Mantle indicator minerals in ant mounds and conglomerates of the southern Green River Basin, Wyoming

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Abstract

Minerals of upper mantle origin occur in conglomerates, pediments and ant mounds in the southern Green River Basin. Chrome-bearing pyrope, pyrope-almandine, diopside, spinel and picroilmenite 6 mm in size are found on ant mounds with similar minerals as large as 12 mm in diameter in the Bishop Conglomerate. The minerals are from disaggregated eclogite and peridotite, similar to minerals found in mantle-derived igneous rocks such as kimberlite and lamproite, the primary host rocks for diamond. No kimberlites or lamproites are known in the immediate area, although the region is above the Archean Wyoming craton which is a conducive setting for diamond occurrence. Current assessment of the minerals using established geochemical criteria suggests that prior coexistence with diamond is unlikely. The ant mound and pediment materials are derived from erosion of the Bishop Conglomerate, whose source area is the Uinta Mountains. Extensive sampling of present-day streams in the central Uinta Mountains produced a few anomalies of only 1-2 grains 0.25 mm or less in size; no continuous mineral train exists between the occurrences in the Green River Basin and those in the Uintas. Minerals of similar chemistry do occur in the Leucite Hills lamproites in the Rock Springs uplift, and in the Kamas lamproites in the western Uinta Mountains. The erosional history of the basin prohibits either area to be the source of the Green River Basin occurrences. Dramatic drainage pattern shifts and extensive glaciation since the Oligocene have hindered efforts to locate a bona fide *igneous host for the minerals.*

Introduction

The heavy minerals (S.G.>2.9) pyrope, pyrope-almandine, chrome diopside, omphacitic diopside, chrome enstatite, chrome spinel, and picroilmenite are referred to as kimberlite indicator minerals, because their presence in secondary environments commonly indicates a nearby kimberlitic source (Gurney and others, 1993). These minerals resist wear during transport and have distinctive colors or surface textures that make visual recognition and evaluation

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possible (Dummett and others, 1987). Some of the minerals share paragenesis with diamond and have unique compositions with respect to certain major and minor elements (Gurney, 1984; McCandless and Gurney, 1989; Dummett and others, 1987; Gurney and others, 1993). The unique compositions make it possible to evaluate the potential diamond association of indicator minerals in secondary environments where the igneous host is undiscovered. In the present study, indicator minerals discovered in the southern Green River Basin and northern Uinta Mountains (**Figure 1**) are described with respect to conditions of formation, diamond potential, and possible source rocks. A detailed discussion of the geochemistry of the minerals can be found in McCandless and Nash (1995).

Geologic setting

The Green River Basin is a broad structural depression located in southwestern Wyoming, with the Uinta Mountains in northeastern Utah forming the southern boundary. The Uintas are an east-west anticline with tilted Phanerozoic rocks flanking a core of flat-lying Proterozoic (1.6-0.9 billion years before present – Ga) metasediments and Archean (2.7 Ga) gneisses (Sears and others, 1982). The Uintas represent an aulacogen, with an original stratigraphic thickness exceeding 15,000 meters (Bradley, 1988).

History of the discovery

In the late 1960s, Heber Campbell, a retired rancher living in Manila, Utah, related the story of finding 'rubies' on anthills in the 1950s in the southern Green River Basin to the local high school teacher, Derell Johnson. Campbell showed Johnson the localities and several years later John-

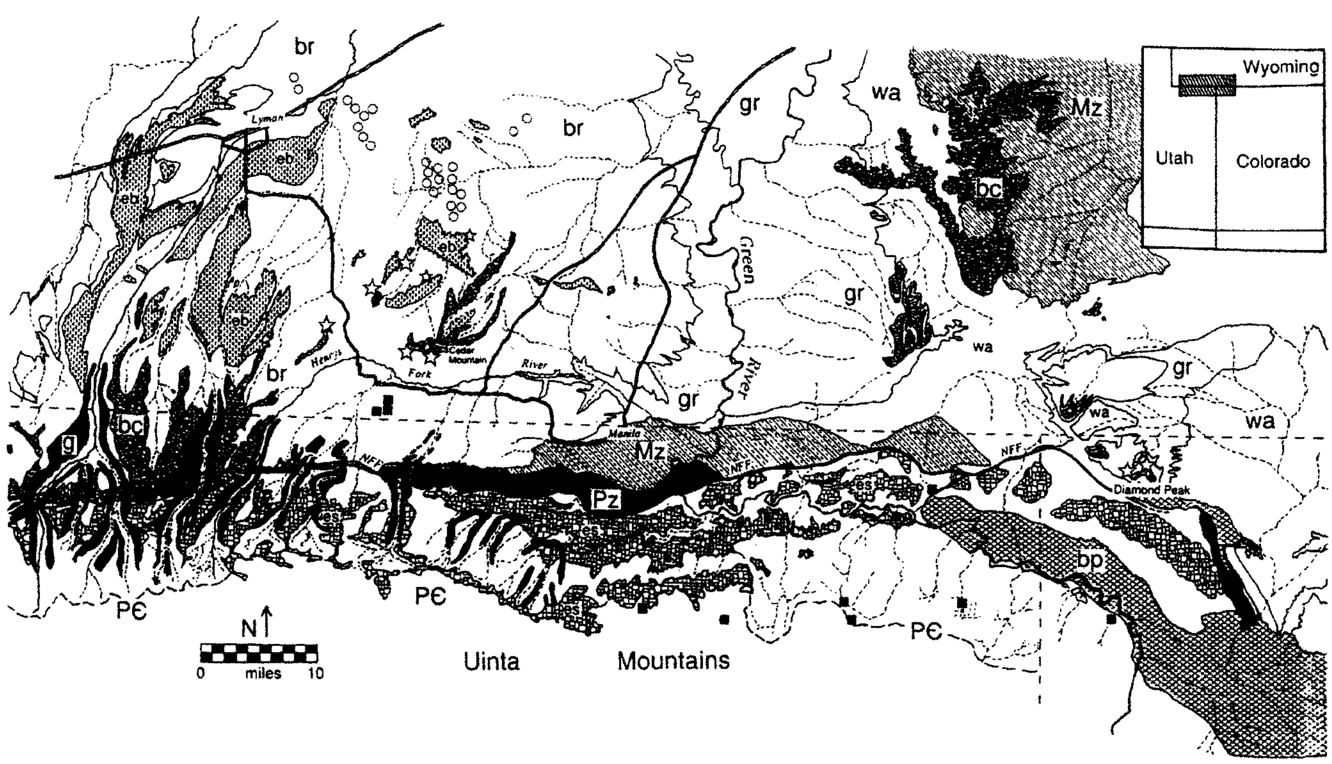

Figure 1. Simplified geologic map of the southern Green River Basin and north slope of the Uinta Mountains. Pre-Tertiary units are Uinta Mountain Group (PЄ), Paleozoic (Pz) and Mesozoic (Ma) sedimentary rocks. Tertiary lacustrine-fluvial units are Green River Formation (gr), Bridger Formation (br), and Wasatch Formation (wa). Erosional surfaces are (es) earliest unit, capping the Uinta Mountains, followed by the Bishop Conglomerate (bc), erosional surfaces later than the Bishop Conglomerate (eb), and the Browns Park Formation (bp). Pleistocene glacial moraines are (g). The projected trace of the North Flank Fault (NFF) is shown as a solid black line. Flaming Gorge Reservoir is omitted for clarity. Long dashed line indicates the present crest line in the Uinta Mountains, short dashed lines are drainages in the Uinta Mountains and the Green River Basin. The towns of Lyman and Manila, and local roads (bordered dashed lines) are included for reference. Indicator mineral anomalies are ant mounds (circles), pediments and conglomerates (stars) and streams (squares). Map modified after Bradley (1936, 1964), Atwood (1909), and Winkler (1970).

son took some of the 'rubies' and associated minerals with him when he moved to Salt Lake City. The senior author met Mr. Johnson at the University of Utah and in 1978 correctly identified some of the minerals as pyrope garnet and chrome diopside after reading of similar occurrences in eastern Colorado and Wyoming (McCallum and Mabarak, 1976). Mr. Johnson made several attempts to relocate the occurrences in 1978, doing so when a puncture on his field vehicle forced him to stop near an anomalous ant mound. The locations were revealed to Superior Minerals in 1980 and an exploration program began shortly thereafter. Extensive sampling established that the minerals are eroded from the Bishop Conglomerate, which is in turn derived from Uinta Mountains, and indicator-bearing streams were subsequently located in the Uinta Mountains (**Figure 1**). Exploration ended in 1985 with the sale of Superior Minerals to Mobil Oil. The earliest scientific description of the minerals is found in McCandless (1982).

Ant mounds and ant behavior

Detrital indicator minerals were first discovered in the Green River Basin on the mounds of *Pogonomyrmex occidentalis*, the western harvester ant (McCandless, 1982). Burrowing ants in southern Africa are believed to carry pyrope and ilmenite up from their kimberlite source through 50 meters of Kalahari sands (Wilson, 1982), and a similar situation was initially considered for *Pogonomyrmex occidentalis*. However, soil auger samples 2-6 m in depth taken near the ant mounds revealed no indicator minerals. Ant mounds of *Pogonomyrmex occidentalis* consist of soil particles (<1 mm diameter) excavated during the digging of the nest below the surface (maximum depth 3 m), and gravel (2-6 mm diameter) collected from the surface and placed on the mound, possibly to prevent erosion (**Figure 2**; Wheeler and Wheeler, 1963). The surface concentrating behavior was confirmed when all the 2-6 mm fragments of a broken glass bottle placed 10 m from an ant mound appeared on the ant mound after only a few weeks. The pyrope, pyrope-almandine, chrome diopside, omphacitic diopside, and chrome enstatite that occur on the ant mounds are also in this size range, and evidently were collected from the surface rather than brought up from depth.

The ant mounds are commonly situated near thin gravel pediments shed from mesas capped by the Bishop Conglomerate (**Figure 2**). Pyrope, pyrope-almandine and chrome diopside were located in paleopediments topographically above the ant mounds, and in the Bishop Conglomerate itself. The Bishop Conglomerate consists of poorly sorted gravels with boulders several meters in diameter. Indicator minerals are in both the coarse and fine layers of the conglomerate, and diopsides up to 12 mm in diameter are in the coarser layers. Cobbles of the Proterozoic Uinta Mountain Quartzite that core the Uinta Mountains to the south confirm that the Bishop Conglomerate is derived from the Uinta Mountains (Bradley, 1936). It is possible that the indicator minerals have experienced at least three cycles of transport before being concentrated onto the ant mounds.

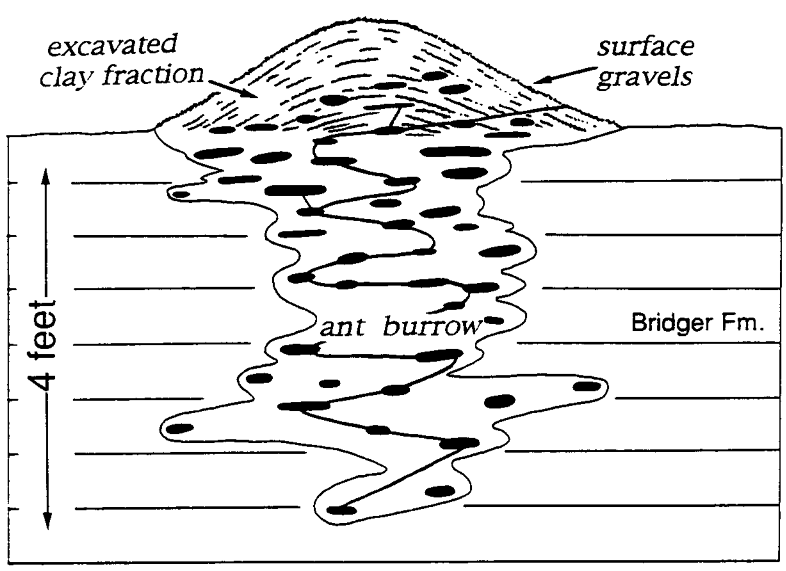

Figure 2. Simplified cross-section of an ant mound, showing the extent of disturbance beneath the surface, geometry of the burrows, and the stratigraphy of the mound, comprising a clay fraction excavated from the tuffaceous clays of the Bridger Formation to a depth of 4 ft, and surface gravels collected from the surface. Modified from Wheeler and Wheeler (1963).

Mineral chemistry

The microprobe is the industry standard for geochemical evaluation of indicator minerals, and most schemes utilize major and minor element abundances obtained by this method. Analyses were obtained on electron microprobes located at the University of Utah, University of Arizona, University of Cape Town, and the Superior Oil Geophysical Lab (Houston, Texas). Natural and synthetic standards were used and matrix corrections made according to the methods of Bence and Albee (1968) and Albee and Ray (1970). All oxide percentages are reported in weight percent.

In a situation where detrital minerals are evaluated, it is important to first establish a mantle derivation. The classification schemes of Dawson and Stephens (1975) and Stephens and Dawson (1976) are useful schemes for establishing mantle derivation. The garnets could be divided into two groups on the basis of magnesium number (Mg# = atom Mg/Mg+Fe) and Cr_2O_3 content (**Figure 3**). The chrome pyropes with 1.09-7.78% Cr_2O_3, Mg# = 0.700-0.874, fit into group 9 of Dawson and Stephens (1975). This group was found by Dawson and Stephens (1975) to consist of garnets from kimberlite and from garnet-bearing lherzolite, websterite, and harzburgite. The pyrope-almandine garnets are lower in Cr_2O_3 (0.09-0.86%), with Mg# from 0.534-0.838. A few unusual pyrope-almandines have low Mg# (0.556-0.652) and high Cr_2O_3 content (1.28-3.08%). The pyrope-almandine garnets classify as group 3 or 6, derived from eclogites (Dawson and Stephens, 1975). These and the high Cr_2O_3 pyrope-almandines are most similar to garnets in chrome-rich eclogites from the Roberts Victor kimberlite (Hatton, 1978; McCandless and Gurney, 1989), and to chrome-rich pyrope-almandine megacrysts from the Dullstroom kimberlite (L.R. Daniels, unpublished analyses). Color intensity varies with grain size, but three color shades can be recognized; chrome pyropes are purple, pyrope-almandines are orange to pink, and the high chrome pyrope-almandines are dark red.

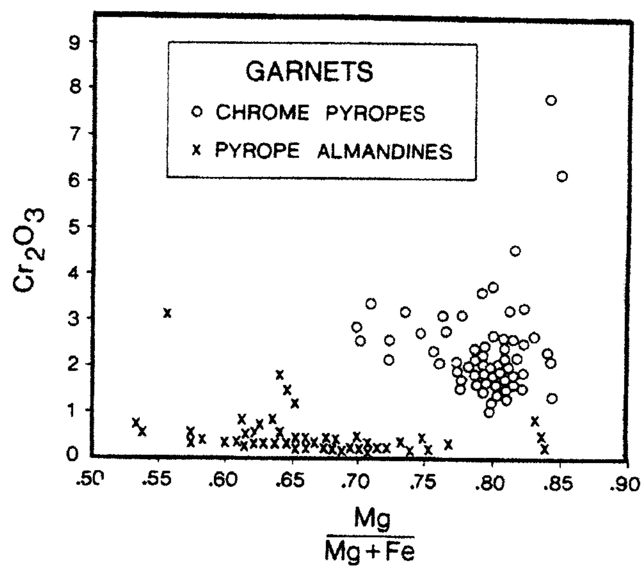

Figure 3. Magnesium number plotted against Cr_2O_3 content for garnets. Crosses are eclogitic garnets and circles are peridotitic garnets (after Dawson and Stephens, 1975). Solid circles are garnets from the Uinta Mountains.

Clinopyroxenes

The clinopyroxenes form a subcalcic group (Ca# = Ca/Ca+Mg = 0.431-0.475) with low FeO (2.23-4.64%), a calcic group (Ca# = 0.480-0.535) with 0.04-3.31% FeO, and a subcalcic to calcic group (Ca# = 0.441-0.521) with 4.07-9.09% FeO (McCandless and Nash, 1995; **Figure 4a**). The last group is termed salitic and requires as much as 3.21% Fe_2O_3 to attain charge balance. The subcalcic group is chrome diopside with 0.65-1.85% Cr_2O_3. The calcic group is omphacitic diopside with high Na_2O (0.16-3.33%). In the pyroxene quadrilateral the chrome diopsides and omphacitic diopsides lie in region C, corresponding to lherzolite and websterite (Dawson and Smith, 1977; Emeleus and Andrews, 1975; Nixon and Boyd, 1973; **Figure 5**). Some salitic diopsides lie in region A, defined for pyroxenes from mica-amphibole-rutile-ilmenite-diopside (MARID) suite xenoliths (Dawson and Smith, 1977; Waters, 1987). Most of the salitic diopsides lie in region D, defined in this study using clinopyroxene megacrysts from the Hatcher Mesa lamproite in the Leucite Hills, Wyoming (Barton and van Bergren, 1981; McCandless, unpublished data). The megacrysts are interpreted to represent metasomatized portions of the mantle which were sampled by the lamproite during ascent (Barton and van Bergren, 1981). They are not reported from kimberlite, and may represent a new indicator mineral in the exploration for lamproites (McCandless and Nash, 1995). Salitic diopsides were not recovered in the Uinta Mountains, but are present in the Bishop Conglomerate (**Figure 4b**).

Orthopyroxenes

Eleven orthopyroxenes have been analyzed and are classified as Cr-Al enstatites after Stephens and Dawson (1977), with 0.32-1.36% Cr_2O_3, 1.42-3.50% Al_2O_3 and 0.40-2.28% CaO. Cr-Al enstatites are dominantly from lherzolites and harzburgites lacking an aluminous phase (garnet or aluminous spinel: Stephens and Dawson, 1977). The Cr-Al enstatites therefore probably did not coexist with the spinels or garnets, but may have coexisted with one of the diopside groups.

Oxides

Hundreds of oxide minerals from numerous sample localities were analyzed; only 203 ilmenites had >0.10% Cr_2O_3 or >2.00% MgO. Less than 10% have Cr_2O_3 over 0.30%. Thirty spinels were analyzed: MgO ranges from 5.26-15.3% and Cr_2O_3 ranges from 34.0-66.3%, only four are considered to be potentially from an ultramafic source, with Cr_2O_3 >55.0%.

Mineral inclusions

Numerous inclusions in the garnets and diopsides establish the coexistence of many mineral phases prior to disaggregation in the secondary environment. The inclusions are euhedral to subhedral, average 0.05 mm in diameter, or in the case of ilmenite and rutile form needles as much as 4.0

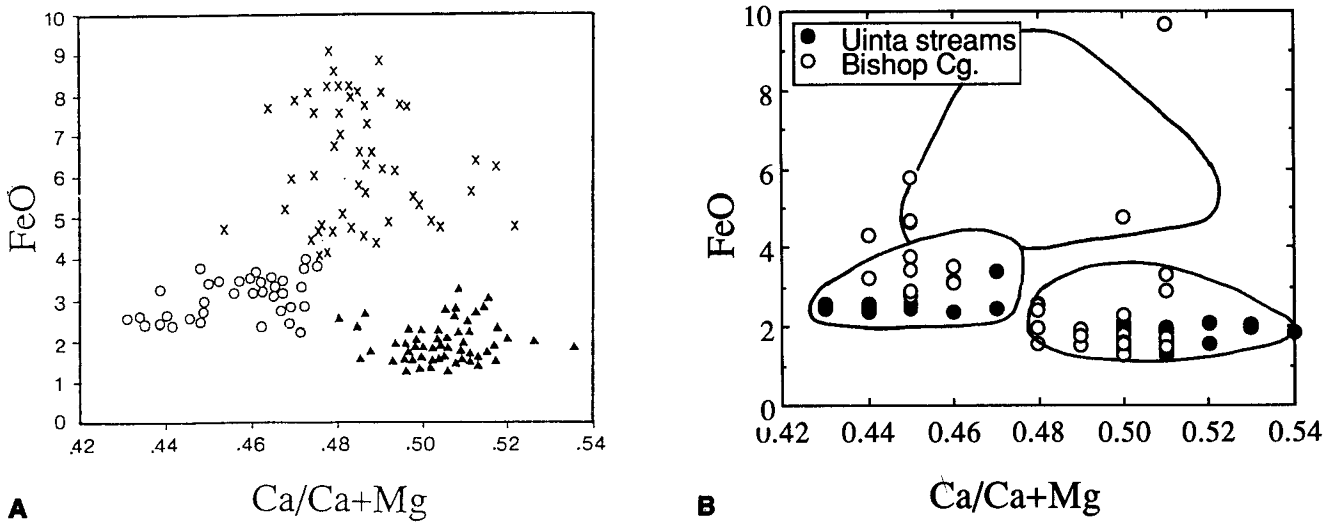

Figure 4. Calcium number plotted againt FeO for clinopyroxenes. Circles are chorme diopsides, crosses are salitic diopsides, triangles are omphacitic diopsides. A. Clinopyroxenes from ant mounds. B. Clinopyroxenes from the Bishop Conglomerate and from Uinta Mountain streams. Note that omphacitic diopsides are absent in the Uinta Mountains.

mm in length. A silica phase occurs in several pyrope-almandine garnets and in one chrome pyrope. Rutile is present in several omphacitic diopsides, and ilmenite in salitic diopside but is low in MgO and Cr_2O_3, unlike kimberlitic ilmenite grains. Carbonate occurs with the ilmenite and is nearly pure calcite (McCandless and Nash, 1995). Chrome spinel in salitic diopside is chemically similar to the single chromite grains and suggests coexistence.

Pyrope-almandine garnets were found in four omphacitic diopsides, and one chrome pyrope had a two-phase inclusion of omphacitic diopside with K-feldspar (Or_{67} $Ab_{24}An_9$ mole%). The omphacitic diopside inclusions are similar in composition to single omphacitic diopsides, and the garnet inclusions span the range of Ca-Mg-Fe values for the single garnet grains. This demonstrates a prior coexistence of omphacitic diopside with pyrope and pyrope-almandine garnet in rocks of eclogitic composition, similar to Cr-rich eclogites found at the Roberts Victor kimberlite, South Africa (Hatton, 1978). One diopside was a miniature xenolith, containing 26 subhedral garnets 0.1-0.2 mm in diameter.

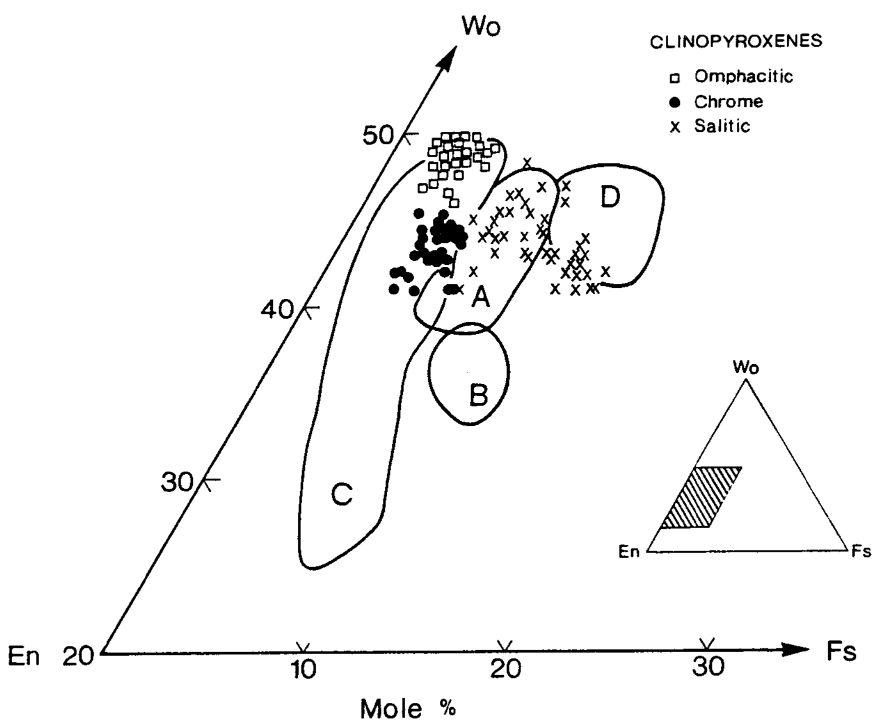

Figure 5. Mole percent Wollastonite-Enstatite-Ferrosilite (Wo-En-Fs) for clinopyroxenes. Regional clinpyroxenes from MARID xenoliths (A), from clinopyroxene-ilmenite intergrowths (B), from garnet peridotite and websterite (C), and from the Hatcher Mesa lamproite (D). Data from Dawson and Smith (1977), Waters (1987), Barton and van Bergren (1981), and this study.

Conditions of formation

The coexistence of garnet and clinopyroxene can provide an estimation of the equilibration temperature at a given pressure, using the distribution coefficient of Fe and Mg in garnet and diopside (Ellis and Green, 1979). A temperature range of 620-730 °C is obtained over an assumed pressure range of 20-30 kilobars (kb) for the pyrope-almandine garnets. In comparison, temperatures calculated for the chrome pyrope-diopside intergrowth C113 are 790-840 °C over an assumed pressure range of 20-40 kb, suggesting different conditions of formation for the two intergrowth types. Coupling these estimates with heat flow data can constrain the pressure-temperature (P-T) regime of the source region. Current heat flow in southwestern Wyoming and northeastern Utah is about 60 mWm^{-2} (Bodell and Chapman, 1982). Assuming a heat flow between 40 and 60 mWm^{-2}, the omphacitic diopside/pyrope-almandine intergrowths would be derived from a region at or just below the crust-mantle boundary (40-65 km), with the chrome pyrope/chrome diopside pair derived from between 55 and 80 km (McCandless and Nash, 1995).

The inclusion assemblage and P-T estimates provide a model of the mantle region from which the minerals are derived. The lower crust/upper mantle boundary region is represented by omphacitic diopside and pyrope-almandine in eclogite and/or websterite with accessory SiO_2 (coesite?), rutile, and ilmenite. Volatiles are indicated by the presence of carbonate (CO_2). Salitic diopside megacrysts indicate metasomatism in the higher part of the upper mantle by analogy to MARID xenoliths; Fe_2O_3 required for charge balance and inclusions of Fe-ilmenite suggest relatively oxidizing conditions. MARID xenoliths, with Fe-ilmenite and salitic diopside, form under oxidizing conditions within the higher part of the upper mantle (Dawson and Smith, 1977), and have affinities suggesting high-pressure crystallization from lamproitic magma (Waters, 1987). The upper mantle region contains lherzolite and/or chrome-rich eclogite of pyrope garnet and omphacitic diopside with accessory SiO_2 and feldspar. Because phlogopite is the common K_2O-bearing mineral in the mantle, the feldspar may have formed owing to an absence of H_2O, or under conditions where phlogopite is unstable.

Economic potential

In addition to serving as physical tracers for a mantle-derived igneous host, some indicator minerals form at pressures and temperatures where diamond is stable and have unique chemistries (Gurney, 1984; McCandless and Gurney, 1989; Gurney and others, 1993). When found in secondary environments, locating their host becomes a priority. The indicator minerals of this study are therefore evaluated with respect to diamond potential in the following discussion.

Gurney (1984) and Gurney and others, (1993) have shown a correlation between pyrope garnet chemistry and the presence of diamond in kimberlites. It is known that G10 pyropes (not equivalent to group 10 of Dawson and Stephens, 1975) originated within the stability field for diamond (Gurney, 1984). None of the pyropes analyzed from the Green River Basin or Uinta Mountains have G10 chemistry, with the exception of two garnets on the boundary (**Figure 6a**). Elevated levels of Na_2O in garnet (>0.07%) are present in diamond-bearing eclogites (Sobolev, 1974; McCandless and Gurney, 1989), and similar enrichment in eclogitic garnets from detrital sources concentrates is believed to indicate the presence of eclogitic diamond in the igneous host (Gurney and others, 1993). In the eclogitic garnets of this study, Na_2O is below the limit of detection (<0.01%), precluding eclogitic diamond potential. Chromite (i.e. chrome spinel) is widely accepted as a useful diamond indicator mineral because it is a common inclusion in diamond (Gurney and others, 1993; Waldman and others, 1987; Dummett and others, 1987). Chromite inclusions in diamond are relatively restricted in Cr_2O_3-MgO space (**Figure 6b**). None of the chromites from this study have MgO and Cr_2O_3 similar to those from diamond inclusions. In ilmenites, the presence of elevated MgO and Cr_2O_3 com-

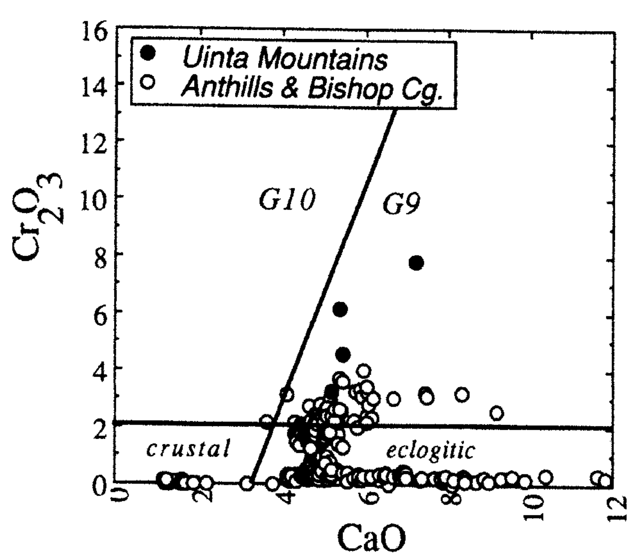

Figure 6A. Weight percent CaO and Cr_2O_3 for garnets. Garnets from the Uinta Mountains are solid circles. A lack of garnets in the diamond favorable G10 region of Gurney and others, (1993) indicates a poor diamond potential.

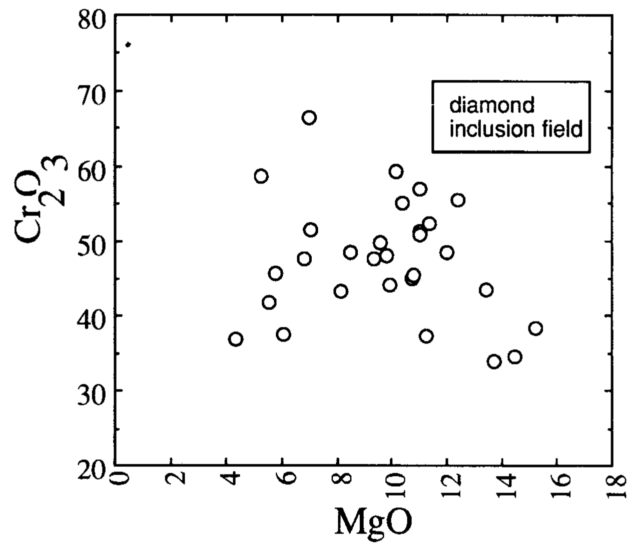

Figure 6B. Weight percent MgO and Cr_2O_3 for chromites. No chromites are in the diamond inclusion field, indicating a poor probability for diamond cogenesis.

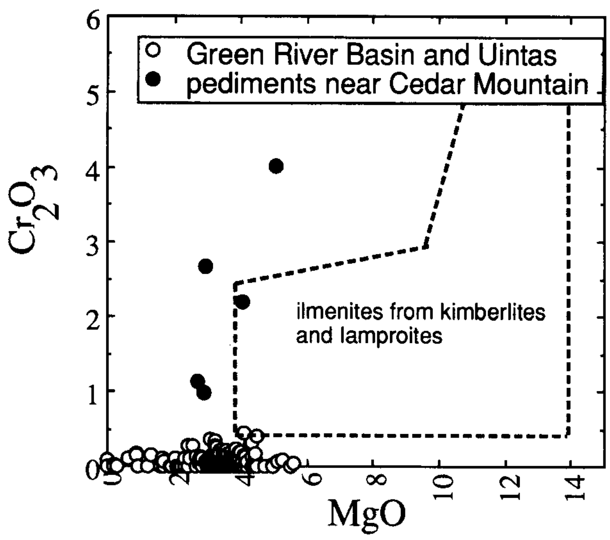

Figure 6C. MgO and Cr_2O_3 for ilmenites, with ilmenites eroded from Bishop Conglomerate as solid circles. None are similar to ilmenites commonly found in diamondiferous kimberlites and lamproites.

bined with low FeO (as total Fe) is usually considered a positive indication of diamond potential (Gurney and others, 1993). A few ilmenites with moderate Cr_2O_3 and MgO were recovered in the vicinity of Cedar Mountain (**Figure 1**), but represent less than 10% of all ilmenites analyzed and lie outside the area for ilmenites associated with diamondiferous diatremes (**Figure 6c**). Finally, estimated temperatures for garnet/clinopyroxene intergrowths that are well outside the diamond stability field indicate that the potential for diamonds in the host rocks of these minerals is very low.

Potential source locations

Evaluating the diamond potential of detrital minerals has built into it the assumption that the detrital suite is representative of the primary host rock from which it has been derived. This assumption has been confirmed with some certainty in areas of continental glaciation, and was the driving force behind the discovery of diamondiferous kimberlite in the North West Territories of Canada (Krajick, 1994; Gurney, 1995). It has also been successful in arid areas where movement of indicators has not been far from the igneous host (Gurney and others, 1993) The assumption remains to be tested in areas where temperate climate and multiple cycles of fluvial transport dominate, such as in mid-continent Canada (Swanson and Gent, 1993). It is known that silicate indicator minerals do not survive lateritization (Garvie, 1981), which complicates application of the complete diamond indicator suite in tropical regions (Gurney and others, 1993). Therefore, an igneous host must be located and tested for diamond cogenetic minerals before it can be established with certainty that diamonds are not present.

Green clinopyroxene megacrysts similar to the high-Fe salitic diopsides of this study are present at the Hatcher Mesa lamproite in the Leucite Hills (Barton and van Bergren, 1981). The restriction of these unusual diopsides to the Leucite Hills and to the Green River Basin occurrences tempts the conclusion that the detrital grains must be derived from the Leucite Hills. This is geologically impossible as the Leucite Hills erupted ~ 1 million years ago (Ma) (Bradley, 1964), whereas indicator minerals are in the 29 Ma Bishop Conglomerate (Hansen, 1986).

Deposition of the Bishop Conglomerate occurred in the Oligocene as a result of renewed uplift of the Uinta Mountains, and detritus reached as far north as the southern edge of the Rock Springs uplift (Hansen, 1986; **Figure 1**). The Bishop Conglomerate is poorly sorted, with boulders several meters in diameter in the coarser layers. Indicator minerals are in both the coarse and fine layers of the conglomerate, with diopside up to 12 mm in diameter in the coarser layers in the vicinity of Cedar Mountain (**Figure 1**). The Bishop Conglomerate can contain over 50% clay-sized particles, which has been shown experimentally to inhibit mineral wear in high energy fluvial systems (McCandless, 1990). Chrome diopside is also considered a poorly traveled indicator mineral (Mosig, 1980), but its survival during transport is strongly dependent on its primary morphology. For example, chrome diopside megacrysts kimberlites of eastern Colorado and southwestern Wyoming have a high cleavage density that allows for rapid disaggregation to produce anomalies consisting of hundreds of very angular grains (Dummett and others, 1987). With the exception of the orthopyroxenes, none of the pyroxenes of this study have well-developed cleavage. Thus, the occurrence of 12-mm-diameter omphacitic diopside in the Bishop Conglomerate does not necessarily infer a short transport distance.

Although the Uinta Mountains south of the Green River Basin anomalies are devoid of igneous rocks excepting a few diorite dikes (Ritzma, 1974, 1980), peridotite and lamproite occur in the western Uintas, with ages from 11.7 to 40.4 Ma (Best and others, 1968; Best, personal communication, 1987). These occurrences could have shed detrital minerals into the Green River Basin during the late Oligocene or early Miocene when transport directions in this area were to the northeast (Hansen, 1969), but indicator minerals have not been recovered from these source localities.

In contrast, indicator minerals recovered from streams in the Uintas are smaller (<0.25 mm) than those in the Bishop Conglomerate. Transport directions of the Bishop Conglomerate are north and northeast (Hansen, 1986), contrary to a northwest direction required to link the minerals in the Green River Basin with the greatest concentration of anomalies in the Uinta Mountains (**Figure 1**). Indicator minerals were not recovered in the Uintas directly south of the Bishop Conglomerate occurrences, despite concentrated sampling in this region. A continuous mineral train may have been severed by the east flowing Henrys Fork River, which established its course in the late Pliocene (Hansen, 1969). Locating source rocks in the immediate vicinity of the Uinta anomalies is also complicated by extensive glaciation of the western and central Uinta Mountains in the Pleistocene (**Figure 1**; Atwood, 1909), which may have eroded sources such that only a few minerals are shed into streams.

It is also possible that the Uinta and Green River Basin occurrences represent two separate regions for potential kimberlite or lamproite, since they are separated by drainages which do not bear indicator minerals (**Figure 1**) and the Green River Basin occurrences are to the northwest, contrary to the north and northeast paleocurrent directions of the Bishop Conglomerate (Hansen, 1986). This is supported to some degree by the presence of higher chrome-bearing pyrope and the near absence of salitic diopsides relative to those in the Bishop Conglomerate (**Figure 4b**).

Throughout this discussion we have assumed that the igneous host for these minerals is either kimberlite or lamproite. However, mantle-derived xenoliths do occur in other igneous rocks and could be disaggregated into secondary environments to produce indicator mineral anomalies, such as alkali-olivine basalts. If basalts were the host for the Green River Basin minerals, one would expect to find basalt clasts in the Bishop Conglomerate, as basalts are very resistant to weathering. No clasts of extrusive igneous rock have been found in the Bishop. Lamproite and kimberlite weather very quickly and would not survive as clasts in a conglomerate. Interestingly, clasts of granulite 5 cm in diameter and granitic clasts as much as 15 cm in diameter have been recovered in the Bishop Conglomerate, whereas no granitic rocks or granulites are exposed in the Uinta Mountains. This supports a model in which an eruption of sufficient magnitude brought these rocks up from the basement. The model does not exclude other types of lamprophyric extrusive rocks, but until kimberlite and lamproite can be ruled out as potential igneous hosts, locating the source of the minerals is important both for academic and economic reasons.

The "Great Diamond Hoax of 1872"

A final complication in the search for the igneous host of Green River Basin minerals is the anthropomorphic factor, more commonly known as the "Great Diamond Hoax of 1872." In 1871 and 1872, an area in the extreme northeastern corner of Colorado was salted with diamonds and assorted gemstones. Clarence W. King and associates exposed the fraud, using fundamental geological deductions (Faul, 1972). The hoax area is located near Diamond Peak on the northeastern flank of the Uinta Mountains (**Figure 1**). Diamonds, with ruby and pyrope garnet, can be found on the surface of a sandstone outcrop of the Wasatch Formation northeast of Diamond Peak. The peak itself is capped by the Bishop Conglomerate and contains indicator minerals (**Figure 1**). Salting with rubies in the original hoax is documented in history, as is the purchase of 50 lbs of pyrope garnet from native Americans in northeastern Arizona (Hausel and Stahl, 1995, this volume). However, the juxtaposition of salted diamonds in an area that, 100 years later, would become part of a diamond exploration program required further investigation. Samples collected from ant mounds in the hoax area produced two diamonds and several rubies and garnets, but no clinopyroxenes were recovered. In contrast, the Bishop Conglomerate at Diamond Peak contains chrome and omphacitic diopsides, but no garnets were recovered. This provided some evidence, albeit circumstantial, that the area was the hoax locality and not a natural occurrence of detrital diamond.

More conclusive evidence was obtained by comparing the surface textures of garnets from the hoax area with garnets from ant mounds in the Green River Basin. Several unusual surface textures develop on mantle minerals prior to and during their ascent in the kimberlitic magma. These morphologies and surface features are primary growth features, formed on the minerals before weathering or erosion. Mantle garnet, pyroxene, and ilmenite often develop kelyphitic rims, due to rapid, incomplete reaction of the minerals with mantle fluids prior to or during entrainment in the igneous host. Beneath the kelyphitic rim, a hummocky surface texture develops commonly referred to as orange peel (McCandless, 1990; Garvie, 1982). Prolonged reaction with mantle fluids can develop sculpturing, which is an ordered pattern that favors cleavages in the case of pyroxenes, or crystallographic planes in the case of garnets. The Green River Basin garnets exhibit an unusual intermediate stage between orange peel and sculpturing, in which oriented planes in deep recessions are separated by orange peel surfaces. The net effect is a texture resembling the surface of a brain rather than an orange peel.

In contrast, the hoax area garnets have a distinctly different, subdued orange peel surface texture. Subdued orange peel texture is present on all of the hoax area garnets, whereas none of the Green River Basin garnets exhibit this surface texture. Significantly, pyrope garnets collected from an ant mound near the Mule Ear diatreme in northern Arizona also have subdued orange peel textures identical to those from the hoax area. These observations suggest that the garnets, rubies and the diamonds found in the hoax area are very likely part of the "Great Diamond Hoax of 1872," and do not represent a natural occurrence. In diamond exploration, stranger things have happened, and until the *bona fide* igneous host can be located, this unusual association should not be totally dismissed.

Conclusions

Detrital mantle-derived minerals found in secondary environments in the Green River Basin and Uinta Mountains are similar to minerals in peridotite and eclogite. Coexisting phases and P-T estimates suggest formation in the upper mantle under oxidizing, volatile-rich conditions over 620-820 °C for an assumed pressure range of 10-30 kb. The overall chemistry of the minerals indicates a very low potential for prior coexistence with diamond. The detrital minerals comprise part of the "indicator" mineral suite that is used worldwide in the exploration for kimberlite and lamproite, and a similar igneous host is envisioned based on the occurrence of lamproite and peridotite in the region. The primary source of these minerals is presently unknown, but is believed to be kimberlite or lamproite located in the southern Green River Basin and northeastern Uinta Mountains.

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