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Detrital Mantle Indicator Minerals in Southwestern Wyoming, U.S.A.: Evaluation of Mantle Environment, Igneous Host, and Diamond Exploration Significance

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Abstract — Detrital minerals of upper mantle origin occur in Holocene conglomerates, pediments and antmounds, and in Oligocene conglomerates in the Green River Basin of southwestern Wyoming. The minerals are derived from disaggregated eclogite and peridotite, and are similar to minerals found in mantle-derived igneous rocks such as kimberlite and lamproite, the primary host rocks for diamond. The Green River Basin bisects the lamproite fields of the Leucite Hills (Wyoming) and Kamas (Utah) and overlies the Archean Wyoming craton, which makes the region favorable for diamond occurrence. Lamproite is the favored host for the indicator minerals because unusual high-Fe salitic diopside macrocrysts from the antmounds and from the ~1 Ma old Leucite Hills lamproites to the northeast are compositionally identical. The salitic diopside is uncommon in other igneous rocks and could be a new indicator mineral for use in lamproite exploration. The Green River Basin indicator minerals are eroded from the ~29 Ma old Bishop Conglomerate, whose source area is the Uinta Mountains to the south. No continuous mineral train exists between the Uinta occurrences and those in the Green River Basin. Assessment of the minerals using established and new geochemical criteria indicate that they formed in an oxidizing upper mantle and were transported to the surface by either kimberlite or lamproite. The existence of cogenetic diamond is unlikely based on the mineral chemistry, but selective removal of diamond indicator minerals through weathering and fluvial transport cannot be excluded. Diamonds and mantle garnets are present in the northeastern Uinta Mountains, but they originated as part of the Diamond Hoax of 1872. A representative mineral assemblage from the true igneous host of the indicator minerals is required to fully establish their economic potential. Post-Oligocene erosion and Pleistocene glaciation have hindered efforts to locate a bona fide igneous source for the indicator minerals. Copyright © 1996 Canadian Institute of Mining, Metallurgy and Petroleum.

Résumé — Des minéraux détritiques provenant du manteau supérieur ont été trouvés dans des conglomérats, des pédiments et des monticules de fourmis de l'Holocéne, et dans des conglomérats de l'Oligocène du Green River Basin, au sud-ouest du Wyoming. Ces minéraux sont issus du démantellement d'éclogites et de péridotites, et sont semblables aux minéraux trouvés dans des roches ignées provenant du manteau, comme les kimberlites et les lamproïtes. Ces dernières sont les roches hôtes du diamant. Le Green River Basin recoupe les champs à lamproïte de Leucite Hills, au Wyoming et de Kamas en Utah, et il surmonte le craton archéen du Wyoming, connu pour la présence de ses diamants. La lamproïte est l'encaissant privilégié pour les minéraux «indicateurs» parce que des macrocristaux exceptionnels de diopsides salitiques riches en Fer, provenant des monticules de fourmis et des lamproïtes de Leucite Hill (~1 Ma), au nord-est, ont des compositions identiques. Le diopside salitique est rare dans les autres roches ignées et peut être un nouvel indicateur dans la recherche de lamproïte. Les minéraux «indicateurs» du Green River Basin proviennent de l'érosion du Conglomérat de Bishop, daté de ~29 Ma, dont la zone source est représentée par les Uinta Mountains, au sud. Aucune continuité des gisements minéraux n'existe entre les affleurements de Uinta et ceux du Green River Basin. La détermination des minéraux, utilisant des critères géochimiques anciens et récents, indique qu'ils se sont formés dans un manteau supérieur oxydant et qu'ils ont été remontés à la surface soit dans des kimberlites, soit dans des lamproïtes. La genèse du diamant n'est que peu liée à la chimie minérale, mais le lessivage d'indicateurs minéraux du diamant par altération météorique et par transport fluviatile ne peut être exclu. Des diamants et des grenats mantelliques sont présents dans le nord-est des Uinta Mountains, mais sont associés avec le Diamant Hoax de 1872. Un assemblage minéral représentatif de l'encaissant igné est nécessaire pour établir pleinement le potentiel économique. L'érosion post-Oligocène et la glaciation Pléistocène ont géné les recherches pour localiser une source ignée fiable pour les minéraux «indicateurs». Copyright © 1996 par l'Institut canadien des mines, de la métallurgie et du pétrole.

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Introduction

The heavy minerals (S.G.>2.9) pyrope, pyrope-almandine, chrome diopside, omphacitic diopside, chrome enstatite, chrome spinel, and picroilmenite are considered kimberlite indicator minerals because their presence in secondary environments commonly indicates a nearby kimberlitic source (Gurney et al., 1993). Indicator minerals are derived from disaggregation of mantle rocks that are caught up in the ascending kimberlite magma. These minerals also resist wear during transport and have distinctive colors or surface textures that make visual recognition and evaluation possible. Some of the minerals have a shared paragenesis with diamonds, and have unique compositions with respect to certain major and minor elements (Gurney, 1984; McCandless and Gurney, 1989; Dummett et al., 1987; Gurney et al., 1993). Indicator minerals recovered from secondary environments can be analyzed to evaluate their potential diamond association where the source is undiscovered. Indicator minerals also develop distinctive surface features during transport in kimberlitic magmas; these features are modified in the weathering and fluvial environment in a manner that can offer qualitative information as to distance from source (McCandless, 1990). Thus, indicator minerals are the most significant tool in the exploration for kimberlites and lamproites. In this paper, indicator minerals in southwestern Wyoming are evaluated with respect to possible source rocks, conditions of formation, and diamond potential.

Geologic Setting

The Green River Basin is a broad structural depression located in southwestern Wyoming, bounded to the south by the Uinta Mountains in northeastern Utah (Fig. 1). The Uintas are an east-west anticline with tilted Phanerozoic rocks flanking a core of flat-lying Proterozoic (1.6 Ga to 0.9 Ga) metasediments and Archean (2.7 Ga) gneisses (Sears et al., 1982). In the Eocene, the Uinta Mountains became an area of positive relief, depositing sand and gravel into the basin; in some places the mountains were eroded to the Precambrian core (Hansen and Bonilla, 1954). Lake Gosiute occupied the basin, which by late Eocene was filled with sediments. Further movement of the Rock Springs Uplift tilted the basin deposits to the southwest, forming an extensive erosional plain and removing several thousand feet of sediment (Bradley, 1964). Deposition of the Bishop Conglomerate followed as a result of renewed uplift of the Uinta Mountains, with detritus reaching as far north as the southern edge of the Rock Springs Uplift (Hansen, 1986). Cenozoic lamproites are found in the Leucite Hills (Wyoming) and the Kamas (Utah) areas, which lie respectively to the northeast and southwest of the Green River Basin. The detrital mantle mineral occurrences are in the southern Green River Basin, which bisects the Leucite Hills and Kamas lamproite fields. Pleistocene glaciation greatly facilitated erosion of the Uinta Mountains and the Tertiary deposits in the basin (Atwood, 1909).

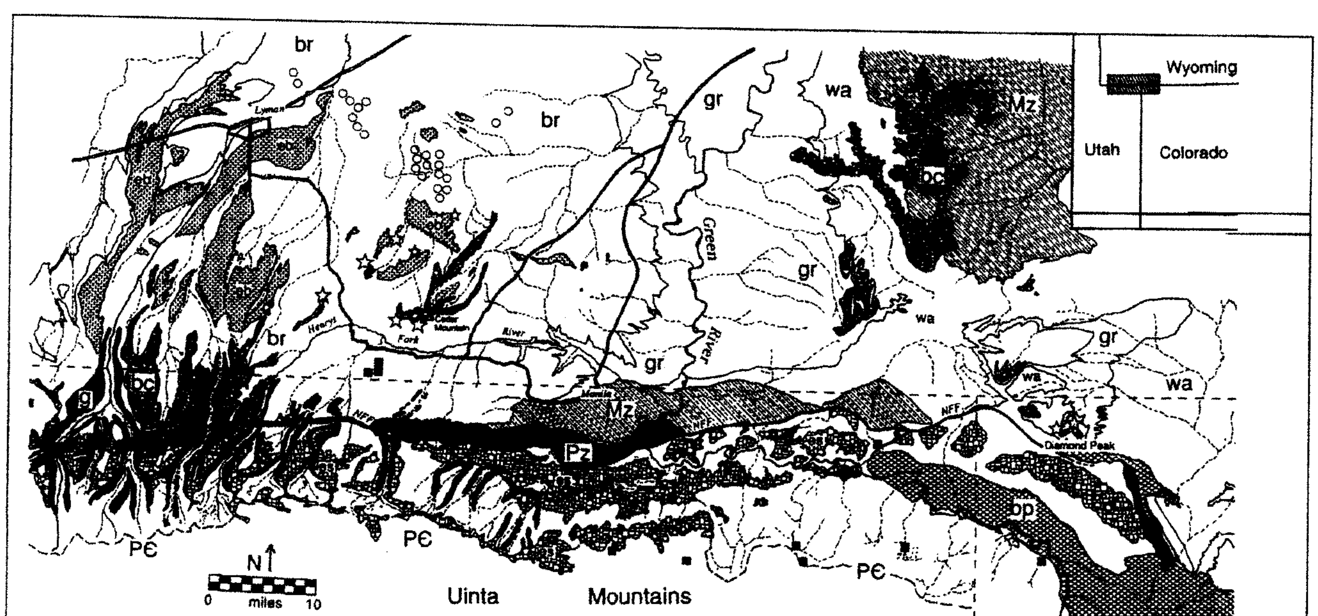


Fig. 1. Simplified geologic map of the southern Green River Basin and north slope of the Uinta Mountains. Pre-Tertiary units are represented by the Uinta Mountain Group (PC), and Paleozoic (Pz) and Mesozoic (Mz) sedimentary rocks. Tertiary lacustrine-fluvial units are represented by the Green River Formation (gr), Bridger Formation (br), and Wasatch Formation (wa). Erosional surfaces (es) are the 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. 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 antmounds (circles), pediments and conglomerates (stars) and streams (squares). Map modified after Bradley (1936, 1964), Atwood (1909), and Winkler (1970).

Previous History

Detrital indicator minerals on antmounds in the Green River Basin were first discovered in the late 1960s by a retired rancher living in Manila, Utah. The minerals were considered by locals to be rubies and emeralds, and were identified as mantle garnets and pyroxenes only in 1978 (McCandless, 1982). The antmound locations were revealed to the Minerals Division of Superior Oil in 1980, which led to an exploration program. In 1985, Mobil Oil purchased Superior Oil and terminated all mineral exploration programs. No bona fide igneous host was located, and the ultimate source of the indicators is still an open question.

Antmounds

The western harvester ant *Pogonomyrmex occidentalis* is a common species that occurs throughout western North America (Wheeler and Wheeler, 1963). In southern Africa, burrowing ants carry pyrope and ilmenite up from their kimberlite source through 50 m of Kalahari sands (Wilson, 1982). A similar situation was considered possible for the antmounds of *Pogonomyrmex occidentalis.* However, the antmounds consist of soil particles (<1 mm diam.) excavated during the digging of the nest below the surface (maximum depth 3 m), together with gravel (2 mm to 6 mm diam.) collected from the surface and placed on the mound, possibly to prevent erosion (Scott, 1950). This surface-related behavior was confirmed when all the 2 mm to 6 mm fragments of a broken glass bottle placed 10 m from an antmound appeared on the mound after only a few weeks. Soil auger samples taken at 2 m to 6 m depth near the antmounds also contained no indicator minerals. Indicator minerals that occur on the mounds evidently were collected from the surface, rather than brought up from depth.

Pediments and Conglomerates

The antmounds are situated on erosional remnants of the Eocene Bridger Formation (Fig. 1) and are associated with thin gravel pediments shed from mesas capped by the Oligocene Bishop Conglomerate (~29 Ma; Winkler, 1970). Pyrope, pyrope-almandine and chrome diopside are also located in paleopediments topographically above the antmounds, 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 are up to 12 mm in diameter in the coarser layers. Cobbles of Proterozoic Uinta Mountain Quartzite in the Bishop Conglomerate confirm that its provenance is the Uinta Mountains (Bradley, 1936). This suggests that some indicator minerals may have experienced at least three cycles of transport before being concentrated onto the antmounds.

Mineral Chemistry

The microprobe is the industry standard for analysis of indicator minerals, and most evaluation schemes utilize major and minor element abundances measured by this method. Analyses were obtained carried out 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).

In evaluating detrital indicator minerals, it is important to establish mantle derivation for the minerals. Although color is distinctive for pyrope garnet and chrome diopside, other minerals that are similar in appearance will report to the same density and electromagnetic fractions in sample processing (e.g., fluorite with 'G10' color, and uvarovite garnet with chrome diopside color). Evidence for more than one suite of mantle derivation, such as the presence of both peridotitic and eclogitic minerals, is also better than either suite alone. A useful first step is to classify the minerals by color. Although color classification is subjective, it is useful in helping the explorationist establish which minerals are the most distinctive in a particular project area. Color intensity also varies with grain size, but color shade does not, and indicator minerals are usually recovered in narrow size fractions that allow for color grouping by shade. Mineral grains from the antmounds fall in a narrow size range due to the preference of *Pogonomyrmex Occidentalis* for grains 2 mm to 6 mm in diameter.

Garnets

Three color shades can be recognized in the antmound garnets: purple, orange-pink, and dark red. The orange-pink and purple garnets are divided into two groups on the basis of magnesium number (Mg# = atom Mg/Mg+Fe) and Cr_2O_3 content (Table 1; Fig. 2). The purple garnets are chrome

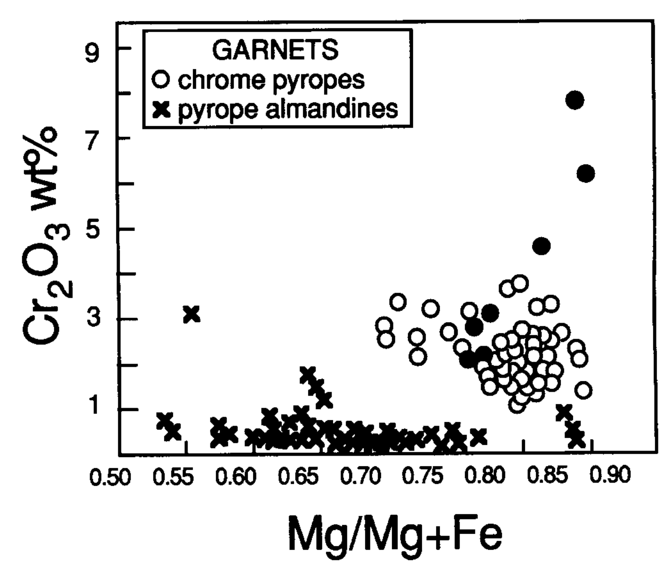


Fig. 2. Cr_2O_3 content versus Mg number for garnets. Crosses are eclogitic garnets, circles are peridotitic garnets (after Dawson and Stephens, 1975). Solid circles are garnets from the Uinta Mountains.

Table 1. Representative analyses of garnet, clinopyroxene and orthopyroxene from the Green River Basin, southwestern Wyoming.

					Garnets						
(wt%)	CM06 (cp)	HK20 (cp)	L3054 (cp)	C113 (cp)	CM91 (cpa)	L2619 (cpa)	CM33 (cpa)	CM98 (cpa)	CM01 (pa)	CM35 (pa)	CM95 (pa)
SiO_2	41.4	42.2	41.7	42..0	40.7	39.5	41.0	41.3	41.2	40.8	41.5
TiO_2	0.07	0.11	0.08	0.05	0.03	0.00	0.11	0.05	0.05	0.02	0.00
Al_2O_3	21.9	22.3	16.7	21.7	21.5	20.5	21.6	21.2	23.0	22.9	23.2
Cr_2O_3	2.60	2.39	7.78	2.47	1.29	3.03	2.14	2.62	0.27	0.32	0.18
FeO	8.16	9.37	6.72	7.91	14.7	17.3	11.5	12.4	13.7	12.0	11.9
MnO	0.49	0.56	0.29	0.45	0.67	0.84	0.52	0.72	0.42	0.42	0.35
MgO	20.1	19.7	20.0	21.1	15.4	12.1	16.9	16.4	16.3	15.9	16.9
CaO	5.28	5.07	7.24	5.05	5.48	6.20	6.03	5.38	6.14	6.86	5.91
Na_2O	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Total	100.0	101.7	100.5	100.7	99.8	99.5	99.8	100.1	101.1	99.2	99.9
					Clinopyroxenes						
(wt%)	CE60 (cd)	CE55 (cd)	LY3205 (cd)	CE21 (sa)	CE47 (sa)	BCG6 (sa)	CE18 (sa)	CTD5 (om)	CE15 (om)	LY3176 (om)	TP97A (om)
SiO_2	53.2	52.5	55.0	53.7	53.5	53.1	53.1	54.5	54.8	54.7	54.0
TiO_2	0.08	0.07	0.18	0.04	0.15	0.15	0.16	0.03	0.03	0.02	0.15
Al_2O_3	1.45	1.30	1.18	0.59	0.79	0.72	0.59	4.48	2.65	1.41	5.14
Cr_2O_3	1.11	0.30	1.24	0.02	0.06	nd	0.05	0.19	0.34	1.07	0.41
FeO	3.40	4.54	2.46	6.77	8.01	7.75	7.97	1.72	2.91	1.55	1.40
MnO	0.12	0.17	0.07	0.42	0.31	0.28	0.33	0.04	0.08	0.08	nd
MgO	18.2	17.7	17.9	15.8	16.1	15.9	16.6	14.7	15.7	16.3	15.2
CaO	21.82	22.6	21.9	23.1	20.7	23.1	21.0	21.7	23.2	22.9	21.6
Na_2O	0.62	0.57	0.92	0.62	0.82	0.62	0.64	2.52	1.91	1.37	2.02
Total	100.0	99.8	100.9	101.1	100.4	99.7	100.4	99.9	101.6	99.4	99.9
					Orthopyroxenes						
(wt%)	OPX1	OPX2	OPX3	OPX4	CE10	CE22	TP1001	TP1002	TP1003	TP1004	TP1005
SiO_2	55.0	55.5	55.1	56.0	57.2	55.4	54.8	55.6	55.7	55.3	55.9
TiO_2	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd	nd
Al_2O_3	3.50	2.22	2.59	2.88	1.42	2.51	2.36	2.95	2.82	3.13	2.73
Cr_2O_3	0.82	0.98	1.36	0.84	0.41	0.81	0.32	0.54	0.54	0.60	0.36
FeO	4.49	4.76	4.40	4.77	4.72	5.11	7.47	4.70	4.75	4.67	4.60
MnO	0.13	0.14	0.12	0.11	0.15	0.12	0.14	0.20	0.12	0.09	0.14
MgO	33.1	34.6	34.6	34.5	35.1	33.6	31.8	34.7	34.4	34.3	34.9
CaO	2.23	1.23	1.03	0.79	0.40	1.29	2.28	0.57	0.97	1.19	0.49
Na_2O	0.05	nd	0.04	0.04	0.08	0.04	0.86	0.76	0.63	0.77	0.83
Total	99.3	99.4	99.2	99.9	99.5	98.9	100.0	100.0	99.9	100.1	99.9

Key: nd = not detected; cp = chrome pyrope; cpa = chrome pyrope-almandine; pa = pyrope-almandine; cd = chrome diopside; sa = salitic diopside; om = omphacitic diopside.

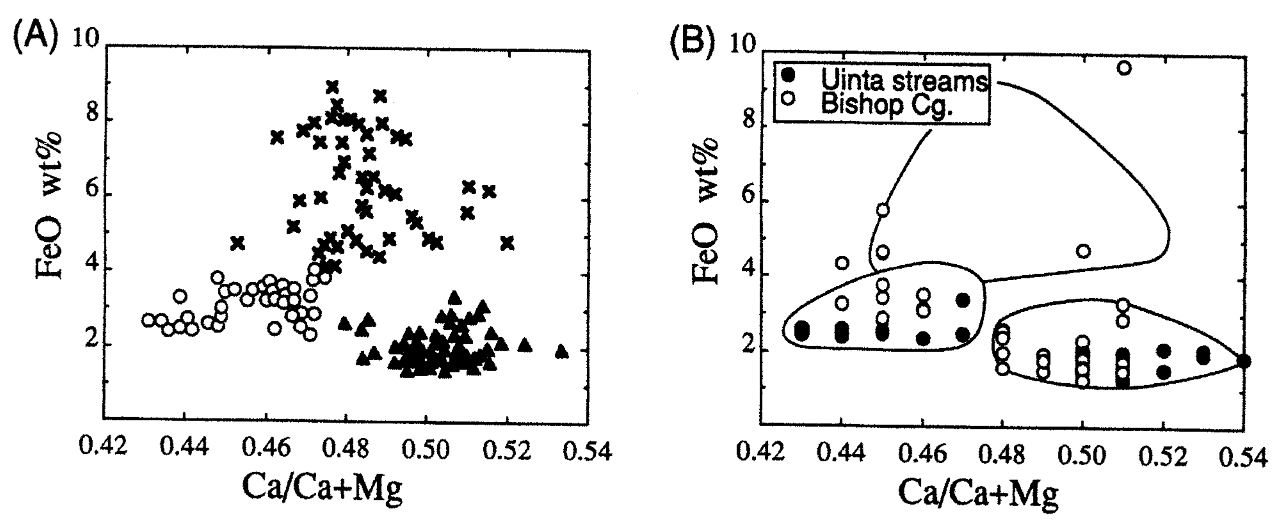


Fig. 3. (a) FeO versus Ca number for clinopyroxenes. (a) Clinopyroxenes from antmounds. Circles are chrome diopsides, crosses are salitic diopsides, triangles are omphacitic diopsides. (b) Clinopyroxenes from the Bishop Conglomerate and from Uinta Mountain streams. The three fields correspond to the data clusters in (a). Note that salitic diopsides are absent in the Uinta Mountains.

pyropes with 1.09% to 7.78% Cr_2O_3 and Mg# = 0.700-0.874, whereas the orange-pink garnets are pyrope-almandines with 0.09% to 0.86% Cr_2O_3 and Mg# = 0.534-0.838. The red garnets are pyrope-almandine with low Mg# (0.556-0.652) but high Cr_2O_3 content (1.28% to 3.08%). The chrome pyropes fit into group 9 of Dawson and Stephens (1975), defined using garnets from kimberlite and from garnet-bearing lherzolite, websterite and harzburgite. The pyrope-almandine garnets are classified as group 3 or 6, which comprise garnets from eclogites (Fig. 3; 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 unusual chrome-rich pyrope-almandine megacrysts from the Dullstroom kimberlite (L.R.M. Daniels, unpubl. analyses).

Clinopyroxenes

Clinopyroxenes can also be divided into emerald green, dark green, and blue green groups (McCandless, 1982). The emerald green clinopyroxenes form a subcalcic group (Ca# = Ca/Ca+Mg = 0.431-0.475) with low FeO (2.23% to 4.64%), whereas the blue green group is calcic (Ca# = 0.480-0.535) with 0.04% to 3.31% FeO, and the dark green group subcalcic to calcic (Ca# = 0.441-0.521) with 4.07% to 9.09% FeO (Table 1; Fig. 3a). The lattermost is a high-Fe salitic diopside group that requires as much as 3.21% Fe_2O_3 to attain charge balance. The subcalcic group is chrome diopside with 0.65% to 1.85% Cr_2O_3. The calcic group is omphacitic diopside with high Na_2O (0.16% to 3.33%). Salitic diopsides were not recovered in the Uinta Mountains, but are present in the Bishop Conglomerate (Fig. 3b).

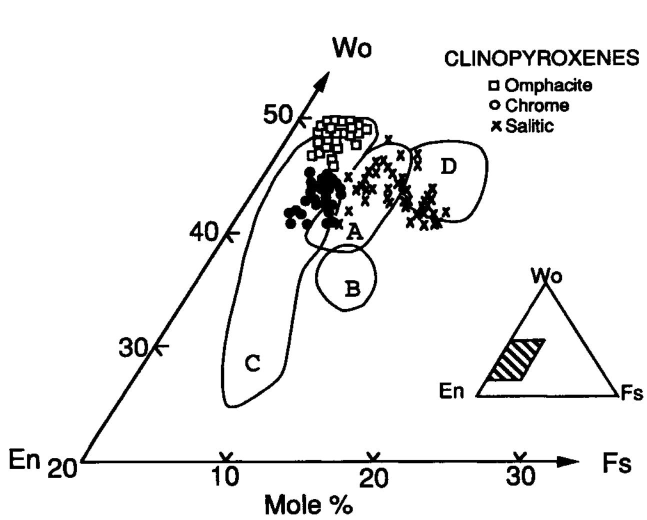


Fig. 4. Clinopyroxene compositions in terms of mole percent Wollastonite-enstatite-ferrosilite (Wo-En-Fs). Outlined areas represent clinopyroxenes from mica-amphibole-rutile-ilmenite-diopside xenoliths (A), from clinopyroxene-ilmenite intergrowths (B), from garnet peridotite and websterite (C), and from the Leucite Hills (D). Data are from this study and from Dawson and Smith (1977), Waters (1987) and Barton and van Bergren (1981).

In the pyroxene quadrilateral (Fig. 4), 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). Some salitic diopsides lie in region A, defined for pyroxenes from xenoliths of the mica-amphibole-rutile-ilmenite-diopside suite (Dawson and Smith, 1977; Waters, 1987). Most of the salitic diopsides lie in region D, defined in this paper using clinopyroxene megacrysts from the Hatcher Mesa lamproite in the Leucite Hills, Wyoming (Barton and van Bergren, 1981; McCandless, unpubl. 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.

Orthopyroxenes

Eleven orthopyroxenes were analyzed and are classified as Cr-Al enstatites after Stephens and Dawson (1977); they contain 0.32% to 1.36% Cr_2O_3, 1.42% to 3.50% Al_2O_3 and 0.40% to 2.28% CaO (Table 1). 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. The orthopyroxenes are also distinguished by a greenish-brown color and a high cleavage density. This latter characteristic allows for faster mineral wear with transport, and may account for the relative scarcity of orthopyroxene in the antmounds.

Oxides

Ilmenites in lamproites and kimberlites can exhibit a wide compositional range with respect to Cr_2O_3 and MgO because they are derived from a variety of mantle and crustal rocks sampled by the ascending magma. Mantle-derived ilmenite has elevated Cr_2O_3 indicative of ultramafic derivation; elevated MgO suggests low oxidation potential in the magma, which appears to be crucial for diamond preservation (Gurney et al., 1993; Fipke et al., 1995). Although hundreds of oxide minerals from the antmounds were analyzed, only 203 ilmenites had >0.10% Cr_2O_3 or >2.00% MgO, considered in this paper as arbitrary minimum values for ultramafic affinity, and less than 10% had Cr_2O_3 over 0.30% (Table 2). Thirty spinels had MgO contents ranging from 5.26% to 15.3% and Cr_2O_3 from 34.0% to 66.3%, but only four, with Cr_2O_3>55.0%, are considered to be potentially from an ultramafic source (Waldman et al., 1987; Gurney et al., 1993).

Mineral Inclusions

Garnets and pyroxenes from the antmounds contain a variety of inclusions. The inclusions are euhedral to subhe-

Table 2. Representative analyses of oxide grains, and of inclusions in garnet and clinopyroxene from the Green River Basin, southwestern Wyoming.

	Oxide Grains										
(wt%)	LY2572 (il)	BCG35 (il)	BCG36 (il)	LY2618 (il)	LY3043 (il)	LY3179 (il)	LY3194 (ch)	LY3092 (ch)	BNS1 (ch)	LY2572 (ch)	LY2572 (ch)
TiO2	33.11	40.1	38.5	39.8	46.3	45.5	0.12	0.16	0.18	nd	0.30
Al2O3	0.48	0.32	0.32	0.64	0.44	0.59	8.11	9.68	31.9	19.4	21.4
Cr2O3	2.66	3.77	3.28	4.01	0.45	nd	66.3	59.4	35.4	48.0	37.4
FeO	58.8	48.4	50.6	48.8	47.1	47.2	14.3	19.4	18.5	21.5	28.3
MnO	nd	0.25	0.22	0.35	0.37	0.46	1.57	0.30	0.31	0.34	0.35
MgO	2.95	4.79	4.12	5.07	4.08	5.46	7.00	10.2	14.3	9.79	11.3
Total	98.0	97.6	97.0	98.7	98.7	99.2	97.4	99.1	100.6	99.0	99.1

	Mineral Inclusions										
(wt%)	CO12 (si/pa)	H201 (si/cp)	C111 (fsp/cp)	C112 (cd/cp)	P151 (pa/om)	P152 (pa/om)	TP98 (ru/om)	P211 (cc/sa)	P212 (il/sa)	CD51 (ru/sa)	BCG37 (ch/sa)
SiO2	99.5	97.0	59.8	54.2	40.3	40.6	0.12	1.27	0.30	1.12	nd
TiO2	0.01	0.02	0.07	0.06	0.06	0.04	95.9	nd	26.1	95.8	0.53
Al2O3	0.51	0.31	22.5	2.85	22.3	22.7	0.32	nd	0.31	0.32	12.4
Cr2O3	0.01	0.07	0.02	1.71	0.50	0.53	2.15	nd	0.07	0.79	47.3
FeO	0.74	0.62	0.78	1.44	17.3	18.1	1.43	0.35	66.2	0.81	26.3
MnO	0.02	0.07	0.01	0.06	0.02	0.63	nd	0.01	0.19	nd	0.40
MgO	0.23	0.22	0.75	16.7	14.7	13.9	nd	0.55	1.34	0.29	8.65
CaO	0.19	0.22	1.75	22.4	5.05	5.09	nd	59.2	1.02	1.17	nd
Na2O	—	—	2.64	1.92	0.01	0.01	—	—	—	—	—
K2O	—	—	10.9	—	—	—	—	—	—	—	—
Total	101.2	98.6	99.4	101.3	100.9	101.6	99.9	64.4	95.6	100.3	95.6

Key: nd = not detected; — = not analyzed for; il = ilmenite; ch = chromite; si = silica phase; fsp = feldspar; ru = rutile; cc = carbonate; diagonal line indicates host mineral where /pa = pyrope-almandine; /cp = chrome pyrope; /om = omphacitic diopside; sa = salitic diopside.

dral, and average 0.05 mm in diameter; in the case of ilmenite and rutile, the inclusions form needles up to 4.0 mm in length. A silica phase occurs in several pyrope-almandine garnets and in one chrome pyrope. Rutile is present in salitic and omphacitic diopsides. Ilmenite occurs in salitic diopside, but it is low in MgO and Cr_2O_3, unlike kimberlitic ilmenite grains. Carbonate occurs with the ilmenite and is nearly pure calcite. Chrome spinel in salitic diopside is chemically similar to the single chromite grains (Table 2), which suggests that they coexisted.

Pyrope-almandine garnets were found in four omphacitic diopsides; one chrome pyrope had a two-phase inclusion of omphacitic diopside with K-feldspar ($Or_{67}Ab_{24}An_9$ mole %). The garnet inclusions span the range of Ca-Mg-Fe values for the single garnet grains, and the omphacitic diopside inclusions are similar in composition to single omphacitic diopsides. This demonstrates a prior coexistence of omphacitic diopside with pyrope and pyrope-almandine garnet in rocks of eclogitic composition, similar to the situation in Cr-rich eclogites found at the Roberts Victor kimberlite, South Africa (Fig. 5; Hatton, 1978). One diopside is a miniature xenolith, containing 26 subhedral garnets 0.1 mm to 0.2 mm in diameter.

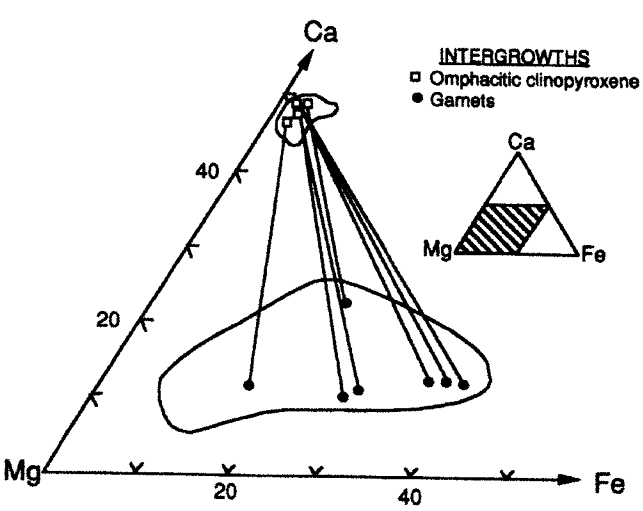


Fig. 5. Composition of garnet/clinopyroxene intergrowths in terms of mole percent Ca-Mg-Fe. Tie lines connect coexisting phases. Outlined areas indicate compositional regions for discrete omphacitic diopside (upper) and pyrope and pyrope-almandine grains from antmounds (lower).

Discussion

Conditions of Formation

The coexistence of garnet and clinopyroxene can be used to estimate the temperature of equilibration for a given pressure, using the distribution coefficient of Fe and Mg in garnet and diopside (Ellis and Green, 1979). A temperature range of 620°C to 730°C is obtained over an assumed pressure range of 20 kb to 30 kb for the pyrope-almandine garnets in omphacitic diopside. In comparison, temperatures calculated for the chrome diopside in pyrope are 790°C to 840°C over 20 kb to 40 kb, suggesting different conditions of formation for the two types of intergrowths. These estimates can be coupled with heat flow data to constrain the 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). Using a heat flow between 40 mWm^{-2} and 60 mWm^{-2}, the intergrowths of omphacitic diopside and pyrope-almandine would have been derived from a region near the crust-mantle boundary (40 km to 65 km), with the chrome

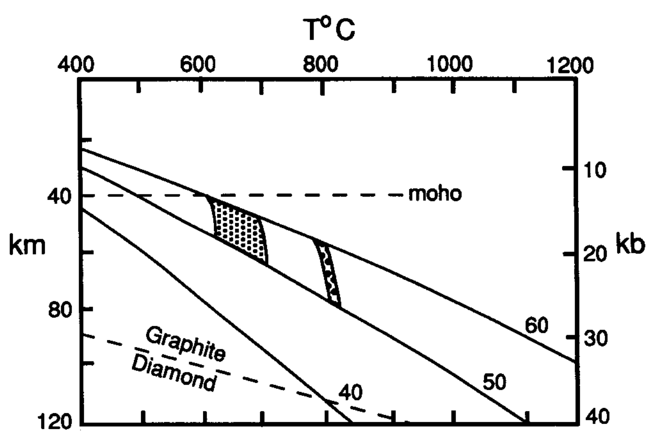


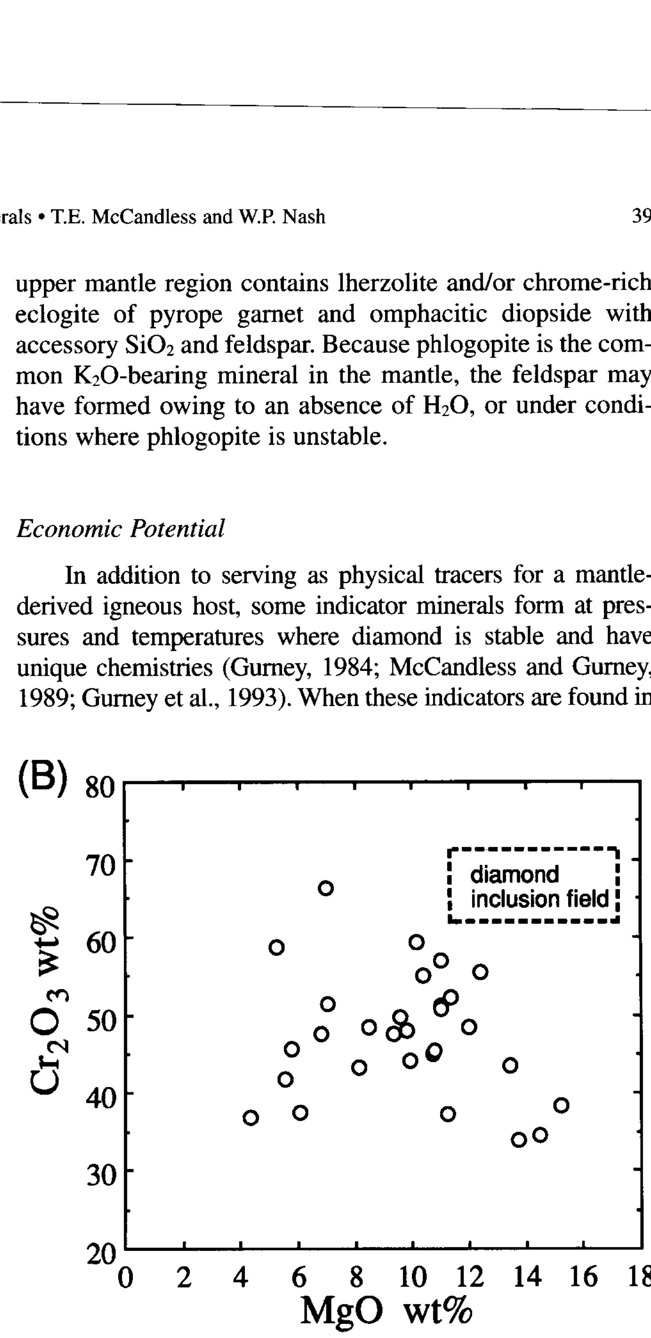
Fig. 6. Estimated pressure-temperature conditions for intergrowths of omphacitic diopside and pyrope-almandine (fine stipple) and chrome pyrope and omphacitic diopside (coarse stipple). Temperatures calculated using the method of Ellis and Green (1979), with all Fe as FeO. Geotherms are in mWm^{-2}. Diamond-graphite phase boundary is from Kennedy and Kennedy (1976).

pyrope/chrome diopside pair derived from between 55 km and 80 km (Fig. 6).

The inclusion assemblage and P-T estimates provide a model of the mantle region from which the minerals were 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, rutile and ilmenite (Table 2). 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 mica-amphibole-rutile-ilmenite-diopside xenoliths. Fe_2O_3 required for charge balance and inclusions of Fe-ilmenite suggest relatively oxidizing conditions within the higher part of the upper mantle (Dawson and Smith, 1977). 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 et al., 1993). When these indicators are found in



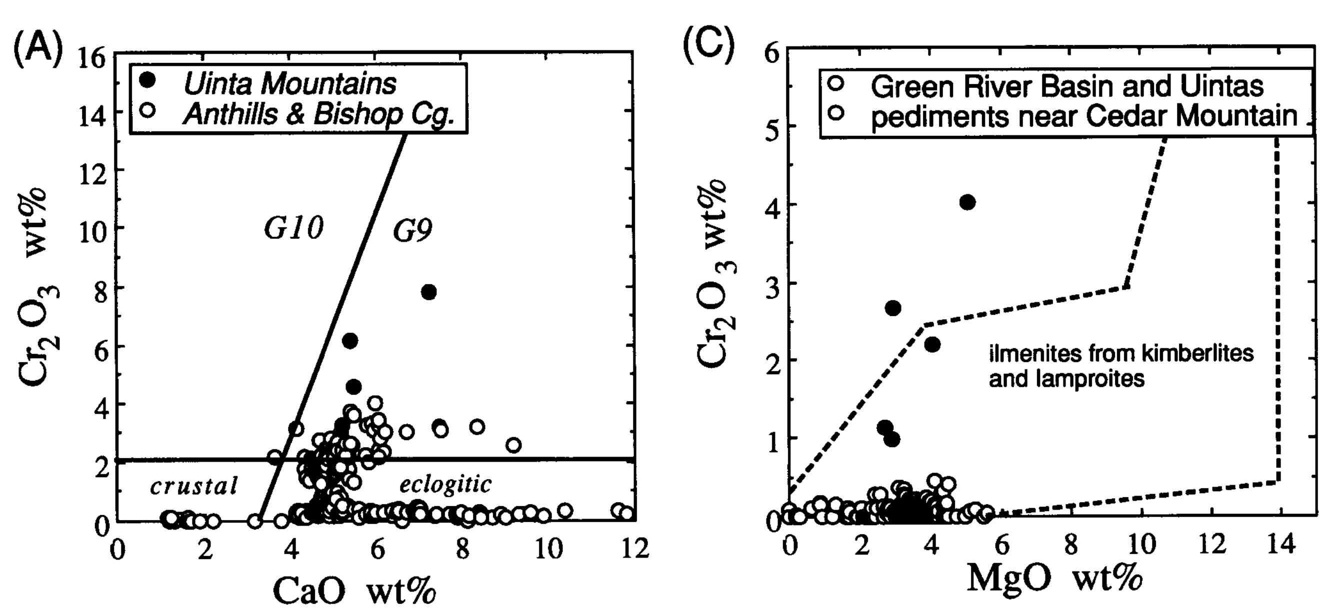


Fig. 7. (a) Cr_2O_3 and CaO contents of garnets. A lack of garnets in the diamond-favourable G10 field indicates a low diamond potential. (b) Cr_2O_3 and MgO content of chromites. No chromites plot in the diamond inclusion field, indicating a poor probability for cogenetic diamond. (c) Cr_2O_3 and MgO content of ilmenites. For some grains high Cr_2O_3 contents indicate a mafic affinity, but the majority of the grains have low MgO which suggests poor diamond preservation potential (fields modified from Gurney et al. 1993).

secondary environments, locating their host becomes a priority. The indicator minerals of this paper are therefore evaluated with respect to diamond potential in the following discussion.

It is known that G10 pyrope as defined by Gurney (1984) originates within the stability field for diamond. No pyropes analyzed from the Green River Basin or Uinta Mountains have G10 chemistry, except for two garnets near the boundary (Fig. 7a). 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 source concentrates is believed to indicate the presence of eclogitic diamond in the igneous host (Gurney et al., 1993). In the eclogitic garnets of this paper, Na_2O is below the limit of detection (<0.01%), and as such precludes eclogitic diamond potential. Chromite (i.e., chrome spinel) is widely accepted as a useful indicator mineral because it is a common inclusion in diamond (Waldman et al., 1987; Dummett et al., 1987; Gurney et al., 1993). Chromite inclusions in diamond are relatively restricted in Cr_2O_3-MgO space (Fig. 7b). None of the chromites from this paper have MgO and Cr_2O_3 contents similar to those from diamond inclusions. In ilmenites, the presence of elevated MgO and Cr_2O_3 values combined with low FeO (as total Fe) is usually considered a positive indication of diamond potential (Gurney et al., 1993). Ilmenites with moderate contents of Cr_2O_3 and MgO were recovered in the vicinity of Cedar Mountain (Fig. 1), but they represent less than 10% of all ilmenites analyzed, and lie outside the area for ilmenites associated with diamondiferous diatremes (Fig. 7c).

As an alternative to using minerals diagnostic of diamond cogenesis as established in previous studies (Gurney, 1984; McCandless and Gurney, 1989; Gurney et al., 1993), the coexistence of garnet and diopside can be used to establish the P-T conditions of the mineral phases prior to disaggegation. This approach may be used in other exploration projects that lack diamond cogenetic indicators. As pointed out previously, estimated temperatures for the garnet/clinopyroxene intergrowths examined in this paper are outside the diamond stability field, and indicate that the potential for diamonds in the host rocks of these minerals is very low.

Potential Source Locations

In using indicator mineral chemistry to prioritize target areas, the explorationist must assume that the detrital suite is representative of the primary host rocks from which it is derived; that is, weathering or transport has not selectively removed G10 garnets, high Cr picroilmenites, etc., from the suite. This assumption has been confirmed with some certainty in northern Canada, where continental glaciation has dispersed indicator minerals up to 750 km from their kimberlitic source (Krajick, 1994; Gurney, 1995). It has also been successful in arid areas where movement of indicators from the igneous host has not been significant (Gurney et al., 1993). However, the assumption remains to be tested in areas where temperate climate and multiple cycles of fluvial transport are involved, such as in mid-continent Canada (Swanson and Gent, 1993). In tropical regions, silicate indicator minerals do not survive lateritization (Garvie, 1981), which complicates application of the complete diamond indicator suite (Gurney et al., 1993). In any exploration program, therefore, the igneous host must be located and tested for diamond cogenetic minerals before it can be established with certainty that diamonds are not present.

Deposition of the Bishop Conglomerate occurred as a result of renewed uplift of the Uinta Mountains, with detritus reaching as far north as the southern edge of the Rock Springs Uplift (Hansen, 1986; Fig. 1). The Bishop Conglomerate is poorly-sorted, with boulders several meters in diameter in the coarser layers. Indicator minerals occur 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 (Fig. 1). The Bishop Conglomerate can contain over 50 vol.% clay-sized particles; such particles have been shown experimentally to inhibit mineral wear in high energy fluvial systems (McCandless, 1990). Chrome diopside travels poorly in fluvial systems of tropical regions (Mosig, 1980), but its survival during transport is also dependent on its primary morphology. Chrome diopside megacrysts in kimberlites in eastern Colorado have a high cleavage density that allows for rapid disaggregation, producing anomalies consisting of hundreds of very angular grains (Dummett et al., 1987). With the exception of the orthopyroxenes, none of the pyroxenes of this paper have high cleavage density. Thus, the occurrence of 12 mm diameter omphacitic diopside grains in the Bishop Conglomerate does not necessarily require a proximal igneous host.

Although the Uinta Mountains south of the Bishop Conglomerate indicator mineral anomalies are devoid of igneous rocks excepting a few diorite dikes (Ritzma, 1974, 1980), peridotite and lamproite do occur in the western Uintas, with ages from 11.7 Ma to 40.4 Ma (Best et al., 1968; M.G. Best, pers. comm. 1987). Outcrops of peridotite and lamproite may 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 no indicator minerals have been recovered from the known exposures.

Indicator minerals have been recovered from streams in the Uinta Mountains, but all of the grains are small (<0.25 mm) compared to 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 anomalies in the Uinta Mountains (Fig. 1). Indicator minerals were not recovered in the Uinta Mountains directly south of the Bishop Conglomerate occurrences, despite concentrated sampling in this region. A continuous mineral train may have been removed 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 (Fig. 1; Atwood, 1909), which may have eroded or obscured exposures such that only a few minerals were shed into streams.

It is also possible that the indicator mineral occurrences in the Uinta Mountains and the Green River Basin represent two separate igneous host regions (Fig. 1; 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 (Fig. 3b). This difference in chemistry is not likely due to grain size bias during sampling, as salitic diopsides in the <0.25 mm size range are also found in the Bishop Conglomerate, and no indicators are present in the +5 mm size fractions in the Uinta Mountains. The alternate possibility, that some indicators have been removed during transport, was alluded to previously.

Throughout this discussion we have assumed that the igneous host for these minerals is either kimberlite or lamproite. Mantle-derived xenoliths do occur in other igneous rocks such as alkali-olivine basalts, and could be disaggregated into secondary environments to produce indicator mineral anomalies. If a basalt originally hosted the Green River Basin indicator minerals, basaltic clasts should be present in the Bishop Conglomerate, as they are relatively resistant to weathering. In contrast, lamproite and kimberlite weather quickly and would not survive as clasts in a conglomerate. No clasts of extrusive igneous rock have been found in the Bishop Conglomerate. Clasts of granulite and eclogite 5 cm in diameter and granitic clasts as much as 15 cm in diameter have been recovered in the Bishop Conglomerate, which supports a model wherein an eruption of sufficient magnitude brought these rocks up from depth. However, this does not exclude other types of lamprophyric extrusive hosts. Until kimberlite and lamproite can be ruled out as potential igneous hosts, locating the source of the indicator minerals remains an important task both for academic and economic reasons.

A final complication in the search for the igneous host of the indicator minerals is illustrated by the Great Diamond Hoax of 1872. In 1871 and 1872, an area near Diamond Peak on the northeastern flank of the Uinta Mountains was salted with diamonds and assorted gemstones. Clarence W. King and associates exposed the fraud, using fundamental geological deductions (Faul, 1972). Diamonds, rubies, and pyrope garnet still can be found on a sandstone outcrop northeast of Diamond Peak; the peak itself is capped by the Bishop Conglomerate. Salting with rubies during the original hoax is historically accurate, as is the purchase of 50 lbs of pyrope garnet from native Americans in northeastern Arizona (Hausel and Stahl, 1995). The garnets from Arizona are derived from lamprophyric and kimberlitic exposures in southeastern Utah and northern Arizona. An antmound in the hoax area produced two diamonds and several rubies and garnets, but no clinopyroxenes were recovered.

Major element chemistry cannot be used to distinguish the garnets in the hoax area from those occurring naturally, because mantle garnets exhibit indistinguishable compositions worldwide except with respect to diamond potential. (It is this consistency that makes it possible to prospect using indicator minerals.) To establish that the garnets were part of the hoax and not a natural occurrence, their surface textures were compared with garnets from antmounds in the Green River Basin. Distinctive surface textures develop on mantle minerals prior to and during their ascent in the kimberlitic magma, the most common of which is the kelyphitic rim. Kelyphitic rims are composed of alteration minerals that form on the surface of mantle minerals due to reaction with mantle fluids during entrainment in the igneous host. Beneath the kelyphitic rim, a hummocky surface texture develops which is commonly referred to as an orange peel, or sub-kelyphitic, surface (McCandless, 1990; Garvie, 1981). The Green River Basin garnets exhibit an unusual, deeply pitted, orange peel surface (Figs. 8a and b). In contrast, garnets from lamprophyric/kimberlitic rocks in southern Utah have a distinctive, low relief orange peel surface texture (Figs. 8c and d). A similar, low relief texture is present on most of the garnets in the hoax area, whereas none of the Green River Basin garnets exhibits this surface texture (Figs. 8e and f). These observations suggest that the garnets, rubies and diamonds found near Diamond Peak are very likely related to the original diamond hoax of 1872, and do not represent a natural occurrence. However, until *bona fide* igneous host for the indicator minerals 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 the Uinta Mountains are similar to minerals comprising peridotite and eclogite. However, diagnostic diamond indicators are absent. P-T estimates from inclusions in the detrital minerals suggest that they originally formed in the upper mantle under oxidizing, volatile-rich conditions at 620°C to 820°C for an assumed pressure range of 10 kb to 30 kb, outside the conditions of formation for diamond. The detrital minerals comprise part of the "indicator" mineral suite that is used worldwide in the exploration for kimberlite and lamproite. A similar igneous host is envisioned based on the occurrence of lamproite and peridotite in the region. The dark green, high-Fe salitic diopsides of this paper are restricted in occurrence to the Hatcher Mesa lamproite in the Leucite Hills and the antmounds in the Green River Basin, and may represent a new indicator mineral that can be used in the exploration for lamproites. The primary source of these mantle-derived indicator 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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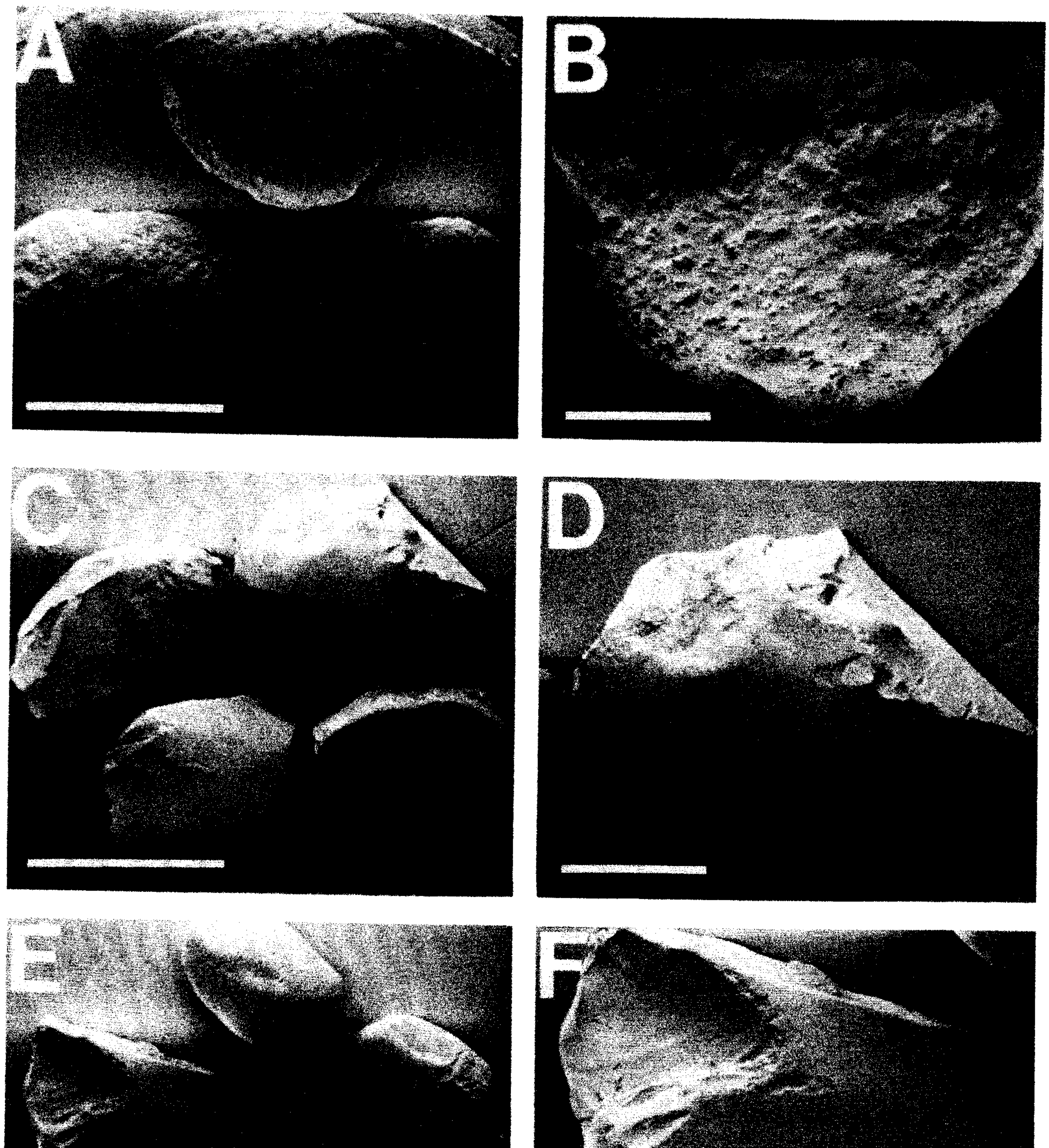

Fig. 8. Photomicrographs of garnets from antmounds in the Green River Basin. (a) Characteristic subrounded shapes. (b) Close-up of upper centre grain in (a) showing pitted orange-peel surface. (c) Garnets from the Mule Ear Diatreme, Utah with typical subangular shapes. (d) Close-up of upper right grain in (c) showing low relief, orange-peel texture. (e and f) Garnets from near Diamond Peak, Colorado. A few garnets are similar to those from the Green River Basin antmounds (e.g. upper centre grain), but most are similar to the Mule Ear garnets and probably originated as part of the Diamond Hoax of 1872. Scale bars are 4 mm in (a), (c), (e), and 1.5 mm in (b), (d), and (f).

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