Microtextures and fluid inclusions from vein minerals hosted in the Pillow Lavas of the Troodos supra-subduction zone

D. Quandt, P. Micheuz, W. Kurz, and K. Krenn
INSTITUTE OF EARTH SCIENCES, UNIVERSITY OF GRAZ, NAWI-GRAZ GECENTER, HEINRICHSTRASSE 26, A-8010 GRAZ, AUSTRIA

ABSTRACT

This study deals with microtextures and fluid inclusions from veins and vesicles hosted in the Troodos Pillow Lavas that enable a conclusive model for vein formation during the post-magmatic stage of the Troodos supra-subduction zone. Three different types of veins from the Upper and Lower Pillow Lavas are distinguished and imply different modes of fracturing, fluid flow, and precipitation. (1) Syntaxial calcite-, quartz-, and zeolite-bearing veins are interpreted as mineralized extension fractures that were pervaded by seawater. This advective fluid flow in an open system changed later into a closed system characterized by geochemical self-organization. (2) Blocky and (3) antitaxial fibrous calcite veins are associated with host rock brecciation due to hydrofracturing and diffusion-crystallization processes, respectively. Based on aqueous fluid inclusion chemistry with seawater salinities in all studied vein types, the representative fluid isotherms crossed with minimum hydrostatic pressure conditions yield vein mineral precipitation temperatures between 180 and 210 °C at 250 bar, independently of the Pillow Lava units. This points to a heat source for the circulating seawater and implies that vein and vesicle minerals precipitated shortly after pillow lava crystallization under dominant isobaric cooling conditions. Compared to previous suggestions derived from secondary mineral parageneses, significant higher temperatures of vein formation in the Troodos Pillow Lavas are proposed.

INTRODUCTION

The Troodos ophiolite, Cyprus, is one of the best preserved ophiolites and exposes a complete sequence of ultramafic rocks, layered gabbros, isotropic gabbros, sheeted dykes, pillow lavas, and marine sediments representing the uppermost mantle and oceanic crust (e.g., Gass, 1968; Moores and Vine, 1971; Pearce, 2003; Dilek and Furnes, 2009). Previous studies on the Troodos ophiolite focused in most cases on the geochemistry of the pillow lavas, their implications for plate tectonics (e.g., Schmincke et al., 1983; Pearce et al., 1984; Rautenschlein et al., 1985b), and processes related to alteration in oceanic crust (e.g., Gillis and Robinson, 1985, 1990). Based on geochemical analyses of the pillow lavas and sheeted dykes, a supra-subduction zone (SSZ) setting consisting of a mid-ocean spreading ridge above a nascent subduction zone (subduction initiation) followed by a slab-rollback was proposed (Pearce et al., 1984; Dilek and Flower, 2003; Pearce and Robinson, 2010). The detailed post-magmatic structural evolution of the Troodos SSZ, however, is widely unknown. Preceding studies quantified the extent of rotation and uplift (Robertson, 1977; Clube et al., 1985; Morris et al., 1990) and determined the temperature conditions of alteration. Thus, secondary mineral parageneses, oxygen isotope compositions of secondary carbonates, and fluid inclusions (FIs) trapped in vein quartz indicate an abrupt change in alteration temperatures from >300 °C in the Sheeted Dyke Complex to <100 °C in the Pillow Lavas (Spooner and Bray, 1977; Gillis and Robinson, 1990). These temperatures are linked with alteration processes but lack any structural information on vein and mineral growth.

Mineralized fractures and vesicles hosted in the Pillow Lavas have the potential to record this post-magmatic evolution in the tectonic framework of a SSZ and are useful indicators of paleo-fluid flow through young oceanic crust. Microtextures and crystallographic orientation of vein minerals provide information on the mode of fracturing. The mineralogy of veins and vesicles has important implications for fluid-rock interaction and environmental conditions during precipitation and growth. Fluid inclusions trapped in vein and vesicle minerals give information on the chemistry of the fluid system and fluid density evolution (e.g., Bons and Montenari, 2005, and references therein). Textural features of re-equilibrated fluid inclusions as a result of differences between internal and confining pressures or ambient temperature changes may contribute to the understanding of the post-magmatic isochoric or non-isochoric pressure-temperature (P-T) evolution path of the Troodos SSZ (Vityk and Bodnar, 1995).

In this study we present microtextures and fluid inclusion data from fracture-related veins and vesicles of the Troodos Pillow Lavas that shed light on this missing piece of post-magmatic evolution. Both approaches, supported by cathodoluminescence (CL) imaging as well as Raman microspectroscopy, contribute to the understanding of conditions of fracturing, fluid flow, vein and vesicle mineral precipitation and growth in the specific Pillow Lava unit. Particular emphasis is put on fluid inclusion petrography and microthermometry in order to analyze the chemical composition of the fluid system, to constrain P-T conditions of vein and vesicle mineral formation, and to track a P-T evolutionary path of veins and vesicles after...
their formation in a SSZ setting. Since structures of the Troodos ophiolite, including veins and vesicles, lack regional metamorphic overprints and uplift-related deformation (Gass and Masson-Smith, 1963; Moores and Vine, 1971), this approach constitutes an efficient tool in order to decipher post-magmatic microstructural as well as geochemical processes in this SSZ setting.

**GEOLOGICAL SETTING AND STRATIGRAPHY**

The Troodos massif is part of a chain of Neotethyan ophiolites that extend along the Alpine-Himalayan orogenic system (Dilek and Flower, 2003). The Troodos ophiolite shows a complete and well-preserved Penrose-type ophiolitic stratigraphy reflecting a section through oceanic lithosphere (Gass, 1968; Moores and Vine, 1971; Anonymous, 1972) that started to form ca. 92 Ma (Mukasa and Ludden, 1987).

Harzburgites and dunites constitute the base of the Troodos ophiolite and represent the sub-oceanic lithospheric mantle. They are overlain by ultramafic cumulates and gabbros, sheeted dykes and extrusive rocks, which are covered by Cretaceous pelagic sedimentary rocks (Moores and Vine, 1971; Rautenschlein et al., 1985b; Dilek and Furnes, 2009). Doming exposed mantle peridotites and plutonic rocks in the center of the Troodos massif that are enveloped by the Sheeted Dyke Complex and the extrusive volcanic sequence (Wilson and Ingham, 1959; Carr and Bear, 1960; Bear, 1960; Bear and Morel, 1960; Gass, 1960; Bagnall, 1960; Gass and Masson-Smith, 1963) (Fig. 1).

The extrusive volcanic sequence was subdivided into the Basal Group (BG), Lower (LPL), and Upper Pillow Lavas (UPL) (Wilson and Ingham, 1959; Bear, 1960; Bear and Morel, 1960; Gass, 1960). The BG constitutes a transition zone from the Sheeted Dyke Complex into the LPL (van Everdingen and Cawood, 1995). It is composed of altered dykes

![Simplified geological map of the Troodos ophiolite](https://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/10/4/566/4262506/566.pdf)
and pillow lava screens (Bear, 1960). The abundance of dykes, which represent a feeder system for the pillow lavas, decreases from the BG to the UPL (Bear, 1960).

In the Akai Canyon, an interlayering of evolved andesites and dacites of the LPL with primitive lavas (basaltic andesites) of the UPL is exposed, suggesting a temporal intersection in volcanic activity (Rautenschlein et al., 1985a, 1985b; Thy and Esbens, 1993). Chemical compositions of pillow lavas and volcaniclastic mate-
tion zone (Pearce et al., 1984), terminated by two transform fault zones related LPL originated from a spreading axis above a nascent subduc-
tion zone (Schmincke et al., 1983; Rautenschlein et al., 1985a, 1985b; Thy and Esbens, 1993). The boundary of these two chemically subdivided suites coincides locally with the LPL-UPL boundary described in earlier studies (Wilson and Ingham, 1959; Bear, 1960; Bear and Morel, 1960; Gass, 1960) or appears at some lower stratigraphic levels (Schmincke et al., 1983).

The UPL consist of silicified aphyric basaltic andesites, andesites, and dacites (Bednarz and Schmincke, 1994; Dilek and Furnes, 2009). They are distinguished from the UPL by their high TiO₂ contents and evolved island arc tholeiitic (IAT) character. Volcanic glass compositions show geochemical similarities with the underlying sheeted dykes (Thy and Esbens, 1993).

The UPL is composed of aphyric and olivine-phyric basalts, andesitic basalts, and picrites (Schmincke et al., 1983; Bednarz and Schmincke, 1994; Dilek and Furnes, 2009). Plagioclase phenocrysts are restricted to basalts, and picrites (Schmincke et al., 1983; Bednarz and Schmincke, 1994; Esbensen, 1993).

Trace element patterns and the presence of boninites in the extrusive sequence argue for a subduction-related setting (Cameron et al., 1979; Schmincke et al., 1983; Rautenschlein et al., 1985b; Pearce and Robinson, 2010). Thus, the Troodos massif is interpreted as a supra-subduction zone ophiolite in which the Sheeted Dyke Complex and the geochemically related LPL originated from a spreading axis above a nascent subduction zone (Pearce et al., 1984), terminated by two transform fault zones (MacLeod and Murton, 1993; Morris and Maffione, 2016). In contrast, the UPL formed due to subsequent magmatic underplating of the juvenile oceanic lithosphere during subduction initiation in a proto-fore- or back-
arc setting (Moore and Vine, 1971; Miyashiro, 1973; Pearce et al., 1984; Thy and Esbens, 1993; Dilek and Furnes, 2009; Pearce and Robinson, 2010). Pillow lava magmatism commenced ca. 91 Ma and persisted until ca. 75 Ma although ca. 56 Ma old depleted boninite eruptions provide a larger time range (Osozawa et al., 2012). The lack of volcaniclastic material indicates that mature arc volcanism was absent (Robinson et al., 1983).

Between Campanian and early Eocene times, the Troodos microplate rotated 90° anti-clockwise (Clube et al., 1985; Morris et al., 1990). Gyps-
sum and palygorskite veins in the uppermost Pillow Lavas are related to Miocene and renewed Pleistocene uplift of the Troodos ophiolite due to the collision with the Erosophenes seamount and serpentinitization of ultramafic rocks (Robertson, 1977; Gillis and Robinson, 1990; Robertson, 1998).

METHODS

Rock Sampling

Samples of veins and vesicles were taken within the LPL and UPL units (Table 1). Due to the occurrence of pillow lavas in the BG and the absence of a clear upper boundary to the LPL, it cannot be excluded that pillow lavas from the BG were sampled as well. The samples originate from pillow lava outcrops distributed throughout the Troodos ophiolite (Fig. 1). The intention was to sample single veins hosted by preferentially unaltered and undeformed pillow lavas that provide empirically more microstructural and -textural information than veins associated with host rock brecciation or alteration. For consistency, thin and thick sections for vein and host rock petrography, CL, as well as FI study were taken from the same rock chip samples.

Fluid Inclusion Microthermometry

FIIs were investigated in double polished thick sections (thickness ~0.15 mm) using a Linkam THSMG600 heating and freezing stage with an operating range from −196 to +600 °C equipped with an Olympus 80x ULWD objective at the NAWI-Graz Geocenter, Institute of Earth Sciences, University of Graz. The Synthetic Fluid Inclusion Reference Set (Bubbles Inc., Blacksburg, Virginia, USA) was used for stage calibration. Temperature measurements are reproducible to 0.2 °C at a heating rate of 0.1 °C/min. The heating-freezing stage was used to obtain eutectic melting temperatures of the aqueous ice phase (Ice₅V → IceLV) as well as final melting of ice (IceLV → LV) [Tₑᵢce(ice)]

TABLE 1. SAMPLE LOCATIONS

| Outcrop number | Coordinates | Location | Stratigraphic unit |
|----------------|-------------|----------|-------------------|
| CY1            | 34°56.158'  | Troodos NW-flank | LPL               |
| CY2            | 34°56.204'  | Troodos NW-flank | LPL               |
| CY4            | 34°58.638'  | Troodos NW-flank | LPL               |
| CY5            | 34°57.672'  | Troodos NW-flank | UPL               |
| CY7            | 35°09.880'  | N-Troodos, interior | LPL/BG            |
| CY8            | 35°11.490'  | N-Troodos, coast  | LPL               |
| CY10           | 34°45.638'  | S of Arakapas fault, S-Troodos | UPL       |
| CY11           | 34°46.359'  | S of Arakapas fault, S-Troodos | UPL       |
| CY12           | 34°44.742'  | S of Arakapas fault, S-Troodos | LPL       |
| CY13           | 34°46.673'  | S of Arakapas fault, S-Troodos | LPL       |
| CY14           | 35°00.681'  | NW cape, interior, NW-Cyprus | LPL/BG   |
| CY15           | 35°02.555'  | NW cape, coast, NW-Cyprus | UPL       |
| CY17           | 35°01.132'  | Near Margi, E-Troodos | Uppermost UPL (+ LPL?) |

Note: Stratigraphic unit is based on the geological map and thin section petrography of host rocks. See also figure 1 for sample location. GPS coordinates refer to WGS84. BG—Basal Group; LPL—Lower Pillow Lavas; UPL—Upper Pillow Lavas.
properties and isochores have been calculated using the software BULK and ISOC, respectively (Bakker, 2003).

**Raman Spectroscopy**

Raman spectra of minerals were performed in confocal mode using a Jobin Yvon LabRam HR800 microspectrometer equipped with an Olympus BX41 optical microscope and a Si-based CCD (charged-coupled device) detector at the NAWI-Graz Geocenter, Institute of Earth Sciences, University of Graz. The instrument uses a 100 mW Nd:YAG laser (532 nm emission), a grating of 1800 grooves/mm, and a slit width of 100 μm. The spectral acquisition time was set to 10–20 seconds for all measurements in the range between 100 and 1200 cm⁻¹ for silicate phases but also between 1100 and 3800 cm⁻¹ for gases and OH⁻ phase. Raman spectroscopy was conducted on the same thin sections as used for petrographic and CL study.

**Cathodoluminescence Microscopy**

The CL study was conducted with a hot cathode CL microscope (Lumic HC5-LM) at the NAWI-Graz Geocenter, Institute of Earth Sciences, University of Graz. Thin sections were polished and carbon coated. Electrons were accelerated with 13–14 kV under a vacuum of <10⁻⁶ mbar. Beam current varied from 0.2 to 0.6 mA. True-color CL in the visible spectrum was observed and imaged in real time with an attached digital camera.

**Petrography**

Vein Macro- and Microtextures

Based on mineral shapes and microstructures, veins and mineralized vesicles are distinguished using the classification by Bons et al. (2012, and references therein). Veins and vesicles differ in their mineralogy. Structural and textural similarities, however, enable a subdivision into (1) syntaxial veins and vesicles, (2) blocky veins, and (3) antitaxial veins. Textural transitions occur between syntaxial and blocky veins.

**Syntaxial Veins and Vesicles**

Syntaxial veins occur throughout the whole Troodos ophiolite within the LPL and UPL. They form single veins with relative uniform thickness of ~1–3 mm. Calcite, quartz, and zeolites are the main mineral phases. Samples can be subdivided into completely sealed veins that display a median line and incompletely sealed veins and vesicles with late-stage precipitates (Table 2).

**Completely sealed syntaxial veins with a median line.** Completely sealed veins of the UPL are pure calcite veins that are characterized by growth competition between adjacent elongate-blocky crystals and a median line (Fig. 2A). Crystals show strong undulatory extinction, while twinning is rare (type I and II after Burkhard, 1993).

**Syntactical zeolite veins with a median line belong exclusively to the LPL.** They are composed of small euhedral tabular heulandite followed by mordenite growth that terminates at the median line (Fig. 2B), or elongate-blocky analcite (Fig. 2C). These zeolites are distinguished by their characteristic Raman spectra (Fig. 3).

**Incompletely sealed syntaxial veins and vesicles with late-stage precipitates.** Incompletely sealed syntaxial veins of the UPL are composed of blocky to elongate-blocky analcite (Fig. 2D) containing up to three wall rock–parallel solid inclusion bands indicative of repeated crack and sealing events. Non-sealed vein sites are filled with twinned (type I and II twins after Burkhard, 1993) anhedral calcite (Fig. 2D), euhedral acicular natrolite, and blocky analcite that are distinguished by their characteristic Raman spectra (Fig. 3).

The occurrence of incompletely sealed syntaxial quartz veins is restricted to the LPL. Quartz develops euhedral crystal habits. Late-stage anhedral calcite and accessoarly goethite precipitated in non-sealed vein sites (Fig. 2E). The proportion between quartz and calcite varies from pure quartz to calcite-dominated veins. Additionally, quartz crystals are located between grain boundaries of late-stage calcite. Late-stage calcite shows type I and II twinning (Burkhard, 1993), and quartz exhibits undulatory extinction.

In large vesicles (up to 25 mm) of quartz is coarse grained (up to 10 mm) and has elongate-blocky crystal shapes with undulatory extinction and subgrain boundaries. Quartz precipitation along the vein margins is characterized by growth competition and followed by formation of late-stage quartz in non-sealed sites of the vesicle (Fig. 2F).

**Blocky Veins**

Blocky veins occur in the LPL as well as UPL and are associated with host rock breccias and branching vein networks. Calcite is the major vein component and develops anhedral blocky crystal aggregates (Figs. 2G–2L, Table 2). Micocrystalline calcite (micrite), euhedral quartz, and angular host rock fragments occur subordinately in some blocky vein samples. Calcite shows undulatory extinction and subgrain boundaries together with type I and II twinning (Burkhard, 1993). Additionally, growth zones are outlined by decrepitated FIs (Fig. 2G). Twin lamellae are partly tapered and slightly bent. Individual euhedral quartz crystals are present along vein margins and between calcite grain boundaries (Fig. 4D).

**Antitaxial Veins ± Median Line**

Antitaxial veins occur exclusively in the UPL (location near Margi) where up to 50-mm-thick veins run parallel to each other or branch out and are exposed over several meters along strike. This vein type consists...
Figure 2. Photomicrographs of all vein types. (A–C) Completely sealed syntaxial veins composed of elongate-blocky calcite (Cal), mordenite (Mor) succeeding heulandite (Hul) and analcime (Anl) whose growth terminates at a median line (# nicols). (D–F) Incompletely sealed syntaxial veins without median line: (D) analcime and late-stage calcite (# nicols), (E) euhedral quartz (Qtz) and late-stage calcite with accessory goethite (Gth, # nicols), and (F) vesicle with elongate-blocky and late-stage quartz (/// nicols). (G–I) Blocky calcite veins are characterized by (G) growth zones, best distinguished by inclusion-rich (fluid inclusion [FI] decrepitation, // nicols, see Fig. 5G for close-up of decrepitated FIs) and inclusion-poor domains or (H) microcrystalline calcite and host rock fragments (# nicols). (I) Blocky calcite vein crosscut by incompletely sealed syntaxial analcime vein (// nicols). (J–L) Antitaxial calcite veins (J and L, # nicols, K, // nicols). (J) Curvature of calcite fibers decreases toward the vein center where the fibrous crystal habit is only recognizable as a relict. Fibers contain multiple wall rock-parallel solid inclusion bands (arrows in J, K, and L).
fibrous calcite veins are visible due to different CL intensities (Fig. 4G). Additionally, crosscutting relationships between two antitaxial cent bands pervade calcite fibers and appear parallel to the median line is only visible under CL, toward the margins (Figs. 4E and F). Luminescence along fiber growth direction from the median line, which in incompletely sealed syntaxial calcite veins, high-luminescent Mn-rich domains exhibit highly repetitive oscillatory growth zoning, including sectoral and intrasectoral zones and luminescent calcite twins (Fig. 4C).

In blocky veins, calcite displays comparatively broad growth zoning. Mn-rich high-luminescent growth zones, rich in decrepitated fluid inclusions, alternate with Mn-poor low-luminescent fluid inclusion-poor areas (Fig. 4D).

Antitaxial calcite veins show abruptly as well as gradually decreasing CL intensities due to elevated Mn/Fe ratios (Fig. 4D). FIs are large in size (up to 100 µm) and in most cases decrepitated and empty. They occur predominantly as clusters and/or as single inclusions (Figs. 5F–5H). Temperature estimates of fluid entrapment ($T_h$) for type A FIs, which show a wide range from 173.6 to 227.3 °C in quartz (peak frequencies between 180 and 190 °C) and 147.3–217.8 °C in calcite (peak frequencies between 190 and 210 °C) (Figs. 6A and 6B). This corresponds to densities in quartz and calcite from 0.85 to 0.92 g/cm³ and from 0.87 to 0.94 g/cm³, respectively. Analcime hosts minor type A FIs (Fig. 5C) with $T_h$ ranging from 183.5 to 210.7 °C (Fig. 6C). Corresponding densities are between 0.88 and 0.91 g/cm³.

**Fluid Inclusion Type B**

Type B FIs are arranged along intragranular (grain boundary to grain boundary or grain boundary to interior) trails and planes (Figs. 5D and 5E) and in general, if not elongated in shape, smaller (in most cases <10 µm) than type A FIs. Elongated FI trails in quartz and calcite, however, reach long axis diameters of 20–30 µm. Degree of fill is almost comparable to type A FIs between 0.90 and 0.95. Minimum temperature estimates for vein formation are linked with homogenization temperatures $T_L$ of primary type A FIs, which show a wide range from 173.6 to 227.3 °C in quartz (peak frequencies between 180 and 190 °C) and 147.3–217.8 °C in calcite (peak frequencies between 190 and 210 °C) (Figs. 6A and 6B). This corresponds to densities in quartz and calcite from 0.85 to 0.92 g/cm³ and from 0.87 to 0.94 g/cm³, respectively. Analcime hosts minor type A FIs (Fig. 5C) with $T_h$ ranging from 183.5 to 210.7 °C (Fig. 6C). Corresponding densities are between 0.88 and 0.91 g/cm³.

**Fluid Inclusion Type C**

Type C FIs in quartz and calcite are in most cases restricted to inclusion-rich growth domains. In case of calcite, these growth zones can be clearly distinguished from areas containing type A and B FIs by increased CL intensities due to elevated Mn/Fe ratios (Fig. 4D). FIs are large in size (up to 100 µm) and in most cases decrepitated and empty. They occur predominantly as clusters and/or as single inclusions (Figs. 5F–5H). $T_h$ range from 197.5 to 239.1 °C in quartz, 219.4–253.0 °C in calcite, and 196.3–262.3 °C in analcime (Figs. 6A–6C). However, due to re-equilibration and density loss, densities are not calculated (Table 3).
Figure 4. Cathodoluminescence (CL) images of vein types. (A) Calcite from completely sealed syntaxial veins with median line and (B) late-stage calcite (Cal) from incompletely sealed syntaxial veins display high-luminescent areas enriched in decrepitated fluid inclusions (FIs). Qtz—quartz. (C) Late-stage calcite associated with analcime in incompletely sealed syntaxial veins exhibits highly repetitive oscillatory growth zoning with sectoral and intrasectoral zones. (D) In blocky calcite veins, Mn-rich luminescent growth zones enriched in decrepitated FIs alternate with low-luminescent growth zones without decrepitated FIs. (E) Fibrous calcite from antitaxial veins reveals a decrease in CL intensities in fiber growth directions (arrows). Note median lines are only recognizable under CL (E, F). (G) Late calcite generation with low CL intensities crosscuts an earlier calcite generation (arrows).

### TABLE 3. MICROTERMOREMISTRY DATA OF FLUID INCLUSIONS

| Sample  | Vein type        | Stratigraphic unit | Host mineral | Fluid inclusion type | n    | Texture          | Size (µm) | Phases | \(T_h\) (°C) | Density (g/cm³) |
|---------|------------------|--------------------|--------------|----------------------|------|------------------|-----------|--------|---------------|-----------------|
| CY5     | Syntaxial vein   | UPL                | Anl          | A                    | 4    | Cluster, single  | 1 to 10   | L + V  | 183.5 to 210.7 | 0.88 to 0.91    |
|         |                  |                    |              | Anl                  |      | Re-equilibrated  | 5 to 30   | L + V  | 196.3 to 262.3 | N.D.            |
|         |                  |                    |              | Cal                  | 3    | Single           | 5 to 20   | L + V  | 1473 to 172.3  | 0.92 to 0.94    |
|         |                  |                    |              | Cal                  | 4    | Trails           | 5 to 20   | L + V  | 90.0 to 139.9  | 0.95 to 0.98    |
| CY7     | Syntaxial vesicle| LPL/BG             | Qtz          | A                    | 5    | Cluster, single  | 5 to 10   | L + V  | 183.0 to 227.3 | 0.85 to 0.91    |
|         |                  |                    |              | Qtz                  | 31   | Trails, planes   | 5 to 30   | L + V  | 92.9 to 177.5  | 0.91 to 0.98    |
| CY8     | Blocky vein      | LPL                | Cal          | A                    | 39   | Single, cluster  | 5 to 30   | L + V  | 169.4 to 217.8 | 0.87 to 0.92    |
|         |                  |                    |              | Cal                  | 13   | Trails, planes   | 1 to 10   | L + V  | 139.7 to 176.2 | 0.95 to 0.98    |
| CY12    | Syntaxial vein   | LPL                | Qtz          | A                    | 19   | Cluster, single  | 5 to 10   | L + V  | 173.6 to 203.9 | 0.88 to 0.92    |
|         |                  |                    |              | Qtz                  | 9    | Trail            | 1 to 10   | L + V  | 149.0 to 175.5 | 0.92 to 0.94    |
|         |                  |                    |              | Qtz                  |      | Re-equilibrated  | 5 to 50   | L + V  | 197.5 to 239.1 | N.D.            |

Note: Anl—analcime; BG—basal group; Cal—calcite; LPL—Lower Pillow Lavas; L + V—liquid plus vapor; n—amount of measured fluid inclusions; N.D.—no data; Qtz—quartz; \(T_h\)—homogenization temperature; UPL—Upper Pillow Lavas.
It is assumed that veining occurs steadily after solidification of newly formed pillow lavas providing sufficient heat for the observed high-temperature precipitates of seawater origin. This implies firstly that veins from the UPL and LPL are not contemporaneous and veining initiated in the LPL before emplacement of the UPL, and secondly solely hydrostatic pressure applies to all host minerals during FI entrapment. The lack of veins in the studied outcrops with consistent mineralogy and microtextures that crosscut the whole volcanic pile supports this assumption.

Pillow lava extrusion spans a wide time range from 91 to 75 Ma with some limited depleted boninite eruptions until 56 Ma along the southwestern flank of the Troodos ophiolite (Osozawa et al., 2012). It is therefore supposed that veins and vesicles have formed within this temporal framework.

P-T conditions for vein and vesicle mineral precipitation are based on isochore calculations of type A and B FIs combined with a hydrostatic pressure estimate of 250 bar that corresponds to a seawater column of ~2500 m, analog to average mid-ocean ridge depth below sea level (Sclater et al., 1971; Spooner, 1980). This hydrostatic pressure, however, represents a minimum estimate, since it cannot be excluded that ongoing pillow lava emplacement on top of the volcanic pile led to an additional load.

Calculated steep isochores of aqueous type A and B FIs from each vein type represent approximate geothermometers (Figs. 7A–7D). Isochores are constrained by ± 20% of the estimated minimum hydrostatic pressure (200 and 300 bar).

Type A FIs entrapped in quartz and calcite from syntaxial and blocky veins of the LPL including vesicles show a clear overlap of P-T conditions
Figure 6. Histograms of type A, B, and C fluid inclusions in (A) quartz, (B) calcite, and (C) analcime.

Figure 7. Calculated isochore fields of type A and type B fluid inclusions (FIs) in calcite, quartz, and analcime hosts from the Lower Pillow Lavas (LPL) and Upper Pillow Lavas (UPL). (A and B) Isochore fields of primary type A FIs entrapped in veins and vesicles from the LPL and UPL. (C and D) Isochore fields of secondary type B FIs entrapped in veins and vesicles from the LPL and UPL. Fields result from steep isochores from aqueous FIs crossed with minimum hydrostatic pressure. Gray arrows indicate isobaric cooling path. Numbers refer to FI densities of respective isochores in g/cm³. Anl—analcime; Cal—calcite.
Mechanisms for Vein Mineral Precipitation and Fluid Flow

The occurrence of quartz and zeolites with high Si/Al ratios (mor-delineite: 5, heulandite: 2.8–4) (Khodabande and Davis, 1996) is restricted to the silica supersaturated LPL and indicates that the chemical composition of the host rock is a crucial factor for mineralogy of vein precipitates. In contrast, veins within the basaltic and boninitic UPL bear calcite and zeolites with low Si/Al ratios (analcime: 2–2.5, natrolite: 1.5) (Chiperia and Apps, 2001).

The dominant fluid system during vein formation is inferred from FIs that indicate a simplified H₂O-NaCl ± KCl ± MgCl₂ chemistry, independent of the host mineral. Regarding microtextures, syntaxial and blocky veins give evidence for advective fluid flow through fractures in an open system. Early-stage quartz, detached from vein walls and located between grain boundaries of late-stage calcite, argues additionally for destructive fluid flow. Oscillatory growth zoning of late-stage calcite in cavities of incompletely sealed syntaxial veins points to a change into pervasive porous flow or even diffusion through host rock and vein minerals. A pure diffusive regime is proposed for antitaxial fibrous calcite veins (Means and Li, 2001; Bons et al., 2012). The fine nature of zonation in late-stage calcite can be related to geochemical self-organization of a Mn-bearing fluid in a closed system. Furthermore, the presence of Mn decreases the precipitation rate and modifies the crystal lattice and its vacancies (Meyer, 1984; Dromgole and Walter, 1987; Reeder et al., 1990). This modification enables the entrapment of numerous large fluid inclusions, which are susceptible to re-equilibration and decrepitation, considering a non-isochoric P-T evolution (e.g., isobaric cooling and/or isothermal decompression) (Vityk and Bodnar, 1995). This explains the coincidence of high-luminescent growth domains with high accumulations of decrepitation textures in blocky and elongate-blocky calcite (Figs. 4A and 4D). Syntaxial veins differ by their degree of sealing that is determined by the ratio between mineral growth and fracture opening. In incompletely sealed syntaxial veins fracture opening exceeds mineral growth rate and results in growth within a fluid-filled space, whereas completely sealed syntaxial veins are characterized by almost equal rates between mineral growth and fracture opening (Fisher and Brantley, 1992). The latter also represents a crack and sealing process (Ramsay, 1980; Bons et al., 2012).

Tectonic Implications—Modeling Vein Precipitation in a SSZ Setting

The mineralogy and high formation temperatures of veins investigated in this study are distinct from uplift-related gypsum and palygorskite veins within the UPL (Gillis and Robinson, 1990) that usually form at lower temperatures (<100 °C) (Church and Velde, 1979; Garcia-Ruiz et al., 2007). Therefore, the studied veins are probably related to pre-uplift processes.

Syntactical veins are the most common vein type and are distributed throughout the whole volcanic sequence (Figs. 2A–2F). They can be related to mode I fractures after Bons et al. (2012), i.e., extensional gashes, characteristic for a tectonic origin. In contrast, blocky veins (Figs. 2G–2I) are associated with brecciated host rocks. They are interpreted as hydrofractures associated with fluid overpressure and prevent a tectonic origin (Bons et al., 2012). Antitaxial fibrous veins (Figs. 2J–2L) result from enhanced crystallization processes and are not related to tectonic processes (Taber, 1916; Means and Li, 2001). Fiber curvature points to changing stress orientations or rotations within the stress field during growth. Solid inclusion bands in these antitaxial fibrous veins most likely document changes in fluid chemistry rather than multiple crack and sealing processes.

Subduction initiation linked with slab-rollback has a significant extensional component. Slow spreading ridges like the Troodos SSZ are in particular associated with tectonic extension and normal faulting compared to fast spreading ridges like the Semal SSZ (Pearce et al., 1984; Dilek and Eddy, 1992; Mutter and Karson, 1992; Searle and Cox, 1999; Abelson et al., 2001; Dilek and Furnes, 2009). Hence, syntactical veins, which are the most common vein type in the whole lava pile, may have a tectonic origin and can be linked with this extensional component.

A tectonic sketch explains formation and precipitation processes of the different studied vein types together with FI evolution in the proto-fore-arc setting of the N-S striking Troodos SSZ (Pearce et al., 1984; Maffione et al., 2017) exemplarily for the UPL (Fig. 8). Completely sealed syntactical veins suffered fracturing and subsequent seawater infiltration accompanied by immediate mineral precipitation (crack and sealing process, Fig. 8A). Incompletely sealed syntactical veins underwent fast opening rates and comparatively slow mineral growth rates (Figs. 8C and 8D). Late-stage minerals filled open cavities and mark a change from an open into a closed system that is characterized by geochemical self-organization (Fig. 8E). FIs entrapped in host minerals of both syntactical vein types underwent isobaric cooling that finally resulted in re-equilibration and decrepitation of especially large-scale FIs (Figs. 8B and 8F). Blocky veins are associated with host rock brecciation and subsequent mineralization (Figs. 8G...
Figure 8. Tectonic sketch of the Troodos supra-subduction zone (SSZ) presenting a model for multiple vein mineral precipitation and fluid inclusion (FI) evolution exemplarily for the Upper Pillow Lavas. Veining of the Lower Pillow Lavas occurred analogously. Spreading axis strikes N-S parallel to the subduction zone and is bounded by the Northern (NTTFZ) and the Southern Troodos Transform Fault Zone (STTFZ). (A and B) Model for completely and (C–F) incompletely sealed syntaxial veins; (G–I) model for blocky veins with mineral precipitation after brecciation; (J) antitaxial veins as a result of diffusion-crystallization processes. (E) Late-stage precipitates in incompletely sealed syntaxial veins indicate initial isobaric cooling that continues over a more extended temperature and time range (B, F, and I) resulting in decrepitation and local re-equilibration of FIs in syntaxial and blocky veins. FIs in antitaxial veins lack or are insufficiently small.
and 8H) Entrapment of Fs and their re-equilibration and decrепitation are comparable to syntaxial veins (Fig. 8I). In both blocky and syntaxial veins, the host rock delivered the heat to increase the temperature of the pervading seawater. In antitaxial veins, fracturing is caused by mineral precipitation due to diffusion (Fig. 8J). Fs in this vein type are insufficiently small.

CONCLUSIONS

Previous studies on secondary minerals in the Troodos Pillow Lava sequence provide insights into temperature and chemical conditions during formation but give no information on vein mineral growth (Gillis and Robinson, 1985, 1990; Staudigel et al., 1986). This study deals with veins that reveal microtextures that are (1) possibly related to tectonic extension (syntaxial veins), (2) originated from hydrofracturing (blocky veins), and (3) formed due to diffusion-crystallization processes (antitaxial veins). The syntaxial veins are in good agreement with the typical tectonic regime observed in proto-fore-arc areas within supra-subduction zones. This area is dominated by extension and associated with slow spreading rates in combination with a potential subduction initiation-coupled slab rollback (Pearce et al., 1984; Dilek and Eddy, 1992; Muller and Karsen, 1992; Searle and Cox, 1999; Abelson et al., 2001; Dilek and Furnes, 2009).

The chemical composition of the host rock is the crucial factor for the vein mineralogy as also observed in the case of low-temperature alteration minerals of the Troodos ophiolite (Gass and Smewing, 1973; Gillis and Robinson, 1985, 1990). Initial mineral precipitation in fractures took place in an open system with advective fluid flow but changed later after incomplete sealing into a closed system characterized by geochemical self-organization (growth zones late-stage calcite). Antitaxial fibrous veins give additionally a clear evidence for a diffusive regime.

Consistent $T_m$ and $T_h$ of calcite, quartz, and analcime veins ($T_m$ from 180 to 210 °C) suggest precipitation from seawater shortly after pillow lava crystallization that agrees with earlier studies by Spooner and Bray (1977) and Staudigel et al. (1986).

Isobaric cooling of the entire Troodos oceanic crust resulted in equilibration and decrепitation of former large Fs entrapped in Mn-rich growth zones that facilitate their modification. These results from veins throughout the Troodos ophiolite are independent of mineralogy as well as spatial or stratigraphic units and argue for a Troodos-wide coherent isobaric cooling process that preceded the uplift of the Troodos ophiolite.

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