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Emerald Mineralization during Regional Metamorphism: The Habachtal (Austria) and Leydsdorp (Transvaal, South Africa) Deposits

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Abstract

Classic schist-hosted emerald deposits are generally attributed to interaction between invading pegmatitic magma or vapor phases with preexisting metasediments, metavolcanics, and/or ultrabasic rocks. Detailed studies of the Habachtal (Austria) and Leydsdorp (South Africa) emerald deposits indicate that the emerald formation was not due to a single-stage contact metamorphism at the border zone of magmatic intrusive bodies. On the contrary, the emeralds are found in metasomatic zones (blackwall zones) associated with ultramafic bodies formed by syn- to post-tectonic reactions during low-grade regional metamorphism. These reactions occur at the contact of preexisting beryl- and phenakite-bearing pegmatites and albitite pegmatoids with biotite-talc and actinolite schists (Leydsdorp), or at the contact of beryllium-rich garnet-mica schists and biotite-plagioclase-gneisses with serpentinites and talc-schists (Habachtal).

Critical evidence for the regional metamorphic origin of emeralds includes the following points:

1. Augen textures occur in the metagranitic and metapegmatitic country rocks.
2. Polystage growth with simultaneous deformation is indicated by sigmoidal inclusion trails in the strongly zoned emeralds.
3. No evidence of contact metamorphism is found in the emerald ores, nor in the immediate country rocks.
4. Both the rare earth element fractionation pattern and the Yb/Ca vs. Tb/Ca ratios of the fluorites which coexist with the emeralds indicate the existence of an abundant nonpegmatitic fluid phase during the formation of the emeralds.

The following reactions for the formation of emerald from preexisting phenakite are suggested by textural evidence:

$$3 \text{ phenakite} + 2 \text{ biotite} = 2\text{MgFe-beryl} + 2K^{+} + 4Mg^{+2}$$

$$3 \text{ phenakite} + 3 \text{ albite} + 2Mg^{+2} = 2\text{MgNa-beryl} + Na^{+} + Al^{+3}.$$

The high magnesium and chromium contents—mainly in the rims of the emeralds—reflect their formation in an increasingly Mg- and Cr-rich environment. Microthermometric studies of fluid inclusions in the emeralds of both deposits suggest that they formed from an H_2O-CO_2 mixed fluid phase, with a salinity of 1 to 9 equiv wt percent NaCl and homogenization temperatures around 300°C.

Mass balance and geochemical analyses suggest that the transformation of serpentinites and Be-rich country rocks into blackwall zones liberates significant amounts of Be and enables the formation of emerald in the presence of Cr. Due to unfavorable growth conditions during regional metamorphism, the extent of emerald mineralization never exceeds that of pegmatites.

Introduction

THE well-known schist-hosted emerald deposits of the Habachtal (Austria) and of Leydsdorp (Transvaal, South Africa) have been generally attributed to an interaction between invading granitic to pegmatitic magma and/or its derived fluid phases with preexisting mafic to ultramafic rocks (Fersman, 1929; Leitmeier, 1937; Beus, 1960; Sinkankas, 1981; Robb and Robb, 1986). This commonly accepted genetic concept was introduced in the classic paper of Fersman (1929, p.96) on the famous Takowaja emerald deposits in the Ural Mountains (USSR): "It is now established that the deposit can be explained by a mutual interaction of a granitic fluid phase on the one hand and a melanocratic, basic rock complex on the other." According to the genetic classification suggested by Sinkankas (1981), this type of deposit is therefore called exometamorphic since the major compounds necessary for the formation of beryl appear to be derived from granitic pegmatites and then transferred to the adjoining basic rocks.

Petrologic and geochemical studies of the emerald deposits of Habachtal (Grundmann and Morteani, 1982) and Leydsdorp (Grundmann and Morteani, 1987) have shown that the exometamorphic concept is not adequate to explain their genesis in either case. On the contrary, the microtextures of the emeralds and their host rocks show unambiguously that the emeralds formed during regional metamorphism in lithologies in which chromium-rich ultrabasic rocks and beryllium-rich metapelites and metavolcanics had been juxtaposed by pre- or synmetamorphic tectonism, or preexisting beryl- and phenakite-bearing pegmatites or pegmatoids intruded chromium-rich ultramafic rocks or had been juxtaposed by postmagmatic tectonism. Recognition of this interplay of tectonism, rock chemistry, and regional metamorphism is very important since only the correct genetic concept (and/or good luck) leads to successful exploration for emerald deposits and avoids a disregard of highly promising terranes.

Regional Geology of the Emerald Deposits

Habachtal

The emerald deposit of the Habachtal is situated in the Tauern window (eastern Alps, Austria) near the tectonic contact between ortho-augengneisses of the Zentralgneiss (central gneisses) and a series of amphibolites, mica schists, and black phyllites with interlayered serpentinites which is called the Habach Formation (Höck et al., 1982; Fig. 1).

Both the central gneisses and the Habach Formation belong to the lowest tectonic unit of the eastern Alps (Penninicum), exposed in the Tauern window. The Habach Formation is part of the Untere Schieferhülle (lower schist cover) which is interpreted as an alpine nappe above the central gneisses. The local geology is described in detail by Grundmann and Morteani (1982) and Grundmann (1983).

The emerald mineralization itself occurs in a sequence of metasomatic blackwall zones developed between a series of metamorphosed pelites, mafic volcanics, and ultramafics, including serpentinites. The emeralds occur predominantly in biotite, chlorite, actinolite, and talc schists. Intense Be, W, Cu, Mo, Bi mineralization, which produces scheelite, beryl (var. aquamarine and emerald), fluorite, phenakite, chrysoberyl, bavenite, chalcopyrite, and Cu-Bi sulfosalts has also been observed within the Habach Formation at the nearby tungsten deposit in Felbertal (Höll, 1975). Figure 2 gives a detailed cross section of the lithologic sequence in the emerald deposit. The entire sequence is interpreted as a tectonic mélange, produced mainly by alpine tectonism.

According to Hoernes and Friedrichsen (1974), Grundmann (1980, 1989), and Grundmann and Morteani (1982), both a pre-Alpine and an Alpine meta-

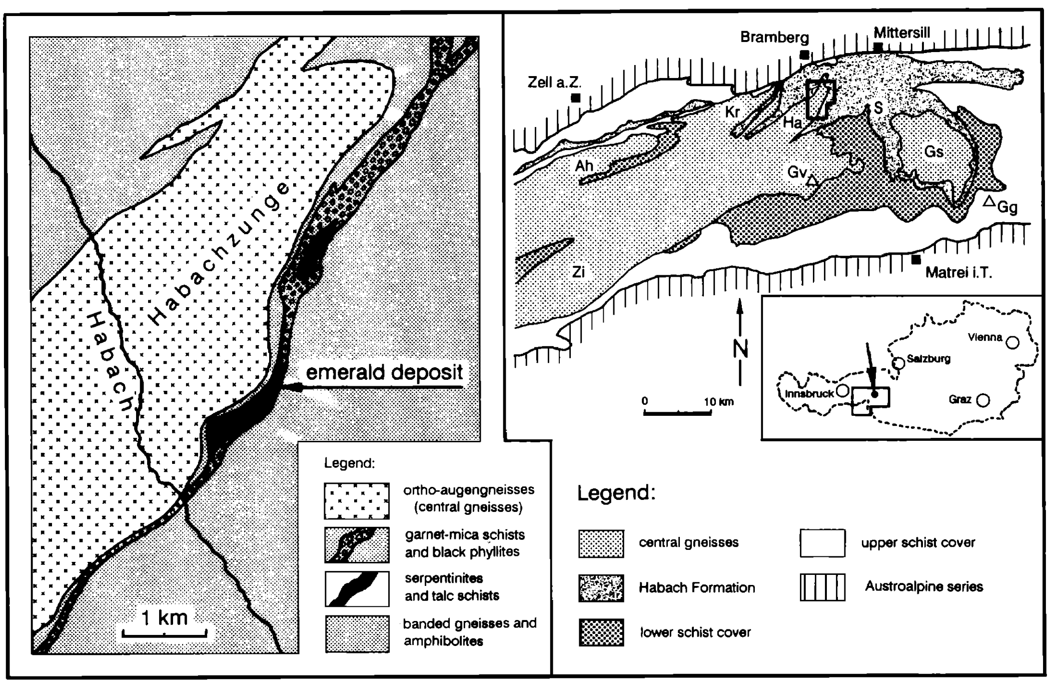

FIG. 1. Habachtal. Geologic sketch map of the area surrounding the Habachtal deposit. Inset shows the regional setting in relation to the Tauern window. Ah = Ahorn core, Gg = Grossglockner, Gs = Granatspitz core, Gv = Grossvenediger, Ha = Habachzunge, Kr = Krimmler Gneiss core, S = scheelite deposit in Felbertal.

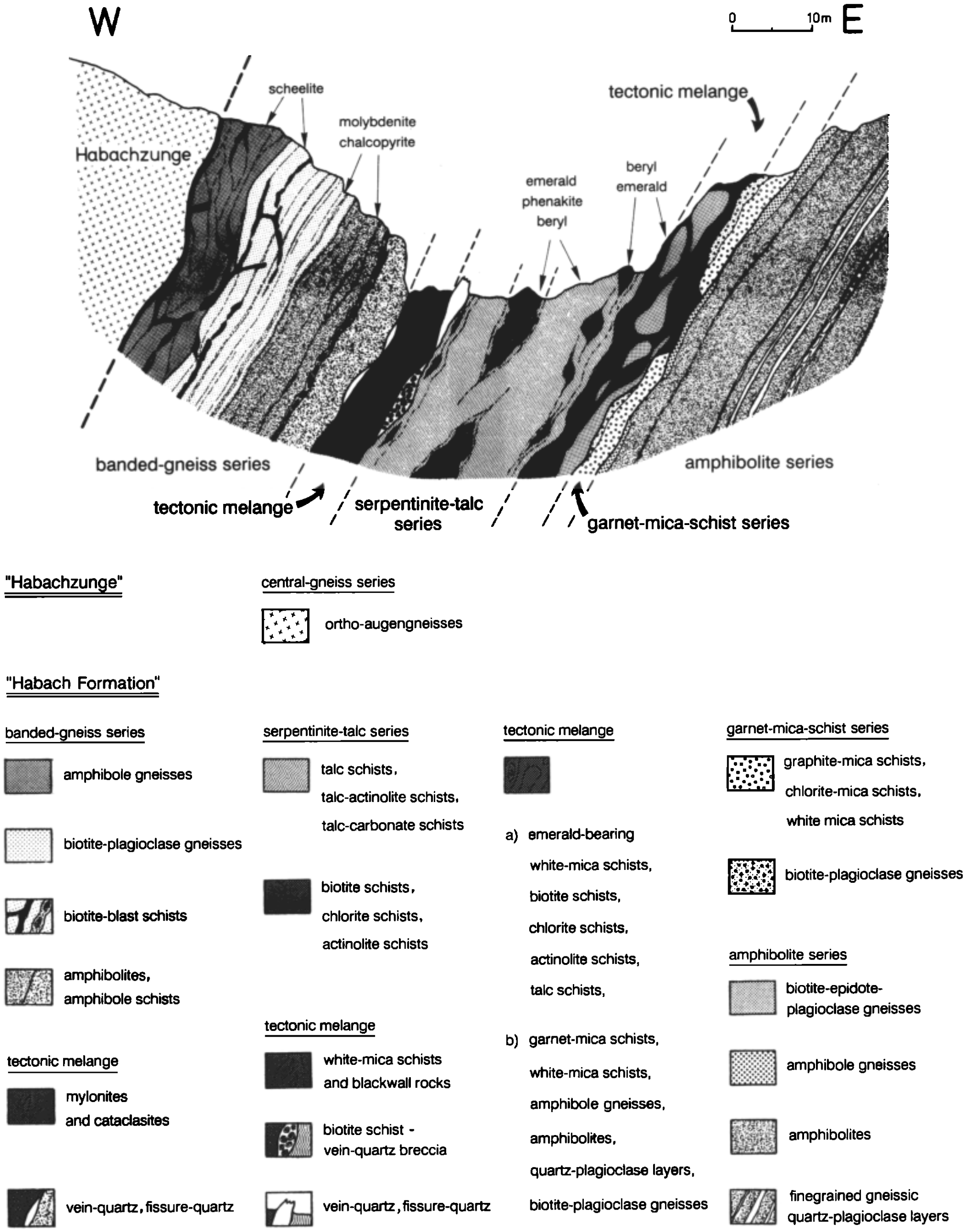

FIG. 2. Habachtal. Geologic cross section of the Habachtal deposit (Grundmann, 1983).

morphic event are documented in the mineral parageneses of the emerald-bearing series. The PT conditions of the pre-Alpine event were <3 kbars and <450°C, those of the Alpine event were 4.5 to 6 kbars and 500° to 550°C.

Leydsdorp

The Leydsdorp emerald district is situated in the Murchison greenstone belt (Republic of South Africa), which is bound on the north by the Rooiwater Igneous Complex and in the south by Archean gneisses (Fig. 3). The Gravelotte emerald mines (Cobra Pit and B.V.B. Ranch) are situated at the contact of Archean tonalitic gneisses with talc-chlorite, actinolite, and biotite schists (Le Grange, 1929; Van Eeden et al., 1939; Robb and Robb, 1986).

Several gneissic metatonalitic to metagranodioritic bodies are associated with an albitite-pegmatoid, which shows a tectonometamorphic overprint within a sequence of biotite, actinolite, chlorite, and talc-

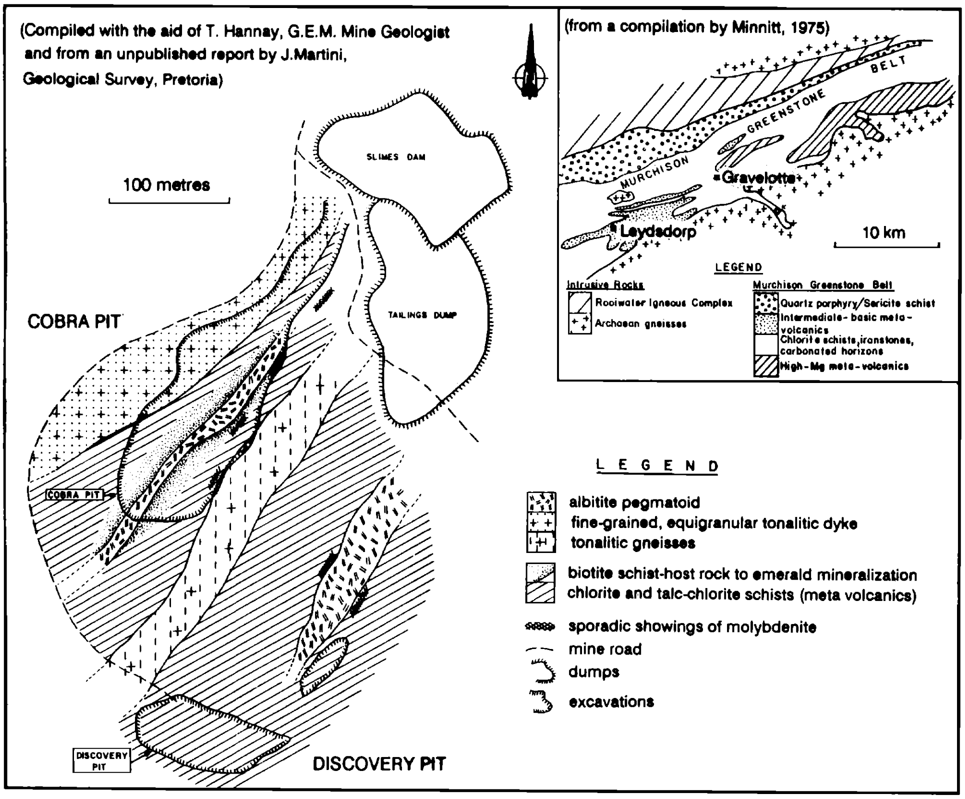

FIG. 3. Leydsdorp-Gravelotte emerald mine. Geologic map of the area surrounding the Cobra pit, Leydsdorp district, South Africa, from Robb and Robb (1986). Inset shows the regional setting in relation to the Murchison greenstone belt.

chlorite schists (Fig. 3). It must be emphasized that the magmatic bodies including the pegmatoid rocks show clear gneissic textures, indicating that the whole series has been affected by deformation and greenschist facies metamorphism. Figure 4 shows deformed pegmatoid bodies in the Cobra pit.

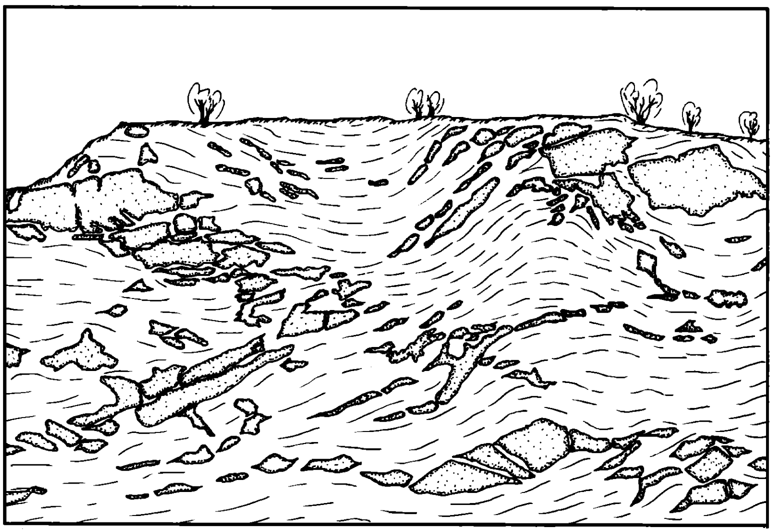

FIG. 4. Sketch from a photograph of the northern face in the Cobra open pit, Leydsdorp district. The pegmatoid bodies (dotted) are clearly deformed.

Macroscopic Features of the Emerald Mineralization

Habachtal

The most intense emerald mineralization of the Habachtal is associated with muscovite-plagioclase-chlorite boudins surrounded by biotite schists. These emerald-bearing boudins formed within a tectonic mélange composed mainly of garnet-mica schists, muscovite schists, biotite-plagioclase gneisses, and talc schists (Fig. 2).

The predominant Be mineral is beryl ($Be_3Al_2Si_6O_{18}$). Other Be minerals occur locally as accessories including phenakite (Be_2SiO_4), chrysoberyl ($BeAl_2O_4$), and milarite ($K_2Ca_4Al_2Be_4Si_{24}O_{60} \cdot H_2O$) (Niedermayr and Kontrus, 1973). Phenakite and chrysoberyl (but no coexisting quartz) are observed only within the serpentinite-talc series (Fig. 2), in massive lenses of chlorite and biotite-phlogopite (Koller et al., 1983) within talc schists.

Figure 5 shows an example of the emerald mineralization associated with boudins of muscovite-albite-schist. Growth of emeralds was syn- to post-tectonic with respect to the formation of the boudins (Grundmann, 1983). The muscovite-albite rock was formed during the metasomatic reactions which produced emeralds from a garnet-mica schist protolith. The emerald-bearing biotite schist is the product of a metasomatic reaction between the garnet-bearing mica schist protolith and the adjoining ultrabasic protolith, e.g., serpentinite and talc schist. This type of metasomatic reaction is characteristic on a larger scale of a typical blackwall zoning. The most common pattern of zoning, from the serpentinites into the tectonic mélange, is talc, talc-actinolite, chlorite, biotite, and muscovite-albite. For a detailed discussion of blackwall zoning and associated mineral reactions see Phillips and Hess (1936), Curtis and Brown (1969), and Brady (1977).

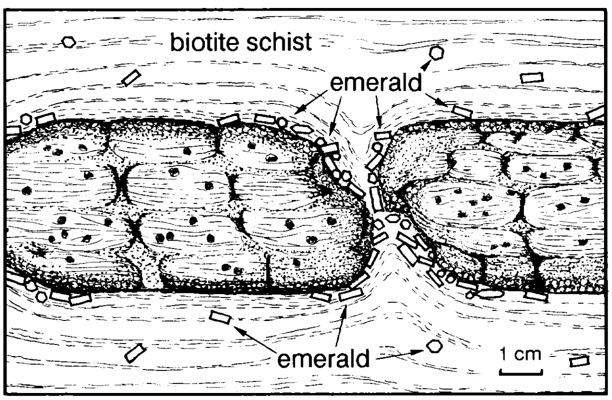

FIG. 5. Habachtal. Schematic section through muscovite-plagioclase-chlorite boudins surrounded by biotite schists. Emerald porphyroblasts grew posttectonically mainly at the contact of the boudins with the biotite schists.

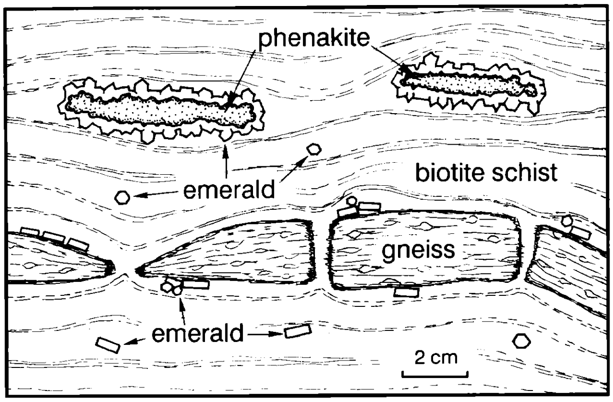

FIG. 6. Leydsdorp. Schematic section through fractured muscovite-rich quartzofeldspathic lenses surrounded by biotite schists. Emerald porphyroblasts occur prevalently on the surface of the boudins and only subordinately in the biotite schists, and at the contact between preexisting phenakite aggregates and biotite schists.

Leydsdorp

The emerald distribution in the Leydsdorp district is very similar to that of the Habachtal. The emeralds are often observed at the margins of or within strongly deformed and often boudinaged rock layers, which Robb and Robb (1986) have classified as albitite pegmatoid rocks. Figure 4 shows the face of the Cobra pit with the intensely deformed pegmatoid bodies. These are embedded in a tectonic mélange of talc, biotite, chlorite, and actinolite schists. The question exists as to whether the metamorphosed albitite pegmatoid rock and pegmatitic rocks were introduced to their present positions tectonically or by magmatic intrusion.

Unlike Habachtal, phenakite is a rather common mineral in the Leydsdorp deposit. It is observed both within the albitite pegmatoid and in pegmatitic schlierenlike bodies. Both types are foliated and are therefore of pretectonic origin.

Figure 6 shows examples of two types of emerald mineralization which occur at Leydsdorp: euhedral emerald porphyroblasts at the margins of thin layers of a strongly foliated and boudinaged augengneiss and in the surrounding biotite schist, and emerald formation on preexisting phenakite crystals in biotite schist.

Similar to the Austrian deposits at Habachtal and Felbertal, the Be mineralization at Leydsdorp is associated with molybdenite, chalcopyrite, scheelite, and bismuth minerals, as reported from the Cobra pit and B.V.B. Ranch mine. Hand specimens from the Cobra pit contain aggregates of biotite-fluorite partly intergrown with emerald. Mining for gold in sulfide deposits in the emerald-bearing series began recently.

Microstructural features

In the following paragraphs, the microstructures of emeralds and their country rocks from the Habachtal and the Leydsdorp district are examined to determine whether the emeralds formed in a pegmatitic environment or due to regional metamorphism. In the case of magmatic crystallization, the emerald and cogenetic minerals should show features indicating static growth, with unoriented inclusions of (possibly) magmatic minerals and a matrix with unoriented hornfelsic contact metamorphic texture.

In the case of regional metamorphic origin, the growth of minerals is typically syn- to post-tectonic within a foliated and/or lineated matrix. The emerald and cogenetic phases should in this case show deformation and oriented inclusions.

Habachtal

Figure 7a and b are sketches made from photomicrographs of emerald set in a biotite matrix. The core of the emerald crystal in Figure 7a shows randomly oriented inclusions of biotite, muscovite, chlorite, quartz, apatite, sphene, and epidote. The core is surrounded by a zone showing a sigmoidally oriented inclusion trail indicating a syntectonic growth of the emerald. The outer rim has few inclusions. In Figure 7b the core contains fine-grained actinolite in tightly folded layers. The outer rim is inclusion poor, but sigmoidal orientation of the trails is still clear.

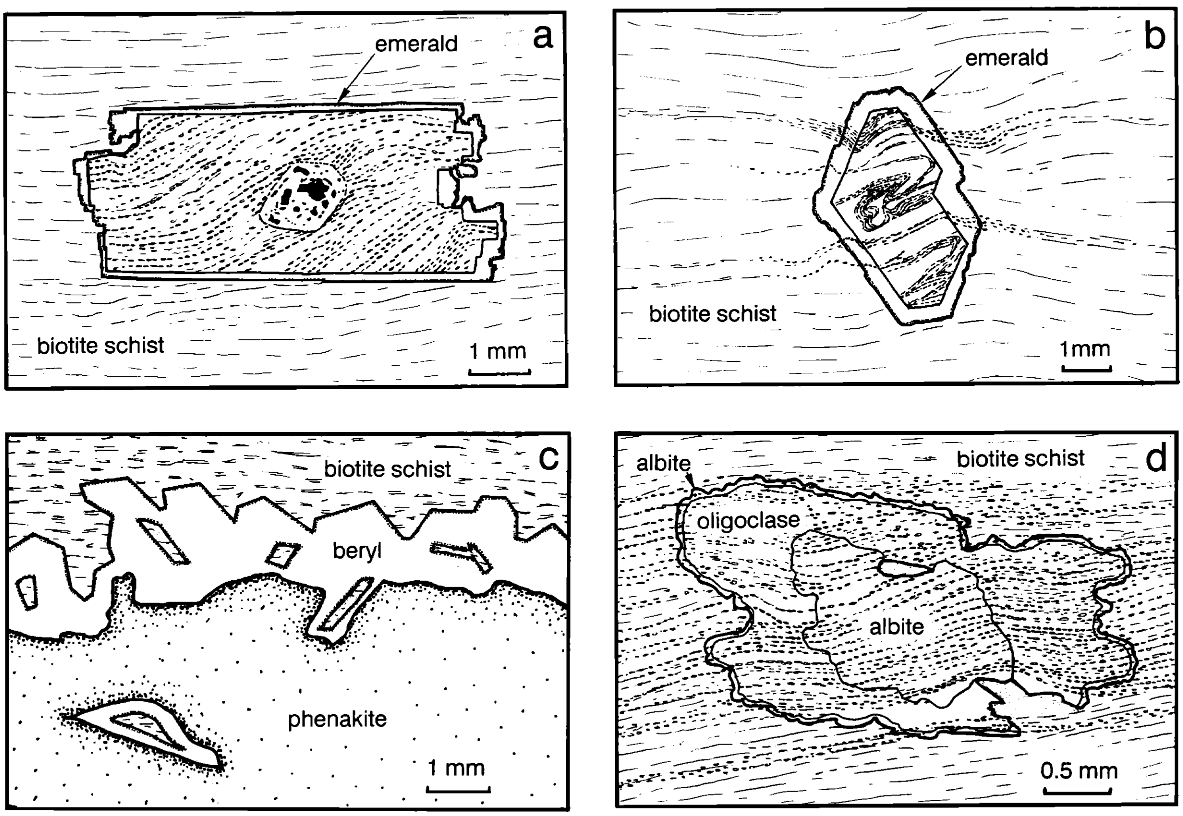

FIG. 7. a. Habachtal. Sketch of a photomicrograph of a zoned emerald porphyroblast in biotite schist. The inner core shows unoriented inclusions of chalcopyrite, pyrrhotite, apatite, and biotite. The inner rim shows sigmoidal orientation of the inclusion trails, indicating syntectonic growth of the emerald. The thin outer rim is very inclusion poor. b. Habachtal. Sketch of a photomicrograph of a zoned emerald porphyroblast in biotite schist. The very fine grained actinolite inclusion trails in the inner zone are tightly folded. The sigmoidal structure in the inclusion-poor outer rim demonstrates syn- to posttectonic growth. c. Habachtal. Sketch of a photomicrograph of a phenakite grain with a reaction rim of beryl at its contact with the biotite schist host. A biotite inclusion in the phenakite also shows a reaction rim. Beryl is clearly post-tectonic with respect to the biotite foliation. d. Habachtal. Sketch of a photomicrograph of a zoned plagioclase porphyroblast (from core to rim: albite, oligoclase, albite) showing clear evidence of syntectonic growth.

The schistosity of the biotite matrix is subhorizontal and contains the c-axes of the emeralds in Figure 7a and b.

A second but very rare type of beryl (rarely emerald) is found as a reaction product at the contact between phenakite megablasts and biotite in biotite-chlorite lenses (Niedermayr and Kontrus, 1973; Franz et al., 1986; Fig. 7c).

Inclusion trails in garnet and plagioclase porphyroblasts from the same metapelitic rock type have orientations similar to those of the emeralds (Fig. 7d; Grundmann and Morteani, 1982), i.e., all indicate multistage syn- to post-tectonic growth which is characteristic of regional metamorphism. A more detailed description and discussion of the structures and of the geochemistry of the emerald-bearing series at the Habachtal is given by Grundmann and Morteani (1982) and Grundmann (1983).

Leydsdorp

Figure 8a is a sketch of a thin section of emerald mineralization in biotite schist, with a quartz lens from within the contact zone between boudinaged albitite pegmatoid (not shown in sketch) and biotite schist. The emerald cores have inclusion trails of mica showing a completely different orientation from those of the surrounding biotite schist. As already mentioned, phenakite is often observed as a relict in the center of emerald (crystal) aggregates (Fig. 8b). A microscopic study shows that emerald replaces the phenakite and biotite and exhibits euhedral terminations against biotite. Some of the emerald porphyroblasts exhibit tensional cracks.

It is important to note that the emeralds at Leydsdorp generally display a more or less sharply zoned structure, with colorless inclusion-rich cores and in-

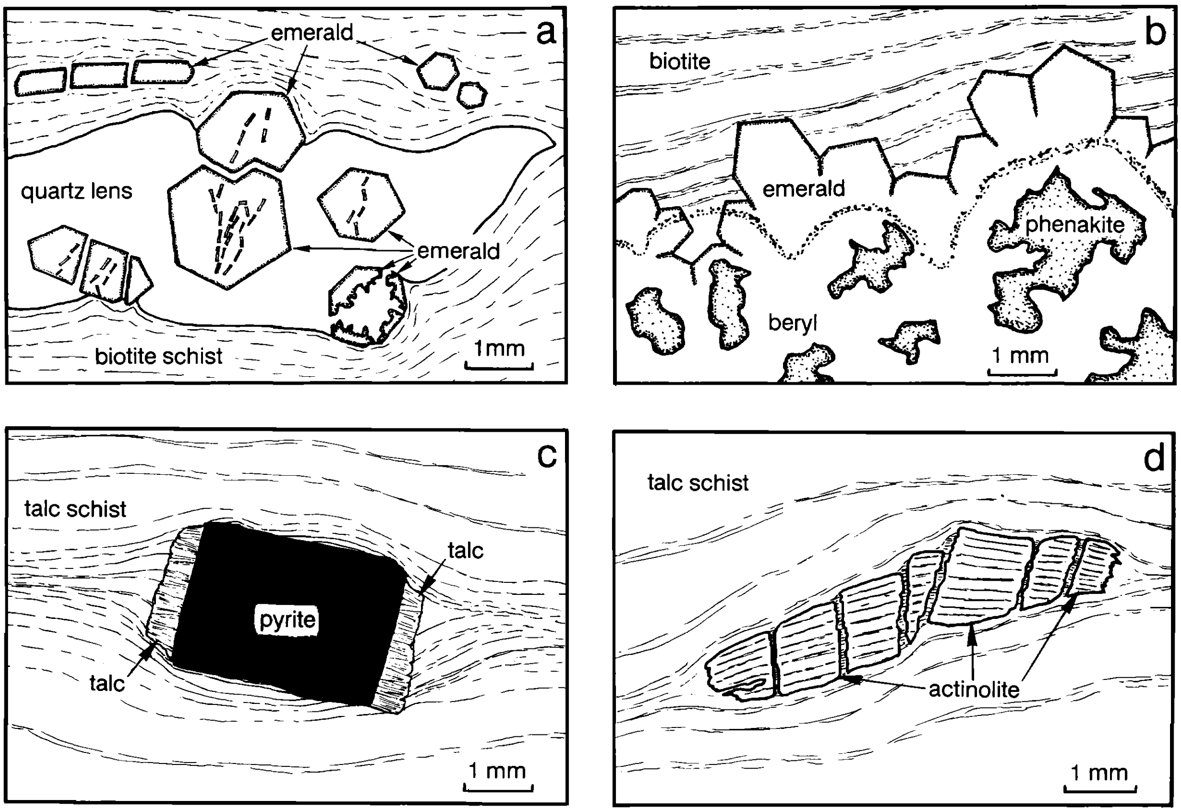

FIG. 8. a. Leydsdorp. Sketch of a photomicrograph. Emerald porphyroblasts (dotted rims) embedded in a quartz lens and in biotite schist. Emerald crystals may show cataclastic deformation both within and outside the quartz lens. The orientation of the biotite inclusions in the emeralds differs from the orientation of the surrounding biotite schist foliation. b. Leydsdorp. Sketch of a photomicrograph of euhedral emerald porphyroblasts growing on colorless beryl which contains relicts of phenakite, in a biotite schist matrix. The original outline of the phenakite is shown by fine-grained opaque inclusions. Emerald in this case is clearly posttectonic with respect to biotite foliation. c. Leydsdorp. Sketch of a photomicrograph of a pyrite crystal in talc schist showing a second generation of undeformed talc growing in the pressure shadows. d. Leydsdorp. Sketch of a photomicrograph of a ruptured actinolite crystal in talc schist showing a second generation of talc growing in the cracks.

TABLE 1. Be Contents of Selected Samples from the Habachtal Area (in ppm) as Determined by AAS and N-Gamma Spectrometry (reported in Grundmann, 1983)

No. of analyses	Rock type	Range (ppm)	Mean (ppm)	Standard deviation
12	Muscovite schists	10.7 to 71.9	35.9	18.6
4	K feldspar gneisses	13.7 to 37.6	27.8	10.3
14	Plagioclase gneisses (see below)	0.5 to 50.7	17.5	17.6
3	Talc-actinolite schists	9.0 to 18.4	15.2	5.4
7	Biotite schists	7.3 to 18.0	9.6	3.0
5	Amphibole gneisses	2.6 to 14.6	5.6	5.1
3	Chlorite schists	2.8 to 5.1	4.1	1.2
6	Garnet-mica schists	2.2 to 5.9	3.2	1.4
9	Amphibolites	0.8 to 4.9	3.1	1.6
1	Talc schist	0.1 to 0.5		
1	Antigorite-serpentinite	1.9		
1	Calcite-epidote rock	4.9		
7	Plagioclase gneisses (type I, prealpine plagioclase generation)	0.1 to 6.7	4.2	2.5
7	Plagioclase gneisses (type II, alpine plagioclase generation)	10.8 to 50.7	30.7	15.8

creasingly green-colored inclusion-poor rims. This observation, also reported by Le Grange (1929) and Van Eeden et al. (1939), emphasizes the preferential incorporation of chromium during a late stage of the growth of emerald. Robb and Robb (1986) have shown that the green coloration in the Leydsdorp emerald is directly proportional to its chromium concentration. Very commonly the green coloration cuts diagonally across the emerald crystals, suggesting that its development is controlled by a local supply of chromium. Microstructures indicate that chromium could originate from the partly or completely replaced (diagonally crosscutting) actinolites and biotites.

Pyrite is often found in association with emerald. Figure 8c shows a pyrite crystal embedded in a talc-actinolite-biotite schist with a very strong subhorizontal foliation. After the formation of the pyrite porphyroblast, a second oriented but undeformed generation of talc was formed in the pressure shadows of pyrite.

Figure 8d shows an actinolite crystal set in a talc-actinolite schist. The crystal is crosscut by numerous cracks. Some segments of the crystal are slightly rotated and displaced along these cracks. In the cracks, opened by tensional movements, a second talc generation has formed.

Plagioclase from both the albitite pegmatoid and the graphic-textured granites has been intensely deformed, leading to augen gneiss textures.

The textural evidence presented above for emerald, plagioclase, pyrite, and actinolite clearly demonstrates a multistage syn- to post-tectonic regional metamorphic growth history for the emerald deposits of Leydsdorp which is strikingly similar to that of the Habachtal deposit.

Geochemistry

An important aspect of the genesis of the emeralds is the source of the beryllium needed for their formation. In the case of Leydsdorp the primary source of beryllium is clear. Premetamorphic albitic and granitic pegmatoid rocks contain colorless beryl and/or phenakite. The macroscopic and microscopic observation of phenakite and colorless beryl in these rocks at Leydsdorp and their replacement or overgrowth by emerald makes the pegmatoid the most likely source of the beryllium.

A completely different situation is found at Habachtal. In this case no pegmatitic rocks of any kind are found; this is in direct contradiction to what was assumed by Leitmeier (1937). Geochemical investigations by Grundmann (1983) of the country rocks and minerals at Habachtal show that the entire rock series is enriched in Be. Table 1 presents the beryllium contents of the most important country rocks and those of the blackwall zones. The average of 68 Be whole rock analyses of the metamorphosed volcanic and sedimentary rocks is 16 ppm Be, well above the average Be value of 3.5 ppm for metamorphic rocks (range 0.5–8 ppm) as given by Hörmann (1969).

In Figure 9 the Be contents of biotite and of muscovite concentrates are plotted vs. the Be content of the host rocks. The Be content of the host rocks ranges up to 60.5 ppm, whereas the Be content of the biotites

TABLE 2. Maximum Measured Be Contents of Minerals (in ppm) According to Hörmann (1969) and Rösler and Lange (1976)

Mineral	Be
Muscovite	120
K feldspar	80
Albite	50
Actinolite	50
Tschermakitic amphibole	20
Biotite	20
Quartz	7
Chlorite	5
Talc	2
Antigorite	1

Table 2 presents the maximum measured Be contents reported worldwide in different minerals by Hörmann (1969) and by Rösler and Lange (1976). All mineral species listed in Table 2 are found as major components in the rocks of the Habachtal deposit. Biotite and muscovite are the most frequent rock-forming minerals in the emerald-bearing rocks as well as in the emerald-free rock series

Table 3 presents the mean Be contents in biotite, muscovite, and their host rocks from the Habachtal emerald deposit and neighboring rock series

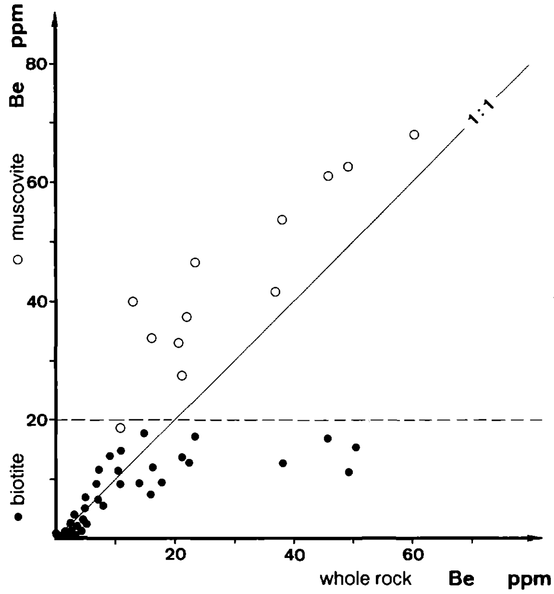

FIG. 9. Habachtal. Solid symbols = plot of the Be contents in biotite and the corresponding host rocks. A close correlation can be seen up to approx. 20 ppm Be in biotite. This can be considered to be the saturation limit. Open symbols = plot of Be contents in muscovite and the corresponding host rock. A good correlation exists, with no indication of Be saturation in the muscovite.

never exceeds 18 ppm and that of the muscovite reaches 68 ppm at the most (cf. Table 3).

The Be content of the biotite asymptotically approaches approx 20 ppm. This value is the maximum concentration of Be in biotite reported by Hörmann (1969) and Rösler and Lange (1976) (Table 2) and can be interpreted as the upper limit of Be uptake by biotite under the given metamorphic conditions. The Be content in the muscovites increases in proportion with the Be content of the whole rocks, and in contrast to the biotites no saturation value is reached in the studied range of Be concentrations.

Until now, no information exists about Be contents in minerals and host rocks from other emerald deposits. Further investigation is badly needed in order to define the saturation limits for Be in rock-forming minerals.

Blackwall Zones and Metasomatic Alteration

Field studies show that the emerald mineralization is associated with the metasomatic blackwall zoning. In order to investigate the site of the emerald mineralization, a detailed geochemical profile of the blackwall zoning including Be contents was made by Grundmann (1983).

The blackwall zoning is represented by this sequence: unaltered serpentinite, talc schist, actinolite schist, chlorite and/or biotite schist, and unaltered metapelites-metavolcanics. Similar sequences have been described and discussed in detail by Phillips and Hess (1936), Curtis and Brown (1969, 1971), and Brady (1977). In Figure 10 the bulk composition of the rocks across a typical blackwall zoning profile from the Habachtal emerald deposit (solid symbols) is plotted in the AMS diagram of Brady (1977). The blackwall zoning from Habachtal correllates reasonably well with the sequences (open symbols) suggested by Phillips and Hess (1936) and Curtis and Brown (1969) (in Brady, 1977).

Figure 11 shows the variation in Mn, Fe, Mg, Co, Cr, and Ni across the same blackwall profile after Grundmann (1983). The data from Habachtal also correlate well with the trace metal behavior in the zoned metasomatic bodies of Unst, Shetland Islands (Curtis and Brown, 1969, 1971). A clear discontinuity can be seen in Cr, Co, and Ni between the chlorite and the talc-actinolite zone. This discontinuity probably marks the original premetamorphic position of the boundary between the serpentinite and the pelitic schists. This assumption is also supported by the rare earth element pattern of the metasomatic zones (not shown) reported in Grundmann (1983). The most extensive emerald mineralization occurred within the biotite and chlorite zones. Biotite and chlorite, with maximum Be contents of 18 ppm and 5 ppm, respec-

TABLE 3a. Mean Be Contents of 38 Biotite Concentrates and Their Host Rocks from the Habachtal Emerald Deposit and Neighboring Rock Series (St.d. = standard deviation)

Number of samples	Rock type	Biotite Mean	Biotite St.d.	Be contents in host rock Mean	Be contents in host rock St.d.
6	Muscovite schists	13.6	(2.4)	28.8	(20.8)
7	Plagioclase gneisses	10.7	(2.5)	20.9	(16.9)
5	Biotite schists	10.1	(3.6)	10.1	(3.6)
1	K feldspar gneiss	9.8	—	15.2	—
4	Amphibole gneisses	4.9	(4.7)	5.8	(4.9)
8	Amphibolites	3.2	(2.1)	3.4	(1.8)
3	Quartz-albite layers	1.7	(1.1)	2.3	(2.1)
4	Garnet-mica schists	1.1	(1.0)	2.6	(0.5)

TABLE 3b. Be Contents of 12 Muscovite Concentrates and Their Host Rocks from the Habachtal Emerald Deposit and Neighboring Rock Series (St.d. = standard deviation)

		Be content			
		Muscovite		Host rock	
Number of samples	Rock type	Mean	St.d.	Mean	St.d.
1	Epidote-muscovite schist	67.9	—	60.5	—
2	Biotite-muscovite schist	61.8	(0.6)	47.6	(1.8)
1	Chlorite-muscovite schist	53.2	—	38.2	—
1	Albite-muscovite schist	41.6	—	37.1	—
2	Epidote-biotite schists	40.1	(6.5)	19.7	(3.6)
1	K feldspar gneiss	39.7	—	13.8	—
1	Quartz-muscovite schist	33.8	—	20.3	—
3	Muscovite-biotite schist	27.8	(9.4)	18.0	(4.1)

tively, replaced relatively Be-rich muscovite (maximum contents up to 68 ppm). The remaining Be was therefore liberated to form a beryllium phase such as emerald (see below).

Composition of Emeralds

Microprobe and wet chemical analyses by Franz (1982), Hänni (1982), Grundmann (1983), and Franz et al. (1984, 1986) show that the emeralds of Habachtal and Leydsdorp contain significant traces of Mg, Fe, Cr, Na, K, and H_2O. Grundmann (1983) and Franz et al. (1986) show that the trace element content of emeralds from Habachtal is controlled by the bulk-rock chemistry and by the composition of the fluid phase.

Using the results of Grundmann (1983), Franz (1982), Franz et al. (1986), and our data, it can be shown that the rims of the Habachtal and Leydsdorp emeralds contain high amounts of Mg (max MgO, 3.21%) and Cr (max Cr_2O_3, 0.91%). Both the magnesium and the chromium contents of the emeralds generally increase from the core to the rim. In contrast, the aluminum generally decreases from the core to the rim. Sodium and iron vary irregularly (max Na_2O, 2.46%; max FeO, 0.91%).

The correlation (Al + Cr)-(Mg + Fe + Na) has been used to distinguish between emeralds from different occurrences. The different compositional fields are shown in Figure 12, based on microprobe analyses of emeralds from Leydsdorp (South Africa), Habachtal (Austria), Takowaya (USSR), Muzo and Chivor (Colombia), Carnaiba and Socoto (Brazil), and Emmaville (Australia) (cf. Franz, 1982).

The Fluid Phase

Analysis of the associated fluorite provides supporting evidence for a nonpegmatitic origin of the emerald deposits. The rare earth element fractionation pattern of fluorite can be used to distinguish between pegmatitic, sedimentary, and hydrothermally formed fluorites (Möller and Morteani, 1983). Here the term hydrothermal is used to denote fluorites precipitated from hot solutions, not necessarily derived from a magmatic source.

The fluorite intergrown with emerald from the Leydsdorp deposit shows a rare earth element fractionation pattern characterized by low light rare earth element contents and enriched heavy rare earth elements (Fig. 13). In a Tb/Ca versus Tb/La atomic ratio diagram from Möller and Morteani (1983), the fluorite which coexists with the emerald from Leydsdorp (L) plots in the hydrothermal field (Fig. 14). No fluorite has been found at the Habachtal deposit, but fluorite

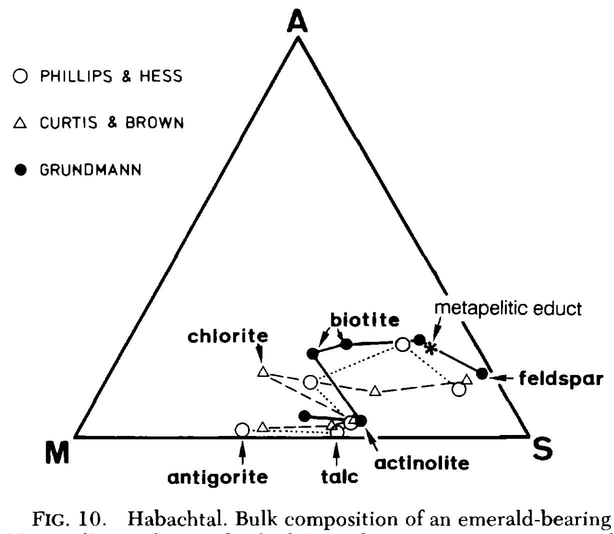

FIG. 10. Habachtal. Bulk composition of an emerald-bearing blackwall zone from Habachtal (Grundmann, 1983), consisting of the following rock types: K feldspar gneiss–white mica schist–biotite schist–chlorite schist–actinolite-talc schist–talc schist–serpentinite, plotted in the AMS diagram (A = Al_2O_3 + Fe_2O_3; M = MgO + FeO + MnO; S = SiO_2) according to Brady (1977). The blackwall sequences from Phillips and Hess (1936) (open circles) and Curtis and Brown (1969) (open triangles) show a good coincidence with the Habachtal sequence.

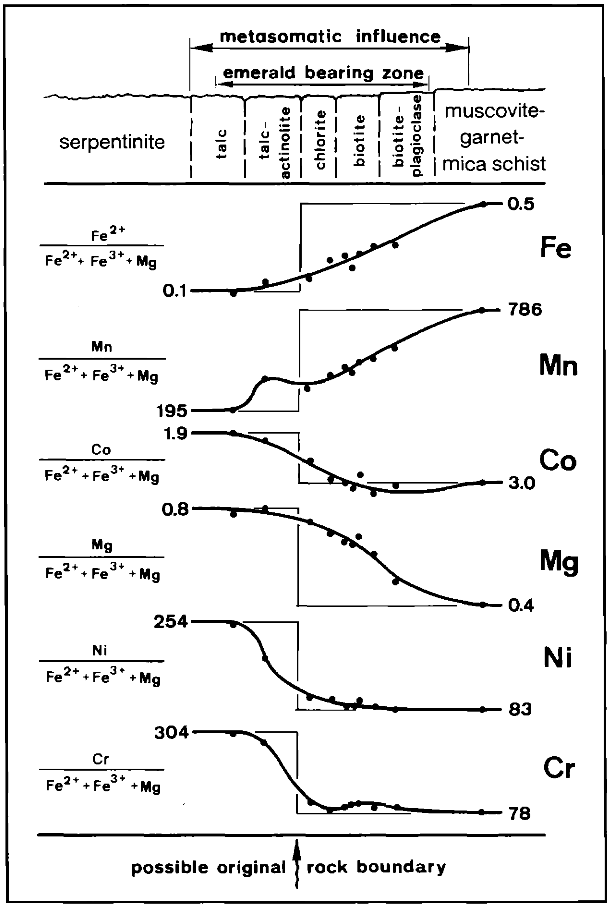

FIG. 11. Habachtal. Distribution of some main and trace elements through the Habachtal blackwall sequence, as shown in Figure 10, plotted as ionic fractions. The diagram was designed originally by Curtis and Brown (1971). Numbers on solid lines indicate minimum and maximum values outside the zone of metasomatic influence.

which coexists with emerald from the nearby scheelite mine Felbertal (F) in the Habach Formation also plots in the hydrothermal field in Figure 14, and shows a rare earth element fractionation pattern very similar to that of the Leydsdorp district.

Fluid Inclusions

Emeralds from Habachtal commonly contain primary fluid inclusions in cavities parallel to the c-axis. At room temperature the inclusions contain three fluid phases (liquid H_2O, liquid CO_2, and vapor; Fig. 15a). Daughter crystals are found locally but have not been identified. Microthermometric measurements, using a Chaixmeca heating-cooling stage, give an average homogenization temperature of 300°C and salinity of 6 to 9 equiv wt percent NaCl. The visually estimated CO_2 content in the three-phase inclusions is 15 to 20 vol percent. These fluid characteristics

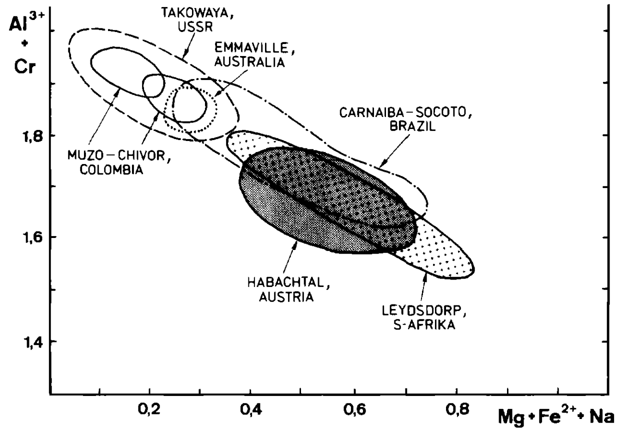

FIG. 12. Variation of Al + Cr with Mg + Fe^{+2} + Na in emeralds from Leydsdorp (South Africa), Habachtal (Austria), Takowaya (USSR), Muzo and Chivor (Colombia), Carnaiba and Socoto (Brazil), and Emmaville (Australia).

closely match those in synmetamorphic Alpine fissures in the Habach Formation reported by Luckscheiter and Morteani (1980), which further supports a regional metamorphic origin of the Habach emeralds.

Emeralds from Leydsdorp contain fewer fluid inclusions than those from Habachtal. Microscopic

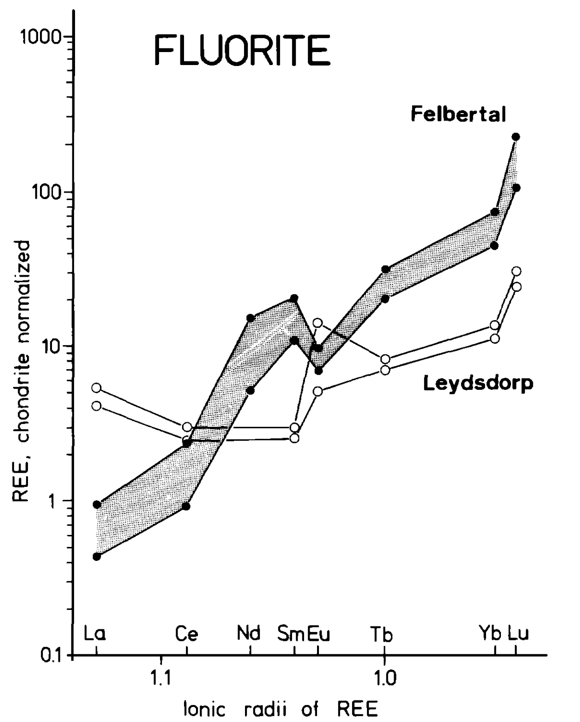

FIG. 13. Chondrite-normalized rare earth element distribution pattern of fluorites which coexist with emeralds from Leydsdorp and of fluorites which coexist with beryl from the scheelite deposit of Felbertal near the emerald deposit of the Habachtal. Both pattern are typical of remobilized fluorites.

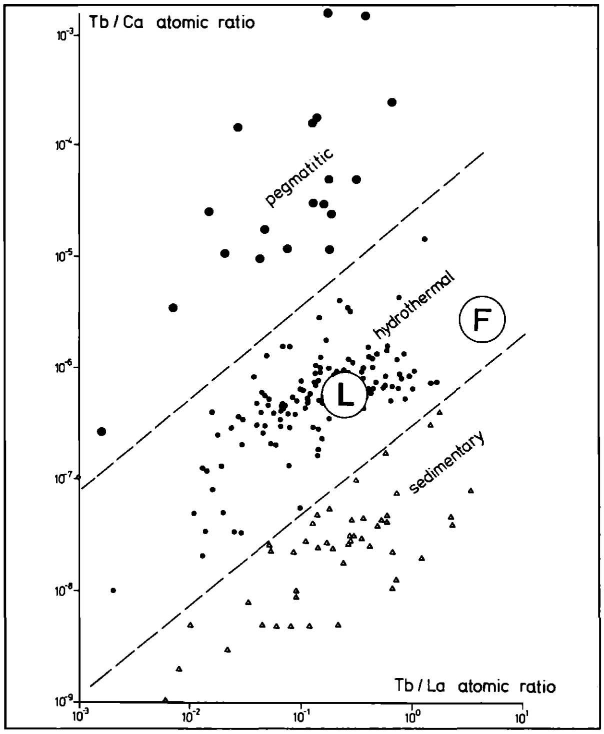

FIG. 14. Tb/Ca versus Tb/La atomic ratio diagram from Möller and Morteani (1983) of the fluorites which coexist with beryl from Leydsdorp (L) and Felbertal (F) emerald deposits.

studies show primary inclusions consisting of three fluid phases at room temperature (liquid H_2O, liquid CO_2, and vapor). Many inclusions also show unidentified daughter crystals (Fig. 15b). Microthermometric measurements using a Linkham S.O. Chaixmeca heating-cooling stage give an average homogenization temperature of 250°C, and salinity of 1 to 7 equiv wt percent NaCl. The visually estimated CO_2 content is 5 to 15 vol percent. A detailed study of the fluid inclusions in both deposits is in preparation.

Discussion

The formation of emeralds requires a combination of beryllium and chromium, two elements with very different geochemical behavior, which are generally concentrated in very dissimilar geochemical environments.

In the case of Habachtal, Be is enriched in the Habach Formation, as demonstrated in Table 1, probably due to submarine volcanic exhalation. This submarine exhalative origin is suggested by the following points: (1) a rich strata-bound Fe-Cu-Mo sulfide mineralization, (2) the very rich strata-bound scheelite deposit of Felbertal near the emerald deposit, and (3) the intercalation of metavolcanics and metapelites. The rare phenakite crystals at Habachtal can safely be excluded as a main source of beryllium for beryl formation.

Beryllium is mainly diadochially bound in silicates such as actinolite, biotite, muscovite, and feldspar. The transformation of the metapelites-metavolcanics and serpentinites into biotite-chlorite-actinolite-talc schists in blackwall zones would liberate significant amounts of Be and allow the formation of distinct Be species such as beryl or, in the presence of Cr liberated from the metasomatized ultrabasic rocks, emerald.

A calculation of mass balance for Be shows that the observed concentration of emerald is consistent with the amount of Be released by metasomatic reactions. Referring to Figure 16, the amount of Be in one metric ton of protolith muscovite schist is 36 g (mean of 12 muscovite schists, see Table 1). The amount of Be which remains in the unequal mass of the monomineralic biotite zone is at most 18 g. Thus the transformation to biotite schist liberated 18 g/metric ton. This would be enough to form 350 g of beryl or emerald.

In general, the contrast in size of emeralds and pegmatitic beryls is striking—they differ by orders of magnitude. Beryl forms larger crystals than any other mineral species—up to 200 metric tons. Aquamarines, for example, may reach 100 kg and retain their gem quality, whereas emeralds rarely if ever exceed 1 kg. The emeralds of Habachtal and Leydsdorp weigh on average less than 1 g.

The metasomatic process which produces a blackwall zoning sequence is therefore of essential importance for the crystallization of emerald in rocks containing no significant amounts of earlier Be minerals such as beryl or phenakite.

The following sequence of events is summarized in Figure 16: (1) juxtaposition of very Be-poor and Cr-rich serpentinite and Be-rich and Cr-poor metasediments, (2) the formation of the blackwall zones, and (3) material transfer. This sequence is supported by the occurrence of emerald mineralization between the boudinaged muscovite-plagioclase-chlorite schist and the biotite schist (see Fig. 5).

In the case of the Leydsdorp deposit, the Be occurs probably as a trace element bound in silicates within preexisting pegmatites, albitic pegmatoids, and biotite schists, and in preexisting beryl and phenakite. In the case of emerald formation from reaction of the preexisting phenakite with micas or plagioclase, the following reactions (1 and 2) are suggested by textural evidence:

$$3 \text{ phenakite} + 2 \text{ biotite} = 2\text{MgFe-beryl} + 2K^{+} + 4Mg^{+2}, \quad (1)$$

$$3 \text{ phenakite} + 3 \text{ albite} + 2Mg^{+2} = 2\text{MgNa-beryl} + Na^{+} + Al^{+3}. \quad (2)$$

In both Leydsdorp and Habachtal, the formation of blackwall zoning during regional metamorphism played a significant role in emerald genesis by pro-

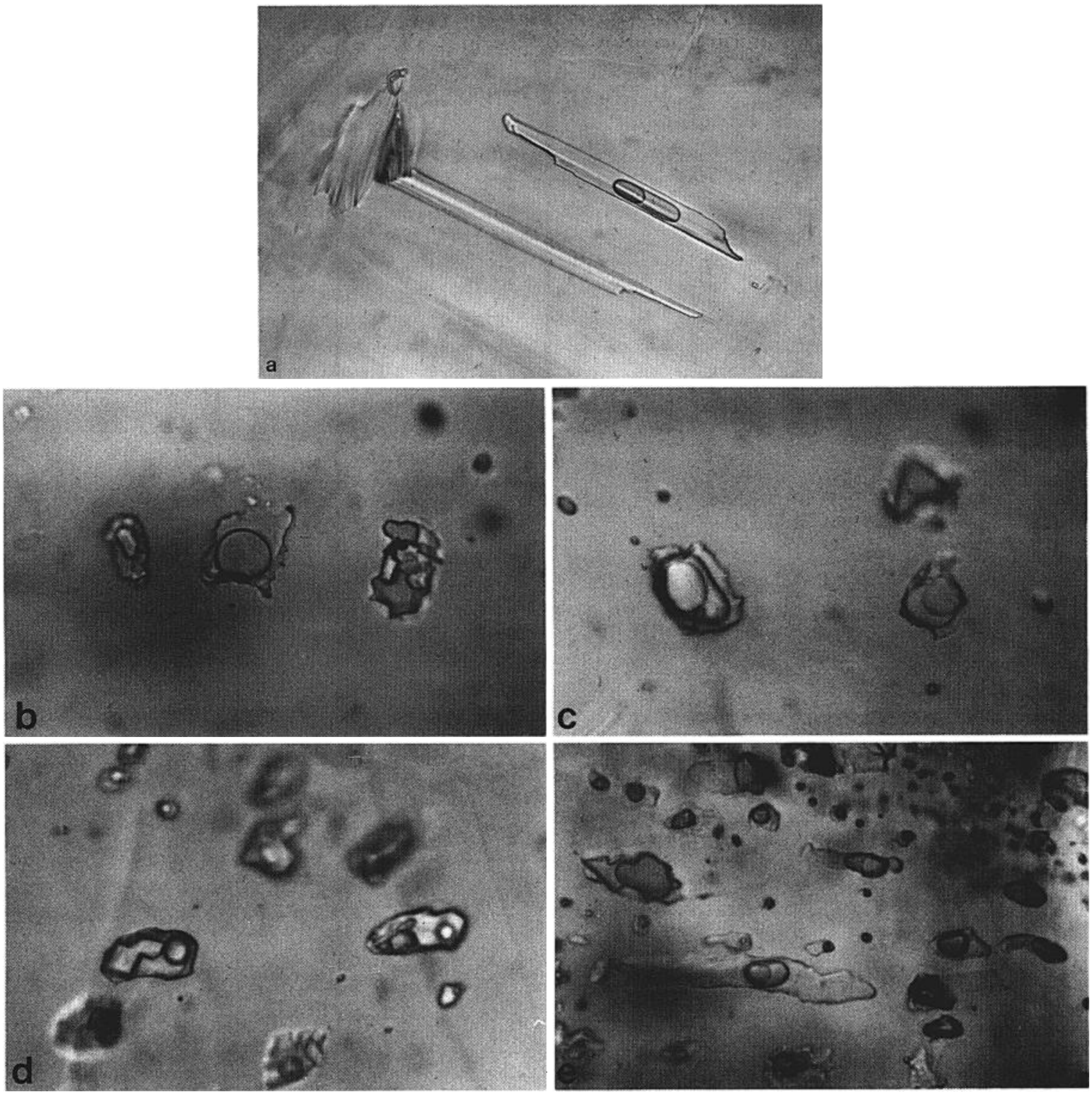

FIG. 15. a. Photomicrograph showing fluid inclusions of liquid H_2O, liquid CO_2, and vapor in an emerald from Habachtal. b. and d. Photomicrographs showing fluid inclusions of liquid H_2O, and vapor and unidentified daughter crystals. c. and d. Photomicrographs showing liquid and vapor CO_2 in emeralds from the Cobra pit, Leydsdorp, from the specimen shown in the sketch of Figure 8a.

viding a source of Cr (mobilized from serpentinites) and Be (mobilized from feldspar and mica-rich rocks). Intense schistosity in both deposits indicates that the blackwall zoning was formed during regional metamorphism involving an intense deformation of the rocks and minerals. Contact metamorphism alone as an emerald-producing process can be excluded in both deposits.

The high magnesium content of the emeralds reflects their formation in an Mg-rich environment. Beryls from pegmatites are very Mg poor (Franz et al., 1984).

The hydrothermal signature of fluorite demonstrates the influence of an abundant fluid phase during the crystallization of the emeralds and is strong evidence against a pegmatitic origin of the fluorite-emerald paragenesis. The composition of the fluid phase can be deduced from microthermometric studies. In both deposits an H_2O-CO_2 mixed fluid occurs with a salinity of about 1 to 9 equiv wt percent NaCl. The homogenization temperatures are around 300°C.

Conclusions

The prevalent theory that schist-hosted emeralds are formed in the context of a contact metasomatic

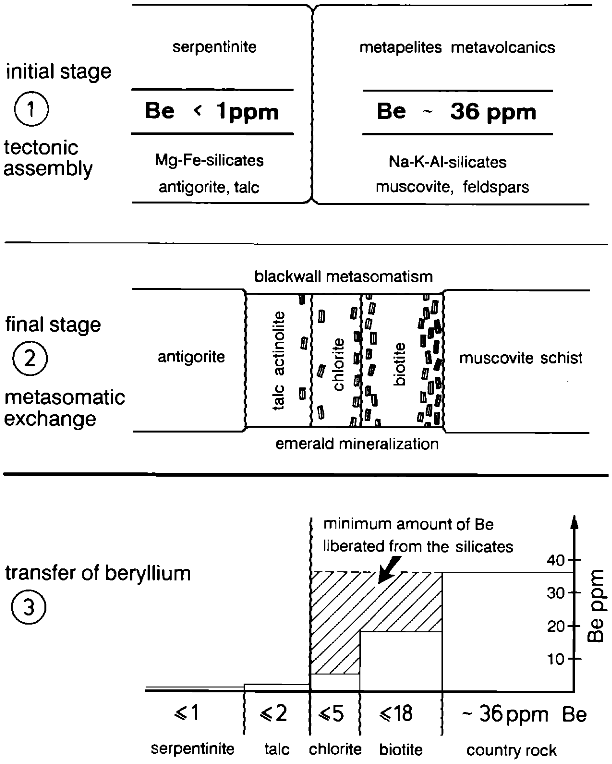

FIG. 16. Emerald mineralization due to regional metamorphism, shown in a schematic sequence of events: (1) initial stage, tectonic assembly; (2) final stage, metasomatic exchange and emerald formation; and (3) the amount of beryllium liberated from the Be-rich silicates such as muscovite (36 ppm Be = mean of 12 muscovite schists, for example, from Habachtal) is incorporated into the emerald porphyroblasts.

reaction of pegmatites or pegmatitic fluids with ultrabasic rocks cannot explain the tectonic, microstructural, and geochemical features of the emerald deposits of Habachtal or Leydsdorp. Microstructural evidence clearly proves that emeralds from both deposits formed in a multistage development of syn- and post-tectonic growth due to metasomatism at the contact between ultramafic rocks and mica-rich quartzofeldspathic rocks during regional metamorphism. The regional metamorphism caused reactions between the ultrabasic rocks and the surrounding silica-, alumina-, and alkali-rich country rocks leading to a classic sequence of schistose blackwall zones. During the formation of blackwall zones, relatively Be-rich Al silicates such as feldspar and muscovite were replaced by Be-poor minerals, such as biotite, chlorite, or talc, causing the liberation of Be and the subsequent crystallization of emerald as an individual Be phase. Alternatively, where phenakite and colorless beryl were present (Leydsdorp), during blackwall metasomatism a Cr-rich biotite was initially produced, then phenakite and colorless beryl reacted with the biotite to produce emerald. The necessary Cr and Mg originated from the simultaneous metasomatic alteration of serpentinites to talc schists.

We suggest that when prospecting for emeralds one should look not only for Be-rich pegmatites but also for Be-rich metavolcanic and sedimentary series in regional metamorphic tectonic mélange zones rich in ultramafic bodies. The grade of metamorphism where emeralds might be expected ranges from greenschist to amphibolite facies conditions. During a high-grade metamorphic event, with PT conditions approaching the anatectic regime, beryl and feldspar are not stable and decompose into chrysoberyl and quartz ± sillimanite (Franz and Morteani, 1981, 1984). Under greenschist facies PT conditions beryl and phenakite (and other silicates) are stable instead (Franz et al., 1984).

Examination of many emerald-bearing samples from other classical emerald occurrences, such as the Urals, Brazil, and Mozambique, revealed features very similar to those described in this paper from the Habachtal and/or Leydsdorp deposits.

It is suggested, therefore, that most if not all of the schist-type emerald deposits in metamorphic rocks may be the product of a deformational event which juxtaposed Be- and Cr-rich rocks, accompanied or followed by regional metamorphism which provided heat and the abundant fluids necessary for reactions. It follows that nomenclature of the emerald deposits, as suggested for example by Sinkankas (1981), must be revised to include a new regional metamorphic type of mineralization. We would like to encourage investigation in other areas and deposits to test our hypothesis.

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