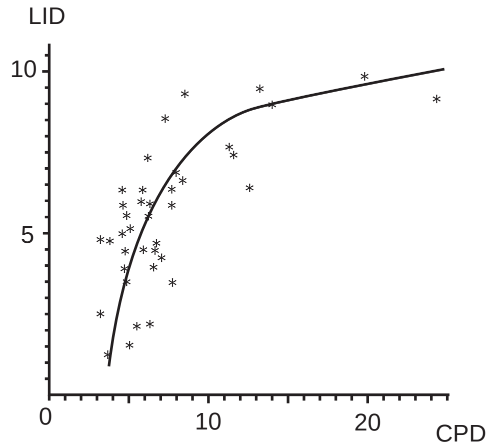
where:

y = percentage of diamonds with yellow luminescence
b = percentage of diamonds with bluish luminescence
g = percentage of diamonds with greenish luminescence
$others$ = percentage of diamonds with "other" colors of luminescence

On the basis of data obtained from 37 kimberlite pipes, the correlation coefficient between the latter formula and Fe/Ti ratio is 0.713 ± 0.081. An equation for the calculation of the luminescence indicator for diamonds (LID) is shown below:

$$\mathrm{LID} = \frac{f(b, y, g, others)}{\psi(b, y, g, others) + \mathrm{constant}} \quad \textbf{(EQ 16.9)}$$

As a first approximation, $f(b, y, g, others)$ can be considered equal to the ratio of Equation 16.8 and $\psi[b, y, g, others] \approx y + others + b + 0.5g$. However, diamond's luminescence is associated not only with iron content in the melt but also with content of many other components including aluminum and potassium. In this case $\psi(b, y, g, others)$ may



LID = Luminescence Indicator of Diamond Content

CPD = Coefficient of Potential Diamond Bearings

FIGURE 16.3 A dependence of the quantity of diamonds with various colors of luminescence from chemical composition of kimberlite for 37 kimberlite pipes in Yakutia (modified from Milashev and Sokolova 2000)

be tentatively considered as equal to y + 0.5(*others*) + $\sqrt{b + 0.5g}$ while simultaneously taking the constant out of the denominator. After all appropriate transformations the final equation, for convenience, is multiplied by 100. Thus, the final equation will take the following form:

$$\text{LID} = \frac{100 \times (y + \sqrt{others})/\sqrt{(b + 0.5g)}}{y + 0.5\, others + \sqrt{(b + 0.5g)}} \qquad \textbf{(EQ 16.10)}$$

Using material from 37 kimberlite pipes from Yakutia a dependence between LID and CPD has been established with a positive correlation coefficient of +0.66 ± 0.09 (Figure 16.3).

It is suggested that the diamond potential of a lode source (CPDf) can be assessed using the luminescence of diamonds (LID) obtained from panned samples and placers (Figure 16.4).

All of the formulas need to be verified using worldwide data. If they can be confirmed, their use will provide an important tool for evaluating diamond potential and economic value of diamond deposits at a comparatively early stage in exploration.

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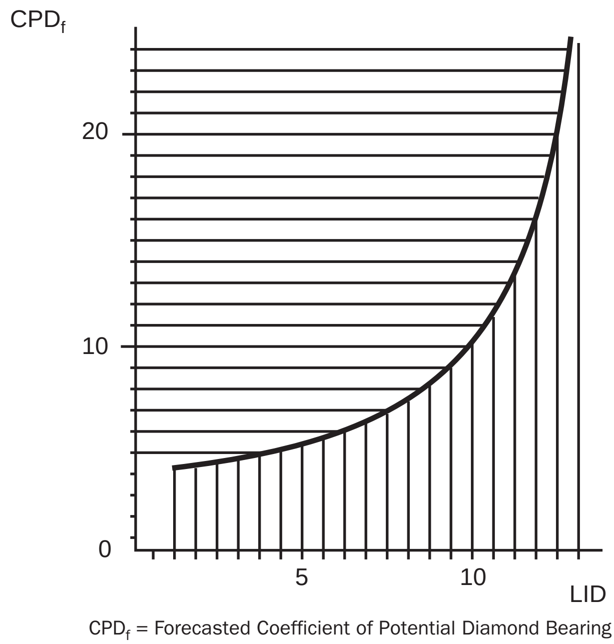

FIGURE 16.4 A nomogram for graphical determination of forecasted coefficient of potential diamond content of kimberlites using meaning of luminescence indicator of diamonds (modified from Milashev and Sokolova 2000)

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CHAPTER 17

Mining and Recovery of Diamonds

ECONOMIC DIAMOND DEPOSITS

Most diamonds are assumed to crystallize beneath cratons at depths of 150 to 200 km at temperatures of 1,050° to 1,200°C and pressures on the order of 45 to 55 kbar. At these depths, the source of most diamonds in kimberlite, lamproite, and lamprophyre lies near the base of thick, cool Archean or Proterozoic cratonic lithosphere. Diamonds are transported from these depths as accidental xenocrysts in kimberlitic or lamproitic magma.

Economically significant diamondiferous kimberlite and lamproite are extremely rare and appear to be restricted to ancient cratons or cratonized provinces older than 1.8 Ga (Jaques 1998). Kimberlites and lamproites occur in clusters of a few to more than 100 intrusives. Within a specific region of a craton, several generations of kimberlite and lamproite may occur. Some may be diamondiferous and others may be barren. Statistics indicate that about 10% of all kimberlites contain diamond. Less than 1% contain diamonds in economic concentrations of <1 ppm (Lampietti and Sutherland 1978).

Other possibilities for economic diamond deposits are entrapment of diamonds in magma erupted from Benioff zones along plate margins, in obducted peridotite or eclogite complexes, in ultrahigh pressure metamorphic schists, and in impact diatremes. Information on economic diamond occurrences in these types of deposits is unavailable as of 2001, but some of them may be a source of economic deposits of diamonds in the future.

Economic diamond deposits depend upon many factors, such as the average price of the stones, the amount of waste material to be removed, methods of mining, and the socioeconomics of the area of interest. For example, some diamond deposits can be mined relatively inexpensively in some regions of India owing to the availability of an inexpensive labor force, whereas in the United States high labor and mining costs require a higher-grade ore for commercial operation. Diamond deposits can be mined at a much lower cost in developed areas where infrastructure is already in place than in remote areas such as Lac de Gras in the Northwest Territories of Canada.

More than half of the world's natural diamonds is mined from kimberlite and lamproite, and the other half is obtained from placers and paleoplacers. The ore grade of kimberlites is generally so low that diamond concentrations rarely exceed 500 carats per 100 tonnes (5.0 carats/tonne). The ore value depends on the value of the contained diamonds, and thus higher-grade kimberlites with low-quality diamonds may be less valuable and even uneconomic when compared with a deposit having lower-grade ore but a large proportion of high-quality gemstones.

The majority of commercial diamond mines process from 1 million to 350 million tonnes of ore, and the richest deposits contain reserves of 500 million carats of diamonds valued in the billions of dollars.

Annual mine production may range from 3 to 6 million tonnes of ore per year for larger mines. Smaller mines may process less than 1 million tonnes per year. Annual diamond production may range from several thousand carats to a few million carats. For example, the Finsch mine, South Africa, produced 5 million carats annually between 1981 and 1991. The Venetia mine, South Africa, produces about 5.9 million carats per year. Annual diamond production for the extremely rich Argyle lamproite, Australia, reached as much as 39 million carats. Open pit diamond mines may produce as little as 100,000 tonnes per year or as much as 10 million tonnes of ore per year.

Diamondiferous pipes with preserved crater facies can be significant sources of diamonds. The difference between small and large kimberlitic diamond deposits is commonly a function of the original diatreme size and erosional level.

Ore grades vary considerably. Average ore grades are reported as 680 carats/100 tonnes for Argyle to as low as about 5 carats/100 tonnes for Kelsey Lake, Colorado (MacKenzie Bay International, Press release, April 18, 2000). In addition, Argyle has produced zones of crater facies that yielded as much as 2,000 carats/100 tonnes. Most economic deposits produce more than 30% gem-quality diamonds. The cutoff grade for economic pipes is typically >10 carats/100 tonnes (Jaques 1998), but this is highly dependent on mining costs and the value of the recovered diamonds.

Currently, the Argyle mine produces about 30% of the world's diamonds and some rare pink diamonds that are the most expensive gemstone in the world. Some "Argyle Pinks" have been valued at $1 million (Australian) for a 1-carat brilliant cut (Rock et al. 1992).

Economic amounts of diamond in lamproite appear are generally confined to volcaniclastic crater-facies olivine lamproite. Magmatic olivine lamproite and magmatic and volcaniclastic leucite lamproite have so far not been found to contain economic amounts of diamond (Scott-Smith 1996), even though some of these rocks do contain diamonds.

Diamondiferous lamproites have been known for many years, although these rocks did not attract serious exploration interest until an exploration program in Western Australia proceeded on the concept that lamproites were differentiates of kimberlite (Prider 1960; Carmichael 1967). This program led to the discovery of diamondiferous lamproite in the Ellendale field in the Fitzroy basin in 1976 through 1977 and led to the discovery of diamondiferous olivine lamproite at the Argyle vent in the East Kimberleys in 1978 (Mitchell and Bergman 1991).

Prior to 1978, kimberlite was considered to be the only host rock for commercial amounts of diamond. However, the discovery of diamondiferous lamproite in the Kimberley province in Western Australia generated interest in these unusual and extremely rare rock types. Following this discovery, a few other diamondiferous rocks were reevaluated and found to be lamproites rather than kimberlite. These deposits included the Majhawan olivine lamproite in India that had been discovered and mined for diamonds as early as 1827, long before diamonds were found in kimberlite in South Africa in 1866, and the Prairie Creek olivine lamproite in Arkansas, which was mined for diamonds as early as 1842.

Where lamproites contain diamond, the diamond is generally restricted to pyroclastics: the magmatic phases (and sills) are notoriously diamond poor because diamond will burn or revert to graphite in the high temperatures sustained in the flows. Thus, the available minable tonnage is limited to the vent facies rocks. Typically, diamond grades are higher in olivine lamproites than in leucite lamproites. Some petrologists now recognize that Group II kimberlites (orangeite) chemically overlap olivine lamproites. Group II kimberlites have been a significant source of gem diamonds (Peterson 1996).

Diamonds from commercial lamproites are typically smaller than from most commercial kimberlites. The average size of Argyle diamonds is <0.1 carat, whereas those from Ellendale are 0.1 to 0.2 carat. The largest reported stones from these lamproites are 6 and 16 carats, respectively (Mitchell and Bergman 1991). The largest reported diamond from Prairie Creek, Arkansas, is 40.42 carats (Bolivar 1984).

Kimberlites, on the other hand, are known for some very impressive stones, including several very large diamonds weighing more than 100 carats. The largest diamond ever recovered from a kimberlite was the Cullinan, which weighed 3,106 carats. Larger diamonds may be found in lamproites in the future, but the general nature of lamproites (smaller xenoliths and lack of megacrysts) suggests that there may be a limit to the size of diamonds that lamproites transport to the surface.

Lamproites also produce many frosted diamonds with resorbed textures. Such diamonds are for the most part poor quality, and the majority of diamonds that have so far been recovered from lamproite are gray, brown, and yellow and are dominantly industrial quality.

Many kimberlites contain more than one diatreme facies along with associated hypabyssal facies and possibly some crater facies. Each facies may possess different ore grades. In most cases a commercial pipe will occur within a large cluster of subeconomic and barren intrusives.

Where there is evidence of considerable erosion of the upper portions of a diamondiferous kimberlite pipe (and lamproite), exploration for commercial placer deposits should be a high priority. Even subeconomic pipes may produce nearby commercial placers.

MINING

In the evaluation of kimberlite pipes, indicator-mineral studies require initial samples that may weigh several kilograms to a few hundred kilograms. If the initial samples are favorable, further samples are collected for diamond verification and characterization. Three phases of sampling for diamond are considered: diamond verification, character sampling, and bulk sampling.

In the initial phase, diamond verification may require a sample of at least 65 kg to several hundred kilograms to extract microdiamonds using caustic fusion or acid digestion to remove the kimberlite or lamproite gangue from the diamonds. Care must be taken during this phase to retain any microdiamonds during processing, because these diamonds are easily lost owing to their small size and their hydrophobic and electrostatic properties. During this process, clays must be completely broken down, because they hinder the extraction of diamonds from the test sample.

Following verification, the next phase (diamond characterization) will require the recovery of many microdiamonds and macrodiamonds. To recover this parcel, a small bulk sample of the intrusive, typically 100 to 500 tonnes is required. At this stage, the sample must be classified to greatly reduce the gangue. To recover diamonds, sample reduction will require a small testing facility, which may use jigs, heavy media cyclones, and/or rotary pans. Diamonds are extracted from the final concentrates by the use of x-ray fluorescence and/or grease tables or some other extractive technique.

If the second phase is favorable, the next phase in evaluation will require the recovery of a large parcel of diamonds (1,000–10,000 carats) from a 5,000 to 100,000 tonne bulk sample. This stage is on the scale of a small mining operation (Coopersmith 1993).

Commercial diamondiferous kimberlite and lamproite may range from narrow dikes a meter or so wide to pipes of about 30 m to about about 1,500 m across. The pipes range in surface area from about 1 ha to more than 150 ha. The average size of a commercial diamond pipe is reported to be 12 ha (Jaques 1998). Even some narrow blows and dikes may host economic quantities of diamond. For example, the Bobi lamproite dike in the Ivory Coast of Africa locally yields grades up to 1,000 carats/100 tonnes (Helmstaedt 1993).

Many kimberlites contain a thin layer of yellow ground, highly oxidized and weathered kimberlite that has the appearance of a tawny gossan exposed at the surface. Depending on the climate, which controls the amount of weathering and erosion, the yellow ground may or may not be present. Beneath the yellow ground, a layer of clay-rich blue ground may persist. The depth of blue and yellow ground will vary from region to region. Resistivity sounding surveys over some Wyoming kimberlites indicated that blue ground continued to as much as 36 m depth (Hausel, McCallum, and Woodzick 1979). Thus initial open pit mining in this blue ground would be relatively inexpensive, because blasting would not be necessary until the hard rock kimberlite was intersected at depth. This was the case at the Kelsey Lake mine in Colorado where the initial open pit operations in 1998 required scraping but no blasting (Howard Coopersmith, personal communication, 1997).

Potentially commercial pipes are sampled in detail, and a map of the diamond grades determined from sampling is produced to assist in a mining plan. Sampling is done both on the surface and by drilling in order to achieve a three dimensional view of ore grades. If the pipe is considered to be economic, planning is completed for the initial open pit design as well as for a mill to be located near the pipe. Open pit mining may proceed into the pipe from a spiral road developed from the rim of the pit toward its center. As work proceeds, the country rock is cut back in steps to aid in supporting the highwalls of the open pit. Mining in the pit may occur in an oval pattern, or in a polygonal pattern, because it is easier to mine in a straight line (Bruton 1979).

Following the depletion of the overlying yellow and blue ground, the open pit will reach the hard rock kimberlite. This kimberlite is serpentinized, but still relatively hard rock.

Most kimberlites (and lamproites) are initially mined by open pit operations that progress to depth. As the pipe narrows at depth the open pit will shrink to a smaller and smaller diameters. Mining operations may ultimately continue as an underground mine using bulk mining recovery. However, less than 30% are developed into underground mines. The diamond ore must be relatively high value, because the cost of underground mining is considerably higher and the amount of ore recovered is considerably lower than from an open pit. Some kimberlites in Siberia and South Africa have been mined to depths of 1,080 m. The larger open pits may have mine lives of 2 to 50 years (Jaques 1998).

Underground mines commonly use block caving for ore extraction. According to Bruton (1979), block caving was first used in diamond mining in 1955 in South Africa, and within a few years, all underground operations in the Kimberley pipes were converted to block caving. This process considerably reduced the number of levels and shafts in the mine, as is needed for continuous mining and mechanization, and it was safer for mining.

Typically, block caving establishes a haulage mine level 122 to 183 m below the top of the hard rock kimberlite. At intervals along the haulage level, raises are driven upward for ore extraction. After the raises are established, ore is removed above the raises, and funnels are cut above draw points at the base of the raises. Then the entire area above the funnels is mined out to a height of about 2 m to produce a large opening, in which the roof peels and slowly collapses under its own weight. The collapsing kimberlite will fill the draw points in the raises until the entire opening is filled with broken ore.

The ore is then extracted through draw points into the underlying drifts. Parallel to these drifts, scraper tunnels are established at 13.7-m intervals, and scraper stations are located at the end of the tunnels. These stations are cut into the adjacent country rock, and winches are housed in the stations. The winches pull scraper buckets that drag the diamond ore to drawpoints leading to the primary crusher.

Scrapers are driven back and forth through the level by winches with chains. The ore is scraped to a decline that allows the ore to be hauled to an underground crusher station prior to extraction to the surface through the production shaft.

Placer diamond mining is simpler than hard rock mining and requires a different type of mining operation. In many cases, placer diamond mining is similar to placer gold operations.

The size of the placer mine operation will vary from a single person operation to a full-scale operation using bulldozers and scrapers. Some marine placer mining operations are more labor and equipment intensive and may use offshore dredges.

In general, the placer operations attempt to define paystreaks in creeks, rivers, or beaches. The paystreaks are then mined using small- or large-scale earth moving equipment. The non-productive overburden is removed to expose paystreaks, which are carefully mined (Bruton 1979).

MILLING

Beneficiation of diamond ore may require different processes depending on the nature of the ore and the type of mining used to extract diamonds. Mining hard rock diamond ore requires a mill for ore dressing.

During the planning stages of an open pit or underground diamond mine, a mill site is established near the ore deposit in order to reduce the cost of transporting ore from the mine to the mill. During this phase of operation, care is taken to ensure that the mill site does not interfere with the development (and potential expansion) of the mine. The mill site is drilled to ensure that it is not constructed on a blind diamond pipe, sill, or dike. This is an important consideration, as it is costly to move a mill.

Diamond ore mined from a pipe must be treated to extract the diamonds. Once the hard rock kimberlite or lamproite is mined, it may be treated on the surface by sprinkling water on the material after it is received at the primary crusher.

The diamond ore is beneficiated to free diamonds. There are several methods used to upgrade the ore that are considered environmentally safe (toxic chemicals are rarely involved in the extraction of diamonds in a full-scale mining operation). After the hard rock ore is mined, it is crushed and then screened. The screening process separates midsize material from larger material rejects and from material too small to contain commercial diamonds. Decisions on the maximum screen size must weigh the cost of processing additional material with loss of potentially priceless large diamonds.

A typical mill uses a basic flow sheet that begins at primary milling and continues on to primary gravity concentration, secondary concentration, magnetic separation, attrition milling, and to the final diamond extraction stage using grease tables, electrostatic separation, and/or x-ray fluorescence extraction (Bruton 1979). Gravity separators are used to separate diamonds from minerals of lower specific gravity. Many of these concentrators are used for alluvial mining (although some are also used in hard rock milling) and may include riffled sluices, spiral concentrators, jigs, shaking tables, rotating cones, and heavy media separators (HMS).

Crushing

Crushing is the first stage of extracting diamonds from hard rock kimberlite, lamproite, or cemented conglomerate. Crushing operations must take care to avoid breaking diamonds. In some cases, the diamond ore will be processed in a ball-mill with a rotating drum or cylinder filled with water and rocks or ball bearings. In other operations, the ore may go directly to crushers or to some similar milling facility to break the ore down to a manageable size.

Washing and Screening

Washing is usually combined with screening to remove fine material. The screening process uses a grizzly (the coarsest screen) formed of parallel steel rods that are designed to reject coarse rock. The material passing through the grizzly will enter a set of screens. The finest screen is designed to filter and reject very fine sand and any diamonds that are too small to facet.

The screened material is washed. Rotary washing pans are often used to separate waste from concentrates. The rotary pan consists of a large circular pan (≥1 m diameter) that has a series of rotating rakes. The water slurry placed in the pan effectively separates denser material from less dense material. The waste flows over the inner edge of the pan while diamonds and other material with relatively high specific gravity settle to the bottom of the pan. After washing, magnetic separation may be used reject any magnetic material in the concentrates.

A pulsator is used in a diamond mill to further upgrade the concentrate. The concentrate is placed on top of a bed of “bullets” in water lying on a sieve. The pulsator produces a rapid pulsating movement causing the material with higher specific gravity to descend between the “bullets” while material of lower specific gravity floats off into a waste pile.

A jig is a unit in which the feed is stratified in water by means of a pulsating motion. The pulsating motion is obtained by alternate upward and downward plunging action in the water. This motion tends to preferentially draw heavier particles downward while keeping the lighter particles suspended to be drawn off the top of the jig.

Heavy media separator cones are used to process a slurry of heavy liquid. This slurry consists of fine ferrosilicon powder and water, which is mixed to produce a density of 2.7 to 3.1. Material processed in the HMS cone is fed at the top of the cone, and material with densities greater than 3.2 (including diamonds) will sink to the bottom of the cone. Much of the remainder of the concentrate (as much as 95%) is rejected at the top of the cone. The concentrate extracted from the HMS may proceed to a secondary

HMS cone containing a slurry with density of 3.25, to further reduce the amount of concentrate (Bruton 1979). HMS treatment may achieve reduction as great as 1,000:1 of ore to concentrate.

A hydrocyclone is used to process gravels or broken kimberlite, and it employs the same method as the HMS. The ferrosilicon slurry is continuously circulated in a hydrocyclone. A centrifuge effect results in the denser fraction moving outside and down into the conical shaped tank. The lighter material is forced upward and rejected.

Spiral concentrators are a form of sluice that consists of a column of tight spirals in which centrifugal force aids in separating material by specific gravity. Heavier minerals ride near the center of the spirals, while the lighter minerals ride higher on the outside of the spirals. At selected points, the heavy mineral concentrates are extracted from the spirals.

Extracting Diamond

Following beneficiation, the concentrates are further processed to extract diamonds. At this point, only a few options are possible for extraction. These may include skin flotation, electrostatic separation, grease tabling, and/or x-ray fluorescent separation (SORTEX™).

During the initial phases of testing a pipe, another extraction method, caustic fusion, may be used. Caustic fusion is a method used by many exploration groups to extract microdiamonds (≤2 mm) from potentially diamondiferous host rock. In this technique, concentrates are added to caustic soda and heated. This mixture typically consists of 10 parts of caustic soda that is heated in a electric furnace to 650°C for 45 min. The process destroys the gangue and leaves microdiamonds intact.

Other methods of caustic fusion are also used (see Bruton 1979). In some early testing phases by the Wyoming State Geological Survey, acid dilution was used for extraction of microdiamonds. This method required samples to be processed in concentrated hydrochloric acid, prior to a final bath of hydrofluoric acid. Although this method is useful in extracting microdiamonds, it is costly, time consuming, and dangerous.

Because diamonds are hydrophobic, they float on water. This method of diamond extraction is seldom used. When used, it requires the concentrate to be dropped from a very low height onto a glassy water surface. When diamonds contact the surface, they float while much of the remaining concentrate will sink in the water. The diamonds are then extracted from the surface.

This process is effective only on smaller diamonds. Unfortunately, many small diamonds are difficult to extract with a grease table, even after treating the concentrates with a reagent, because much of the fine gangue is also made nonwettable and will stick to the table. Such fine material is processed in an attrition mill, which gradually grinds down the gangue and preserves the diamonds.

According to Bruton (1979), another method of skin flotation treats concentrates in boiling hydrochloric acid, diluted with about three times the volume of water. Decanting and stirring causes the microdiamonds to rise and float, where they are skimmed off the surface of the liquid.

Electrostatic sorting has been used at many alluvial mines. It has also been used to recover microdiamonds that are less than 1.65 mm. In this method, the concentrate is processed on a high-voltage electrostatic unit. The unit consists of a series of horizontal rollers in vertical columns in which the rollers on one side are grounded, while those on the other side and positively charged.

Different minerals act as either conductors or insulators in an electric field, and hence can be separated from one another when charged (MacDonald 1983). Diamonds, being electrical insulators, are unaffected by the charged rollers and fall vertically; however, much of the gangue will achieve a charge and be attracted towards the positively charged rollers (Bruton 1979).

The hydrophobic (non-wetting) property of diamond allows the mineral to stick to nonwettable grease, whereas most of the remaining concentrates will wash over the grease. Many grease tables use about 0.6 to 2.5 cm of axle grease, which is spread on a sloping table (Bruton 1979). Most tables vibrate along a vertical axis.

The grease table used by the Wyoming State Geological Survey is a modified Wilfley Table, which shakes along a horizontal drive. This design is used only because it is less expensive, but it also is effective in extracting diamonds.

Grease tables may contain four or five steps. The grease used by the Wyoming State Geological Survey consists of 1 part paraffin and 10 parts Vaseline™. This grease is melted and poured onto the grease table top to form a relatively smooth surface only a few millimeters thick. When macrodiamonds contact the grease, the diamonds stick. The diamonds are then recovered by scraping the grease off the table and melting the grease in a pot containing water (Hausel, McCallum, and Roberts 1985). The final concentrates may be cleaned by hydrochloric acid prior to the final diamond extraction and sorting.

Some operations use a continuously rotating grease belt. This belt has the advantage of automatically applying grease at one end of the belt, and continuously removing the diamondiferous grease against a stationary knife at the other end.

However, not all diamonds stick to grease. Diamonds extracted from marine deposits are coated with salt making them wettable. Treating the diamond concentrates with fish oil and caustic soda will tend to clean the diamond surfaces making them nonwettable and grease attractive. Some diamonds from certain pipes are also wettable and must be treated. A chemical reagent of maize acid oil with caustic soda is used to treat these diamonds. There are also other soap solutions that will also effectively clean the diamond surfaces (Bruton 1979).

Diamonds fluoresce when irradiated by x-rays. In this process, diamond concentrates are passed through the x-ray sorter. When diamonds fluoresce, the energy is detected by a photoelectric cell that activates a jet of air and blows the diamond into a separate recovery unit. X-ray sorting is very efficient and is now used at all major diamond mines.

The extracted diamonds are cleaned by acid and hand sorted by size, shape, clarity, color, and cutability according to the Central Selling Organization classifications (see Chapter 18, Diamond Markets and Exploration Prejudices). The cutoff size of the diamonds that are recovered is usually about 1 mm. Diamonds equal to or larger than 1 mm are termed “macros”; those smaller than 1 mm are termed “micros.” The mean diamond size may range from 0.01 to 0.7 carats (Jaques 1998).

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CHAPTER 18

Diamond Markets and Exploration Prejudices

INTRODUCTION

The characteristics of diamond provides information about processes deep within the earth. Almost any diamond-related problem creates challenging problems for researchers and has implications for general geological studies in adjacent fields. This started with crystallographic studies of diamond. It is not by chance that decoding diamond's crystalline structure has been considered as a "great triumph" in diamond research (Shafranovsky 1964).

In the previous chapters, we have indicated that there is a consistent association of ultramafic-alkalic magmatic activity (including kimberlitic and lamproitic volcanism) with specific types of platform structures—anteclises and stable blocks. Synchroneity of most types of diamond deposits and a coincidence of the timing of formation of diamond deposits with the timing of global tectonic transformations suggests a common source of volatiles in the earth. Discovery of new types of diamond deposits with fruitful discussions about their origins can have a great impact on concepts of the origin and deepest levels of geologic activities. We saw this in relation specific types ring structures, which are accompanied by great pressure along with the formation of lonsdaleite, and also diamond deposits related to high-pressure metamorphic rocks.

Not until the twenty-first century has the diamond industry come to maturity. The finding of new kimberlite-related deposits and the sharply increased market demand for both industrial diamonds and gems produced a sharp growth in production. At the same time institutionalization of market control and attempts to regulate world diamond markets occurred. The amount of diamonds mined from 1867 to 1984 totals 1,450 million carats (about 290 tonnes). Annual output changed from 2.7 million carats in 1915 to 51.4 million carats in 1984 (Milashev 1989). Of this amount 46% were industrial diamonds, 39% near-gem quality, and 15% gems (Miller 1995).

In 1984, more than 90% of all diamonds were mined in Africa, and about half of the world reserves were thought to be located in the Democratic Republic of Congo (formerly Zaire).

Diamond prices vary considerably. There are approximately 5,000 diamond categories with prices that vary from \$0.5/carat up to several tens of thousands of dollars/carat (for large uncut or colored "fancy" diamonds). Although definition of the term *gem* is partly subjective depending on the size and physical characteristics, the lower boundary of gem diamonds is tentatively recognized at the price level of $80/carat (Miller 1995). Thus the market value of diamonds depends on their quality.

The diamond market is evaluated on the basis of two different figures: total output and the number available gem quality stones. These two figures creates a "dualism" in the market that can result in market volatility, depending on new discoveries of rich gem deposits and the cost of mining the new discoveries.

The diamond market could potentially be affected by the discovery of any new significant sources of gem-quality diamonds. The discovery of a few rich pipes in Botswana resulted in a sharp increase in its national production—from 2,718 million carats in 1975 to 18,911 million carats in 1984. The result of one single discovery at Argyle, Australia, significantly increased the world's diamond production almost overnight. Other discoveries in Zaire increased diamond production in that country from 14,189 million carats in 1970 to 18,470 million carats in 1984 (Milashev 1989). Owing to political instability and a series of civil wars, an enormous amount of smuggling was associated with these diamond deposits.

The dynamism of the diamond market depends on periodic discoveries of new sources. The discovery in 1981 and rapid development of rich gem-bearing kimberlite deposit of Jwaneng, Botswana, which contained 35% gem-quality diamonds, reached production of 5.9 million carats in 1983. At the Venetia Mine in northern Transvaal, South African Republic, in the 1980s, production reached 5.5 million carats/years. The total diamond production in Botswana increased from 544,000 carats (including 60 thousand carats of gem-quality diamonds in 1970) to an estimated production of 12,911,000 carats in 1984 (Milashev 1989).

Other examples of great developments in the diamond industry are provided by the discoveries and consequent developments in Russia. Alluvial gem diamond production was a mere 6% of the total output. Russian diamond production peaked in the mid-1980s to 22 to 25 million carats. The importance of intensive exploration was clearly demonstrated when speculation about difficulties in the Russian diamond industry (falling output and the predicted exhaustion of the Udachnaya pipe, exhaustion of stockpiles at the Mir pipe, and the necessity to go underground on Aikhal; see Miller 1995) were followed by the discovery of the famous Lomonosov deposit in the Arkhangel'sk region northeastern Russian platform. In the same region another very rich pipe was discovered, the Grib pipe.

Similar development may be taking place in Canada. Of interest is the development of the A-154 and Panda kimberlites in the Northwest Territories. Similar developments in the future of the United States will occur only if political agendas again favor mining.

One major way to reduce the cost of diamond production is to decrease the cost of mining. Locally, increased diamond production is related to open-pit operations, and decreased production is related to the high cost of underground mining. The discovery of new diamond deposits and subsequent mining of the deposits depends on the success of exploration programs.

DIAMOND MARKETS

The great value of diamonds created a tendency to establish strict control over diamond production and marketing. In ancient times, the simplest form of market control was beheading those who attempted to smuggle diamonds or infringe upon government regulations. Later efforts to control diamond markets became more sophisticated and took on the form of government regulations and the creation of a multinational organization designed to regulate the number of diamonds that enter the market by imposing quotas on production.

Almost no limitations can be seen in the methods used to maintain restrictions and impose rules. They vary from massive acquisitions of prospective diamond properties to the apparent destruction of mining facilities (see Chapter 1, History of Diamond Discoveries, Ouachita Mountains). In the extreme, wars were initiated or supported (such as in South Africa at the turn of the twentieth century and in Angola in the middle of the twentieth century) (Miller 1987, 1995).

In India, special decrees by sultans forbade, under the threat of execution, the selling of diamonds greater than 10 carats without special government permission. Nevertheless, smugglers flourished in India. Metgold, who described these restrictions, purchased several diamonds that weighed more than 20 carats (Milashev 1989).

The later discovery of Brazilian diamonds led to two events: the first attempt to regulate diamond market prices and mass production of industrial diamonds. By the time of the Brazilian diamond discoveries, most of the Indian diamond deposits were nearly exhausted. Thus the Brazilian discoveries resulted in a major rush. The rush and ensuing mining led to a fall in diamond prices. In order to stabilize prices, rumors were generated suggesting that diamonds had not been found in Brazil but were still being mined in India. To provide impetus for this rumor, several diamond merchants transported Brazilian diamonds to another one of Portugal's colonies, Goa, in India, and then re-exported them to Europe as Indian stones. However, following the discovery of the great diamond deposits in the state of Minas Gerais, it became impossible to hide the discovery.

To prevent a catastrophic fall of diamond prices, the Portuguese government imposed taxes for diamond exportation, imposed a significant tax burden for the leasing of diamond-prospecting grounds, and proclaimed mining to be a monopoly of the Portuguese crown. As a result of these actions, work in the diamond mines practically ceased. The situation changed only after Brazil proclaimed its independence from Portugal in 1822. The Brazilian government stimulated mining activities by allowing private enterprises to mine diamonds. In 1850 Brazil reached its highest level of production and, in the meantime, the Brazilian government attempted to impose strict regulations on diamond production to prevent smuggling. Despite the government's efforts, as much as 30% of total production may have been smuggled out of the country (Milashev 1989).

Following the discovery of diamonds in South Africa, the major diamond deposits in South Africa were initially subdivided into numerous small mining claims that made it nearly impossible and unsafe to work individual claims. Thus, in the 1880s, the miners consolidated their properties and created large companies. The DeBeers Company became especially active in consolidation and purchased most of the claims.

In 1888, DeBeers, headed by Cecil Rhodes, paid £5 million to acquire its major competitor, the Kimberly Central Mining Company, which was founded by his archrival, Barney Barnato. As a result, by 1892 DeBeers owned all of the diamond mines in the Kimberley area. In 1902, the year of Cecil Rhodes' death, Ernst Oppenheimer established another great diamond and gold mining company, the Anglo-American Corporation of South Africa.

In 1929, Oppenheimer was elected chairman of DeBeers and thus gained the control of most of the diamond and gold production in South Africa, which accounted for about 80% of world production of both commodities. This company became so influential that its decisions strongly affected the world diamond markets. To further control the diamond market, DeBeers took steps to control the marketing of diamonds and created the Central Selling Organization (CSO), also known as the Diamond Trading Company (DTC). This organization still controls practically all of the world diamond markets (Miller 1995).

It is still generally accepted that once a producer's annual gem diamond output exceeds 750,000 carats, the only satisfactory long-term sales mechanism is through the CSO. The CSO's control goes through a system of agreements with the major producers. Under such agreements, the CSO takes the burden of stockpiling, advertisement costs, and marketing. DeBeers acts as the equivalent of a world central bank in diamonds. In exchange for taking the greatest chunk of rough diamonds from its clients, the CSO provides the participants with market price stability and a safety net from market fluctuations. This monopoly is similar to some other commodity monopolies (for example, the Organization of Petroleum Exporting Countries in the oil industry) and the great stockpile of beef, wine, and vegetables in the European Community).

DeBeers works to protect its interest, and in the past it apparently attempted to slow exploration and development of some potentially new diamond-producing regions. One example was DeBeers' attempt to inhibit exploration activities in Tanganyika in the 1930s. It took great boldness for a Canadian geologist, John T. Williamson, to purchase a bicycle and undertake diamond exploration, despite DeBeers' resistance, in Tanganyika in 1938. After 2 years of exploration, he discovered the largest known kimberlite deposit, the Mwadui pipe, with reserves estimated at nearly 3.9 million carats (Keller 1992).

From this mine, an exceptional 54.5-carat diamond was recovered in 1947 and was presented by Williamson to Princess Elizabeth (now Queen of England) on the occasion of her marriage. In 1948, the rough diamond was cut by Briefel and Lemer of London into a fabulous 23.6-carat gem. This diamond, which has been referred to as the Queen Elizabeth Pink diamond, is also commonly referred to as the Williamson diamond (Copeland 1974).

By controlling such great wealth according to Miller (1995), DeBeers became powerful enough to play significant roles in fueling wars. Probably the first was the English-Boer war in 1899, in which Transvaal lost its independence. Another was a war in Angola (1950–1960). In a notorious involvement, DeBeers took sides in the civil war (1982–1993) in Angola, one of the more diamond-rich countries in the world. In the very beginning of the war, DeBeers supported the Unita party led by Jonas Savimbi, in an attempt to control developments of the diamond industry. Intervention in Angola coincided with major shakeups in the diamond markets and resulted in the CSO cutting back on contractual purchase obligations by 25%.

Miller (1995) described two events that coincided with complications in the international diamond markets and led to interference in Angola. First, in 1982 and 1983 a major new diamond mine was developed in Botswana—the Jwaneng mine, which was a fabulously rich source of gems (±35% of the general output). Later, the Argyle mine in Australia began producing 7.5 million carats annually (about 6% gems), which coincided with the soaring output of Russian mines that reached 22 to 25 million carats annually.

Second, during 1991 through 1993, the Venetia mine in northern Transvaal was placed in production, and there was a new political development in Angola. Months before the United Nations brokered peace and free elections for its central government, an apparent bribe was made to the public that lifted almost all restrictions on "entrepreneurial activity" and exportation of diamonds. Illegal smuggling of gems to major cutting centers in Europe soared. A cautious estimate of the value of Angolan diamond output was around $900 million (Miller 1995).

Several simultaneous events followed. As a result of the Angolan election, the MPLA (Movement Popular for Liberation of Angola—the local Communist-dominated movement) declared a victory, and Savimbi retreated to his southern strongholds and launched a series of devastating attacks on the diamond-producing areas in the north and east. Within days, the diamond mining areas were overrun, and both legal and illicit production came to a halt.

IMPEDIMENTS TO DIAMOND EXPLORATION

Funding Constraints

Most fundamental studies related to diamond exploration depend on government-funded programs and grants that are necessary to support and entice exploration and development. For example, under the socialistic system of the former Soviet Union, competition and availability of money were relatively unimportant. The major impetus for diamond exploration and mining was determined by the need of state economies. Major obstacles were created by the very nature of a significant bureaucratic machine that would not tolerate a nonstandard approach used by nonconformists.

Diamond research projects in the West historically have been greatly underfinanced, and government funding has been influenced by political considerations. For example, funding for diamond research for the U.S. Geological Survey has been a very low priority, even though significant diamond discoveries could lead to the development of whole new industries. In addition, major cutbacks and redirection of the survey's priorities owing to politics in the 1990s have made this once-preeminent government agency nearly ineffectual. The U.S. Bureau of Mines, considered to be one of the leading research agencies in the world along with the U.S. Geological Survey, had its funding terminated for political reasons in the 1990s. Other examples include state geological surveys in the United States, which are so underfunded that diamond projects are considered to have little or no priority.

One classic example occurred in the 1990s when dozens of kimberlites were discovered by a government agency in the North American craton. The initial research budget for the project, a mere $50,000 per year, was spread thin and used to pay for salaries, geochemical analyses, geophysical surveys, and geological mapping. Yet after only 2 years, politics terminated the project even though the results led to the discovery of several new kimberlites, many of which yielded diamond-stability-field minerals. Moreover, this decision occurred at a time when the state budget had a $700 million budget surplus! Such primitive political logic will ensure that the United States may never become a source of diamonds and many other significant strategic mineral resources.

The unique geology of diamond deposits has made scientific studies one of the major driving forces of exploration. For example, any increase in the understanding of the nature of diamond deposits commonly leads to new major discoveries. This very important concept reaped major rewards for Canada, where the Canadian government, through the Geological Survey of Canada, supported practical research on glacial transport of kimberlitic indicator minerals. Prior to 1998, Canada did not produce any diamonds. After a short period following considerable exploration and expenditure of funds, Canada will be considered a world power in the diamond industry for many decades. This status will result in a dramatic increase in tax revenues and jobs.

Other governments in regions of favorable geology have not been inclined to fund practical research. As such, they may be losing hundreds of millions of dollars in tax revenue. Such is the case for the governments in Alaska, Colorado, Michigan, Montana, Wyoming, and even Arkansas. These states are underlain by favorable terranes where many kimberlites, lamproites, lamprophyres, kimberlitic indicator mineral anomalies, and detrital diamonds have been found, but the state governments have been reluctant to fund projects. The Wyoming Geological Survey, with minimal funding, has mapped 20 diamondiferous kimberlites and has identified several diamond targets and more than 300 kimberlitic indicator-mineral anomalies. Research funding in Montana, Colorado, and Alaska has been minimal. A substantial increase in research funding in these areas could reap substantial monetary benefits for the states and for the entire country.

Other problems that may preclude the discovery of a diamond pipe include land access problems, budget considerations, and other political constraints. In many companies and research agencies, the individuals who make decisions concerning exploration are not exploration geologists or prospectors, but are instead individuals who may be adept at raising funds (such as lawyers, accountants, career politicians, bureaucrats, and political geologists). More often than not, these people make decisions on the basis of factors other than sound geologic facts.

In one example, a consultant for a mining company developed access to several diamondiferous kimberlites along with some untested lamproites and lamprophyres (some of which yielded diamond-indicator minerals). These projects were presented to the company management. The cost of obtaining several proven diamondiferous kimberlites was similar to that of acquiring a single phlogopite-leucite lamproite that was considered to be a poor diamond prospect. However, the management of the company opted to drill the lamproite.

It is unfortunate that management decisions like these are commonplace in industry and government. It is one of the principal reasons that major mining companies fail and that government bureaucracies continue to waste funds and eliminate efficient programs. More often than not, good exploration targets and research programs are eliminated by management and bureaucrats for political reasons, personality conflicts, and many times for no logical reason. Mining and research industries need a dramatic revolution in their upper management practices. Geologists, geophysicists, and mining engineers require 4 to 7 years of college education and many years of experience to do their jobs, yet the majority of managers have no education in management and often little management experience.

As we discussed in the preceding chapter, diamonds were mined in India and Brazil for centuries, but only following the recognition of kimberlite as a source rock of diamond did the modern diamond industry develop. That industry currently (2001) produces more than $25 billion in diamond sales annually. In the same way, research that led to the legitimization of lamproite as another source for diamond resulted in a chain of new discoveries. It is predictable that the legitimization of other potential host rocks and an understanding of their characteristics and geologic environments (such as metamorphic rocks, astrobleme-related deposits, and stratified ultramafic intrusions) will result in new discoveries around the world.

A combination of greed, thirst for knowledge, and the heroic persistence of geologists makes the finding of diamond deposits, fields, and provinces possible. The drama of exploration is especially intense in the business of diamonds, in comparison with any other commodity, because of the intense competition in the mining of diamonds and the struggle for control of the world diamond market.

The concept that the formation of diamond represents an event that originated within the deepest sampling levels (>150 km below the surface) of magmas within the earth led to persistent prejudices that need to be overcome in the course of future exploration. How such prejudices developed is left up to the psychoanalyst, as few of them are well grounded in scientific fact. Some are as follows.

- There is a single source rock for diamonds (i.e., kimberlite).
- The Argyle diamondiferous lamproite is unique.
- Rich diamond deposits are unique, and it is highly unlikely that within the region surrounding such a deposit, another rich diamond deposit will be found.
- Within a specific province, all kimberlites and other diamond-bearing rocks are formed during a single geologic event.
- Diamonds found in stream and glacier deposits are transported from a great distance and from unknown source regions.
- Diamond deposits are restricted to ancient stable cratonic environments.

Only by overcoming these and other prejudices will major new discoveries be made.

For several decades, the concept that kimberlites were a unique source rock governed the diamond industry's entire approach to exploration. This concept has been so ingrained in the industry that prospectors and geologists often have called any diamond-bearing rock “kimberlite.” Historically, this misuse of the term occurred at the Crater of Diamonds in Arkansas, which is now known to be olivine lamproite. Moreover, the name Kimberley has been applied to various regions and properties in the United States and Australia, possibly in hope that the name itself would affect the geologic situation and attract investors. The name “American Kimberlite” was even applied to a lamproite near the Crater of Diamonds, and “West Kimberley” was attached to the region in Western Australia where diamonds were later found.

The influence on orientation surveys concerning exploration programs of the fundamental concepts of diamond source rocks is well illustrated by the history of exploration in India. The oldest known source of diamonds in India were placers. The later discovery of the South African kimberlites reoriented exploration programs and led to many subsequent discoveries. In the third quarter of twentieth century, some kimberlites in India were redefined as olivine lamproites. In recent years, exploration again has turned to placers.

Only because of the discovery of nonkimberlitic, lamproite-hosted, commercial diamond deposits (in the Ellendale field and at Argyle in Western Australia) did geologists reconsider whether kimberlite was the only host rock for commercial diamond deposits. However, even following these discoveries, prejudices persisted. During a diamond-exploration short course taught by one of the authors more than 10 years after the Argyle discovery, a company chief executive officer sitting in the course told him to ignore any discussions about lamproite, as the Australian discovery was considered a fluke. The officer exclaimed that their company “only explores for diamondiferous kimberlite in Elephant Country.” Such statements are best reserved for “gregory” awards. (Refer to Chapter 1, South Africa, for an explanation of “gregory” awards.) This example is just one of many of prejudice that persists in the diamond industry. In spite of such prejudices, the review of lamproitic petrology resulted in a new approach to diamond exploration targets in the North American and Eastern European platforms by some companies.

In recent years, there has been some reference to the possibility of commercial diamond deposits in unusual or unconventional (nonkimberlitic, nonlamproitic) host rocks (e.g., Kaminsky 1984, Raeside and Helmstaedt 1982; Erlich et al. 1989; Nixon 1995; Hausel 1996a). For many years, any reference to studies of unusual source rocks for diamonds was made with contempt. Today we see that not just one but rather several diamond deposits have been found in metamorphic rocks, and we anticipate that the discovery of accessory diamonds in some ultramafic rocks will lead to some major discoveries in what are currently considered unusual host rocks.

Prejudice 1—One Diamondiferous Deposit per District

Many of the prejudices associated with diamond exploration have evolved around the concept that diamonds originate in deep-seated parts of the earth under extreme pressure and temperature. The common concept is that the event that brings the diamonds to the earth's surface is unique and cannot be repeated. This idea takes many forms, all of which have had disastrous implications in exploration.

One example of a near exploration failure, owing to exploration prejudices, occurred in Russia. Following the discovery of the Lomonosov kimberlite pipe in the Arkhangel'sk province of Russia, little effort was made to search for similar deposits in the district. The Lomonosov kimberlite was a world-class discovery and was estimated to contain about half of world's known diamond reserves in this single group of pipes.

Some very experienced Russian geologists, who knew that kimberlite pipes in this province had formed in several stages, were under the erroneous impression that another rich diamondiferous kimberlitic pipe in the immediate vicinity of the Lomonosov was improbable. It was thought that the Lomonosov group of pipes was unique and formed under unique circumstances. However, despite complete ignorance of the geological data by bureaucratic authorities, another very rich kimberlite pipe (the Grib, which had a diamond content even greater than that in the Lomonosov deposit) was discovered in the area during the course of 2 years of exploration. Careful interpretation of complex geophysical data that included aeromagnetic, gravity, and electric surveys was followed by drilling.

Prejudice 2—One Age of Diamond Host Rock per District

Another misconception in the geologic community is that diamond deposits within any single region are formed during a single event. Even though there are numerous examples of multiple episodes of diamondiferous host rock eruptions within cratonic environments, districts, and even within single pipes, this misconception still persists and often means that exploration programs are terminated before regions are thoroughly examined.

Often, following the discovery of diamonds within a specific conglomerate or within a host rock of a specific age, exploration programs are terminated when younger rock formations are intersected. The misconception is that diamonds can be found only in rocks of an age similar to that of the rocks of the initial discovery. Ultramafic-alkaline provinces with associated kimberlites formed throughout 400 to 500 million years, and the emplacement of kimberlites within each province is commonly repeated several times. Up until now, there has been no indication that diamonds from a specific kimberlitic group or body depends on its relative age within each province. Since the majority of diamonds are xenocrysts, they can be transported to the surface by any magma of any age originating from depths great enough to sample the diamond stability field.

The presence of several kimberlitic eruptions within a single region is well established in the Siberian platform (see Chapter 14, Temporal Distribution of Diamond Deposits). This fact was confirmed by the presence of fresh, coarse grains of kimberlitic indicator minerals in conglomerates of Permian, Lower Jurassic, Upper Jurassic-Valanzhinian, and even Upper Cretaceous age within the Permian-Mesozoic marginal depression north of the Anabar anteclise.

In recent years, evidence of several episodes of mantle-derived intrusives has emerged in the Wyoming craton. The initial group of kimberlites identified in the State Line district were identified as Early Devonian in age. Many exploration groups concentrated only on potential Devonian intrusives. No effort was made to find diamond host rocks of any other age until recently. Now, kimberlites, lamproites, and lamprophyres of Precambrian-Cambrian, Devonian, Tertiary, and Recent ages have been identified, and there is even one report of a diamond found in a Middle Proterozoic quartz-pebble conglomerate (Hausel 1996b). Exploration prejudice was even extended to the possible host country rock intruded by kimberlite when one major exploration group refused to explore any regions that did not contain the same dominant host rock (Sherman Granite) as the State Line district.

Prejudice 3—Only Ancient Platforms Host Diamond Deposits

Correspondence between the age of diamond mineralization and the age of cratonization of different stable blocks in Australia clearly shows that a main precondition for development of kimberlitic/lamproitic volcanism is cratonization. The influence of cratonization age is comparatively minor.

The type of tectonic regime, not the age of stabilization, predetermines localization of diamond deposits. Appropriate examples include stable massifs in Europe and Asia (Bohemian, in Central Asia, Russian Far East and other regions of the world). One might predict that in the future proper relations between diamond deposits in stable blocks and adjacent mobile belts will be established.

Prejudice 4—Transport Never Destroys Diamond

Another prejudice overemphasizes the ability of diamond to resist mechanical destruction during surficial transport (mainly in water). This prejudice commonly leads to the conclusion that the diamond source must have been at a great distance from the place where placer diamonds were discovered.

One company that initiated an exploration program in California in the 1980s was unable to locate the source of diamonds found in the region. Thus, it concluded that the numerous placer diamonds must have originated from the nearest diamondiferous kimberlites, which are located in the Colorado–Wyoming State Line district at the easternmost edge of the Rocky Mountain uplift. No consideration was given to the fact that many of the industrial diamonds found in California possess characteristics that were atypical of the Colorado–Wyoming diamonds and that many California diamonds possess characteristics suggesting a nearby source. But even more absurd was that the geography, past geologic events, and known river transportation routes made it impossible for the California diamonds to have originated in the Wyoming craton.

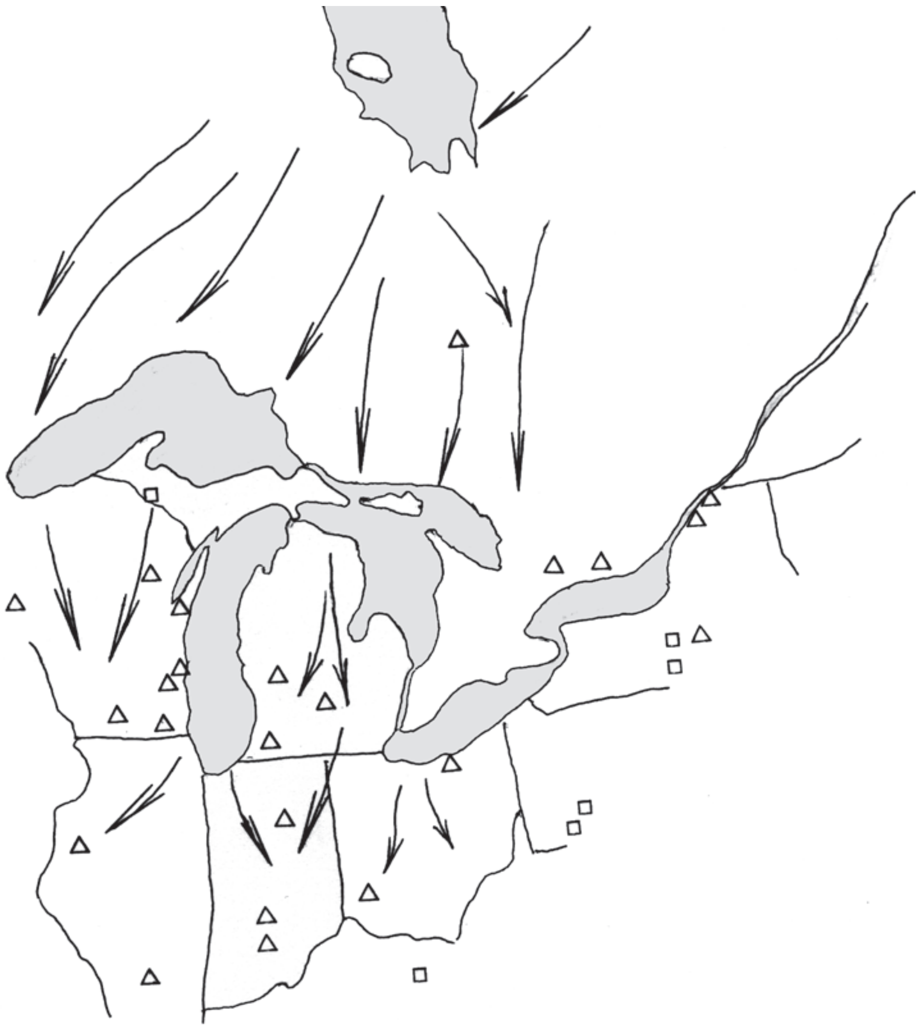

FIGURE 18.1 Distribution of diamonds found between James Bay and the central part of the North American craton (modified from Brummer 1978)

In the Great Lakes region, some diamonds have been found in glacial till and have been interpreted as having been transported by glaciers from great distances originating within the James Bay area in northern Canada (Figure 18.1). However, after long and laborious exploration, it has been shown that almost every diamond that was discovered corresponded to a nearby source rock.

During the initial phase of diamond exploration in Siberia, only placer deposits were found. It was thought that the original source of the diamonds was located somewhere within the mountains surrounding the Siberian platform (in particular within Baikalides of the Patomsky highland). Years later, Milashev (1989) stressed that, in spite of a lot of sophisticated reasons developed by its supporters, this theory could potentially have grave consequences in the search for diamonds within the central and northern part of the Siberian platform, where several great diamond deposits were finally found.

On the other hand, even when evidence suggests transportation from a distant source, some researchers have suggested a proximal source for some diamonds. Between 1982 and 1986, three small diamonds were found in alluvial placers in Crooked Creek within the Circle district of Alaska (Forbes, Kilne, and Clough 1987). The diamonds weighed 0.3, 0.83, and 1.4 carats and were found within a Tertiary basin close to the Denali fault zone. The stones showed percussion marks and fractures. These features and the absence of any kimberlitic indicator minerals led the researchers to conclude that the diamonds probably originated from a distal source. Nevertheless, a major diamond exploration company requested an exclusive right to prospect for diamonds within the state (R. Forbes, oral communication, 1987). (No serious exploration had occurred as of 2001.)

THE IMPORTANCE OF FUNDAMENTAL RESEARCH

Fundamental geological problems related to diamonds embrace practically all aspects of geology. At first glance, these problems can appear only to increase the cost of exploration. However, nothing is further from the truth. Good theoretical studies are practical. One fundamental problem deals with the boundary between conventional and unconventional diamond sources. Currently, minettes are considered to be one of a number of conventional sources. With time, the number of conventional source rocks recognized for diamond will be extended. Recent discoveries of diamond deposits related to specific types of ring structures and to high-pressure metamorphic rocks (such as the Beni Bousera massif, which is incredibly rich in graphite pseudomorphs after diamond) support this assertion, as do recent discoveries of diamond in ultramafic plutons.

Fundamental problems usually start with the most basic aspects of the geology of diamond deposits—the mineral association and diamond morphology and structure. Creating and understanding new data on diamond deposits may provide priceless clues for understanding diamond formation and other processes deep within our planet.

It is no wonder that any progress in diamond exploration gave impetus to many new academic studies. The discovery of diamond-bearing lamproites resulted not only in new exploration programs leading to the discovery of new deposits in the Urals, Fennoscandia, and other places, but also to a flurry of studies such as detailed petrographic and mineralogic descriptions, classification, and geochemistry. The discovery of lonsdaleite within the Popigay ring structure of Siberia generated a successful search for similar structures worldwide that was accompanied by petrologic, mineralogic, and geochemical studies and by many discussions concerning the origin of this structure. There is little doubt that future works on unconventional source rocks for diamond will bring new turns in fundamental studies.

From the dawn of diamond mining there has been the misconception that the diamond market is in danger of imminent collapse owing to oversupply (Miller 1995). However, the whole history of the diamond industry has shown that demand for both gem and industrial diamonds is always ahead of supply. Again and again the real driving force of the industry has been new discoveries—especially discoveries of deposits rich in gem diamonds. Even if strain overcame the exploration geologist, his or her persistence and personal bravery in the process of each new discovery resulted in real progress for the diamond industry.

It is worthwhile to close this book on optimistic note: markets have always had highs and lows, political situations will always affect mining development, but in the diamond industry, all the discoveries are the result of the work of the EXPLORATION GEOLOGIST, to whom this book is dedicated.

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Note: *f* indicates figure; *t* indicates table.

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About the Authors

Edward I. Erlich

A native of Leningrad, USSR, Erlich began taking part in geological explorations in 1951. In 1957, he graduated from Leningrad Mining Institute as an exploration geologist. From 1958 to 1962, he participated in geological mapping and exploration excursions in diamondiferous regions of the Siberian Platform for the Scientific Research Institute of Arctic geology. Erlich's work helped to uncover a new province of ultramafic-alkaline rocks (the Udja Province), along with the greatest ring ultramafic-alkaline complex (with carbonatitic core [Tomtor massif] and associated kimberlites) in the world. In 1962, he successfully defended his Ph.D. thesis on the structural control of kimberlites found in the northeastern part of Siberian Platform. He then went to work for the Institute of Volcanology Academy of Science, USSR, in Petropavlovsk in Kamchatka, where he spent 10 years. For the following 10 years, he continued exploring the Tomtor Massif, which resulted in the discovery of a world-class rare metal deposit.

In 1983, Erlich emigrated to the United States, where he has since worked on National Science Foundation grants, under contract to the U.S. Geological Survey and the Smithsonian Institution, and as a consultant geologist. He has also assessed lamproitic fields in Arkansas and kimberlite-prospective sites in Alaska and California, and has explored for gold in Sonora, Mexico, and in Ecuador. Erlich's main field scientific interest is the relationship between tectonics and magmatic activity.

Erlich has authored more than 100 papers and has had books published in Russia, the United States, Japan, the United Kingdom, France, and South Africa. Throughout his distinguished career, he's participated in numerous scientific congresses and conferences.

W. Dan Hausel

Dan Hausel received BSc and MSc degrees in geology from the University of Utah in 1972 and 1974. Since 1975, he has worked as a consultant for various companies and as a research geologist for the Wyoming State Geological Survey (WSGS). Currently, he is the senior economic geologist for the WSGS. In that capacity, he investigates precious and base metals, gemstones, mineralogy, mining districts, Archean greenstone belts, and diamondiferous host rocks.

In his 25-year career, Hausel has mapped more than 1500 km^2 of complex metamorphic terrain, including the United States' two largest kimberlite districts and its largest lamproite field. His work has led to the discovery of diamond and other gemstone occurrences including several base and precious metal deposits. While on leave from the

WSGS, he launched diamond exploration investigations and consulted on precious metal deposits for various mining companies around North America. In addition, he has taught several diamond exploration and prospecting workshops and short courses for various companies and professional groups.

A prolific writer and public speaker, Hausel has authored or coauthored more than 400 books, professional papers, general interest articles, and geological maps, and lectured to hundreds of groups around North America. In 1992, he was awarded the American Association of Petroleum Geologist Energy Mineral Division's President's Award and the Wyoming Geological Association's Certificate of Appreciation for Outstanding Endeavors and Contributions. In 1994, the Laramie Lyceum recognized Hausel as a distinguished lecturer. In 1998, he was recognized as a distinguished lecturer by the University of Wyoming Department of Geology and Geophysics, presented with the Prospector's Best Friend Award by the Rocky Mountain Prospectors and Treasure Hunters, awarded a Medal of Achievement in Science by the International Biographical Center, and elected to the American Biographical Institute's Millennium Hall of Fame. In 2001, he earned the Education Award and was elected to the National Rockhound and Lapidary Hall of Fame. His achievements have been highlighted in several Who's Who publications, including *Who's Who in Science & Engineering*; *Who's Who in the West*; *Who's Who in America*; *Who's Who in the World*; *2,000 Notable American Men*; and *5,000 Personalities of the World*.