Al-free di-trioctahedral substitution in chlorite and a ferri-sudoite end-

member

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Abstract

A compilation of Fe^{3+} -bearing chlorite analyses is used 1) to investigate the Al-free ditrioctahedral (AFDT) substitution $2\text{Fe}^{3+} + \Box = 3(\text{Mg}, \text{Fe}^{2+})$ in chlorite and 2) to estimate the composition of a ferri-sudoite end-member (Si₃Al) [(Fe²⁺, Mg)₂ Fe³⁺₂ \Box Al) O₁₀(OH)₈ within the chlorite solid-solution domain. According to our observations, up to two Fe³⁺ cations might be allocated in the M2-M3 chlorite sites by the substitution AFDT, which does not involve Al. These unexpected observations were made possible by the development of μ XANES techniques allowing *in-situ* measurements of XFe³⁺ (Fe³⁺/(Fe²⁺+Fe³⁺)) in heterogeneous chlorite. Although, further studies are required to confirm the crystallographic position of Fe³⁺ and refine its ionic/magnetic behaviour in chlorite, it opens perspectives for developing new geothermometers.

Introduction



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Chlorite is an ubiquitous phyllosilicate composed of tetrahedral and octahedral layers according to the 14.2 Å 2:1:1 arrangement (e.g., Meunier, 2005). The tetrahedral sheet hosts trior tetravalent cations (Si⁴⁺, Al³⁺ or rarely Ti⁴⁺, Fe³⁺), while the octahedral one is composed of dior trivalent cations (usually Fe²⁺, Mg²⁺, Mn²⁺, Al³⁺ and Fe³⁺). The general formula of chlorite is: $R^{2+}_{6-x-3y}R^{3+}_{x+2y}\square_y$ (Si_{4-x} R^{3+}_x) O₁₀ (OH)₈, in which R^{2+} represents divalent cations, R^{3+} trivalent cations and \Box octahedral vacancies (e.g., Wiewiora and Weiss, 1990). According to Bailey (1988), the chlorite structure is formed of two types of tetrahedral crystallographic sites (denoted T1 and T2) and two octahedral ones (denoted M1, M2). Two other octahedral sites (M3 and M4) are located in the brucite interlayer. Al^{VI} or trivalent Fe³⁺ cations are preferentially located in the M4 site, while the divalent cations occupy the other sites (Bailey, 1988). There are three categories of chlorite in diagenetic and low to high-grade metamorphic rocks: tri-trioctahedral (sum of octahedral cations closed to 6 a.p.f.u), di-trioctahedral (5 a.p.f.u) and di-dioctahedral (4 a.p.f.u). The main chemical substitutions occurring in chlorite are: Tschermak (TK) $Si_{(IV)}^{4+}$ + $(Fe^{2+}, Mg^{2+})_{(VI)} = Al^{3+}_{(IV)} + Al^{3+}_{(VI)};$ ferromagnesian (FM) $Mg^{2+}_{(VI)} = Fe^{2+}_{(VI)}$ and di-trioctahedral (DT) $2Al^{3+}_{(VI)} + \Box_{(VI)} = 3(Mg^{2+}, Fe^{2+})_{(VI)}$. Two filling models are used in the literature for the distribution of cations on the crystallographic sites. The first model called "disorderly", considers that there is no preferred configuration (e.g., Inoue et al., 2009). The second model called "ordered" is constrained by specific rules for filling cations in the crystallographic sites such as preferred configuration and Fe - Mg equipartition (e.g., Vidal et al., 2001; 2005; 2006; Lanari et al., 2014a). These authors assume, following Holland et al., (1998), Al_{IV} in T2, vacancies in M1, Alvi in M4 then M1 and then eventually in M2-M3, Fe and Mg in M2-M3 then in M1. Following the equipartition assumption, X_{Mg} (Mg²⁺/(Mg²⁺+Fe²⁺)) is considered to be equal between M1 and M2-M3 sites. Several chlorite end-members are classically defined (Table 1): Al-free chlorite (Si₄) [Mg₆] O₁₀(OH), amesite (Si₂Al₂) [(Fe²⁺,Mg)₄ Al₂] O₁₀(OH)₈, clinochlore (Si₃Al) [Mg₅ Al)

 $O_{10}(OH)_8$, corundophilite (Si₂Al₂) [Mg₄Al₂] $O_{10}(OH)_8$, daphnite (Si₃Al) [Fe²⁺₅ Al] $O_{10}(OH)_8$, pyrophyllite-gibbsite (Si₄) [Al₄ \Box_2] $O_{10}(OH)_8$ and sudoite (Si₃Al) [(Fe²⁺,Mg)₂ \Box Al₃] $O_{10}(OH)_8$. These end-members do not include any ferric iron component.

For fifteen years, several authors have emphasised the importance of Fe^{3+} in chlorite, particularly for their use as a geothermometer (Laird, 1988; Vidal et al., 2006; Inoue et al, 2009, Bourdelle et al., 2013, Lanari et al., 2014a). An additional substitution: <u>Al-Fe³⁺</u> that is supposed to occur on the M4 site was defined but without defining a ferric chlorite end-member (Vidal et al., 2006). In this model, the incorporation of Fe^{3+} changes the chlorite composition and the activity of the other end-members. This choice of the M4 site is based on the results of Smyth et al. (1997) suggesting that Fe^{3+} in chlorite shows a strong preference for the M4 site where it substitutes for Al³⁺.

Several methods are available to measure Fe^{3+} in chlorite: chemical titration (Tchermak, 1891; Orcel, 1927; Hallimond et al., 1939; Brindley, 1951), Mössbauer (Pal et al., 1993; Aja and Dyar, 2000; Inoue et al., 2009), XPS (Inoue et al., 2010) and μ -XANES spectroscopies (Wilke et al., 2001; Vidal et al., 2006; Rigault, 2010; Trincal et al., 2015).

Chlorite data from literature

A compilation of almost two hundred published data of chlorite structural formulae containing Fe^{3+} is reported in Table 2. These chlorites come from a wide range of rocks from various localities. Some are of detrital origin; other one experienced diagenetic to metamorphic conditions as oscillatory zoning chlorites from the Pic-de-Port-Vieux outcrop (Figure 1; Trincal et al., 2015). In addition, 202 data from three localities were incorporated in this review (Inoue et al., 2009). Most of the selected chlorite contain (in a.p.f.u, see Table 2): 2.5 to 3.5 Si⁴⁺, 2 to 4

 Al^{3+} (Figure 2), 1.5 to 5.5 R^{2+} (Mg²⁺, Mn²⁺, Fe²⁺) (Figure 3) and less than 1 Fe³⁺ (Figure 4). These analyses are distributed between classical end-members and apparently follow the classical DT and TK substitutions (Figures 2, 3) or the Al-Fe³⁺ substitution (Figure 4).

In diagenetic to metamorphic rocks, the XFe^{3+} content of chlorite depends on the temperature and the oxygen fugacity that is controlled by the buffering assemblage, i.e. the bulk rock composition (Lanari et al. 2014a). This control is supported by the data from Inoue et al. (2009), which show relationships between sampling localities and chlorite structural formulae and oxidation state (Figures 2, 3 and 4). It is therefore critical to look at the variability of XFe^{3+} recorded by chlorite in one rock sample, for a given bulk rock composition. In such cases the observed variability of XFe^{3+} is caused by temperature variations, rather than by changes in the buffering assemblage (Lanari et al. 2014a; Trincal et al. 2015). Most of the spot analyses reported in the literature correspond to different geological environments (Table 2); it is thus very difficult to use them to evaluate the link between the Fe³⁺ behaviour and the evolution of a parameter such as pressure, temperature, f_{02} , etc.

The choice of the analytical method to estimate the XFe^{3+} in chlorite is essential. Chlorite Fe^{3+} measurements by Mössbauer spectroscopy or titration are relatively common but provide little constrains due to poor spatial resolution. Indeed, recent investigations using high-resolution mapping (e.g., deAndrade et al., 2006; Muñoz et al., 2006), demonstrated that chlorite is frequently zoned, recording strong compositional variability even at the crystal scale (e.g. Lanari et al., 2014b). Compositional zoning reflects time series of equilibrium conditions experienced by the rock. Thus in-situ micrometric analyses are required to ensure a precise determination of the Fe^{3+}/Fe_{Tot} of the successive growth zones, which is impossible using Mössbauer spectroscopy. μ -XANES XFe³⁺ spot analyses or mapping allow highlighting iron chemistry heterogeneity of

chlorite, which is critical to see the relations between formation temperature and crystal geometry (Vidal et al., 2006; Trincal et al., 2015).

µXANES data from zoned chlorites

Accurate in-situ μ XANES and electron microprobe analyses performed on zoned chlorite from the Sambagawa metamorphic belt in Japan (Vidal et al. 2006) and on oscillatory zoning chlorite from hydrothermal veins in the Pic-de-Port-Vieux, Pyrenees, Spain (Figure 1; Trincal et al., 2015) show variations in the composition of the Fe³⁺, Mg and Fe²⁺cation proportions, while that of Si and Al remain constant (Table 2, Figure 2). This trend is independently observed in both samples, which have different bulk rock compositions and experienced different metamorphic conditions. Indeed, chlorite from Japan comes from a blueschist facies metapelite made of K-white mica, garnet, albite, quartz and graphite (Vidal et al. 2006), while chlorite from Pyrenees crystallizes in a vein of a greenschist facies metapelite without garnet or graphite (Trincal et al., 2015). The absence of graphite is a good indicator of more oxidizing conditions. In both cases, XFe³⁺ increases as temperature decreases, as suggested by numerous authors (e.g. Lanari et al., 2014a; Vidal et al., 2016). Furthermore, from both cases, there is no significant change in the Al content of the zoned chlorite. This result excludes the hypothesis of DT, TK or Al-Fe³⁺ substitutions being at the origin of the XFe³⁺ variations.

Al-free di-trioctahedral substitution

Based on μ XANES analyses performed on zoned chlorite by Vidal et al. (2006) and Trincal et al. (2015), and in agreement with the literature data (Table 2), we demonstrate that a

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di/trioctahedral substitution implying Fe^{3+} can be used to explain the variations of XFe^{3+} in natural chlorite:

$$2Fe^{3+} + \Box = 3(Mg^{2+}, Fe^{2+})$$
(1)

As this substitution does not explicitly involve any Al cations, it is named Al-free ditrioctahedral substitution (AFDT); it must not be confused with Al-free chlorite end-member.

This substitution is similar to another couple suggested by Billaut et al. (2002) and based on electron microprobe and Mössbauer spectroscopic analyses on sudoite:

 $Fe^{3+} = Al_{(VI)}$ and $Fe^{2+} + Fe^{3+} + \Box = 3(Mg^{2+}).$ (2)

However, chlorite analyses from their study display both Al and X_{Mg} variations, which is not our case.

Chlorite compositions from Vidal et al. (2006) and Trincal et al. (2015) are plotted in a Fe^{3+} versus R^{2+} diagram (Figure 4). In this diagram, data from zoned chlorite are distributed along a line parallel to AFDT substitution vector. The AFDT substitution allows to define a theoretical Fe^{3+} -bearing chlorite end-member; namely ferri-sudoite ((Si₃Al) [(Fe²⁺, Mg)₂ Fe³⁺₂ \square Al) O₁₀(OH)₈) with two atoms of Fe³⁺ (Table 1). Ferri-sudoite can be separated into ferri-sudoite-Fe ((Si₃Al) [Fe²⁺₂ Fe³⁺₂ \square Al) O₁₀(OH)₈) and ferri-sudoite-Mg ((Si₃Al) [Mg₂ Fe³⁺₂ \square Al) O₁₀(OH)₈). As Fe³⁺ replaces Mg and Fe²⁺, it seems convenient to allocate up two atoms of Fe³⁺ to the M2-M3 sites (total multiplicity of 2) leaving Al in the M4 (Table 1).

According to several authors, Fe^{3+} cation seems limited to 1 a.p.f.u and is generally allocated to the M4 site following the Al-Fe³⁺ substitution. This last assertion is supported by Z-contrast images of HAADF-STEM and others studies (e.g. Smyth et al., 1997; Inoue & Kogure, 2016; Vidal *et al.*, 2016). Among data collected in the literature, only 6 analyses show Fe³⁺ content higher to 1 a.p.f.u (Orcel, 1927; Brindley and Youell, 1953; Malysheva et al., 1977;

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Goodman et Bain, 1979; Kodama et al., 1982; Singer et al., 2009 in Table 2). However, these analyses would actually match to other phyllosilicates, suggesting contamination issues. Although AFDT substitution suggests that it is possible, no sudoite with more than one Fe^{3+} atom has been yet discovered; it opens interesting research opportunities.

Conclusion

The compilation of Fe³⁺-bearing chlorite analyses made in this study, especially those acquired by μ XANES spectroscopy in zoned crystals, suggests the existence of a new chlorite substitution, the AFDT substitution that is characterized by $2Fe^{3+} + \Box = 3(Mg^{2+},Fe^{2+})$ and implies a new chlorite end-member: the ferri-sudoite (Si₃Al) [(Fe²⁺, Mg)₂ Fe³⁺₂ \Box Al) O₁₀(OH)₈ with two atoms of Fe³⁺. These results are not in agreement with previous studies that showed Al-Fe³⁺ substitution implying only one Fe³⁺. The absence of constrains on the Fe³⁺ crystallographic position together with the Fe³⁺ ionic properties in chlorite, require further studies to confirm the existence of the AFDT substitution and of the ferri-sudoite end-member, for example with accurate in-situ Fe³⁺ measurements in synthetic or diagenetic chlorite formed in different oxidizing environments. This study opens the door to further research developments that would have strong implications, for example in chlorite geothermometry.

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Figures and Tables

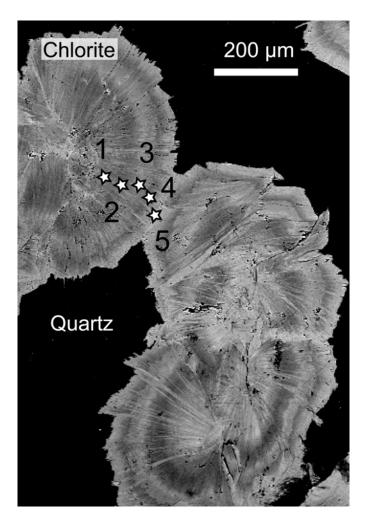


Figure 1 : Backscattered-electron image of an oscillatory zoning chlorite showing Al-Free Di-Trioctahedral substitution, from the Pic-de-Port-Vieux, Pyrenees, Spain. Numbers refer to structural formula reported in Table 2.

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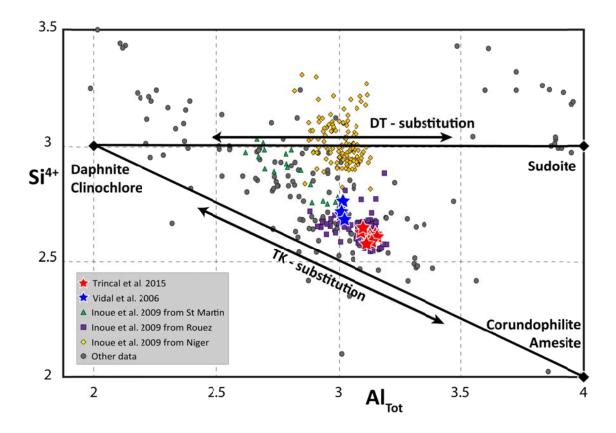


Figure 2: Compilation of the Fe³⁺-bearing chlorite compositions. The Si - Al plot is used to highlight the di-trioctahedral (DT) and Tschermak (TK) substitutions. End-members and structural formulas are reported in tables 1 and 2.

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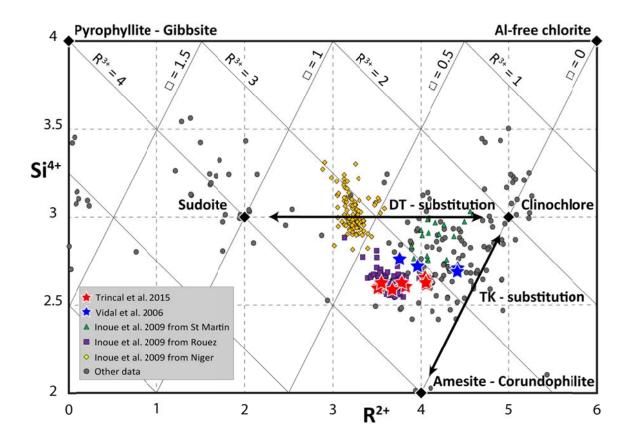


Figure 3: Compilation of Fe³⁺-bearing chlorite compositions. The Si⁴⁺ - R²⁺ plot by Wiewiora and Weiss (1990) is used to highlight the di-trioctahedral (DT), Al-Fe³⁺ and new AFDT substitutions.

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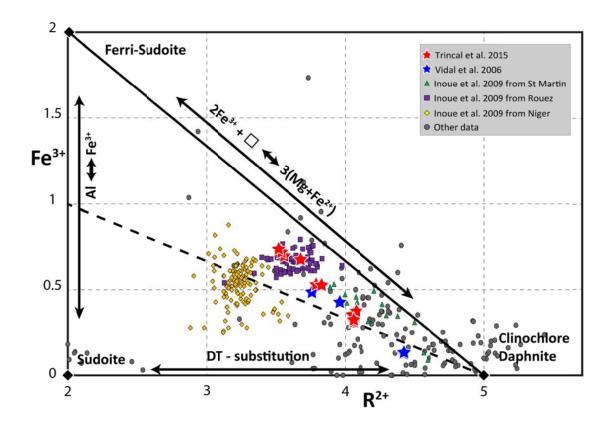


Figure 4: Compilation of the Fe³⁺-bearing chlorite compositions. The Fe³⁺ - R²⁺ plot highlights the di-trioctahedral (DT), Al-Fe³⁺ and new AFDT substitutions.

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Table 1 : Classical chlorite end-members and new ferri-sudoite end-member structural formula. The numbers in brackets refer to the quantity of atoms by structural formula distributed in tetrahedral (T) or octahedral (M) sites.

	T1 (2)	T2 (2)	M1 (1)	M2-M3 (4)	M4 (1)
Amesite-Mg / Corundophilite	Si,Si	Al,Al	Al	Mg,Mg,Mg,Mg	Al
Amesite-Fe	Si,Si	Al,Al	Al	Fe,Fe,Fe,Fe	Al
Clinochlore	Si,Si	Si <i>,</i> Al	Mg	Mg,Mg,Mg,Mg	Al
Chamosite/Daphnite	Si,Si	Si <i>,</i> Al	Fe	Fe,Fe,Fe,Fe	Al
Sudoite-Mg	Si,Si	Si <i>,</i> Al		Mg,Mg,Al,Al	Al
Sudoite-Fe	Si,Si	Si <i>,</i> Al		Fe,Fe,Al,Al	Al
Al-Free Chlorite	Si,Si	Si <i>,</i> Si	Mg	Mg, Mg, Mg, Mg	Mg
Pyrophyllite-Gibbsite	Si,Si	Si,Si		AI,AI,AI,□	Al
Ferri-Sudoite	Si,Si	Si,Al		$(Fe,Mg)_2,Fe^{3+},Fe^{3+}$	Al

Table 2: Fe^{3+} -bearing chlorites structural formula compilation, calculated on the basis of O = 14. Speciation of iron was obtained using chemical titration (CHE) or, since twenty years, using XANES (XAN), Mössbauer (MÖS) or XPS spectroscopies.

Cf. attached xls file

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