Hydrothermal activity on the CV parent body: New perspectives from the giant Transantarctic Mountains minimeeteorite TAM5.29

J. NAVA,*, M. D. SUTTLE, R. SPIESS, L. FOLCO, J. NAJORKA, C. CARLI, and M. MASSIRONI

1Department of Geosciences, University of Padova, Via G.Gradenigo 6, 35131 Padova, Italy
2Dipartimento di Scienze della Terra, Università di Pisa, 56126 Pisa, Italy
3Core Research Laboratories, Department of Earth Science, The Natural History Museum, Cromwell Rd, London SW7 5BD, UK
4IAPS-INAF, Istituto Nazionale di Astrofisica e Planetologia Spaziali, Roma, Italy

*Corresponding author. E-mail: jacopo.nava@phd.unipd.it

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Abstract—TAM5.29 is an extraterrestrial dust grain, collected on the Transantarctic Mountains (TAM). Its mineralogy is dominated by an Fe-rich matrix composed of platy fayalitic olivines and clasts of andradite surrounded by diopside-jarosite mantles; chondrules are absent. TAM5.29 records a complex geological history with evidence of extensive thermal metamorphism in the presence of fluids at \( T < 300 \) °C. Alteration was terminated by an impact, resulting in shock melt veins and compaction-orientated foliation of olivine. A second episode of alteration at lower temperatures (\(<100 \) °C) occurred postimpact and is either parent body or terrestrial in origin and resulted in the formation of iddingsite. The lack of chondrules is explained by random subsampling of the parent body, with TAM5.29 representing a matrix-only fragment. On the basis of bulk chemical composition, mineralogy, and geological history TAM5.29 demonstrates affinities to the CVox group with a mineralogical assemblage in between the Allende-like and Bali-like subgroups (CVoxA and TAM5.29 are rich in andradite, magnetite, and FeNiS, but CVoxA lacks hydrated minerals, common in TAM5.29; conversely, CVoxB are rich in hydrated phyllosilicates but contain almost pure fayalite, not found in TAM5.29). In addition, TAM5.29 has a slightly different metasomatic history, in between the oxidized and reduced CV metamorphic grades while also recording higher oxidizing conditions as compared to the known CV chondrites. This study represents the third CV-like cosmic dust particle, containing a unique composition, mineralogy, and fabric, demonstrating variation in the thermal metamorphic history of the CV parent body(-ies).

INTRODUCTION

The CV (Vigarano-like) carbonaceous chondrites (CCs) are a group of primitive meteorites sharing approximately equal ratios of chondrules and matrix, as well as the highest abundances of refractory phases (CAIs [Ca-Al-rich inclusions] and AOs [ameboid olivine aggregate]) among any chondrite class (~10 vol%; McSween 1977; Brearley and Jones 1998; Weisberg et al. 2006). Chondrules in CVs are typically large (averaging 1 mm in diameter) type I, porphyritic olivine subtypes (~90%; Jones 2012) and often surrounded by thick (~400 μm) accretionary rims (King and King 1981; Tomeoka and Ohnishi 2010). Meanwhile, their matrices are rich in fayalitic olivine, Ca,Fe-pyroxenes, and andradite (Krot et al. 1995, 1998; Brearley and Jones 1998; Weisberg et al. 2006).

The CV chondrite group most likely represent members of a single parent body, derived from an asteroid-sized planetesimal. Thermal remnant magnetism studies reveal a single coherent magnetic field among constituent chondrules (Carporzen et al. 2011; Elkins-Tanton et al. 2011). This requires the presence of a partially differentiated structure with a molten core and
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silicate mantle overlain by a cold chondritic “lid” as described by the model of Weiss and Elkins-Tanton (2013) and Elkins-Tanton et al. (2011). Furthermore, isotopic signatures of CV chondrites ($\varepsilon^{54}$Cr versus $\varepsilon^{56}$Ti and $\varepsilon^{54}$Cr versus $\Delta^{17}$O) fall within the carbonaceous supergroup, requiring an outer solar system origin or late accretion history (Warren 2011). In addition to a partially differentiated interior, the CV chondritic “lid” attests to a protracted episode of parent body evolution with evidence for aqueous alteration (variously characterized by oxidation, hydration, and replacement [e.g., phyllosilicate-rich chondrule rims (Tomeoka and Tanimura 2000) or oxidized metal and mobilized Na and K (Krot et al. 2004)]) as well as recording the highest grade shock deformation among the CCs (commonly S2–S3 shock stages; Scott et al. 1992) and thermal metamorphism (~300 °C; Krot et al. 1995, 1998).

The CV class is divided into two subgroups: a reduced (CV$_{\text{red}}$) and an oxidized group (CV$_{\text{ox}}$)—dependent on their ratio of metal/magnetite and on the Ni content of sulfide phases (McSween 1977). The oxidized CVs are further classified into Bali-like (CV$_{\text{oxB}}$) and Allende-like (CV$_{\text{oxA}}$) subtypes, with the Bali-like population containing high abundances of hydrated minerals (up to 4.2 vol%; Howard et al. 2010) (including Fe-phyllosilicate) and nearly pure fayalite (Fa$_{90}$; e.g., Kaba: F$_{0100} = 20.9$ vol% and F$_{090} = 21$ vol%; Mokoia: F$_{0100} = 19.3$ vol% and F$_{090} = 13.3$ vol%; Howard et al. 2010), while the Allende-like population lacks hydrated minerals altogether and do not contain pure fayalite (Weisberg et al. 2006; MacPherson and Krot 2014). Further differences are observed in the abundance and speciation of secondary minerals formed by metasomatic alteration (MacPherson and Krot 2014). Fe-rich olivine and diopside-hedenbergite are found in all CVs. CV$_{\text{ox}}$ and CV$_{\text{red}}$ groups can be distinguished as CV$_{\text{ox}}$ (especially the Allende-like members) and contain nepheline, sodalite, diopside-diopside-hedenbergite, andradite, grossular, kirschsteinite, and phyllosilicates) is clear evidence of alteration from primary minerals held inside CAIs (i.e., meilite, anorthite, Al-Ti-diopside, hibonite, spinel, and perovskite; Krot et al. 1995) with dissolution of anorthite and albite in chondrule mesostasis releasing CaO and SiO$_2$ for secondary mineral growth (Krot et al. 1995).

Among micrometeorite collections, only two particles have previously been recognized as CV-like (Genge 2010; Van Ginneken et al. 2012), although neither were the subject of a detailed investigation into their parent body processing, a task that we discuss in this work with a more detailed characterization of the CV-like TAM5.29 particle. Furthermore, the classification of these samples as CV-like reveals some issues. The particle described by Van Ginneken et al. (2012) contains wüstite instead of Fe-Ni metal and the magnetite composition does not fall in the CV magnetite field. The sample described by Genge (2010) shows olivines with rounded morphologies and a high density of crystal defects, which are not typical characteristics of CVs. For this reason, Van Ginneken et al. (2012) and Genge (2010) also suggested an affinity with CO chondrites. In addition, TAM5.29 has andradite inclusions, diopside crystals, and oriented petrofabric of olivine that were not detected in the two other CV-like particles described by Genge (2010) and Van Ginneken et al. (2012), making TAM5.29 different and unique with an enhanced metasomatic history and with an undoubted metamorphosed CV-like mineralogy.

Micrometeorites from anhydrous CC groups are relatively rare, although up to 50% of the incoming micrometeorite flux have oxygen isotopic compositions related to the CO/CV/CK anhydrous supergroup (Suavet et al. 2010). Thus, the characterization of unique micrometeorites and those derived from rare parent bodies is crucial for our collective understanding of the near-Earth dust complex and the diversity of the asteroid belt.

Here, we provide a detailed characterization of the third CV-like cosmic dust particle with unique mineralogy and fabric, expanding knowledge in the compositional range of micrometeorites as well as investigating the thermal metamorphism and hydrothermal history of the CV parent body(-ies). Furthermore, given that micrometeorites originate from the asteroid belt (Genge et al. 2008; Van Ginneken et al. 2012) and cometary sources (Noguchi et al. 2015), the study of this sample can be a useful support to recent and upcoming space missions to C-type asteroids (e.g.,
NASA-Dawn mission at [1] Ceres, Hayabusa 2 at [162173] Ryugu, Osiris-Rex at [101955] Bennu) giving possible insights into the surface and subsurface composition and geological events. In particular, TAM5.29 with its particular metasomatic history adds knowledge on aqueous alteration and hydrothermalism, which are known to characterize C-type asteroids, and the relative products: secondary mineral formation in particular hydrous minerals like phyllosilicates, the formation of opaque phases, and postaccretion processing of organic matter, especially aliphatic and aromatic hydrocarbons.

METHODS

Sample Collection

TAM5.29 (~300 μm × ~600 μm) was recovered from Frontier Mountain (72°59'S–160°20'E; at ~2800 m above sea level and ~600 m above the local ice sheet surface), within Victoria Land, Antarctica. TAM5.29 was collected by the Italian Programma Nazionale delle Ricerche in Antartide (PNRA) during the 2003 expedition (Folco et al. 2008). Frontier Mountain consists of igneous rocks belonging to the Granite Harbour Igneous Complex. The top of the mountain is characterized by flat glacially eroded surfaces, created by an overriding ice sheet in the past (Folco et al. 2008). On these flat surfaces, numerous joints and weathering pits are found. These are filled with loose fine-grained bedrock detritus in which thousands of micrometeorites accumulated during the last ~1 to 2 million years along with a relatively small component of background terrestrial sediment (Folco et al. 2008).

Petrography and Major Element Analysis

Particle TAM5.29 was embedded in epoxy resin, sectioned, and polished. The resulting thick section was used for whole-section petrographic analysis using a field-emission scanning electron microscope (ESEM-FEG-STEM FEI Quanta 450), equipped with microanalytical EDS (energy dispersive spectrometry) Bruker, QUANTAX 400 XFlash Detector 6/10, which has a 129 eV spectral resolution and an intermediate size 10 mm² detector plate, capable of rapid data collection and semi-quantitative results at the Centro Interdipartimentale di Scienza e Ingegneria dei Materiali (CISIM), University of Pisa. All analyses were performed under high vacuum and at a fixed working distance of 10.0 mm (the optimal sample-to-pole-piece distance to maximize X-ray counts at the EDS detector on this instrument). Operating conditions are standardized for our lab and use an electron beam accelerating voltage of 15 kV and an unmonitored beam current. Spectra were acquired with an acquisition time of 30 s maintaining a dead time of approximately 10%. EDS data reported are therefore uncalibrated and standardless. Weight totals were determined using the Bruker’s “interactive oxides” and are quoted as weight normalized values in Tables S1 and S2 in supporting information. Elemental detection limits for this instrument are on the order of 0.2–0.5 wt%.

Mineral phases were identified by Raman measurements at the Department of Geosciences, University of Padova, using a Thermo Scientific DXR™ Raman Microscope using a 532 nm laser excitation source. Analyses were performed using a 50× and 100× long working distance objective with ~2.5 cm⁻¹ spectral resolution, ~1 μm spatial resolution, and 25 μm pinhole operating at a minimum of 1 mW to a maximum of 5 mW of power. Low power (1–5 mW) coupled with short exposure times of 3–4 s was essential to avoid damage to minerals and carbonaceous phases. To minimize noise, each spectrum was acquired 10–15 times. Spectra were recorded in the frequency range from 100 to 3500 cm⁻¹. Spectral fitting was carried out using the Thermo Scientific™ OMNIC™ Spectra Software. Rectangular areas were analyzed with the Raman point-by-point mapping technique—again each spectrum in the map was collected 10–15 times using an exposure time of 3–4 s and spectra were obtained from a grid of points spaced 2 μm along X-axis and 2 μm along Y-axis.

EBSD analysis was performed on a CamScan 2500 SEM (Department of Geosciences, University of Padova) equipped with a LaB6 source, a NordlysNano EBSD detector of Oxford Instruments and Channel 5.12 EBSD acquisition, and postprocessing software. The sample was for 1 h Syton polished to remove surface damage related to conventional diamond polishing, and was then carbon coated (few nanometers of thickness) to improve conductivity. Operation conditions were 25 mm working distance, 15 kV beam acceleration, and 10 nA probe current. Considering the small grain size of the olivine crystals, for EBSD mapping in automated mode, a 0.1 μm step size in X and Y directions was applied during acquisition of a 600 × 400 data grid. EBSD does not discriminate among solid solutions of olivine, garnet, and clinopyroxene (Cpx); therefore, forsterite and diopside reflector files of the HKL database were used to index olivine and Cpx, whereas andradite of the American Mineralogist Crystal Structure Database (Downs and Hall-Wallace 2003) was used to index andradite. Indexing of the EBSD patterns was accepted when at least six Kikuchi bands were detected. EBSD data were processed using the Oxford Instruments HKL software.
package Channel 5.12, generating crystallographic orientation maps, band contrast maps, phase maps, and pole figure plots after noise reduction. The latter was done by removing isolated misindexed data points using a wildspike correction, whereas all nonindexed points were infilled to a six nearest-neighbor crystallographic orientation by extrapolation. In the Channel 5.12 software package, grain detection in EBSD maps is based on crystallographic orientation, using a misorientation angle of at least 10° between two adjacent pixels to identify grain boundaries. Grain orientation data from the entire map were plotted onto lower hemisphere equal area projections as one point per grain to avoid grain size-related bias during contouring.

We determined bulk mineralogy through μXRD diffraction methods, using a Rigaku Rapid II microdiffraction system, equipped with a 2-D curved imaging plate detector, at the Department of Earth Science, the Natural History Museum in London. This was employed to collect in-plane diffraction pattern data from the TAM5.29 sample. A Cu X-ray source with an incident beam monochromator provided Kα radiation (1.5418 Å). This was collimated by a pinhole system to a beam spot of 100 μm. Analysis ran for 10 h, after which the 2-D diffraction image was converted to a 1-D XRD pattern following automated removal of the background signal and integration of the Debye rings. Data were collected on cross section samples using a constant θ angle of 16°20 and a rotating φ axis to maximize the number of crystallites and the randomness of their orientations in the X-ray beam. Low-angle (<16°20) diffraction peaks could not be collected because the polished resin block which holds the particles prevented the stage from rotating in the θ plane by <16°20. Peak positions in the converted 1D patterns were identified by comparison against a mineral reference database (PDF-4 database from ICDD).

EMPA spot analyses were carried out at the Department of Geosciences, University of Padova with a CAMECA SX50 instrument with five wavelength dispersive spectrometers. Data were acquired with a beam current of 20 nA, accelerating voltage of 20 kV, defocused beam diameter of 5 μm, and we used for each element an acquisition time of 10 s. Amelia plagioclase, olivine, orthoclase, diopside, sphalerite (blenda), synthetic MnTiO₃, and Cr₂O₃ standards were used for instrumental calibration. The Pouchou–Pichoir procedure (PAP), supplied by the manufacturer, was used for raw data reduction. Detection limits (wt%) are: Na₂O = 0.04; MgO = 0.02; Al₂O₃ = 0.02; SiO₂ = 0.3; SO₃ = 0.03; K₂O = 0.02; CaO = 0.02; TiO₂ = 0.02; Cr₂O₃ = 0.04; MnO = 0.04; FeO = 0.04. We also used EMPA spot analyses to obtain the bulk composition of TAM5.29 by averaging a grid of 122 randomly spaced EMPA analyses in a specific representative area chosen on the basis of EDX maps, especially to avoid areas with major terrestrial alteration. Supporting standardless EDS spot analyses were conducted on the FE-SEM at the University of Pisa.

Reflectance near-IR spectra were acquired at the laboratories at the IAPS-INAF (Istituto di Astrofisica e Planetologia Spaziali — Istituto Nazionale di Astrofisica). We used a microscope Micro-IR Hyperion 2000 FTIR Vertex Bruker®. The spectra were acquired in the spectral range between 1.3 and 22 μm (here we focused and report only the range between 1.5 and 4.2 μm), with an MCT detector. Infragold (LabSphere®) has been used to calibrate spectral reflectance. The spectra were acquired with a spectral resolution of 2 cm⁻¹ and an aperture on the sample of 150 × 150 μm.

RESULTS

Petrography and Mineral Chemistry

TAM5.29 is an ~300 μm × ~600 μm sized particle with a partial magnetite rim only found along the fusion crust. The melt layer is discontinuous and the texture is similar to the fusion crust found on meteorites (see Fig. 5A and Fig. S1 in supporting information). Voids are recognizable in the melt layer (see Fig. 5A). High-resolution FE-SEM imaging demonstrates that the particle is composed primarily of euhedral olivines, with widespread andradite inclusions surrounded by dark halos composed of intermixed pyroxene and jarosite, which also occur as alteration veins (Fig. 1A and 1B). These inclusions form lenses with an augen-like texture (Fig. 1A, 1B, and 1E). A few of these lenses show a distinct asymmetrical shape (Fig. 1B). Olivine is euhedral to subhedral with dimensions from a few micrometers up to ~10 μm. Andradite appears as subrounded crystals 5–10 μm in size, while the diopside and jarosite surrounding these are anhedral and difficult to distinguish from one another in backscattered electron images since they are small crystals (few micrometers) finely mixed and with similar grayscales. EDX spot analyses (Table S1) reveal fayalitic olivines, with heterogeneous grain compositions, ranging from Fa₄₂ to almost pure fayalite Fa₉₂. Rare crystals of forsterite are also present (Fo₆₆). Similarly, pyroxenes have heterogeneous grain compositions (Fs₁₅.₆ W₀₆.₄) Owing to the small grain size, pores, and limitations with the spatial resolution and the interaction volume of the electron microprobe, spot analyses on some phases suffer from beam overlap with adjacent hydrated minerals, oxides, and sulfides; thus, analyses shown in Table S1...
may reveal low weight totals (olivine 89 wt%; andradite 95 wt%). We therefore supported mineral identification with spatially resolved Raman and bulk μXRD measurements.

Raman data revealed pyroxenes to be diopside (Fig. 2) with sparsely distributed isolated enstatite grains. In Fig. 2, a Raman map shows the typical microtexture of this particle. Furthermore, an aggregate of spinel crystals (mean Cr/[Cr+Al] = 0.003 and Fe/[Fe+Mg] = 0.34 based on EMPA analyses) are present (Fig. 1C-10, Tables S1 and S2), these co-occur with micron-sized Fe-oxides (mainly magnetite, spinel, and
Only minor FeNi alloys are found (see Fig. 10 and Table S2) dispersed within the matrix (Fig. 1). Raman analysis together with EDS, EMPA, and EBSD suggests that a considerable portion of the olivine has been altered resulting in a mixture of fine-grained hydrated Mg-Fe-sulfur-rich minerals and minor carbonates (likely Fe-carbonates) as suggested by the 3.9 μm band in TAM5.29 IR spectra (see Fig. 11) that we identify as iddingsite (see the Fine-Grained Material section for a more detailed description). Fine-grained phases are found as weathering films on fayalite and also scattered throughout the whole particle. Given their cryptocrystalline nature, it is difficult to definitely resolve individual mineral phases; however, fibrous
The bulk composition of TAM5.29 compared to bulk composition of carbonaceous chondrites and average composition of unmelted fine-grained Antarctic micrometeorites (UMMs).

|                | TAM5.29 Mean | CV Mean | CM Mean | CO Mean | CI Mean | CK Mean | CR Mean | UMM Mean | CV matrix Mean |
|----------------|--------------|---------|---------|---------|---------|---------|---------|----------|----------------|
| Na₂O           | 0.26         | 0.46    | 0.47    | 1.13    | 0.42    | 0.31    | 0.44    | 0.46     |                |
| MgO            | 12.6         | 23.3    | 23.9    | 19.1    | 24.4    | 17.4    | 16.8    | 18.8     |                |
| Al₂O₃          | 4.93         | 4.02    | 2.55    | 2.02    | 2.98    | 2.31    | 4.28    | 2.49     |                |
| SiO₂           | 29.2         | 33.2    | 30.1    | 27.3    | 34.2    | 31.9    | 33.9    | 28.6     |                |
| P₂O₅           | n.d.         | 0.28    | 0.27    | 0.25    | 0.28    | 0.25    | 0.29    | 0.40     | n.d.          |
| SO₃            | 2.17         | 3.96    | 8.6     | 2.34    | 17.9    | 2.9     | 2.7     | 1.9      | 3.29          |
| K₂O            | 0.28         | 0.04    | 0.06    | 0.11    | 0.04    | 0.04    | 0.13    | 0.02     |                |
| CaO            | 1.41         | 3.50    | 1.96    | 2.26    | 1.71    | 2.39    | 1.91    | 0.72      | 0.85          |
| TiO₂           | 0.12         | 0.24    | 0.13    | 0.13    | 0.10    | 0.18    | 0.10    | 0.11      | 0.07          |
| Cr₂O₃          | 0.53         | 0.50    | 0.47    | 0.52    | 0.42    | 0.53    | 0.53    | 0.49      | 0.36          |
| MnO            | 0.15         | 0.19    | 0.24    | 0.23    | 0.28    | 0.18    | 0.22    | 0.23      | 0.23          |
| FeO            | 40.1         | 28.9    | 29.5    | 31.3    | 28.5    | 30.3    | 30.4    | 29.6      | 37.6          |
| NiO            | n.d.         | 1.37    | 1.19    | 1.66    | 1.13    | 1.44    | 1.72    | 0.37      | 1.65          |
| Total          | 91.8         | 100     | 96.5    | 99.9    | 100     | 100.2   | 89.9    | 89.4      | 94.5          |

n.d. = not determined.  
*From MetBase.  
*From Genge et al. (1997).

Fig. 3. Bulk composition of TAM5.29, average CVs (data from METDB database), average CV matrix, and average unmelted fine-grained Antarctic micrometeorites (UMMs; data from Genge et al. 1997) relative to the CI chondrites composition.

Phyllosilicates are clearly resolved as dark halos around andradite inclusions (Figs. 1A, 1B, and 1F). Micro-XRD revealed that these fibrous phyllosilicates are mainly antigorite and saponite (see Fig. 8).

The bulk composition of TAM5.29 shown in Table 1 is similar to that of CCs, being within two orders of magnitude of CI values, and similar to other unmelted fine-grained Antarctic micrometeorites (UMM; Fig. 3). However, TAM5.29 also demonstrates notable enrichment in Fe and depletion in Mg compared to the CCs, as also suggested by the 2.8 µm band (see Fig. 11) typical of Fe-rich hydrous phases (Takir et al. 2013). The aluminum content of TAM5.29 is similar to that of CV meteorites (Table 1 and Fig. 3) and is significantly higher than other CCs—consistent with the high concentrations of refractory elements found in CV and CK chondrites. Conversely, Ca and Ti are depleted compared to CVs. The spider diagram in Fig. 3 shows a strong enrichment in K in TAM5.29 compared to both CVs and other UMM. If we consider bulk composition of only the matrix of CVs (Table 1 and Fig. 3), we see that FeO-MgO-CaO-TiO₂ values are closer to those of TAM5.29. On the contrary, Al₂O₃ and Cr₂O₃ CVs matrix values are considerably different from the values of TAM5.29.
which are closer to the average CVs (considering also chondrules). Slight depletion of Na₂O in TAM5.29 compared to both CVs and CVs matrix is also found.

Carbon is also ubiquitous; Raman analyses (Figs. 2 and 6) identified characteristic “G” and “D” band peaks (located at ~1350 and ~1590 cm⁻¹, respectively) and associated with disordered carbonaceous phases (Bonal et al. 2006). Poorly graphitized carbon (PGC) occurs as tiny inclusions of 100–200 nm in fayalite and as thin films surrounding crystals (Krot et al. 1998; Abreu and Brearley 2011). It is interesting to note that Raman spectra of the various mineral phases (fayalite, andradite, diopside, jarosite, and fine-grained material) also have peaks around 2680 and 2930 cm⁻¹. These peaks are second-order peaks of C but also attest to the presence of OH as well as S-H and C-H functional groups within organic molecules. Sulfur is detected in all EMPA and EDS analyses, reflecting either S in jarosite, S-bearing organics, or S-rich phyllosilicates. IR reflectance spectra (see Fig. 11) also show the presence of organic matter (i.e., CH compounds, aromatic and aliphatic hydrocarbons) with the 3.3 and 3.4 μm bands; the broad band between 3.6 and 3.8 μm can be instead indicative of S-H compounds. IR spectra also indicate the presence of carbonates with the 3.9 μm can be instead indicative of S-H compounds. IR spectra also indicate the presence of carbonates with the 3.9 μm band (see Fig. 11).

Another important observation is a shock melt vein 136 μm in length and approximately 5 μm in thickness (Fig. 1D). This feature crosses the fayalitic groundmass and either deforms or displaces the primary features (an andradite-diopside vein; Fig. 1D). In addition, we observe a conjugate synthetic fracture and release band (Fig. 1D). The displacement of these features appears to show a dextral shear sense, although this interpretation remains uncertain owing to significant variations in the width of the jarosite/diopside band on either side of the melt vein. The lower margin of the linear feature (as seen in Fig. 1D) shows an abrupt compositional contact with the host groundmass, while the upper margin is transitional over approximately 5 μm. This feature is composed of a nanocrystalline or glassy matrix with a high porosity and hosting anhedral rounded small (<2 μm) olivine crystallites and minor Fe-Ni oxides.

Preferred Orientation of Olivine

Figure 4 reports an EBSD map data acquired on the same site shown in Fig. 1B where olivine crystals wrap around andradite-diopside/jarosite inclusions and form a foliation texture (Fig. 1B). To investigate whether the olivine crystals show a crystallographic preferred orientation (CPO), a detailed EBSD map (Fig. 4) was collected. In Fig. 4A, a phase map, olivine (red), andradite (green), and diopside (yellow), is shown. The dark part of the map is material that was not indexed. It is unlikely that the material between the olivine crystals has not been indexed because of potential polishing problems, since micro-Raman, SEM, EMPA, and micro-XRD analyses show the presence of fine-grained alteration products (see below).

The three pole figures (Fig. 4B), lower hemisphere equal area projections, are obtained considering one point per olivine grain. This has been done to avoid a bias of the data distribution due to grain size effects. Pole figures in Fig. 4B show the pertinent crystallographic orientation of every single olivine crystal. In these pole figures, a clear trend is recognized, with the [100] axis forming a maximum in the NW periphery, suggesting a NW-SE preferred orientation of this direction, hence orthogonally to the foliation trace. Conversely, the [010] and the [001] directions are dispersed along girdles consistent with the overall trend of the foliation trace. This implies that the [100] axis is a rotation axis, and that the differently elongated shape of the olivine crystals is the result of crystals having their [010] and [001] directions at systematically changing angles within the foliation plane.

Fine-Grained Material

With EDS, EMPA, and Raman analyses widespread fine-grained cryptocrystalline phases are found in the TAM5.29 dust grain matrix in addition to the previously described iddingsite. Raman analyses produce mixed spectra suggesting the presence of sulfides, probably related to the S-H bands as well as poorly crystalline hydrated phyllosilicates (see Fig. 7). The Raman spatial resolution was 1 μm (see the Methods section), evidence that the fine-grained material has a grain size below 1 μm. Micro-XRD data (see Figs. 9 and 10) are in agreement with this interpretation, revealing Fe-Ni sulfides, phyllosilicates, and possibly oxides and hydroxides. Carbonates are not clearly detected by Raman and micro-XRD, but their presence is inferred by the 3.9 μm band in the IR spectra (see Fig. 11).

DISCUSSION

TAM5.29: Micrometeorite or Meteorite?

TAM5.29 has a discontinuous melt layer with a texture similar to the fusion crust found on meteorites (Figs. 5A and S1). The presence of such a prominent melt layer is unusual for a micrometeorite exterior, which instead has a thin (~5 to 50 μm) double-layered igneous rim and magnetite rim (Genge 2006). This could imply that TAM5.29 is a fragment of a larger meteorite.
Although meteorite fusion crusts are highly variable in composition, texture, and thickness (Genge and Grady 1998), they are commonly >1000 µm (i.e., 10× thicker than the crust on TAM5.29, Fig. S1). In addition, the bulk composition of CV fusion crust is also markedly different from bulk composition of TAM5.29 (Table 1). As reported by Genge and Grady (1998), they have MgO = 25.65 wt%, SiO2 = 34.5 wt%, FeO = 30.46 wt%, CaO = 2.39 wt%, and Al2O3 = 3.34 wt%. On the contrary, TAM5.29 has major elements bulk composition (Table 1) of MgO = 12.6 wt%, SiO2 = 29.2 wt%, FeO = 40.1 wt%, CaO = 1.41 wt%, and Al2O3 = 4.93 wt%. Furthermore, because the melt layer on TAM5.29 is discontinuous, this requires that the particle fragmented while on the Earth’s surface, as evidenced by the lack of fusion features on the remainder of the particle perimeter. We estimate that the pre-atmospheric particle diameter – assuming a spherical shape (a simplistic approximation given fusion crusts can show protrusions) – was between 1300–2000 µm, approaching the size limit for micrometeorites. Thus, TAM5.29 may represent a different class of extraterrestrial material intermediate in size between micrometeorites (<2000 µm) and meteorites (~>1 cm). This size domain was previously described by Harvey and Maurette (1991), Kurat et al. (1994), and Folco and Rochette (2010) and termed “minimeteorites.”

The hypothesis that TAM5.29 was a small particle in space is further supported by its unique petrology (chondrule-free nature and mixed metamorphic history). Owing to their unique delivery mechanism (P-R drag, operating on grains <1 cm; Gonczi et al. 1982) cosmic dust samples a more diverse collection of asteroid parent bodies than their larger meteorite counterparts. Thus, because TAM5.29 represents a new lithology, this adds to our confidence that this sample is not simply a fragment of a larger meteorite broken off during atmospheric entry but instead existed as a grain of dust in interplanetary space derived from a parent body not currently delivering meteorites to the Earth and otherwise unsampled.
Finally, the preferential fragmentation of a heterogeneous material, with fractures following boundaries of chondrules and CAIs, may explain their absence in this minimeteorite, thereby reflecting the unrepresentative sampling (of coarse-grained features within parent body), as previously argued by Genge et al. (2008).

**Terrestrial Weathering**

Micrometeorites from the Transantarctic Mountains (TAM) often show moderate to significant terrestrial weathering. In particular, jarosite is a common product formed in subaerial Antarctic environments (Van Ginneken et al. 2016). The EDX map in Figs. 5C and 5D reveals co-occurrences of S and K, the main components of jarosite, concentrated along the particle border and in the fractures within TAM5.29. The effect of terrestrial weathering is, however, limited being confined to the fractures and margins. The EDX map in Fig. 5C reveals that high Ca concentrations correspond to andradite inclusions, unaffected by replacement. In addition, Mg and Fe-element maps (Fig. 5B) are well delineated, suggesting that these elements have not been leached by terrestrial fluids. It appears that TAM5.29 has therefore suffered limited weathering, retaining the majority of its pre-atmospheric mineralogy unaltered. Because the melt vein displaces a dark vein of diopside + jarosite (marked in Fig. 1D), we conclude that the shock event postdates the metasomatic event. However, it remains unclear whether iddingsite is also intersected by the melt vein or if alteration that created the iddingsite overprints the shock vein. We favor the latter option assuming that iddingsite formed by fluids released by hydrous minerals as a consequence of the pressure exerted by the impact (because olivine in TAM micrometeorites typically weathers to jarosite and not iddingsite; Van Ginneken et al. 2012). Later fractures instead cut the shock melt vein creating possible localized terrestrial alteration also on the shock melt vein.

**Petrogenesis—Record of Hydrothermal Environment on the Parent Body**

The mineralogy displayed by TAM5.29, primarily composed of fayalitic olivine plus Ca-Fe-rich pyroxenes, andradite, phyllosilicates, and sulfides as well as the close
matrix compositions demonstrate a clear affinity to the CV chondrite group. The mineralogy in TAM5.29 is, however, distinct from any currently reported CVs and also unique among reported micrometeorite studies.

TAM5.29’s most common mineral—fayalite—is associated with Fe-alkali-halogen metasomatism at temperatures <300 °C (Krot et al. 1995, 1998; Zolotov et al. 2006; MacPherson and Krot 2014). The preferred fayalite-growing mechanism in CVs is during thermal metamorphism from an amorphous precursor phase in the presence of fluids (Abreu and Brearley 2011) and Ca-Fe-rich phases are derived from the aqueous alteration of CAIs and anorthitic-albitic mesostasis (Krot et al. 1995).

Fe-rich olivine and diopside–hedenbergite assemblages are common in all CVs, while Ca-rich silicates (e.g., andradite) and Fe-Ni sulfides are almost absent in CVred, which instead contain kirschsteinite (MacPherson and Krot 2014). Nepheline and sodalite are typical of CVoxA and absent in CVoxB, which in turn contain phlogopite and saponite (Krot et al. 1998; MacPherson and Krot 2014). TAM5.29 is therefore consistent with the CVoxB group's mineralogy, having several andradite inclusions plus phyllosilicates. This metasomatic alteration is similar to the terrestrial serpentinitization–rodentization process (Python et al. 2007; Bach and Klein 2009). During serpentinitization at temperatures <385 °C, clinoxyroxene replaces tremolite and, at lower temperatures, below 275 °C, clinoxyroxene is in turn replaced by andradite (Python et al. 2007; Bach and Klein 2009).

In summary, diopside and andradite grow at the expense of serpentine and, at lower temperatures (<275 °C), andradite also replaces diopside (Python et al. 2007; Bach and Klein 2009). Tremolite and disordered biopyroboles have been found in Allende chondrules and their presence represents a peak metamorphic temperature below 340 °C (Brearley 1997). In TAM5.29, tremolite is not found. Furthermore, serpentine and diopside are small anhedral crystals that form dark halos around andradite that appear to consume them. This provides a temperature constraint on the formation of secondary phases in TAM5.29, between 275 and 250 °C, in agreement with temperatures estimated by Krot et al. (1995, 1998). Krot et al. (1998) based the temperature range mainly on the tensile strength. The CV chondrites’ tensile strength considered by Krot et al. (1998) is around 100 bars, which puts a temperature limit at ~310 °C for the existence of liquid water on the CV parent body. This limit of 310 °C is also supported by textural observations of CVs, O-isotopic composition, and thermodynamic analysis (Krot et al. 1995, 1998).

The bulk composition of TAM5.29 supports the Fe-alkali-halogen metasomatism hypothesis. Significant enrichment of Fe and K is observed compared to average CV bulk composition. The enrichment in Fe is indicative of highly oxidizing conditions. However, Fe value of CV matrix is comparable with the concentration of Fe in TAM5.29. This is probably due to the higher concentration of Fe-oxides (i.e., magnetite and alloys) in the matrix of CVs. Instead, K is easily leached from minerals by fluids. The high concentration of K is explained with jarosite formed by terrestrial alteration (even if bulk analyses were made in areas where K enrichment due to terrestrial alteration was as low as possible based on EDX maps). It is also possible that part of the K is representative of hydrothermal fluid circulation rich in alkali elements, even if this alone cannot account for the very high K abundance in TAM5.29. In fact, localized heating on C-type bodies (e.g., Ceres) has been proposed to be due to leaching and redeposition near the surface of long-lived radioisotopes like 40K, which decays in 40Ar (Castillo-Rogez et al. 2008). The Ca and Ti depletion compared to CVs is instead due to the absence of refractory inclusions in TAM5.29. However, Ca and Ti concentration of TAM5.29 is higher than that in the CV matrix. This enrichment is due to the higher concentration of Ca-rich minerals and Ti-oxides in TAM5.29, compared to the CV matrix, formed during metasomatism. In addition, Ca depletion relative to CVs can also be related to Ca leaching that formed carbonate veins (not sampled by TAM5.29) in the parent body, a feature commonly found in terrestrial serpentinitization processes (Python et al. 2007). Mg depletion in TAM5.29 is a consequence of the oxidizing conditions and Fe enrichment. Na is easily depleted during aqueous alteration and TAM5.29 does not have nepheline and sodalite, the main carriers of Na. Al in TAM5.29 is higher than CVs and CV matrix. This high Al content is tricky to explain but is likely given by Al in the phyllosilicates derived from leached CAIs.

However, under these environmental conditions, fayalite is not stable with andradite and Ni-sulfides (Krot et al. 1998; Zolotov et al. 2006; MacPherson and Krot 2014). Andradite is stable at higher temperatures than fayalite and the two phases can only coexist at equilibrium at T < 100 °C (Krot et al. 1998; MacPherson and Krot 2014). Temperature <100 °C has also been proposed for the formation of iddingsite in the Lafayette meteorite (Treiman et al. 1993). Raman spectra and EDS analyses of TAM5.29 also show that andradite retains a certain amount of water (Fig. 2 and Table S1); thus, the OH in andradite could represent a (SiO4)4− ← (O4H4)4+ substitution reaction for which low temperature formations are also expected within a low-pressure postmagmatic environment (Amthauer and Rossman 1998). However, structural (O4H4)4+ is
expected to have Raman peaks at \( \sim 3560 \text{ cm}^{-1} \) (Amthauer and Rossman 1998). In contrast, in TAM5.29 andradite Raman spectra show OH peaks at lower wavelengths between \( \sim 2680 \) and \( \sim 2930 \text{ cm}^{-1} \), implying the presence of nonstructural water (in addition to S-H and C-H bonds).

In support of a low temperature formation, the bulk composition of TAM5.29 shows considerable enrichment in Fe compared to all other CCs (Table 1 and Fig. 3). The majority of this Fe is held within the silicate minerals (rather than reduced Fe-Ni metal) suggesting highly oxidizing conditions at low temperatures.

Serpenitization may also occur at lower temperatures; but in this case, lizardite is also expected, which is not detected in TAM5.29. Instead, in TAM5.29, the presence of antigorite can only be formed by serpenitization at higher temperatures, in particular at \( \sim 310 \text{ °C} \) occurs the chrysotile breakdown to antigorite + brucite at low pressure (1 bar, which is also typical of minor bodies of the solar system; Wunder et al. 2001). In contrast, fayalite can form over a wide range of temperatures from 30 to 300 °C with W/R (water to rock ratio) between 0.06 and 0.2 (Zolotov et al. 2006) and thus does not provide significant temperature restrictions.

Thus, the alteration assemblage in TAM5.29 must have formed across a range of temperatures, with two distinct alteration periods with distinct environmental conditions. An initial alteration regime generated the fayalite, andradite, antigorite, and diopside at temperatures of \( \sim 275-250 \text{ °C} \) under oxidizing conditions. While, later, at a distinctly lower temperature (<100°C), the iddingsite formed. We conclude that fayalite-andradite-FeS-NiS assemblages derive from a retrograde metasomatism, inferred mainly by the andradite crystals consuming diopside and serpentine, at \( \sim 250 \text{ °C} \), suggested by the mineral assemblage typical of serpenitization. Conversely, the formation of iddingsite appears to be related to an independent low-\( T \) event that will be discussed later in the discussion.

### Origin of Preferred Orientation of Olivine

The CPOs seen in the EBSD map (Fig. 4) may at a first glance resemble the terrestrial “type D” fabric of olivines recognized in simple shear experiments by Bystricky et al. (2000). Olivine “Type D” fabric refers to high stress and water-poor conditions (Michibayashi et al. 2016), and Bystricky et al. (2000) associated this fabric with upper mantle conditions, for example, in regions of extensional deformation in major detachment zones near the crust–mantle boundary, or in subduction zones where hot mantle convects past the upper side of cold slabs. Such deformation regimes are inconsistent with that recorded by TAM5.29. We believe that pole figures (Fig. 4), showing a clear maximum in the [100] direction, while the [010] and [001] directions are scattered in two girdles, are describing a kind of olivine fiber texture that relate to an axially symmetric shortening regime in which the (100) plane of olivine orients orthogonally to the maximum compression direction (\( \sigma_1 \)) and relate to compaction rather than to shearing.

It is also interesting to note that the elongated olivine crystals in TAM5.29 are generally not associated with the elongated [100] direction, but are randomly associated with the [010] and [001] directions. Further examination reveals that the longest olivine crystals show clear alignment around andradite inclusions. These preferential CPOs concentrated locally around coarse-grained inclusions were previously described in CV chondrites where olivine crystals wrap around chondrules or dark inclusions (Watt et al. 2006; Forman et al. 2017). These areas, around chondrules in CVs and around andradite crystals in TAM5.29, were probably characterized by a higher porosity compared to the rest of the matrix and thus suffered a more significant compression and pore collapse during impact. Such a scenario results in heterogeneous strain distribution with significant heat production (~575 °C) and locally high stress at the sites or pore collapse while the surrounding low-porosity matrix retains a lower degree of compaction and andradite inclusions do not record any sign of compaction at all (Bland et al. 2014). The matrix therefore remains partially unaffected by the shock wave passage (Bland et al. 2014; Forman et al. 2017), while around andradite inclusions compaction is more enhanced with the creation of augen-like structures as predicted and observed in Allende by Bland et al. (2014). This process also produces a distinct heterogeneity in the CPOs (Forman et al. 2017), which is in agreement with our EBSD data.

Another intriguing fact is the dislocation of the diopside-jarosite vein in Fig. 1D. This vein has a dextral displacement. This kind of deformation is typically non-uniaxial and is in contrast to the theoretical uniaxial nature of the shock compaction. In support of the non-uniaxial deformation, there is also the asymmetric andradite-diopside inclusion (Fig. 1B), proof of a shear stress. It seems thus that the preferred orientation of olivine has been created in the instant when the impact occurred—a moment in which stress is uniaxial. Subsequent to the initial stage of the impact, simple shear deformations form radial to the impact area (Kenkmann et al. 2014). In this second stage, deformation has been accommodated along the shock
vein melt and around andradite, the two areas of TAM5.29 where weakness is higher. In particular, maximum strain occurred along the upper part of the shock vein melt where crystals were forced into the melt vein creating the transitional upper limit, the conjugate synthetic displacement, and the releasing band (Fig. 1D).

Alternatively, non-impact processes may have generated the preferred orientation of olivine seen in TAM5.29. These include: sintering, subgrain recrystallization, lithification, and gravitational compaction (Forman et al. 2016, 2017).

Sintering requires heating at temperatures higher than 360 °C over long (>1 million years) timescales to achieve recrystallization alignment (Gail et al. 2015). However, both the temperature and duration of heating are inconsistent with the formation conditions of TAM5.29, which require lower temperatures (~250 °C)—as determined by the secondary mineral assemblage and shorter durations—as determined by Raman data and olivine crystal morphology. Likewise, the petrofabrics in TAM5.29 cannot have been formed by plastic deformation processes such as subgrain rotation, recrystallization, and diffusion creep as these would result in significantly less elongation of olivine crystals and lower aspect ratios (Forman et al. 2017).

Petrofabrics formed by compression due to lithostatic forces cannot be ruled out, but it is less likely as also suggested by Watt et al. (2006), Forman et al. (2016, 2017), and Bland et al. (2014). This is because the pressure on small protoplanets is negligible, especially on porous water-rich carbonaceous chondritic parent bodies. For example, at the center of a 200 km diameter body, asteroid pressures are thought to reach a maximum of 1 MPa (corresponding to a depth of a few tens of meters on Earth), far too low for lithostatic compaction (Weidenschilling and Cuzzi 2006). Meanwhile, on larger Ceres-like bodies, pressure estimates vary between <0.2 GPa (Neumann et al. 2015) and 1220 MPa (Suttle et al. 2017). Even if sufficient pressures are possible, liberation would require an impact of an enormous magnitude, equivalent to the complete destruction of such a body. In both cases, we would expect a brecciated texture of the resulting asteroid's chunks. In TAM5.29, brecciated texture is not observed. However, TAM5.29 is a small fragment and does not necessarily sample an area with clast boundaries. For this reason, we cannot assume that TAM5.29 is not part of a breccia and we cannot completely rule out the possibility of lithostatic compression.

Finally, a gravitational compaction model, such as that proposed for Allende by Watt et al. (2006) may be possible. Here, a muddy outer layer of a parent body affected by sedimentary processes operating under microgravity result in the alignment of olivine. However, gravity is very low even on a Ceres-like body (Ceres gravity is ~0.28 m s⁻², around 1/35 Earth's gravity). The low gravity coupled with the chaotic [010] and [001] axis distribution of olivines in TAM5.29 (Fig. 4), bring us to prefer an impact-induced compaction (Gattacceca et al. 2005; Bland et al. 2014; Forman et al. 2016, 2017). Consequently, the process that most likely created the preferred orientation of olivines in TAM5.29 is impact compaction, and this is further supported by the presence of the shock melt vein within the micrometeorite.

**Carbon and PGC**

Carbon, which is widely distributed in TAM5.29, is a powerful index of metamorphic grade. In particular, the maturity of the organic matter is influenced by thermal metamorphism and can be used to establish petrologic types for individual meteorites (Bonal et al. 2006). The maturation grade of the organic matter can be determined by the study of the Raman D-band (~1350 cm⁻¹) and the G-band (~1580 cm⁻¹) peak parameters. In the least metamorphosed samples, the intensity of the G-band I_G is higher than the intensity of the D-band I_D, the opposite is seen in samples with higher metamorphic grade (Bonal et al. 2006). When the I_D/I_G ratio of the CV chondrites is compared to the FWHM (full width at half maximum) of the D-band (FWHD-D), two distinct groups are recognized (Fig. 6; Bonal et al. 2006). The oxidized CVs have the highest I_D/I_G ratio (1.05–1.55) and lowest FWHM-D (~60 to ~100 cm⁻¹; lower right, Fig. 6), they are known to have experienced the highest metamorphic grade (Bonal et al. 2006). In particular, these data show that Allende suffered the highest thermal metamorphism (Bonal et al. 2006). An exception to this is the meteorite Kaba, which is a CV_oxB that lies in the upper left group (Fig. 6) showing a minor metamorphic grade typical of the reduced CVs (Bonal et al. 2006). Analysis of the I_d and I_g bands from TAM5.29 (Fig. 7) shows an R1 ratio (I_d/I_g) > 1 and in some other cases I_d/I_g ≤ 1 (but never <1); thus, TAM5.29 does not clearly belong to either group. Instead, TAM5.29 values tend toward the most metamorphosed group (CV_ox) although several values also lie in the less metamorphosed group (Fig. 6). This is proof of the highly unequilibrated nature of this micrometeorite and suggests that TAM5.29 is transitional between the Kaba-like CVs and the more evolved Allende-like CVs. The reason for this unspecified petrologic type is, however, not clear. It is known that within the CV parent body, many different environmental conditions existed from oxidizing fluid-enriched locations to the reducing fluid-poor localities.
The TAM5.29 metamorphic grade is thus representative of a new lithology of the CV parent body that experienced more oxidizing conditions (resulting in significant Fe enrichment) with an incomplete thermal metamorphism terminated by the impact that also created the shock melt vein and preferred orientation of olivine.

**Origin of the Fine-Grained Material**

Iddingsite is a common alteration feature that affects olivine in terrestrial rocks and is also found as a native minor component in chondritic meteorites. Iddingsite forms as a weathering film and represents a complex mixture of secondary hydrated silicates as well.
as carbonates, sulfates halides, and oxides. Iddingsite compositions in TAM5.29 are difficult to interpret because of their very small grain size (<1 μm) and mixed phase composition, which in turn gives averaged data when analyzed with Raman, EDS, and EMPA.

Lee et al. (2015) described iddingsite in the Lafayette meteorite as an alteration sequence affecting olivine and augite concurrent with the formation of hydrous Fe-Mg-phyllosilicates. These newly formed phyllosilicates are then partially replaced by siderite. During the growth of siderite, Fe-oxides also begin to form (Abreu and Brearley 2011). The alteration sequence ends with saponite and other fibrous phyllosilicates replacing siderite. Tomeoka and Buseck (1985) described similar alteration features in the matrix of CM chondrites formed as an intergrowth of Fe-Ni-S-O phases and cronstedtite. Based on these findings, we looked for possible constituent minerals of iddingsite within TAM5.29.

Raman spectra of the fine-grained material of TAM5.29 (Fig. 8) show possible matches to mackinawite, cronstedtite, and chukanovite (from RRUFF database). The best match among these phases is the hydrated Fe-Ni sulfide mackinawite (the two peaks at 209 and 279 cm\(^{-1}\) of TAM5.29 are also well matched by troilite), although this lacks characteristic peaks around 525 and 888 cm\(^{-1}\). Mackinawite is a poorly crystalline precipitate formed by the reaction between HS\(^-\) and Fe (Lennie et al. 1997). In nature, mackinawite occurs as hydrothermal alteration product within serpentinitized peridotites and has also been reported in meteorites. Conditions of formation of mackinawite are in agreement with the conditions of formation of TAM5.29, making this phase a plausible candidate.

Phyllosilicates are also present. Although cronstedtite can be ruled out due to a lack of diagnostics within the Raman spectra, the μXRD data (Fig. 9) revealed the presence of saponite and antigorite in TAM5.29. Phlogopite and clay minerals may also be present but, since diffraction data could not be collected below 16°20 and clay minerals have their main peaks in this region, it is not possible to obtain further details about these phases. In addition, the μXRD data also suggested the presence of Fe-Ni sulfides (Fig. 9) (pentlandite in matrix olivine in Allende was reported by Brearley 1999) as well as tentative evidence of Fe-carbonates.

Collectively, these data imply that the fine-grained material in TAM5.29 is a mixture of fibrous phyllosilicates (antigorite, saponite, and possibly phlogopite-cronstedtite), Fe-Ni sulfides, and possibly Fe-oxy-hydroxides (Fig. 10) with a possibility of rare carbonates inferred from the 3.9 μm band in TAM5.29 IR spectra (Fig. 11). This mineralogy is in agreement with the final stage of alteration described by Lee et al. (2015) from the Lafayette meteorite, and demonstrating that TAM5.29 records a protracted episode of intense postimpact aqueous alteration. Fluids involved in the
formation of iddingsite may therefore derive from the partial dehydration of phyllosilicates (previously formed during metasomatism) liberated after the impact event.

In this scenario, iddingsite formation occurs after the interruption of metasomatic at lower temperatures and in agreement with iddingsite temperature of formation proposed by Treiman et al. (1993; <100 °C).

However, low-temperature aqueous alteration of olivine alone cannot explain the entire fine-grained mineral assemblage. Evidence also exists for the loss of CAIs. Greshake et al. (1996) reported within four CAIs, several crystals of periclase (MgO), rutile (TiO₂), calcium oxide (CaO), and corundum (Al₂O₃) inside, and at grain boundaries of the constituent minerals of the inclusions. These oxides have dimensions of 50–200 nm (most of them under 100 nm), a grain size similar to the fine-grained material of TAM5.29 (<1 μm). Since Ca-rich minerals (andradite and diopside) derive from alteration of CAIs and PRCs (plagioclase-rich chondrules; Krot et al. 2002), it is possible that TAM5.29 preserves some of these residual oxides as described by Greshake et al. (1996). CaO, Al₂O₃, and TiO₂ in fact match some of the strongest peaks in the μXRD pattern of TAM5.29 (Fig. 9). So, in addition to the previously listed minerals, we believe that residual Ca-Al-Ti oxides are present as relicts of the primary parent body CAIs and PRCs. However, there is no evidence of residual CAIs in TAM5.29. We thus infer that the residual CAIs and PRCs minerals were not in situ alteration but were mobilized by fluid circulation. In fact hibonite and spinel are two of the most resistant CAI crystals to metasomatic alteration and in strongly
altered CAIs, Al-Ti-diopside is replaced by ilmenite and phyllosilicates (Krot et al. 1995). Al$_2$O$_3$ may also be indicative of the presence of sericite, an alteration aggregate of fine-grained minerals such as illite, muscovite, and palagonite (Al- and K-rich minerals) formed by hydrothermal fluid circulation. Palagonite and allophane on Earth are also alteration products volcanic glasses and water interaction. TAM5.29 had some glass content before the metasomatic event (that largely created fayalite, see the Petrogenesis—Record of Hydrothermal Environment on the Parent Body section). It is possible that part of it was converted into palagonite-allophane. Allophane can also be enriched in Fe and Ti (Gérard et al. 2007). The Ti enrichment can explain the TiO$_2$ detection in TAM5.29. So, illite and palagonite-allophane can explain the presence of Al$_2$O$_3$ and TiO$_2$, but not the presence of MgO and CaO oxides. Furthermore, these minerals are also K-rich, which can explain part of the high K and Al bulk concentration of TAM5.29. The fine-grained material in TAM5.29 therefore requires two distinct alteration events and is derived from two different processes: Fe-Ni sulfides and oxides are residues of the metasomatic event and iddingsite components (such as saponite) are derived from weathering at lower temperature (<100 °C) in the presence of fluids released from hydrous minerals by an impact.

CONCLUSIONS

We document an unambiguous and unique micrometeorite from the CV chondrite group (a member of the CV$_{OX}$ family), thereby expanding our collective knowledge of micrometeorite parent body diversity.
Primary mineral phases of TAM5.29 are Fe-rich olivine, andradite, and Ca-Fe-rich pyroxenes plus carbonaceous matter containing OH, S-H, and C-H functional groups. Fayalite crystals grew during thermal metamorphism potentially from an amorphous precursor phase in the presence of fluids.

The fine-grained material is derived by two distinct alteration events. The metasomatic process created: Ni-Fe sulfides (e.g., mackinawite), Mg-Ca-Al-Ti oxides partly derived by residual CAIs constituents mobilized by fluids and partly by illite-palagonite-allophane created by aqueous alteration. Low-temperature alteration created: Mg-Fe-phyllosilicates (saponite and possibly phlogopite), possibly minor Fe carbonates and FeO-OH.

TAM5.29 mineralogy lies in between the CVoxA and CVoxB. CVoxA are rich in andradite, magnetite, and FeNiS like TAM5.29 but lacks high abundances of hydrated minerals, common in TAM5.29. Conversely, CVoxB are rich in hydrated phyllosilicates but contain almost pure fayalite not found in TAM5.29. TAM5.29 retains a mineralogical assemblage that might be a link between the CVoxA and CVoxB.

TAM5.29 retains a mineralogy dominated by thermal metamorphism products formed at ~275 to 250 °C within the presence of Fe-alkali-halogen-rich fluids and under highly oxidizing conditions resulting in significant Fe enrichment.

This may represent a newly described alteration environment on the CV parent body, similar to the conditions recorded by Allende-Axtell-Mokoia-Kaba, etc., but with differences. These differences are: higher oxidizing conditions, heterogeneous thermal metamorphism that shows different degrees of alteration within only one micrometeorite, and a different secondary alteration history enabled by a particular impact history.

This is the proof of an even more heterogeneous CV parent body(-ies) thus adding a unique sample to the known CV lithologies.

In conclusion, the hypothesis of formation of the TAM5.29 minimiteteorite may be divided into three main stages (Fig. 12):

- Stage one: metasomatism at ~275 to 250 °C with Fe-alkali-halogens-rich fluids occurred on the parent body.
- Stage two: the particle was involved in an impact that terminated the metamorphic event resulting in a strongly unequilibrated composition with cryptocrystalline and amorphous phases and generating a preferred orientation olivine petrofabric.
- Stage three: characterized by the formation of iddingsite at lower temperatures, possibly from fluid released by hydrated minerals during the impact.

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REFERENCES

Abreu N. M. and Brearley A. J. 2011. Deciphering the nebular and asteroidal record of silicates and organic material in the matrix of the reduced CV3 chondrite Vigarano. Meteoritics & Planetary Science 46:252–274.

Amthauer G. and Rossman G. R. 1998. The hydrous component in andradite garnet. American Mineralogist 83:835–840.

Bach W. and Klein F. 2009. The petrology of seafloor rodingites: Insights from geochemical reaction path modeling. Lithos 112:103–117.

Bland P. A., Collins G. S., Davidson T. M., Abreu N. M., Ciesla F. J., Muxworthy A. R., and Moore J. 2014. Pressure-temperature evolution of primordial Solar System solids during impact-induced compaction. Nature Communications 5:5451.

Bonal L., Quirico E., Bourot-Denise M., and Montagnac G. 2006. Determination of the petrologic type of CV3 chondrites by Raman spectroscopy of included organic matter. Geochimica et Cosmochimica Acta 70:1849–1863.

Brearley A. J. 1997. Disordered biopyriboles, amphibole, and talc in the Allende meteorite: Products of nebular or parent body aqueous alteration? Science 276:1103–1105.

Brearley A. J. 1999. Origin of graphitic carbon and pentlandite in matrix oligines in the Allende meteorite. Science 285:1380–1382.

Brearley A. J. and Jones R. H. 1998. Planetary materials. Reviews in Mineralogy and Geochemistry 36:3–1.

Bystricky M., Kunze K., Burlini M., and Burg J.-P. 2000. High shear strain of olivine aggregates: Rheological and seismic consequences. Science 290:1564–1567.

Carporzen L., Weiss B. P., Elkins-Tanton L. T., Shuster D. L., Ebel D., and Gattacceca J. 2011. Magnetic evidence for a partially differentiated carbonaceous chondrite parent body. Proceedings of the National Academy of Sciences 108:6386–6389. https://doi.org/10.1073/pnas.1017165108.
Ishiwatari A., Hawkins J. W., and Ji S. 2016. Natural olivine crystal-fabrics in the western Pacific convergence region: A new method to identify fabric type. Earth and Planetary Science Letters, 443:70–80.

Neumann W., Breuer D., and Spohn T. 2015. Modelling the internal structure of Ceres: Coupling of accretion with compaction by creep and implications for the water-rock differentiation. Astronomy & Astrophysics 584: A117.

Noguchi T., Ohashi N., Tsujimoto T., Mitsuohara T., Bradley J. P., Nakamura T., Tho S., Stephan T., Iwata N., and Imae N. 2015. Cometary dust in Antarctic ice and snow: Past and present chondritic porous micrometeorites preserved on the Earth’s surface. Earth and Planetary Science Letters 410:1–11.

Python M., Ceuleeneer G., Ishida Y., Barrat J. A., and Arai S. 2007. Oman diopsidites: A new lithology diagnostic of high-temperature hydrothermal circulation in mantle peridotite below oceanic spreading centres. Earth and Planetary Science Letters 255:289–305.

Scott E. R., Keil K., and Stoeffler D. 1992. Shock metamorphism of carbonaceous chondrites. Geochimica et Cosmochimica Acta 56:4281–4293. https://doi.org/10.1016/0016-7037(92)90268-n.

Suavet C., Alexandre A., Franchi I. A., Gattacceca J., Sonzogni C., Greenwood R. C., Folco L., and Rochette P. 2010. Identification of the parent bodies of micrometeorites with high-precision oxygen isotope ratios. Earth and Planetary Science Letters 293:313–320.

Suttle M. D., Genge M. J., and Russell S. S. 2017. Shock fabrics in fine-grained micrometeorites. Meteoritics & Planetary Science 52:2258–2274. https://doi.org/10.1111/maps.12927.

Takir D., Emery J. P., McSween H. Y. Jr, Hibbitts C. A., Clark R. N., Pearson N., and Wang A. 2013. Nature and degree of aqueous alteration in CM and CI carbonaceous chondrites. Meteoritics & Planetary Science 48:1618–1637.

Tomeoka K. and Buseck P. R. 1985. Indicators of aqueous alteration in CM carbonaceous chondrites: Microtextures of a layered mineral containing Fe, S, O and Ni. Geochimica et Cosmochimica Acta 49:2149–2163.

Tomeoka K. and Ohnishi I. 2010. Indicators of parent-body processes: Hydrated chondrules and fine-grained rims in the Mokoa CV3 carbonaceous chondrite. Geochimica et Cosmochimica Acta 74:4438–4453. https://doi.org/10.1016/j.gca.2010.04.058.

**SUPPORTING INFORMATION**

Additional supporting information may be found in the online version of this article:

**Fig. S1.** The melt layer on TAM5.29 developed during atmospheric entry and comparison against fusion crusts and igneous rims on micrometeorites.

**Table S1.** Oxide (wt%) EDS analyses of olivine, pyroxene, andradite, and spinel of the TAM5.29 micrometeorite.

**Table S2.** Oxide wt% EDS analyses of metal oxides of TAM5.29.