

Migration of the Mendocino Triple Junction and the Origin of Titanium-Rich Mineral Suites at New Idria, California

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Abstract

Titanium-rich minerals in peculiar mineral suites are associated with the New Idria serpentinite, California. Previously some of these Ti-rich minerals have been described as metasomatic in origin. This paper presents new data on the compositions and parageneses of these mineral suites, and presents a petrologic model for their formation by essentially isochemical metamorphism from precursor assemblages. All of these mineral suites are manifestations of tectonic uplift, M2 metamorphism, and magmatism associated with the passage of the Mendocino Triple Junction underneath New Idria at about 12 Ma.

Ti-bearing garnets associated with chlorite and diopside occur in metamorphosed pyroxenite veins that have been tectonically dismembered during emplacement of the serpentinite. These veins have been metamorphosed under greenschist or blueschist (M1) and lower greenschist (M2) conditions. Accessory minerals in these rocks include, among others, perovskite, titanite, zircon, Ti-REE-vesuvianite, zirconolite, pyrophanite, and copper sulfides. *Benitoite*, the state gemstone of California, is formed by lower greenschist M2 metamorphism of tectonic blocks of Franciscan greenstone entrained within the serpentinite. Benitoite is found in association with neptunite, joaquinite, natrolite, actinolite and clinzoisite. *Kaersutite* crystals up to 30 cm long, containing up to 4 wt% TiO₂ occur with albite, apatite, and other accessory minerals in the New Idria syenite, a small group of stocks associated with the serpentinite. This syenite is shown to have an intrusive origin and an age of 12.4 ± 0.8 Ma.

Introduction and Summary of Previous Work

THE NEW IDRIA serpentinite, located in the California Coast Ranges, has been recognized as a unique locality since the 1851 discovery of cinnabar ore, which led to over a century of profitable mercury mining (Fig 1A). However, the first published map showing the New Idria district as an extensive *serpentinite* mass was that of Anderson and Pack (1915), who were mainly interested in the oil potential of the area. Subsequent exploration led to mining of other commodities, including magnesite, chromite, gemstones, and most recently, chrysotile asbestos. Since the landmark paper on benitoite by Louderback (1909), the region has also received attention from mineral collectors and academic geologists. However, prior to 1950, most of these investigations were topical and did not attempt overall understanding of the serpentinite and its unique minerals.

In 1950, R. G. Coleman began the field work at New Idria for his dissertation, under the guidance of

Stanford Professor Colin Osborne Hutton. He spent several field seasons there, and completed his thesis *Mineralogy and Petrology of the New Idria District, California*, in 1957. The stated purpose of this thesis was to “study a peculiar suite of minerals found within the serpentinite and the associated rock types.” These peculiar minerals remain a subject of study and controversy to this day. Coleman established that the serpentinite body is a diapir, fault bounded on all sides with the sense of shear, where it could be determined showing upward movement against the country rock. The most prominent of the boundary faults is known as the New Idria thrust fault, marking the eastern boundary of the serpentinite. He also showed that numerous tectonic blocks of varying lithologies are entrained within the diapir.

In 1972 the historic mercury mines closed, due to restrictions imposed by the U.S. Environmental Protection Agency (EPA). Also, starting in 1982 and continuing to the present, EPA has designated portions of the district as Superfund sites, but related to the mining and milling of chrysotile asbestos, rather than mercury. Further investigations, related to remediation of such perceived hazards, have been

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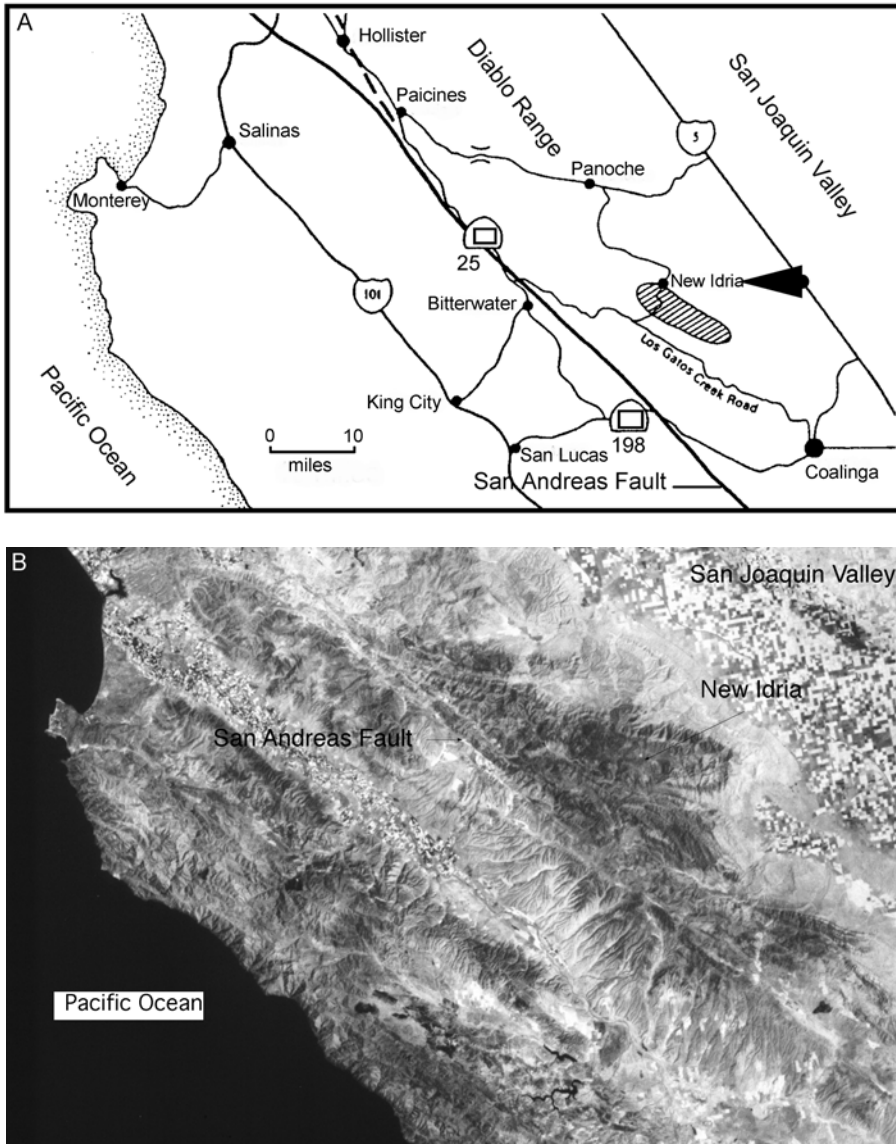


FIG. 1. A. Map of central California showing New Idria (arrow), modified from Coleman (1986). B. Satellite photo of a portion of central California, from Monterey Bay and the Carmel Peninsula to the San Joaquin Valley. The New Idria serpentinite is at the core of an antinodal structure lying east of the San Andreas fault. The structure is rotated $\sim 20^\circ$ relative to the trend of the San Andreas fault.

conducted by consulting firms employed by EPA (e.g. Levine-Fricke, 1989; Woodward-Clyde, 1990).

Regional Geology

The most remarkable feature of the New Idria District is an oval, 23 by 8 km, fault-bounded, ser-

pentinite massif flanked by steeply dipping, locally overturned, sedimentary and metamorphic rocks of the Jurassic–Cretaceous Franciscan and the Cretaceous to Pliocene Great Valley Sequence. The outcrop area of serpentinite rock exceeds 150 km². The New Idria serpentinite lies in the core of one of a series of anticlines that characterize the Coast

Ranges in this part of California (Arnold and Anderson, 1908; Taliaferro, 1943; Eckel and Myers, 1946; Zigler et al., 1986). Numerous other partially to completely serpentinized peridotites lie in similar structural positions along the east side of the San Andreas fault in central and northern California.

The New Idria serpentinite is a part of the Coast Range Ophiolite in California; its emplacement within the Franciscan Formation is related to subduction of the Farallon plate during the Jurassic Period. The age of the ophiolite is 153 to 165 Ma (Hopson et al., 1981). The tectonic setting of the ophiolite was a forearc basin lying between the Sierra Nevada volcanic arc to the east and a trench-subduction complex to the west, represented by the Franciscan Formation (Bartow, 1990). A piece of oceanic crust or underlying mantle was obducted onto the North American continent, later to rise through the Franciscan Formation as serpentinite diapirs (but see Namson and Davis, 1988 for an alternative interpretation). The structural trend of the New Idria serpentinite body and related folds and faults is 300° , and so is inclined 20° to the San Andreas fault system, which trends 320° in this part of California, as shown in Figure 1B.

Based on mineralogy, bulk composition analyses, weathering profiles, and relict textures in the serpentinite, the ultramafic protolith was determined to be a peridotite consisting chiefly of dunite with subordinate harzburgite, together with minor lherzolite and pyroxenite veins (Van Baalen, 1995). The New Idria serpentinite is locally heterogeneous: numerous rock bodies of limited extent and contrasting physical character are included within the mass of the serpentinite. These bodies are isolated from each other, have sheared contacts with the surrounding serpentinite, and show evidence of brittle to ductile deformational processes; they are interpreted as tectonic blocks or "knockers." A consequence of this interpretation is that some tectonic blocks have a history distinct from the rest of the serpentinite. Coleman (1957) noted the lithologic similarity between many of these blocks and the Franciscan rocks found at the margin of the serpentinite; following a suggestion made by Louderback (1909), he proposed that some tectonic blocks were derived from the Franciscan section during emplacement of the serpentinite. The pre-entrapment history of some of the blocks is fairly clear; for others the evidence is equivocal.

The dimensions of the tectonic blocks range from 1 to 1000 meters; many of the larger blocks are

elongate and have a SE-NW orientation. Most of the smaller blocks and some of the larger blocks have hidden margins, so that their shapes and orientations cannot be determined unambiguously. The blocks can be distinguished in the field by distinct soil color and erosion properties, contrasting vegetation above, and anomalous mineral assemblages within.

Chlorite-Diopside-Garnet Association

Titaniferous andradite garnets coexisting with chlorite and diopside occur in isolated patches throughout the New Idria serpentinite. Green to yellow-weathering, blue-grey to white, foliated rocks consist mainly of chlorite and diopside, with melanite garnet (1–5 wt% TiO_2), schorlomite garnet (> 5 wt% TiO_2), variable amounts of serpentine and magnetite, and several Ti-rich accessory minerals, notably vesuvianite (idocrase), perovskite, and titanite (sphene), along with minor chromite, kammererite, chromian garnet, apatite, zirconolite, pyrophanite, pyrolusite, and copper sulfides. Other accessory minerals have also been reported (e.g., Coleman, 1986). Carbonate minerals including calcite and hydromagnesite fill some of the voids in the rock. These surface alteration minerals were probably precipitated from meteoric waters and are not part of the metamorphic phase assemblage. The rocks are poorly exposed in most places; exceptions occur where a slab of more resistant serpentine cap rock overlies the chlorite-diopside-garnet rock.

Garnet in these outcrops occurs in two generations: an early (M1) generation of foliation-parallel blebs, stringers, and isolated grains, and a late (M2) generation of subhedral to euhedral vein-filling crystals. The synkinematic M1 garnets are anhedral to subhedral; the individual grains are typically too small to be seen with a hand lens: using SEM, isolated garnets of micron size can be found. Typically the garnets have a light to deep red color in thin section and are isotropic; they tend to form elongate, polycrystalline aggregates up to 5 mm long. The M1 garnets principally occur in chlorite-rich domains of the host rock, so that the garnets coexist with abundant chlorite and minor diopside, along with accessory minerals. Prior to the advent of the electron microprobe, the M1 garnets generally went unrecognized.

The second generation M2 garnets, which fill crosscutting veins and form drusy coatings on the walls of cavities, have received the most attention by

mineral collectors because of their relatively large size (up to 1 cm) and the relative abundance of gem and museum-quality specimens. M2 garnets are commonly anisotropic, sector twinned, and zoned.

A representative outcrop

An outcrop, originally designated as 34-50 by Coleman (1957), has continued to be a key locality for understanding the formation of Ti-garnets. At this locality a set of four resistant rock towers surrounded by talus sits astride a low N-S ridge at an elevation of 4360 ft. (1329 m) MSL, above the stream bed of the San Benito River, a short distance downstream from the Gem Mine. The four towers are 2–3 m tall and arranged roughly in a 5×10 m rectangle oriented with its long axis SE-NW. The towers and associated talus cover approximately 100 m² and are surrounded by dense brush. Each of the towers is capped by dark green, sheared but relatively resistant serpentinite, under which lies a slab of light green to white, foliated, chlorite-diopside-garnet rock dipping gently to the east. The four towers are apparently the eroded remnants of a once-continuous slab of chlorite-diopside-garnet rock capped by serpentinite. The slab strikes 336° and dips 23° E, subparallel to the New Idria thrust fault. The outer contacts of the slab are not exposed, so its original lateral dimensions are unknown. However, the upper contact of chlorite-diopside-garnet rock against the serpentinite is exposed in each tower, and in the northwest tower the upper and lower contacts are both exposed, so that a complete cross section is visible. Figure 2A is a photograph of the west face of the northwest tower, showing both contacts. Figure 2B shows the essential features of this outcrop, in schematic form (see also Coleman, 1957).

The chlorite-diopside-garnet rock of the 34-50 locality is divided into a central, strongly foliated, chlorite-rich zone and two diopside-rich outer zones, an upper and a lower. The central zone includes the prominent horizontal band marked by the hammer head in Figure 2A, while the outer zones include the white horizontal bands in that photograph. The mineralogy of the three zones is similar, but the modal proportions of the minerals vary. This locality is remarkable for the low abundance of magnetite, a nearly ubiquitous accessory mineral at New Idria.

The contacts between the various zones are abrupt to gradational; locally there is considerable interfingering, especially between diopside and serpentinite in the outer margin. The interfingering has

probably resulted from recrystallization of diopside and serpentinite under shearing stress, and possibly from local mobility of Ca in metamorphic fluids (e.g. Labotka and Albee, 1979).

The M1 fabric at 34-50 is crosscut by veins interpreted as tension cracks. These veins are filled with M2 garnet, diopside, and minor chlorite. The M2 minerals in the veins have beautifully preserved delicate textures with no sign of post-mineralization deformation. Therefore the M2 mineralization was postkinematic.

Worldwide occurrences

Other localities that have Ti-garnets associated with serpentinites include the Sangabawa District of Japan (Onuki et al., 1981), and some of the Alpine serpentinites. Recently, Müntener and Hermann (1994) have described an occurrence of Ti-garnets in the Malenco serpentinite in the Italian Alps. Their paragenesis in chlorite-diopside host rocks related to pyroxenite veins appears to be very similar to the New Idria occurrences. Switzer et al. (1971) reported an oceanic occurrence of Ti-garnet hosted by chlorite-diopside rock in a dredge sample of serpentinitized peridotite from the Mid-Atlantic Ridge near 43° N. Switzer's samples, now residing at the Smithsonian Institution, were examined by the author in conjunction with this study, and were found to be very similar to New Idria samples.

Mineral compositions

Most, but not all, of the New Idria garnets are calcic and belong to the andradite clan. Titanium substitution in these garnets occurs principally along the exchange vector $\text{TiFe}^{2+}\text{Fe}^{3+}_2$, which is algebraically equivalent to TiFe_{-1} . Some of the garnets are also hydrous (see Lager et al., 1989) and vary along the hydrogarnet exchange vector H_4Si_{-1} . Still other garnets are Ti-grossulars, a portion of composition space not yet widely explored (but see Labotka, 1995). Compositional variation in these Ti-grossulars occurs along the vector $\text{TiFe}^{2+}\text{Al}_{-2}$. In the present study, more than 300 electron microprobe analyses of garnets were completed. The mean TiO_2 content of the garnets is 5 wt%. One of the garnets contains 15.62 wt. % TiO_2 , which corresponds to approximately one Ti atom per formula unit, i.e. the schorlomite end member.

Some yellow, Ti-free andradites occur in tectonic blocks of massive antigorite. Outside of the tectonic blocks, widespread but minor emerald green uvarovitic garnet is found as an alteration product of the



B

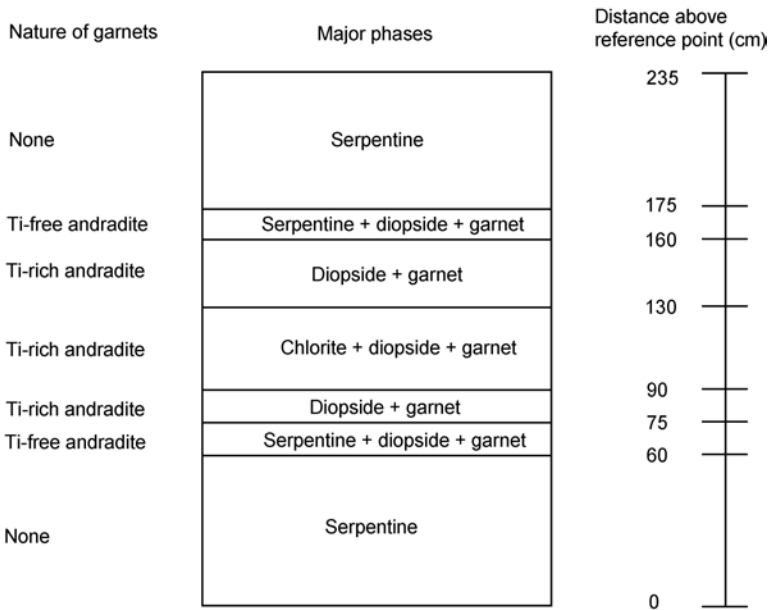


FIG. 2. A. Photo of the northwest tower of the 34-50 locality, showing a metapyroxenite vein. The tower is capped by resistant serpeninite. The hammer handle marks a diopside-rich horizon of chlorite-diopside-garnet rock. The hammer head marks a chlorite-rich horizon. See Figure 2B for a schematic drawing. B. Schematic drawing of 34-50 locality, measured vertical section.

ubiquitous chromite grains in the serpentinite. Uvarovitic garnet inside tectonic blocks may also be titaniferous, however.

Andradite garnets containing up to 38 mole percent uvarovite component are found in slabby serpentized harzburgite blocks. These garnets are anhedral, emerald green, and often associated with relict chromite grains. Chromian andradites are very rare: these are believed to be the first chromian andradites reported from the United States.

The SAG projection

Titaniferous garnets may best be considered as modified andradite and grossular garnets, in which Ti substitutes for Fe and Al, respectively. A convenient projection for this portion of composition space is the SAG projection, using the ternary components schorlomite ($\text{Ca}_3\text{TiFe}^{2+}\text{Si}_3\text{O}_{12}$), andradite ($\text{Ca}_3\text{Fe}^{3+}\text{Si}_3\text{O}_{12}$) and grossular ($\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$), projected through pyralspite and all other components. The transformation equations for this projection, using atom fractions of Ti, Fe, and Al respectively, are as follows:

$$S = \frac{2\text{Ti}}{\text{Ti} + \text{Fe} + \text{Al}} \quad A = \frac{\text{Fe} - \text{Ti}}{\text{Ti} + \text{Fe} + \text{Al}}$$

$$G = \frac{\text{Al}}{\text{Ti} + \text{Fe} + \text{Al}}$$

Garnet compositions

Figure 3A shows 300 electron microprobe analyses of M1 and M2 garnets (undifferentiated) from New Idria. All but one of the garnet analyses in this study plot within the triangle defined by these components. One Ti-rich analysis plots just outside the triangle beyond the S vertex. Compositional variation along the three binaries S-A, A-G, and S-G can be expressed by the exchange vectors TiFe_{-1} , AlFe_{-1} , and TiFeAl_{-2} , respectively. In Figure 3A, most, but not all, of the analyses are relatively low in Al and plot near the S-A join.

Figure 3B presents analyses of Ti-garnets from other localities for comparison with the New Idria samples. The Onuki et al. (1981) data are from the Sangabawa district of Japan; the Müntener and Hermann (1994) data are from Val Malenco, Italy. In both cases the garnets are found in serpentinites. The Switzer et al. (1971) samples are from serpentized peridotite dredged from the Mid-Atlantic

Ridge near 43° N. Ti-andradites of course also have an igneous paragenesis, in association with alkaline rocks such as nepheline syenites. The Flohr and Ross (1989) data are from Magnet Cove, Arkansas, where the garnets have an igneous origin.

Figure 3B shows that the trends of the Onuki et al. (1981) and the Flohr and Ross (1989) analyses are roughly parallel to the S-A join at relatively constant G. One of the Onuki samples is somewhat enriched in Al (4.75 wt% Al_2O_3). The Müntener and Hermann (1994) samples fall into two groups: a low-Ti group that lies along the S-G join, and a Ti-enriched group that lies near the S-A join at low and constant Al.

Comparison of Figures 3A and 3B shows that the New Idria garnets vary over a much wider compositional range than garnets from other localities. This is particularly noticeable in the Al contents: some New Idria garnets are quite aluminous, as expressed by the G component. Labotka (1995) explored aluminous Ti-garnets further, but in a different bulk composition.

Compositions of chlorite and diopside

Chlorite and diopside occur with Ti-garnets; Table 1 provides representative compositions for these minerals from the 34-50 locality. Note that the diopsides are more magnesian than the coexisting chlorites, an exception to the general rule that chlorite is more magnesian than coexisting mafic minerals. Further discussion of chlorite compositions is found below under M1 and M2 metamorphism.

For convenience, the 300 garnet analyses represented in Figure 3A were divided into two groups based on their analysis totals: the first group had totals $\geq 99\%$ and the second had totals $< 99\%$, and may be hydrous. The mean compositions of these two groups are shown in Table 1. A comparison of the two groups shows that the garnets in the second group are slightly lower in Si and Ca, noticeably higher in Ti and Al, noticeably lower in Fe, and higher in the trace constituents Cr and Mg. Mn as a trace constituent is the same in both groups. The lower Si contents of the garnets in the second group correlate in a general way with lower analysis totals, but a plot (not shown) of SiO_2 wt% versus analysis totals shows considerable scatter. Therefore, the simple hypothesis that low analysis totals are mainly due to the hydrogarnet substitution does not fully explain the compositional complexity in these garnets.

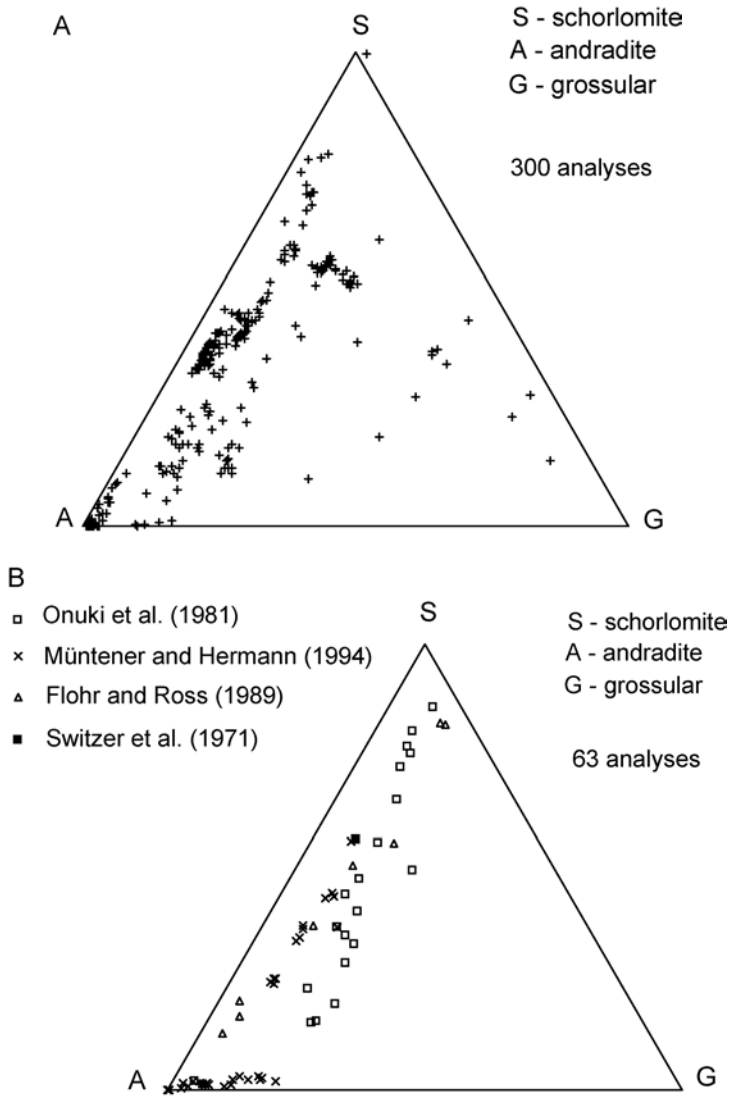


FIG. 3. A. Compositions of New Idria garnets, M1 and M2 (undifferentiated). B. Compositions of garnets from other localities.

Origin of chlorite-diopside-garnet rocks

Chlorite-diopside-garnet rocks are the host rocks for the garnets described in this study. As explained in this section, these rocks were formed by essentially isochemical metamorphism, under greenschist or blueschist-facies conditions, of fertile ultramafic rocks of pyroxenite composition. In pyroxenites Ti would have been present in clinopyroxene and/or spinel. The formation of andradite garnet is probably due to component reactions involving the heden-

bergite component of clinopyroxene (Taylor and Liou, 1973):

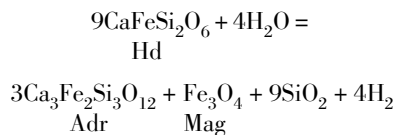


TABLE I. Representative Mineral Compositions¹

Oxide	Chlorite	Diopside	Garnet >99%	Garnet <99%
SiO ₂	29.95	54.47	34.47	34.01
TiO ₂	0.01	0.02	4.40	5.73
Al ₂ O ₃	16.87	0.70	1.85	2.85
Cr ₂ O ₃	0.22	0.04	0.06	0.15
Fe ₂ O ₃ *	n.d.	n.d.	24.08	20.75
FeO*	12.81	2.66	n.d.	n.d.
MnO	0.14	0.14	0.15	0.15
MgO	26.35	16.85	0.27	0.70
CaO	0.04	24.56	34.21	34.06
Na ₂ O	0.01	0.31	n.d.	n.d.
K ₂ O	0.01	n.d.	n.d.	n.d.
H ₂ O (diff)		13.57		
Totals	100.00	99.75	99.49	98.40
Mg#	0.78	0.92		

*All Fe in chlorite and diopside is treated as FeO, in garnet Fe₂O₃.

In these reactions, the SiO₂ component does not appear as free quartz; rather, its effect is equivalent to a decrease in the Ol/Opx ratio and a change in the amounts of modal chlorite, serpentine, and brucite produced during serpentinization. The presence of josephinite at New Idria (Coleman, 1986), suggests that serpentinization may have occurred at very low oxygen fugacities. Peretti et al. (1992) demonstrated that andradite garnet occurs in serpentinites in conjunction with reduced phases such as josephinite and awaruite. The formation of Ti-garnet probably occurs in reactions similar to those above, in which a Ti component of the precursor clinopyroxene, (expressed as the exchange vector TiFe₋₁), enters the garnet phase.

Coleman (1957) referred to the chlorite-diopside-garnet rocks of New Idria as *metasomatic chlorite-rich rocks*, and proposed that they were formed by metasomatic alteration of ordinary serpentinite rock. He proposed a process of alteration of serpentine rock to one containing chlorite, diopside, and garnet, by fluids importing Ca, Al, Ti, and Fe while exporting Mg and Si to the surrounding environment. He suggested that the metasomatic fluids were hydrothermal fluids related to the intrusion of some small intrusive syenite bodies 1 to 2 km distant. The main reason for this hypothesis was to

explain the extreme enrichment of Ti in the “metasomatic” rocks (more than four orders of magnitude over the adjacent rock). He suggested that breakdown of Ti-rich amphibole in the intrusives could have contributed aqueous Ti complexes to the hydrothermal fluids. However, kilometer-scale movement of Ti-rich hydrothermal fluids is problematical given current understanding of the very limited mobility of Ti in crustal rocks (see Van Baalen, 1993).

Ti (and Al) complexes in aqueous fluids at low temperatures have such low concentrations that the required advection of these elements could be accomplished only by massive fluxes of fluids infiltrating through a well developed subterranean plumbing system. At New Idria, however, there is no evidence of a large-scale system of fluid conduits and vein systems. On the contrary, the chlorite-diopside-garnet blocks are relatively small, isolated from one another, and show no obvious spatial relationships to each other. In the present study, no evidence of veins connecting blocks has been found. Furthermore, close examination of serpentinite at the contact with the syenite intrusive shows no sign of metasomatic alteration. The serpentine side of the intrusive contact instead appears to be relatively dry and baked, with an aureole containing in turn

olivine, antigorite, and lizardite/chrysotile at increasing distances from the contact. No mineralization was found. On the syenite side of the contact, however, there is in places a rind of albitite or chloritized amphibole with titanite that suggests deuteric alteration. Formation of titanite as a breakdown product of Ti-amphibole suggests that Ti may have been conserved rather than exported during this process.

MI metamorphism

All of these problems with the metasomatism hypothesis are solved by the simpler proposal adopted here, that the chlorite-diopside-garnet rocks represent nearly isochemically metamorphosed veins of a pyroxenite bulk composition that are included within the larger serpentinite body. Only water is required as an open system component. This interpretation is supported by the field relations and rock textures, by comparison of bulk compositions, and finally by analogy with similar rocks from other localities that do not have the same local geology as New Idria.

If the blocks are indeed metapyroxenites, the question of the origin of the blocks must be addressed; there are several possibilities. It is possible that the metapyroxenite blocks are tectonic inclusions from an unknown source entrained within the serpentinite; tectonic inclusions of a variety of other rock types are present. It is also possible that the metapyroxenite blocks might be primary igneous features and have a cognate origin with the depleted dunites and harzburgites. They may represent pre-serpentinization crosscutting veins and dikes, or alternatively, cumulate layers. Other, less-altered ophiolitic slices in the Coast Ranges, e.g. Elder Creek in Tehama County, have meter-scale veins and dikes of clinopyroxenite (Rynearson, 1946; Shervais and Beaman, 1994). The present random orientation of the blocks may well result from boudinage or other tectonic dismemberment of dikes followed by block rotation during the diapiric ascent of the New Idria serpentinite.

Comparison of bulk compositions

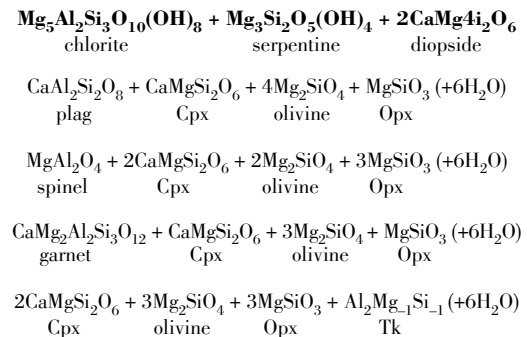
Table 2 provides major- and trace-element whole-rock analyses for chlorite-diopside-garnet rocks from New Idria, as well as pyroxenites from other localities, and in the final column, a harzburgite from New Idria. Comparison of these data shows that, although pyroxenites vary significantly in composition, it would take extreme enrichment in the harzburgite Ti, Al, Nb, Zr, V, and Ce concentrations

to bring these elements up to the range shown by the pyroxenites and chlorite-diopside-garnet (metapyroxenite) rocks.

Component reactions

If the chlorite-diopside-garnet rocks are derived from pyroxenites, then it should be possible to write whole-rock reactions relating the two rock types. These whole-rock reactions in turn must be linear combinations of the component reactions. Since the Ti-garnets, magnetite, and chromite are only accessory minerals to the rock-forming minerals chlorite and diopside, we are justified in ignoring them at first. This simplification allows us to initially employ the model system CaO-MgO-Al₂O₃-SiO₂-H₂O (CMASH).

A representative model composition in CMASH for a pyroxenite is 10MgO + 2CaO + Al₂O₃ + 9SiO₂ + 6H₂O. Under upper mantle conditions, this composition (ignoring the water) would exist as one of the anhydrous phase assemblages shown below. Recast in terms of New Idria minerals, this composition is equivalent to a mixture of chlorite + serpentine + diopside. The composition corresponds to the open triangle symbol in Figure 4. This point falls within the pyroxenite (olivine websterite) field, and lies near the Hawaiian reference sample.



The equivalence of the anhydrous phase assemblages and the metamorphic reactions connecting them were demonstrated by Thompson (1982, 1991) using the concepts of modal space and reaction space. At New Idria, there is no reason to prefer one of these upper mantle assemblages over another; for present purposes they will be considered equivalent. In other words, the conclusions drawn here do not depend on the starting mantle mineralogy.

Figure 4 shows New Idria samples and pyroxenites from other localities for comparison. All samples plot in the pyroxenite field, most in the olivine websterite portion of pyroxenite space.

TABLE 2. Pyroxenite and Metapyroxenite Bulk Compositions

Sample:	990-22	1091-07	890-12	H-1 ¹	97jsb-2 ¹	66SAL-1	Peretti	989-45	Model comp
Lithology:	Chl-Di	Chl-Di	Chl-Di	pyroxenite	pyroxenite	pyroxenite	pyroxenite	harzburgite	Chl-Di
Location:	New Idria	New Idria	New Idria	China	China	Hawaii	Val Malenco ¹	New Idria	New Idria
SiO ₂	39.39	45.35	36.22	48.80	47.24	44.82	45.61	47.89	45.86
TiO ₂	1.25	1.39	0.66	0.42	0.66	0.52	0.18	0.00	0.00
Al ₂ O ₃	11.82	4.14	16.37	6.08	10.28	8.21	4.20	1.45	11.12
Fe ₂ O ₃ *	14.56	11.02	11.81	12.38	10.61	10.12	8.55	6.44	0.00
MnO	0.21	0.26	0.26	0.20	0.14	0.19	0.14	0.16	0.00
MgO	19.24	15.47	30.00	19.89	20.35	26.53	30.55	43.83	30.77
CaO	12.81	21.26	3.57	11.62	8.87	8.12	10.75	0.06	12.23
Na ₂ O	0.21	0.20	0.11	0.81	1.07	0.89	0.00	0.16	0.00
K ₂ O	0.00	0.01	0.00	0.02	0.70	0.03	0.00	0.00	0.00
P ₂ O ₅	0.00	0.09	0.02	0.03	0.10	n.d.	0.02	0.00	0.03
Total	99.49	99.17	99.01	100.25	100.02	99.43	100.00	100.00	100.00
					Trace, ppm				
Nb	10.89	12.42	1.29	0.50	46.40	n.d.	<4	0.00	n.a.
Zr	198.37	198.25	50.46	15.81	20.69	n.d.	<10	4.23	n.a.
Sr	44.73	125.55	1.68	106.50	9.09	n.d.	<15	1.75	n.a.
Zn	211.13	104.58	134.07	80.50	33.57	n.d.	19	55.38	n.a.
Ni	850.85	722.39	1083.27	435.00	735	2000	332	2171.52	n.a.
Cr	1023.11	799.67	1418.00	1085.00	2457	2000	853	2304.95	n.a.
V	167.89	138.01	136.35	199.90	3955	n.d.	64	24.00	n.a.
Ce	56.96	49.93	2.31	5.60	9.66	n.d.	n.d.	0.51	n.a.
Ba	5.49	47.20	3.86	1.00	6.27	n.d.	<10	32.45	n.a.
La	27.04	22.80	0.15	1.47	4.29	n.d.	n.d.	1.14	n.a.

*All expressed as Fe²⁺O₃.¹Samples H-1 and 97jsb-2 are from Xu, 2002; Peretti sample from Peretti et al., 1992.

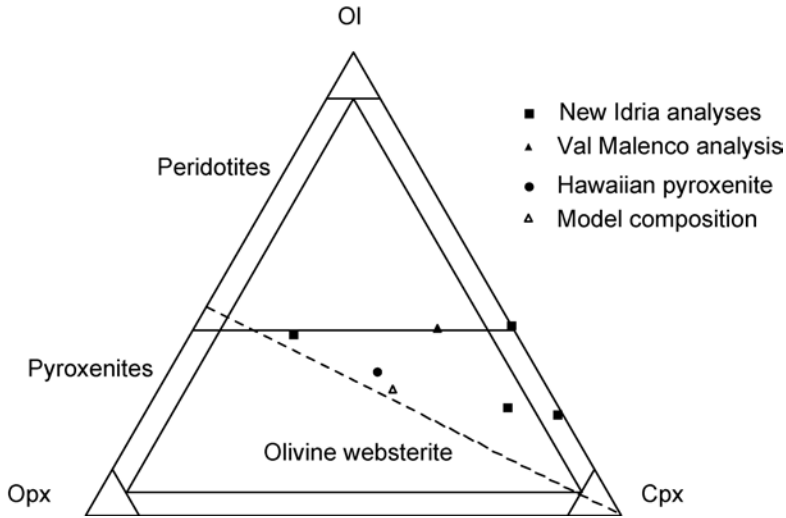


FIG. 4. Projection of chlorite-diopside-garnet compositions into an untramafic ternary diagram. Projection is from spinel and minor components, units are oxy equivalents. Dashed line is the locus of points equivalent to mixtures of chlorite + diopside.

M2 metamorphism

As noted previously, M1 mineral assemblages featured penetrative foliation defined by M1 chlorite, chlorite-rich and diopside-rich domains, and foliation-parallel blebs, stringers, and isolated grains of Ti-garnet, plus accessory minerals. M2 mineral assemblages form in veins and tension cracks that crosscut the M1 foliation, and uniformly strike perpendicular to the anticlinal structures of the region, and to the New Idria thrust fault. These tension cracks are interpreted features related to the uplift and formation of the anticlinal structures of the Coast Ranges that lie to the east of the San Andreas fault. M2 garnets fill some of these veins and cracks and form drusy coatings on the walls of others. The garnets are subhedral to euhedral, and commonly anisotropic, sector twinned, and zoned. Most are deep red in thin section. These garnets coexist with diopside whiskers that are affixed to vein walls, and with chlorite that locally forms fan-shaped bundles. It is here proposed that the M2 minerals have recrystallized from M1 assemblages, with perhaps millimeter-to-centimeter mobility of low-solubility components such as Ti (see Van Baalen, 1993, for discussion of Ti mobility)

The garnets also contain fluid inclusions. A fluid inclusion study (Van Baalen, 1995), showed that the maximum trapping temperature, reflecting

formation of the garnet, was probably less than 400°C. The presence of small amounts of methane in fluid inclusions from the garnets of Santa Rita Peak also implies low oxygen fugacity during M2 metamorphism.

M2 chlorites form a compositional group distinct from M1 chlorites. As shown in Figures 5 and 6, M2 chlorites are more magnesian and less aluminous than M1 chlorites. Because the Fe/Mg ratio in chlorite depends on the local silicate bulk composition, creation of M2 magnetite may effectively remove Fe from the system and result in more magnesian chlorite. The aluminum in chlorite is accommodated by the Tschermak exchange vector $Al_2Mg_{-1}Si_{-1}$ (Tk). Laird (1988) has demonstrated that in mafic and ultramafic rocks, a strong positive correlation exists between progress along the Tk vector in chlorite, and metamorphic grade. We may therefore conclude, based on chlorite compositions, that the M2 event occurred at lower metamorphic grade than M1.

Many M2 garnets are zoned, with Ti-rich cores and Ti-poor rims. This is particularly true in vein-filling subhedral garnets. Van Baalen and Zbinden (1989) proposed that M2 garnets formed in the presence of fluids that evolved from Ti-rich to Ti-poor compositions. Some exceptions to the normal zoning patterns were observed: some garnets varied from Ti-poor to Ti-rich and back to Ti-poor in a traverse

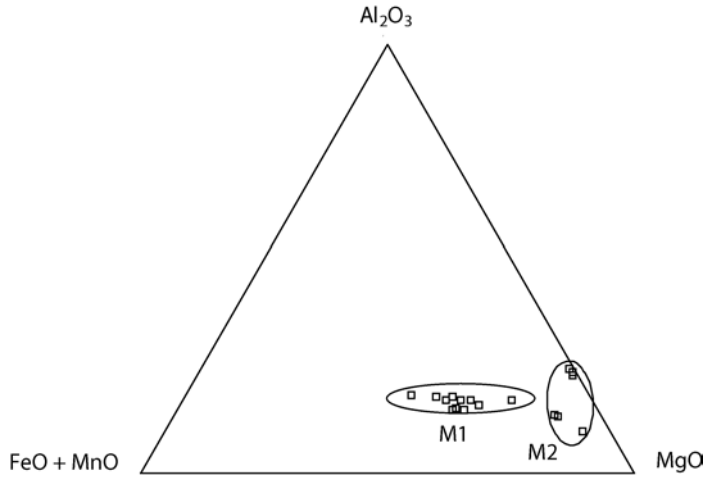


FIG. 5. AFM projection of M1 and M2 chlorite compositions. Projection is from SiO_2 , H_2O ; units are mole fractions.

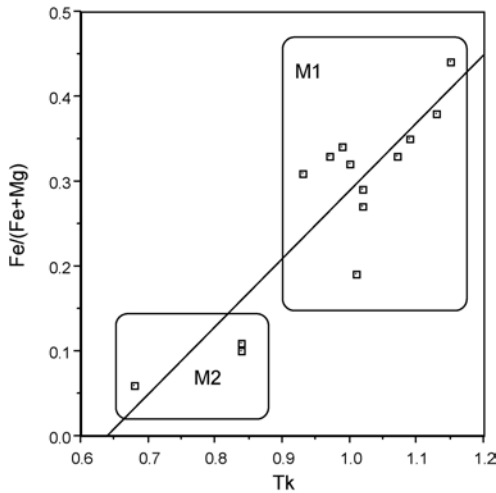


FIG. 6. Tschermak variation in chlorite, showing M1 chlorites distinct from M2.

from core to rim, reflecting a more complex history. M2 garnet zonation makes a comparison with the compositions of unzoned M1 garnets difficult. Although some overall compositional difference between M1 and M2 garnets cannot be ruled out, none was observed in this study.

Figure 7 plots the Ti and Fe contents of vein-filling M2 garnet, showing that the early garnet along the vein walls was richer in Ti than the late-stage garnet at the vein center. The total vein width is

about 3 mm. The inverse correlation between Ti and Fe documents progress along the TiFe_{-1} exchange vector in these garnets.

Tectonic Blocks Containing Benitoite

Blocks of mafic schist and greenstone that contain up to 2 wt% TiO_2 are exposed in the vicinity of the Gem Mine in the southern part of the New Idria District, at the Victor Claim in the Clear Creek Valley, and on the east summit of Santa Rita Peak. The blocks of mafic schist contain blue amphiboles (crossite, glaucophane) and are compositionally similar to Franciscan blueschists; they are interpreted as altered blueschists. Blocks of greenstone contain albite, chlorite, and non-jadeitic pyroxene; they are interpreted as Franciscan metagreywackes and metavolcanics. Mineralized M2 veins crosscut the M1 foliation of the mafic schists.

The Gem Mine is famous as the site of the 1906 discovery of benitoite ($\text{BaTiSi}_3\text{O}_9$), a trigonal blue mineral that is now the state gemstone of California (Louderback, 1907). Neptunite ($\text{Na}_2\text{KLi}(\text{Fe}, \text{Mg}, \text{Mn})_2\text{Ti}_2\text{O}_2\text{Si}_8\text{O}_{22}$, the second reported occurrence) and joaquinite ($\text{Ba}_2\text{NaCe}_2\text{Fe}(\text{Ti}, \text{Nb})_2\text{Si}_8\text{O}_{26}(\text{OH}, \text{F}) \cdot \text{H}_2\text{O}$, a new mineral) from the mine were described along with benitoite by Louderback (1909). Wise and Gill (1977) described the minerals of the Gem Mine and concurred with Coleman (1957) on their probable metasomatic origin.

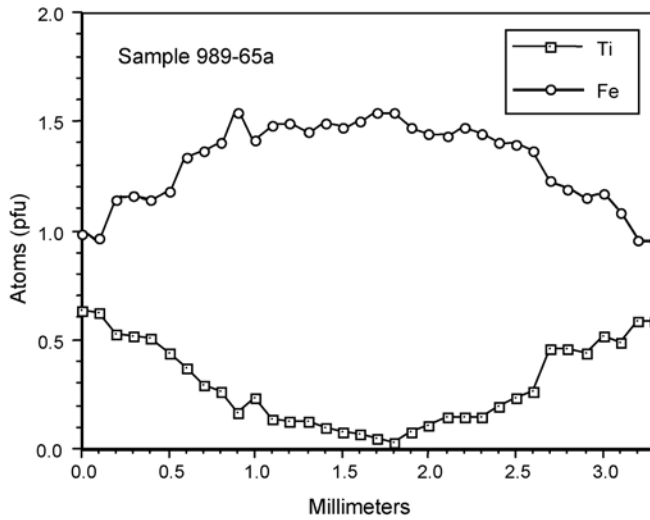


FIG. 7. Garnet vein traverse, showing Ti-rich garnet at walls, Fe-rich in center.

Geological description

The original geologic relations at the Gem Mine have been largely obscured by decades of mining. For this reason, the original field descriptions by Louderback are of paramount importance. Louderback (1909) reported that the brecciated and mineralized zone was 520 ft long and about 400 ft wide in its widest part. Within that zone, the benitoite-bearing rocks were only 64 ft wide and 400 ft long, with the abundance of benitoite decreasing to the west. Benitoite was only found at the surface for 230 ft of the 400 ft. Louderback also noted that the greenstone diminished in abundance to the west. The general nature of the mineralized zone was, from the center outward, drusy cavities containing benitoite, neptunite, and natrolite deposited concurrently; next white, relatively pure natrolite; then bluish natrolite containing abundant amphibole needles; and finally the altered wall rock, in which cavities filled with amphiboles in a felty texture suggested the leaching out of the feldspathic components of the wall rock. Outside the mineralized zone, albite rather than natrolite was the characteristic sodic mineral.

The main feature of the Gem Mine geology is an oval, NW-trending body of mafic schist and greenstone that is embedded in the surrounding serpentinite. This oval body contains two parts: greenstone to the north and mafic schist to the south, as shown in cross section in Figure 8. The mineralized zone lies in a brecciated zone along the contact between

the two parts. The excavated working area of the Gem Mine lies in the mafic schist of the footwall, where veins and fractures contain the benitoite and associated minerals. The greenstone of the hanging wall forms a 10 m high cliff on the north side of the mine.

As noted above, the greenstone and mafic schist at the Gem Mine are interpreted as Franciscan rock entrained within the serpentinite. A large block of Franciscan rock containing a contact between greenstone and mafic schist may have been caught up with the serpentinite during emplacement. Alternatively, a block of Franciscan greenstone and a block of Franciscan mafic schist may have been separately entrained, but moved to their present adjacent positions by tectonic movements during ascent of the diapir. Relative movement between the two blocks would help to explain the brecciated zone along the contact.

Formation of benitoite

One of the enduring mysteries of the New Idria District has been the origin of the rare mineral benitoite, $\text{BaTiSi}_3\text{O}_9$. Coleman (1957) hypothesized that benitoite formed by metasomatic alteration of Franciscan rocks at the Gem Mine as a result of fluids associated with syenite intrusions 1 km or more distant. This hypothesis was analogous to that for the formation of chlorite-diopside-garnet rocks from ordinary serpentinite, as described in the previous section.

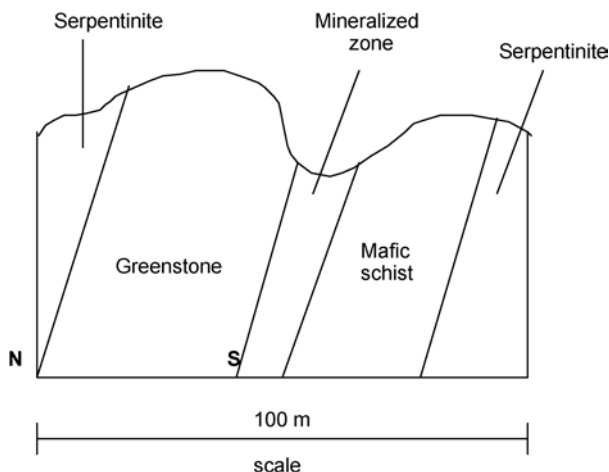


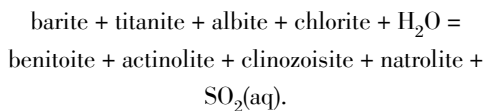
FIG. 8. N-S cross-section at Gem Mine.

An alternative hypothesis is proposed here, that benitoite at the Gem Mine formed by metamorphism and local metasomatism of the contact between the hanging wall greenstone and footwall mafic schist, in the presence of Na-rich, low-silica fluids. Mineralized drusy cavities and extensive development of vein systems certainly suggest hydrothermal mineralization at the Gem Mine, but the chemical components needed to produce benitoite and probably the other exotic minerals are available in the adjacent wall rocks. Therefore, there is no need to advect Ti, Al, and other components from a great distance. Ordinary brines associated with M2 metamorphism probably produced local mobility of all the required chemical components of benitoite and other minerals. The pockets of felty amphiboles that are found throughout the brecciated zone and altered wall rock suggest that sodic feldspar was leached out of this rock and redeposited in the veins as natrolite \pm thomsonite, with attendant loss of silica to the environment. Natrolite must have precipitated from a fluid with low silica activity.

Louderback (1909) commented on the absence of barite or any other apparent source of barium in his original report. Only with the advent of the electron microprobe did it become possible to recognize the relatively abundant, micron-sized grains of this mineral present. A mass balance calculation of the Ba budget for these rocks shows that the abundance of benitoite is consistent with the abundance of barite in the surrounding rocks. A whole-rock analysis of the greenstone (not shown) shows the amount of Ba to be 100–200 ppm, which compares with

150 ppm in the Catalina greenstone (Sorensen, 1986). Assuming that all of the Ba is incorporated in benitoite after the above reaction, the amount of benitoite should be approximately 0.01 wt% in the mineralized zone, or 1 part in 10,000. The actual amount of benitoite present is well under this amount. A similar mass balance of the Ti budget shows adequate titanium to account for abundance of the Ti-bearing minerals such as benitoite and neptunite.

By assuming the end member formulas for required minerals, we then have the following simplified component reaction:



A similar reaction may be written for the formation of neptunite. The actual whole-rock reactions that took place in the presence of hydrothermal fluids are likely to be different than these reactions, particularly in the specific aqueous species present. There is also no guarantee that the system was closed except for water. Extensive wall rock alteration in the mineralized zone suggests large volumes of fluid have passed through. However, all of the reactant phases in this equation are hanging wall phases; all of the product phases occur in the footwall. The above reaction is not unique: other similar reactions could be constructed, e.g., by substituting thomsonite for clinozoisite as a product. In all of these reactions, however, the Ba required to produce

benitoite can be found in barite, a rather common mineral in Franciscan rocks. Likewise, the required Ti can be found in titanite (sphene).

Constraints on M2 metamorphism

The temperature and pressure of formation of benitoite are not well constrained. Millage (1981) suggested that the upper stability limit of benitoite is about 600°C. If benitoite and natrolite formed concurrently, then the conditions must have been within the stability field of natrolite, which is limited to about 250°C at 1 bar (Peacor, 1973). Wise and Gill (1977) argued that the presence of djurleite as a companion phase placed a constraint on the temperature of formation of benitoite. The upper limit of stability of djurleite is about 95°C. However, as pointed out by Ulrich Petersen (pers. commun., 1992), the copper sulfide minerals readily recrystallize and invert to other forms with falling temperatures, unlike most silicates. This means that the crystals that are now djurleite could have originally formed as digenite or another high-temperature sulfide. Therefore, the implied constraint on the temperature of formation of benitoite at or below 100°C should not be considered a reliable limit. Fluid inclusion evidence from garnets in nearby tectonic blocks (Van Baalen, 1991) suggests that M2 metamorphism may have occurred at about 300°C; similar temperatures may have occurred at the Gem Mine at the same time. However, no fluid inclusions have yet been found in benitoite or neptunite.

In an effort to understand the timing of M2 metamorphism at New Idria, a reconnaissance study of the age of crystallization of the benitoite-neptunite crystals was undertaken. These minerals are assumed to have crystallized simultaneously. Using Rb/Sr dating techniques on samples from this study, Marvin Lanphere of the USGS constructed a two-point isochron of a mineral pair, yielding an age of 11.6 Ma (M. Lanphere, 1993, pers. commun.). With a two-point isochron, no reliable error estimate is possible, so the benitoite-neptunite age should be considered preliminary. However, this preliminary age agrees well with the 12.4 ± 0.8 Ma age of the New Idria syenite, discussed below.

Kaersutite in the New Idria Syenite

The New Idria serpentinite is intruded by several small syenite stocks of Miocene age, here collectively named the New Idria syenite. Stream boulders of this syenite were noted by Anderson and Pack

(1915), and their source was traced to the headwaters of White Creek. The original description of the syenite was of this stock in the White Creek valley, hence the body was named the White Creek stock (Robertson, 1940). Since there are additional outcrops not in the White Creek drainage, it is here proposed that that name be abandoned in favor of the name *New Idria syenite*. Significantly, these stocks form a linear array, subparallel to the New Idria thrust fault, on the map; their emplacement may be partially fault controlled. It is significant that the White Creek stock intrudes the boundary between the serpentinite and the adjacent Panoche Shale. Coleman (1957) also described the New Idria syenite.

The New Idria syenite is a soda syenite, composed almost entirely of sodic feldspar and kaersutite (formerly barkevikite) amphibole containing up to 4 wt% TiO₂ (Robertson, 1940, Coleman, 1957, Van Baalen, 1995). The rock is very coarse grained in places, with amphibole crystals up to 30 cm long. Elsewhere the rock is fine-grained with only millimeter-sized amphibole crystals. Enclaves of coarse-grained rock are found in fine-grained rock, and vice-versa. In places, olivine and pyroxene are found; Coleman (1957) described that facies as camptonite. Accessory minerals include biotite, apatite, titanite (sphene), magnetite, ilmenite, pyral-spite garnet, and zircon. Deuteric and later alteration products include zoisite, prehnite, zeolites, white mica, chlorite, and calcite. Table 3 presents some whole-rock analyses of the syenite. A CIPW norm calculation shows that in all cases the rock is nepheline normative. Trace-element concentrations are consistent with expected values for igneous intrusives at convergent margins.

The abundance of tectonic blocks of several types in the New Idria serpentinite raises the question of whether the syenite outcrops are intrusive or not. At a large syenite outcrop informally named The Nose, located west of Perovskite Knob (also an informal name), the serpentinite surrounding the syenite has been contact metamorphosed: both antigorite and prograde olivine in "starburst" texture are present. The aureole grades outward to the "normal" serpentinite consisting of chrysotile and lizardite. This observation shows that the contact is indeed intrusive; the margins of tectonic blocks do not have contact aureoles.

The age of the New Idria syenite has been determined by Marvin Lanphere of the U.S. Geological Survey using the ⁴⁰Ar/³⁹Ar total fusion technique on

TABLE 3. Whole-Rock Analyses of the New Idria Syenite¹

	1	2	3	4	5
Sample:	989-8	989-39	A&A, 1910	Coleman	Coleman
Location:	White Crk.	The Nose	White Crk.	White Crk.	White Crk.
	wt%				
SiO ₂	51.46	47.85	60.00	48.96	51.42
TiO ₂	1.53	2.44	0.42	2.01	2.20
Al ₂ O ₃	18.56	15.38	16.88	15.72	16.58
Fe ₂ O ₃	8.42	10.61	1.83	0.36	1.48
FeO	n.d.	n.d.	3.02	9.65	7.62
MnO	0.14	0.16	0.12	0.17	0.23
MgO	2.92	9.45	1.40	6.52	3.68
CaO	6.11	9.03	3.16	7.76	6.77
Na ₂ O	9.89	3.27	9.31	5.22	6.04
K ₂ O	0.42	1.47	0.94	0.75	1.06
P ₂ O ₅	0.50	0.14	0.14	n.d.	0.49
H ₂ O+	n.d.	n.d.	1.53	2.29	1.41
H ₂ O-	n.d.	n.d.	0.43	0.42	0.34
Total wt%	99.94	99.79	99.18	99.83	99.32
	Trace, ppm				
Nb	39.87	25.30	n.d.	n.d.	n.d.
Zr	260.83	123.50	300	n.d.	trace
Sr	836.15	410.35	200	n.d.	800
Zn	90.41	88.53	n.d.	n.d.	n.d.
Ni	9.28	24.03	n.d.	n.d.	n.d.
Cr	0.00	0.60	0	n.d.	n.d.
V	64.68	314.23	n.d.	n.d.	n.d.
Ce	58.98	30.08	n.d.	n.d.	n.d.
Ba	562.37	441.39	600	500	trace
La	23.00	11.61	n.d.	n.d.	n.d.

¹Analyses 1 and 2 anhydrous; all Fe expressed as Fe₂O₃. In analysis 3, CO₂ = 0.59 wt %; in analysis 5, CO₂ = 0.45 wt%. Sources: Analyses 1 and 2 from this study; analysis 3 from Arnold and Anderson, 1910; analyses 4 and 5 from Coleman, 1957.

an amphibole mineral separate. This unpublished age is 12.4 ± 0.8 Ma (M. Lanphere, pers. commun., 1993). Such an age is consistent with the intrusive nature of the contact, assuming that the serpentinite as a whole has a Late Jurassic age.

The occurrence of the prehnite-zoisite pair in deuterically altered syenite provides a pressure estimate at the time of alteration (Van Baalen, 1997). The stability fields of prehnite and zoisite overlap only between 1 and 3 kb (Frey et al., 1991).

Therefore the outcrop presently at the surface lay at a depth of 3.5 to 10 km at the time of alteration, assuming an average density of 2.8 g/cm^3 for the rock formerly overlying the outcrop (the average density of several syenite samples was determined to be 2.8 g/cm^3 using a Jolly balance; the density of serpentinite was found to be 2.4 to 2.8 g/cm^3).

If the 12.4 Ma age of the syenite measured by Lanphere is close to the time of deuteric alteration, which is reasonable for a small intrusion, then the unroofing rate since the late Miocene may be calculated. The average unroofing rate is calculated to be 0.3 to 0.8 mm/year for the past 12 million years. This calculation agrees well with the estimated rate of 1 mm/year for the Coast Ranges, obtained using other data (e.g., Montgomery, 1993; Ring and Brandon, 1994).

Discussion

The geological and mineralogical descriptions in this paper inevitably lead to the question of the origin of the "peculiar mineral suites," but also of the evolution of the New Idria serpentinite itself. The traditional model, advocated by Coleman and others, has been a metasomatic origin of the mineral suites, suggesting that the missing chemical components from the barren serpentinite rock were advected into (or out of) the mineralized areas, via fluids from an unknown source, perhaps the New Idria syenite. It is of course possible in a chemical sense to adjust the compositions of rocks through fluid interaction, but the number of components involved in this case becomes rather large: not simply Ti and Al, but also Nb, Zr, V, and Ce must be mobile in order to satisfy the large differences in bulk composition shown in Tables 2 and 3. Furthermore, their mobility must be in the exact proportions required to create the geochemical signature of a pyroxenite. Ti, Al, and Zr are considered among the least mobile components in crustal rocks, yet to accept the metasomatism hypothesis they must be very mobile, on a kilometer scale. Van Baalen (1993) showed that Ti mobility is extremely low in crustal rocks due to low solubility in aqueous fluids, and that kilometer-scale transport is therefore unlikely, and not found elsewhere.

The simpler hypothesis adopted in this paper shows that essentially isochemical M1 metamorphism of pre-existing pyroxenite veins brings all of the required components for exotic mineral suites to the site of crystallization, without the need for long-

distance transport of low-solubility aqueous species. Furthermore, the worldwide occurrence of chlorite-diopside-garnet rocks, in geological settings different from New Idria, argues for a more general process. Prior to the advent of the electron microprobe, M1 metamorphism went generally unrecognized.

The discovery of M1 and M2 metamorphism at New Idria, with M2 mineral assemblages forming by recrystallization of M1 minerals, including local mobility of Ti and probably Al, perhaps on a centimeter scale, is consistent with the known behavior of Ti in crustal rocks undergoing metamorphism. One may then conclude that the metasomatism hypothesis actually does have a place, but only as a local process around veins, fractures, and contacts.

The chlorite-diopside-garnet rocks at New Idria bear a superficial resemblance to a coarse-grained rock type known as rodingite. Rodingites are metasomatic calc-silicate rock bodies associated with certain serpentinites, such as the one at Belvidere Mountain, Vermont (Van Baalen et al., 1999). The characteristic minerals of rodingites are hydrogrossular, idocrase, diopside, and chlorite.

Rodingites were first described at exposures along the Roding River in the Dun Mountain ultramafic complex, New Zealand (Bell et al., 1911). Mafic dikes were intruded into the ultramafite prior to serpentinization, so that when serpentinization occurred, chemical potential gradients and incompatible mineral assemblages provided the driving force for metasomatic exchange of Ca (out of the ultramafite) and Si (into the ultramafite), producing the characteristic suite of calc-silicate minerals. The term rodingite has since been used more generally (e.g., Coleman, 1977) to describe metasomatic rinds that form during serpentinization, by reactions either with intrusive dikes or siliceous country rock.

Figure 9 is an ACF diagram representative of bulk compositions of chlorite-diopside-garnet rock from New Idria compared with the rodingite field from Coleman (1977). Although the New Idria rocks plot near a tie line between chlorite and diopside, the rodingite field sits astride the grossular-diopside join.

The chlorite-diopside-garnet rocks of New Idria did not form as rodingites because the bulk composition of the metapyroxenite blocks and the enclosing serpentinite were *both* silica-poor. Chlorite, diopside, and titaniferous andradite garnets are all stable phases in association with serpentine minerals. However, Ca mobility may be locally important (e.g., the 34–50 locality described above,

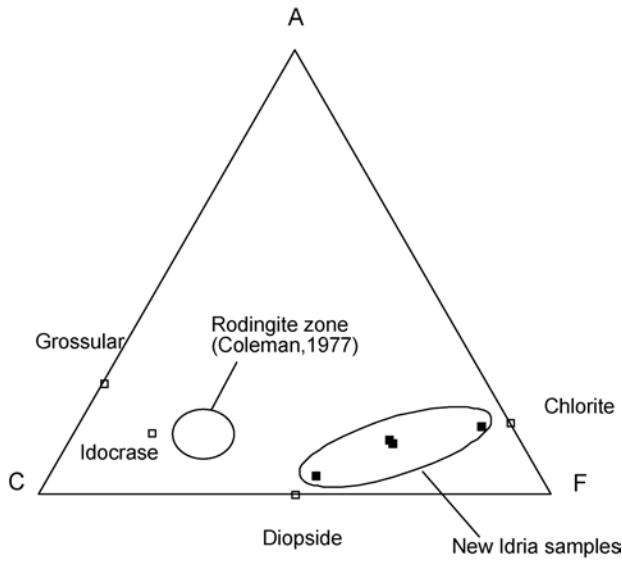


FIG. 9. ACF projection showing comparison between bulk compositions of chlorite-diopside-garnet rocks from New Idria with the rodingite field of Coleman (1977).

where recrystallized diopside interfingers with serpentinite).

There is, however, a possible rodingite body at New Idria, at locality 990-1 near the Gem Mine, where the rock contains both vesuvianite and grossular garnet, the characteristic rodingite minerals. The bulk composition of this outcrop is unknown, but could lie in the rodingite field: it is not a chlorite-diopside-Ti-garnet rock.

What, then, is the relationship between M2 metamorphism and the New Idria syenite? The crystallization ages of benitoite and neptunite are very close to the cooling age of the amphiboles in the syenite, 11.6 and 12.4 Ma, respectively. The most probable explanation of this close association is that the M2 metamorphism and the igneous intrusion are both effects of a common cause, which is a heat pulse generated by the passage of a slab window underneath New Idria at approximately 12 Ma. This slab window, first proposed by Dickinson and Snyder (1979), is now generally accepted (e.g., Atwater and Stock, 1998). The window results from the abrupt change in plate boundary geometry at the Mendocino Triple Junction as it migrates northwest at the speed of transform movement along the San Andreas fault. According to Stock and Hodges (1989), the location of the Mendocino Triple Junction was underneath New Idria at 12.9 Ma, as shown in Figure 10. The plate configurations were recently

re-evaluated by Atwater and Stock (1998), with a similar position for the Mendocino Triple Junction. Dickinson (1997) recently revisited the northerly progression of magmatic activity along the trace of the triple junction for the period since the Miocene, and correlated this with the geometry of plate boundaries during this interval. The New Idria syenite, intruded at 12.4 Ma, lies along the spatial and temporal trend shown by Dickinson.

Another effect of the passage of the Mendocino Triple Junction was the tectonic uplift of the serpentinite massif, which reached the surface in the late Miocene, as documented by the sudden appearance of clastic serpentinite in the Great Valley Sequence (Big Blue Formation). This formation, including serpentinite sandstone and serpentinite boulder conglomerate, was first described by Anderson and Pack (1915).

The tectonic uplift is most likely the cause of the extensive fractures and tension gashes in the chlorite-diopside-garnet rock, which are nearly all oriented perpendicular to the axis of the regional anticlinal structure. These fractures are filled by the M2 minerals that are clearly postkinematic.

A recent study by Vermeesch et al. (2003) of the uplift history of Great Valley Sequence rocks on the Joaquin Ridge near New Idria used apatite fission track methods to show abrupt uplift at 13 Ma; the authors suggest that a heat pulse caused by passage

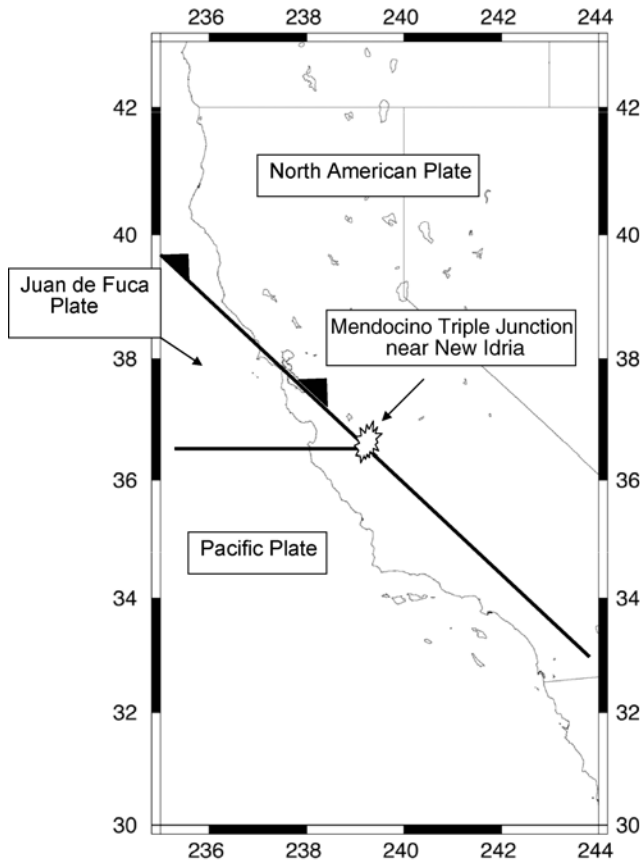


FIG. 10. Configuration of plates, relative to a fixed North America, in late Miocene, ca. Anomaly 5AA, 12.9 Ma. Note position of Mendocino Triple Junction near New Idria. Plate boundaries schematic only.

of the Mendocino Triple Junction resulted in annealing of apatite fission tracks at that time. The Vermeesch et al. study and the present study are mutually supporting, using independent methods, giving increased confidence in their results.

The question of the formation of benitoite has been one of the enduring mysteries of the New Idria District, owing to the lack of any apparent source of Ba and Ti. Again, the advent of the electron microprobe made it possible to identify barite and titanite as hanging wall phases at the Gem Mine, allowing M2 metamorphism of an assumed contact between Franciscan greenstone and mafic schist to become the mineralized zone of the Gem Mine. Local mobility in low-temperature metamorphic fluids of all required components is sufficient to explain the mineral suites there. There are so few benitoite localities worldwide that it is not possible to gener-

alize the mineral paragenesis shown at the Gem Mine. However, searches for benitoite elsewhere should certainly look for tectonic blocks in serpentinites!

Finally, the spatial and temporal association of the New Idria syenite with the "peculiar mineral suites" described in this paper has previously been interpreted as a cause-and-effect relationship involving metasomatism by hydrothermal transport of Ti. It is noteworthy, but not conclusive, that of the several separate stocks of the syenite, only one (the Nose) is within even 1 km of outcrops containing Ti-rich mineral suites, whereas no such mineral suites are associated with the other exposed stocks of the same syenite. Given the rock and mineral compositional data presented in this paper, and the demonstrated plausibility of isochemical metamorphism of pyroxenite veins or tectonic blocks of

known lithologies, the simplest interpretation is that the tectonic uplift, the Ti-rich mineral suites, and the syenite itself are multiple effects of a single underlying cause. That cause is the thermal pulse caused by Mendocino Triple Junction passage. As shown in Van Baalen (1993), the solubilities of titanium, zirconium, and certain other chemical components in aqueous fluids are simply too low to sustain the metasomatism hypothesis.

Conclusions

1. Ti-garnets and associated titaniferous accessory minerals, in a rock consisting mainly of chlorite and diopside, are the products of metamorphism of pyroxenite veins. These veins were tectonically dismembered during diapiric emplacement of the New Idria serpentinite.

2. Two generations of metamorphism, M1 and M2, have affected the New Idria serpentinite and its included tectonic blocks and dismembered pyroxenite veins. The metamorphic grade of M1 was higher than that of M2, based on study of chlorite compositions. In the ultramafic system, chlorite is less magnesian than coexisting diopside, an exception to the general rule that chlorite is more magnesian than coexisting mafic minerals. During M2 metamorphism, fluids in veins and fractures evolved from relatively Ti-rich to Ti-poor, as documented in garnet zonations.

3. Benitoite and other Ti-bearing minerals have formed by M2 metamorphism of the contact between tectonic blocks of Franciscan greenstone and blueschist. The age of crystallization of benitoite and neptunite is ~11.6 Ma.

4. The passage of the Mendocino Triple Junction underneath New Idria at about 12 Ma resulted in M2 metamorphism, intrusion of the New Idria syenite at 12.4 ± 0.6 Ma, and tectonic uplift of the serpentinite, which breached the surface in the late Miocene as documented by the nonmarine BigBlue Formation of the Great Valley Sequence.

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