

THE GAMAYE PLUTON: A TECTONIC MARKER OF THE LATE-EBURNEAN TRANSPRESSIONAL OROGENIC EVOLUTION OF THE K'EDOUYOU- K'ENI'ÉBA INLIER (SOUTHERN WEST AFRICAN CRATON)

Aliou Dembele^{a,b,d,*} , Olivier Bolle^a, Sarah A. Friedman^c, Moussa Dabo^d, Mamadou Lamine Bouaré^e

^aDépartement de Géologie, Université de Liège, Allée du Six Août 12, 4000 Sart Tilman, Belgium

^bLaboratoire de Géodynamique et de Cartographie, Département de Géologie, Faculté des Sciences et Techniques, Université des Sciences, des Techniques et des Technologies de Bamako (USTTB), BP E3206, Bamako, Mali

^cSchool of Earth Environment and Sustainability, Georgia Southern University, GA, USA

^dDépartement de Géologie, Faculté des Sciences et Techniques, Université Cheikh Anta Diop de Dakar, BP 5005, Dakar-Fann, Sénégal

^eDépartement de Géologie et Mines, Ecole Nationale d'Ingénieurs Abderhamane Baba Touré (ENI-ABT), 410 Avenue du Général Moussa Traoré, BP 242, Bamako, Mali

* Corresponding author at: Département de Géologie, Université de Liège, Allée du Six Août 12, 4000 Sart Tilman, Belgium. E-mail addresses: aliou.dembele@uliege.be, aliou.m.dembele@gmail.com (A. Dembele).

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ABSTRACT

Fabric of the calc-alkaline, high-K Gamaye granitic pluton which crops out in the Paleoproterozoic Kédougou- Kéniéba inlier (southern West African craton) has been investigated by means of the anisotropy of magnetic susceptibility technique. Thermomagnetic curves and low values of the bulk magnetic susceptibility ($\leq 0.5 \times 10^{-3} \text{SI}$) reveal a paramagnetic signature, with biotite as the main magnetic carrier according to the granite petrography. A strain gradient is highlighted by field observations and a microstructural analysis, from an almost unmodified magmatic fabric in the north-east of the pluton to a mylonitic fabric in the south-west. Mylonitization occurred in the ca. N-S-striking, sinistral strike-slip Fekola shear zone, a high-strain zone which hosts a world-class gold mineralization, the Fekola deposit and is related to a late-Eburnean transpressional tectonic event, referred to as D₂. Magnetic foliations in the Gamaye pluton define a zone axis, parallel to a gently-plunging mean lineation and also concordant with the D₂ folding direction recorded in the host-rocks of the granite. A model of *syn*-D₂ emplacement and deformation of the Gamaye pluton in a partitioned transpressional system is proposed, with a synmagmatic, contraction-dominated deformation away from the Fekola shear zone and shearing along this high-strain zone down to lower temperature conditions. This model highlights the importance of regional strain partitioning in shaping parts of the southern West African craton during the late-Eburnean, transpressional evolution of this crustal domain.

1. Introduction

The West African craton (WAC) has recorded major Precambrian orogenic episodes, with distinct tectonometamorphic and magmatic events integrated into geodynamic models of the Paleoproterozoic Eburnean orogeny (ca. 2200–2060 Ma) that provided new insights into ancient orogenic systems (e.g. Grenholm, 2019; Grenholm et al., 2019; Traoré et al., 2022; Perret et al., 2025). Understanding such orogenic systems requires detailed tectonic reconstructions which can be done using for example granitoids which are powerful structural markers capable of recording regional deformations in the course of their crystallization (e.g. Clemens and Mawer, 1992). The generally ill-defined fabric of granitoids may in turn be accurately quantified using the anisotropy of magnetic susceptibility (AMS) technique (Bouchez, 1997, 2000), which has become widely applied for the structural mapping of granitic plutons and batholiths (e.g., Archanjo et al., 2002; Dawaï et al., 2017; Kleuser et al., 2021; Bolle et al., 2023; among many others), including in the WAC (Naba et al., 2004; Vegas et al., 2008; Yaméogo et al., 2023). When combined with a detailed microstructural study, AMS data enable in particular to decipher the respective role played by syn- to post-emplacement regional tectonics and igneous processes (magmatic flow or magma chamber dynamics) in the acquisition of a granitic fabric.

The focus of the present study is the Gamaye pluton, a representative of the post-collisional, late-Eburnean granitoids (Ndiaye et al., 1997; Dioh et al., 2006) emplaced in the Paleoproterozoic Kédougou-Kéniéba inlier (KKI), one of the westernmost unit of the southern WAC. The sedimentary, tectonometamorphic and magmatic evolutions of the KKI have recorded different phases of the Eburnean orogeny in the area (Bassot and Dommange, 1986; Ledru et al., 1991; Dia et al., 1997; Gueye et al., 2007; Lambert-Smith et al., 2016a; Masurel et al., 2017c; among many others). The KKI also hosts numerous Eburnean gold deposits, mainly near terrane boundaries generally defined by large-scale faults or shear zones like the Senegal-Mali shear zone (SMSZ) (e.g. Milési et al., 1992; Diene et al., 2015; Masurel et al., 2017c; Allibone et al., 2020; Diallo et al., 2024).

The Gamaye pluton crops out close to a terrane boundary marked by the SMSZ along almost its entire length and also in the immediate vicinity of a major gold mineralization, the Fekola deposit. It is a potentially key marker of the late-Eburnean tectonics generally considered to be transpressional (e.g. Dabo and Aïfa, 2010, 2011; Diene et al., 2012; Masurel et al., 2017c). However, it remains poorly studied: there are only preliminary structural, petrographic and geochemical data (Bassot and Caen-Vachette, 1984; Ndiaye et al., 1997; Lambert-Smith et al., 2016a), as well as a Rb-Sr date of 2045 ± 27 Ma on whole rocks and alkali feldspars interpreted as the emplacement age of the intrusion (Bassot and Caen-Vachette, 1984). An AMS study of the Gamaye pluton, supported by thermomagnetic experiments, and combined with field and microstructural observations is presented here. The main goal is to highlight the internal fabric of the pluton, and to investigate the structural relationships with the country rocks in the late-Eburnean tectonic framework. In particular, the structural and timing connections between the Gamaye pluton and a strike-slip shear zone hosting the Fekola gold deposit will be explored and envisaged at a broader scale in the late-Eburnean transpressional context of the southern WAC.

2. Geological setting

2.1. THE WEST AFRICAN CRATON

The WAC (Fig. 1a) is made of an Archean-Paleoproterozoic basement overlain by Mesoproterozoic to Phanerozoic sedimentary rocks accumulated mainly in the Taoudéni basin (e.g., Abouchami et al., 1990; Rocci et al., 1991; Boher et al., 1992; Feybesse et al., 2006a). The basement of the WAC crops out in the Reguibat shield in the north, the Léo-Man shield in the south, and the KKI and Kayes inlier in between. Both the Reguibat and Léo-Man shields consist of an Archean block to the west and a Paleoproterozoic domain to the east called the Kenéma- Man domain and the Baoule-Mossi domain for the Léo-Man shield, respectively (Bessoles, 1977; Rollinson, 2018).

The WAC has a polycyclic genesis (e.g. Ledru et al., 1991; Milési et al., 1992) which began with the creation of an Archean crust at 3.5–2.5 Ga, now cropping out as blocks alongside Paleoproterozoic domains (Kouamelan et al., 2015). These Paleoproterozoic domains correspond to an accretionary-collisional belt formed in the interval 2.5–2 Ga, during a diachronous event (Hirdes et al., 1996; Hirdes and Davis, 2002), the Eburnean orogeny (e.g. Baratoux et al., 2011; Ledru et al., 1991). In the Eburnean orogenic evolution, a distinction can be made between an early phase termed Pre-Eburnean or Eoeburnean and the main phase referred to as Eburnean *sensu stricto* (e.g. Grenholm et al., 2019). Perret et al. (2025) have recently defined six key time slices in the Paleoproterozoic evolution of the southern WAC (ca. 2350–2265 Ma, 2265–2200 Ma, 2200–2135 Ma, 2135–2120 Ma, 2120–2095 Ma and 2095–2060 Ma). The third time slice (2200–2135 Ma) and the next three (from 2135 to 2060 Ma) represent the Eoeburnean phase and the Eburnean *sensu stricto*, respectively (the last time slice, 2095–2060 Ma, being referred to as late-Eburnean; e.g. Masurel et al., 2022a). “Birimian” is commonly used instead of “Eburnean”, but this term does not refer to the same age range and may also designate lithologies, in particular in the Baoulé-Mossi domain (Abouchami et al., 1990; Milési et al., 1992).

2.2. THE KÉDOUGOU-KÉNIÉBA INLIER

2.2.1. MAIN SUBDIVISIONS

The KKI is the northern prolongation of the westernmost part of the Paleoproterozoic Baoulé-Mossi domain (Fig. 1a). It can be subdivided into four ca. N-S-trending lithological units, from west to east the Mako volcanic belt, and the Diale-Daléma, Falémé and Kofi series (Fig. 1b). These units are separated by two major discontinuities, the Main transcurrent zone (Ledru et al., 1991) and the SMSZ (Lawrence et al., 2013; this structure was first described as the Senegalo-Malian fault by Bassot and Dommange, 1986). According to the latter authors, the SMSZ extends over 150 km into both Malian and Senegalese territories, possibly reaching 300 km if considering its continuation to the north and south beneath the KKI sedimentary cover and its possible southern reappearance in a small Paleoproterozoic inlier in Guinea, and its width varies from about one hundred of meters in northern Mali up to 2–5 km to the south (or at least 10 km, as more recently suggested; Dabo and Aïfa, 2010;

Lambert-Smith et al., 2016a). The SMSZ has been described by various researchers as a brittle-ductile corridor made of a network of shear zones of varying importance, straddling at many places rather than strictly marking the boundary between the Falémé and Kofi series (Dabo and Aïfa, 2010, 2011; Diene et al., 2015; Masurel et al., 2017c). More recently, Allibone et al. (2020) and Lambert-Smith et al. (2020) noted that there is no major shear zone at the contact between the Falémé and the Kofi series in the Loulo mining district (Gara-Yaléa-Goukoto gold deposits) located at 25–50 km to the north–north-west of the Gamaye pluton, and concluded that the SMSZ was absent from this region, Allibone et al. (2020) suggesting moreover that it might also be the case elsewhere in the eastern KKI.

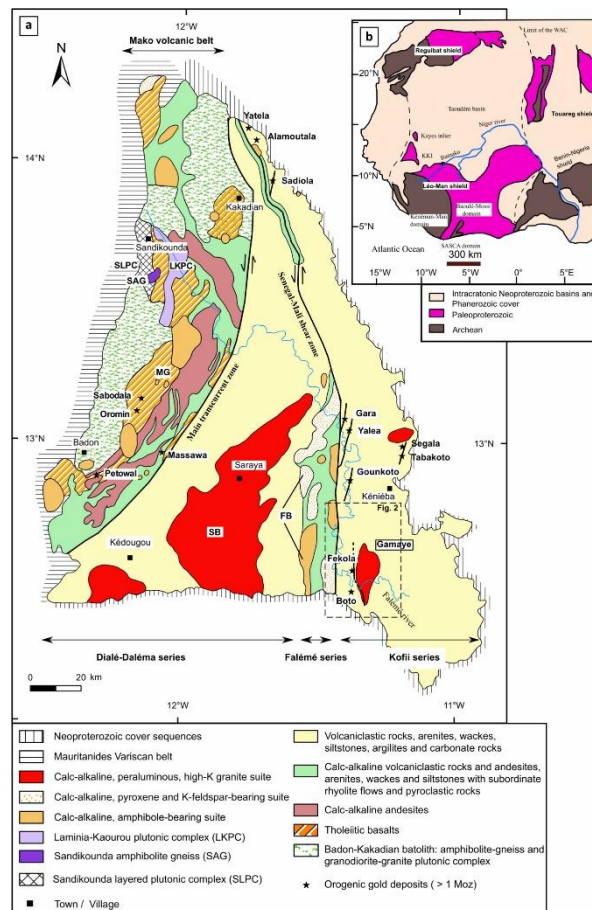
2.2.2. LITHOSTRATIGRAPHY AND MAGMATIC ACTIVITY

The Mako volcanic belt consists of tholeiitic basalts capped by calcalkaline andesitic lavas intercalated with sedimentary rocks, that have been intruded at various stages by plutons (Bassot, 1987; Dia et al., 1997; Dioh et al., 2006; Dabo et al., 2017). The Dialé-Daléma series, separated from the Mako volcanic belt by the Main transcurrent zone, the Kofi series, separated from the Dialé-Daléma and Falémé series by the SMSZ (except maybe locally, where this structure would be absent; Allibone et al., 2020; Lambert-Smith et al., 2020), and the Falémé series are composed of siliciclastic and volcanoclastic rocks with intercalated lava flows and minor carbonates, all these sequences being crosscut by igneous rocks, mostly granitoid plutons (Fig. 1b; Bassot, 1987, 1997; Lambert-Smith et al., 2016a).

The KKI has recorded a polycyclic magmatic activity during its Eburnean evolution (*sensu lato*), lasting from ca. 2210 to 2060 Ma (Dia et al., 1997; Gueye, 2001; Dioh et al., 2006; Feybesse et al., 2006d; Gueye et al., 2007, 2008; Masurel et al., 2017c). The oldest (Eoeburnean to slightly older) volcanic and igneous rocks are found in the Mako volcanic belt: these are the tholeiitic basalts and calc-alkaline andesites, as well as intrusions formerly considered as belonging to the same large plutonic complex (the Badon-Kakadian batholith; Bassot, 1966), namely the Sandikounda amphibolite gneiss, the Sandikounda layered plutonic complex and the Badon granodiorite which have magmatic emplacement/protolith crystallization ages of 2205 ± 15 Ma to 2194 ± 4 Ma, 2158 ± 8 Ma, and 2213 ± 3 and 2198 ± 2 Ma, respectively (Pb-Pb and U-Pb zircon dates; Dia et al., 1997; Gueye et al., 2007). A series of granitoids with mostly (grano)dioritic compositions, as well as volcanic rocks were emplaced throughout the KKI from ca. 2145 to 2100 Ma (Masurel et al., 2017c; and references therein), as exemplified by the oldest components of the Laminia-Kaourou plutonic complex in the Mako volcanic belt (Pb-Pb and U-Pb zircon ages of 2138 ± 6 , 2127 ± 6 Ma and 2105 ± 8 Ma; Dia et al., 1997). The youngest magmatic phase is a post-collisional, late-Eburnean event marked by the emplacement of large volumes of biotite-, amphibole- and pyroxene-bearing granitoids from ca. 2095 Ma to 2060 Ma, coeval with a felsic volcanism (e.g. Hirdes and Davis, 2002; Feybesse et al., 2006e; Lambert-Smith et al., 2016a; Masurel et al., 2017c). These granitoids are widespread in the KKI, being found in the Mako volcanic belt (e.g. the Mamakono granodiorite dated at 2076 ± 3 Ma; U-Pb zircon age; Hirdes and David, 2002), the Dialé- Daléma series (e.g. the Saraya batholith dated from 2079 ± 2 Ma to 2061 ± 15 Ma; U-Pb monazite and zircon ages; Hirdes and Davis, 2002; Delor et al., 2010b; Goujou et al., 2010), the Fal'eme series (e.g. the Falémé batholith dated from 2089 ± 9 Ma

to 2076 ± 25 Ma; U-Pb zircon ages; Hirdes and Davis, 2002; Lambert-Smith et al., 2016a; Allibone et al., 2020) and the Kofi series (e.g. the Gamaye pluton with a Rb-Sr age on whole-rocks and alkali feldspars of 2045 ± 27 Ma; Bassot and Caen- Vachette, 1984). These post-collisional granitoids are characterized by a calc-alkaline, mostly high- to medium-K, per- to metaluminous geochemical signature (Ndiaye et al., 1997; Dioh et al., 2006; Lambert- Smith et al., 2016a; Faye et al., 2023).

Fig. 1.



Regional geology. (a) Geological map of the Kedougou-Kéniéba inlier (after Gueye et al., 2007; Lambert-Smith et al., 2016a, b, c; Lawrence et al., 2013; Masurel et al., 2017c). Trace of the Senegal-Mali shear zone, reproduced here from the above references, is actually the original one of Bassot and Dommange (1986) and it should be revised according to recent mapping of some portions of this high-strain zone, for example by Diene et al. (2015) and Allibone et al. (2020) (see e.g. Fig. 2 of the present study). The Falémé river forms the international border with Senegal to the west and Mali to the east and north. Abbreviations refer to plutons mentioned in the text: LKPC = Laminia-Kaourou plutonic complex, Mg = Mamakono granodiorite, SAG = Sandikounda amphibolite gneiss, SLPC = Sandikounda layered plutonic complex (Mako volcanic belt); SB = Saraya batholith (Diale-Daléma series); FB = Falémée batholith (Falémé series). (b) Synthetic map showing the geological setting of the West African craton (modified after Boher et al., 1992; Perret et al., 2025).

2.2.3. TECTONO-METAMORPHIC EVOLUTION OF THE KKI

Several successive tectonic phases have been reported in the KKI (Ledru et al., 1989, 1991; Dabo and Aïfa, 2010, 2011; Diene et al., 2012; Masurel et al., 2017c; Allibone et al., 2020; Diallo et al., 2024), corresponding to two major periods in the tectonic evolution of the inlier (Masurel et al., 2022a, 2017c): an early period of contractional and thrusting tectonics accommodating orogen-

perpendicular E-W shortening and responsible for crustal thickening, and a subsequent period of transpressional tectonics accommodating oblique convergence. The first period, which corresponds for most authors to a single event called D_1 , may be subdivided in some regions into two successive phases (Masurel et al., 2017c). In the sedimentary rocks of the Dialé-Daléma, Falémé and Kofi series, the D_1 structural components are NNW- to, predominantly, NNE-trending overturned to upright folds (F_1) associated with an axial planar cleavage (S_1) cutting across or, commonly, subparallel to bedding (S_0) and an intersection lineation (Ledru et al., 1991; Dabo and Aïfa, 2010; Masurel et al., 2017c). The second tectonic period, here named D_2 in accordance with the classical terminology, is largely dominated by a sinistral transpressional phase. The D_2 event is mostly characterized by sinistral strike-slip movement along ca. N-S shear zones, either newly formed or reactivating inherited planar structures (sinistral transcurrent displacement along the SMSZ is particularly worth noting) and folding (F_2 folds with an axial planar cleavage S_2) (Ledru et al., 1991; Dabo and Aïfa, 2011; Diene et al., 2012). The S_2 cleavage/foliation plane commonly contains a predominantly shallowly plunging mineral/ stretching lineation (L_2). Example of a large-scale F_2 fold is found in the Dialé-Daléma series where a mega-antiform is occupied by the Saraya ' batholith (Bassot and Dommanget, 1986; Ledru et al., 1989, 1991), emplacement of this granitic body being coeval with the D_2 phase (Pons et al., 1992). An additional, late and minor transcurrent phase has been evidenced locally (Dabo and Aïfa, 2011; Diene et al., 2012; Masurel et al., 2017c). This phase, here called D_3 following the terminology of Dabo and Aïfa, (2011) and Diene et al. (2012), is associated with dextral strike-slip reactivation of older structures, formation of steep SW-NE to NW-SE dextral shear zones or faults and locally asymmetric (Z) folds. In that tectonic evolution, the deformation regime progressively changed from ductile to brittle. A similar polyphase tectonic evolution, i.e. contractional and thrust tectonics followed by transcurrent/transpressive movements, has been reported in other parts of the WAC (Milési et al., 1992; Feybesse et al., 2006a; Masurel et al., 2022a; Perret et al., 2025).

Emplacements of the Sandikounda layered plutonic complex at ca. 2160 Ma (2158 ± 8 Ma; Dia et al., 1997) and part of the Laminia-Kaourou plutonic complex at ca. 2130 Ma (2138 ± 6 Ma and 2127 ± 6 Ma; Dia et al., 1997) in the Mako volcanic belt, previously considered as having been accommodated by D_2 ductile shearing (Gueye et al., 2008), are actually coeval with the D_1 deformation phase, according to Masurel et al. (2017c). The latter also distinguished three groups of magmatic rocks in the Dialé-Daléma and Kofi series, at the northernmost tip of the KKI (Sadiola mining district, covering the Yatela, Alamoutala ' and Sadiola gold deposits), dated using zircon U-Pb geochronology: a pre- to $\text{syn-}D_1$ intrusion, the Finman dioritic to granodioritic complex dated at 2147 ± 8 and 2142 ± 6 Ma, various $\text{syn-}D_1$ igneous bodies (in particular, diorite, granodiorite and monzonite plutons) dated from 2118 ± 9 Ma to 2081 ± 3 and some $\text{syn-}D_2$ biotite (monzo)granites dated from 2074 ± 8 Ma to 2066 ± 8 Ma. The $\text{syn-}D_2$ Saraya batholith (Pons et al., 1992) and Fal'em'e batholith (Diene et al., 2015), which emplaced in the Dialé-Daléma series at ca. 2080–2060 Ma and in the Falémé series at ca. 2090–2075, respectively (Hirdes and Davis, 2002; Delor et al., 2010b; Goujou et al., 2010; Lambert-Smith et al., 2016a; Allibone et al., 2020), are other major time markers of the D_2 phase. Based on all these results, the lower limit of the D_1 deformation phase may be placed at ca. 2160 Ma (see also Masurel et al., 2017c) and the limit between the D_1 and D_2 phases, hence the switch from

contractional to transpressional tectonics in the KKI, may be put at ca. 2090 Ma (ca. 2085 Ma according to Allibone et al. 2020; and Masurel et al., 2022a). Absolute age constraints are not available for the D₃ deformation phase.

The D₂ time interval (from ca. 2085 Ma to, at least, 2060 Ma) coincides with the late-Eburnean period (ca. 2095–2060 Ma), whereas D₁ (ca. 2160–2085 Ma) straddles the Eoeburnean (ca. 2200–2135 Ma) and the early Eburnean (ca. 2135–2095 Ma), if we refer to the time slices of Perret et al. (2025). However, these authors stressed that these intervals should not be taken as a rigid time frame, because of the diachronous timing of the geological events in the southern WAC (for example, the onset of the Eburnean orogeny *sensu stricto* varies from ca. 2160–2150 Ma in the north-east to ca. 2120 Ma in the south-east and north-west; references in Grenholm et al., 2019). Moreover, the D₁ event, as envisaged here, may indeed be subdivided, at least locally, into two phases, as stated above, including a pre-Eburnean (Eoeburnean) period of early convergence (ca. 2160–2120 Ma; Masurel et al., 2017c).

A regional metamorphic phase under greenschist-facies conditions is testified, in the sedimentary and volcanic sequences of the KKI and the Baoulé-Mossi domain, by the widespread occurrence of metamorphic parageneses with minerals such as biotite, chlorite, sericite, epidote and albite (Ledru et al., 1991; Liégeois et al., 1991; Debat et al., 2003; Allibone et al., 2020). In the KKI, greenschist-facies metamorphism likely began during sediment burial (Feybesse et al., 2006c) and peaked during the D₁ and D₂ tectonic phases (S₁ and S₂ cleavages are defined by the shape-preferred orientation of greenschist-facies phyllosilicates; Ledru et al., 1991; Allibone et al., 2020). Greenschist-facies assemblages may also reflect “pervasive retrogression and hydrothermal overprint of older amphibolite-facies assemblages” (Masurel et al., 2017c). A static metamorphic phase, induced in particular by the emplacement of the late-Eburnean granitoids in the KKI (Feybesse et al., 2006c), caused local occurrence of amphibolite-facies mineral assemblages around the intrusions (Pawlig et al., 2006; Masurel et al., 2017b).

2.3. GOLD DEPOSITS

2.3.1. GOLD MINERALIZATIONS IN THE KKI

The WAC is well known for hosting gold mineralizations which are particularly abundant in the Baoulé-Mossi domain (including the KKI), the world’s premier Paleoproterozoic gold province (~10.000 t gold endowment; Masurel et al., 2022a). Re-Os dating of sulfides from gold deposits in the eastern part of the Baoulé-Mossi domain highlighted two major metallogenic episodes (Le Mignot, 2014): an Eoeburnean (2200–2120 Ma) episode characterized by low-grade mineralization disseminated in the country rocks or hosted in deformed quartz veins and a late-Eburnean (2050–2040 Ma) episode whose mineralizations are hosted by brittle structures. Thébaud et al. (2020), Masurel et al. (2022a) and Perret et al. (2025) proposed a roughly similar history, also for the Baoulé-Mossi domain (and the KKI), with three episodes of gold mineralization: a minor Eoeburnean phase (ca. 2200–2135 Ma), and two peaks of gold mineralization, respectively early Eburnean (ca. 2110–2095 Ma) and coeval with inversion of intra-orogenic basins, and late-Eburnean (ca. 2095–2060 Ma)

and coeval with transpressional tectonics and emplacement of the post-collisional granitoids. The Yatéla-Alamoutala-Sadiola (Sadiola district), Gara-Yaléa-Gouunkoto (Loulo district), Tabakoto-Segala and Fekola gold deposits, as mined in the Kofi series (Fig. 1b) were formed during the late-Eburnean mineralization phase. They are mostly fault or shear zone-controlled (Lawrence et al., 2013; Rhys, 2015; Treloar et al., 2015; Lambert-Smith et al., 2016b; Masurel et al., 2017a, 2017b).

2.3.2. THE FEKOLA GOLD DEPOSIT

The Fekola deposit is the southernmost mined gold mineralization of the Kofi series and also the nearest (~1.5 km) to the Gamaye pluton (Fig. 1b). The probable mineral reserves of this deposit totalize 1.79 Moz (50.75 t), comprising the open pit with 1.5 Moz (42.52 t) at 1.82 g/t and the stockpiles containing 0.22 Moz (6.24 t) at 0.93 g/t (B2gold, 2025). In the following, the Au mineralization is described using the technical report of Garagan et al. (2019), unless otherwise stated.

The Fekola gold deposit is elongated along a ca. N-S direction for about 3 km, with a maximum width of 200 m and a downward extension up to at least 440 m depth. The country rocks consist of various metasedimentary, mostly detrital rocks and diorite sills, all these lithologies being cut by quartz-feldspar porphyry dykes and breccia zones. The deposit in itself is hosted by a bedded siltstone-mudstone formation.

The host rocks of the gold deposit are foliated (cleavage) and folded, and cut by an anastomosing network of shear bands belonging to the Fekola shear zone (FSZ) (Rhys, 2015), a ~250 m thick and ca. N-S high-strain zone with a steep westerly dip, whose kinematics will be analyzed below. The Au mineralization is located within this shear zone and was formed mostly by fluids collected through the network of shear bands (Garagan et al., 2019; Brown, et al., 2023). The mineralization and its country rocks are also cut by late brittle structures, the most significant being the Fekola fault, a N-NW-striking, W-dipping fault, with a small offset and an apparent normal displacement (Garagan et al., 2015, 2019).

3. Material and methods

3.1. SAMPLING AND AMS MEASUREMENTS

Sampling according to a regular grid, with an mean density of one sample per km², as recommended for fabric analysis in granites (Bouchez, 2000), proved to be impossible here due to the scarcity of outcrops and the bad quality (weathered) of some of them. Our sampling grid is therefore irregular, implying that some portions of the Gamaye granite were not investigated, especially in the east and north-west of the pluton. Oriented cores were collected with a portable drill coupled with a generator on outcrops close to roads (22 sites) and oriented blocks were taken instead of cores on less accessible outcrops (32 sites). These oriented samples were then drilled and reoriented in the laboratory. In total, 54 sites were sampled, with two or three cores obtained per site. All these cores, 1 in. in diameter, were further cut into two or three 22-mm-high cylinders, providing four to six

cylinders per sampling site. The AMS measurements were conducted on these small cylinders ($n = 237$). AMS measurements were performed at the Laboratory of Petrology, Geochemistry and Petrophysics of the Liège University (Belgium), using a MFK2-FA Kappabridge susceptometer (Agico Inc.) characterized by a high sensitivity of 2×10^{-8} SI (976 Hz, 400 A/m). The AMS measurements provided, for each small cylinder, the length and orientation of the AMS ellipsoid principal axes ($K_1 \geq K_2 \geq K_3$) and were averaged for each sampling site. The length of the AMS principal axes enabled calculation of the bulk magnetic susceptibility $K_m = (K_1 + K_2 + K_3)/3$, as well as the corrected anisotropy degree, P_j , and shape parameter, T_j , of Jelínek (1981), describing respectively the strength and shape (prolateness or oblateness) of the magnetic fabric. The orientation of the AMS principal axes represents that of the magnetic fabric, with a magnetic foliation and a magnetic lineation defined by the K_1K_2 plane and the K_1 axis, respectively.

3.2. THERMOMAGNETIC MEASUREMENTS

Magnetic susceptibility versus temperature curves, intended to characterize the magnetic mineralogy, were measured on crushed scraps from five AMS samples (GM29, GM34, GM36, GM38 and GM54) selected to cover the range of facies types, grain sizes and K_m values. These thermomagnetic measurements were done using a CS-L Cryostat and a CS4 Furnace (Agico Inc.), for low- and high-T quantifications respectively, coupled with the MFK2-FA Kappabridge of the Liège University, following a standard Agico procedure, as detailed in Bolle et al. (2023).

4. Observations and results

4.1. FIELD OBSERVATIONS

4.1.1. THE HOST ROCKS OF THE GAMAYE PLUTON

The Gamaye pluton was emplaced in metasediments of the Kofi series, which form N-S- to NE-SW-trending units and have mostly sandstone, siltstone and argillite compositions (Figs. 2, 3a). The bedding (S_0) in all these metasediments is affected by a cleavage (S_1) defined by the shape-preferred orientation of sericite and chlorite. However, it is generally difficult to differentiate S_1 and S_0 , suggesting that bedding has been transposed and is now sub-parallel to the S_1 cleavage, forming a composite planar structure (S_{0-1}), as commonly observed for the oldest cleavage generation in the KKI metasediments (Ledru et al., 1991; Dabo and Aïfa, 2011; Masurel et al., 2017c). Hereafter, this S_{0-1} structure is called S_1 cleavage for the sake of simplicity. Some S_1 cleavages we have measured near the northern margin of the Gamaye pluton during the AMS field campaign and S_1 measurements made by Feybesse et al. (2006e) are compiled on an equal-area plot on Fig. 2 (rough data are available in Supplementary Material 2). This compilation indicates that S_1 is dominantly NNE-striking, with either W or E dips and a large range of dip values from 15° to 90° , poles displaying a girdle distribution on the equal-area plot, with a cylindrical best-fit axis of 195/02. This reflects folding of the S_1 cleavage around a NNE-SSW-trending mean axis, as a consequence of the D_2

deformation. In an argillite close to the northern tip of the Gamaye pluton, we have observed an outcrop-scale example of such a fold affecting S_1 (F_2 fold; Figs. 2, 3b-c): an upright, gently N-plunging antiformal anticline, with parasitic crenulations defining a NNE-striking axial-planar cleavage (S_2) and a lineation (L_2) oriented at 010/20, parallel to the fold axis (Fig. 3d). S_2 cleavages measured by Feybesse et al. (2006e) predominantly to the north of the pluton (Supplementary Material 2) are also mostly subvertical, with an average orientation (weighted mean) of N025-86WNW (Fig. 2). The metasediments are frequently affected by E-W-striking fractures (Fig. 3a), probably related to a late brittle event which could be referred as D_3 .

4.1.2. THE GAMAYE PLUTON

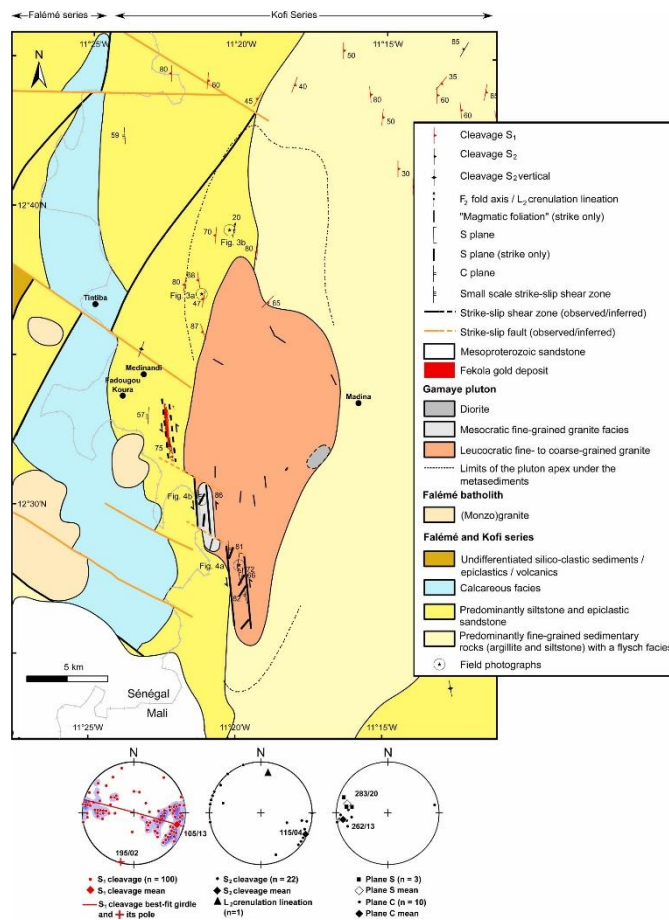
In map view, the Gamaye pluton is elongated along the Eburnean N-S direction, with dimensions of ~24 km x 9 km (Fig. 2). However, contact metamorphism, hydrothermalism and aeromagnetic data indicate that the pluton apex extends to the north and south under the metasediments and have a total length of ~33 km (Feybesse et al., 2006e), meanwhile a highly negative gravity anomaly associated with the pluton suggests higher dimensions at depth, ~40 km x 15 km (Diallo et al., 2020). Outcrops are relatively rare, especially in the eastern and north-western parts, due to a thick lateritic cover, the best outcrops being found near permanent or temporary streams. Large outcrops occur also along the Falémé river and its affluents which cut across the pluton southern part, but the granite is very weathered there.

Our field investigations revealed the presence of two main petrographic facies in the Gamaye pluton. The dominant facies, present throughout the intrusion, is a leucocratic biotite granite whose granulometry progressively increases from north (fine-grained granite) to south (medium-grained granite, locally porphyritic with alkali feldspar megacrysts). Such a southward increase of the grain size was already noticed by Bassot and Caen-Vachette (1984) and Ndiaye et al. (1997). However, in both studies the northern, finer-grained facies was also described as being darker, which we do not confirm. The second main, however less abundant facies is a fine- to very-fine-grained, mesocratic biotite granite which forms a N-S elongated body (~5 km x 3.5 km) close to the western margin (Fig. 2). A macroscopically similar mesocratic facies crops out sporadically (total outcrop surface of ~22,000 m²) in the east of the pluton and a small elongated body (~1 km x 2 km) made of a dioritic facies is present along the eastern pluton margin (Fig. 2). We have not observed the contact between these different minor facies and the dominant leucocratic granite, except for the fine-grained mesocratic facies found locally in the east (sharp contact). Mafic microgranular enclaves (MMEs) are abundant near the SW edge of the pluton, both in the fine-grained mesocratic facies and the coarser-grained leucocratic granite. Fine-grained, felsic enclaves of unknown affinity may also occur very locally, in the leucocratic granite. The different granitic facies of the Gamaye pluton are cross-cut by aplite and tourmaline-bearing pegmatite dykes, as well as quartz-tourmaline veins.

A foliation defined by the shape-preferred orientation (SPO) of the rock-forming minerals, mostly feldspars and biotite, and locally underlined by flattening of the MMEs is visible on many outcrops. In most cases, this mesoscopic foliation is ill-defined and has a magmatic aspect. Only its strike has been measured with confidence (Fig. 2): it is ca. N-S, hence parallel to the pluton elongation, except in the north (NW-SE to WNW-ESE).

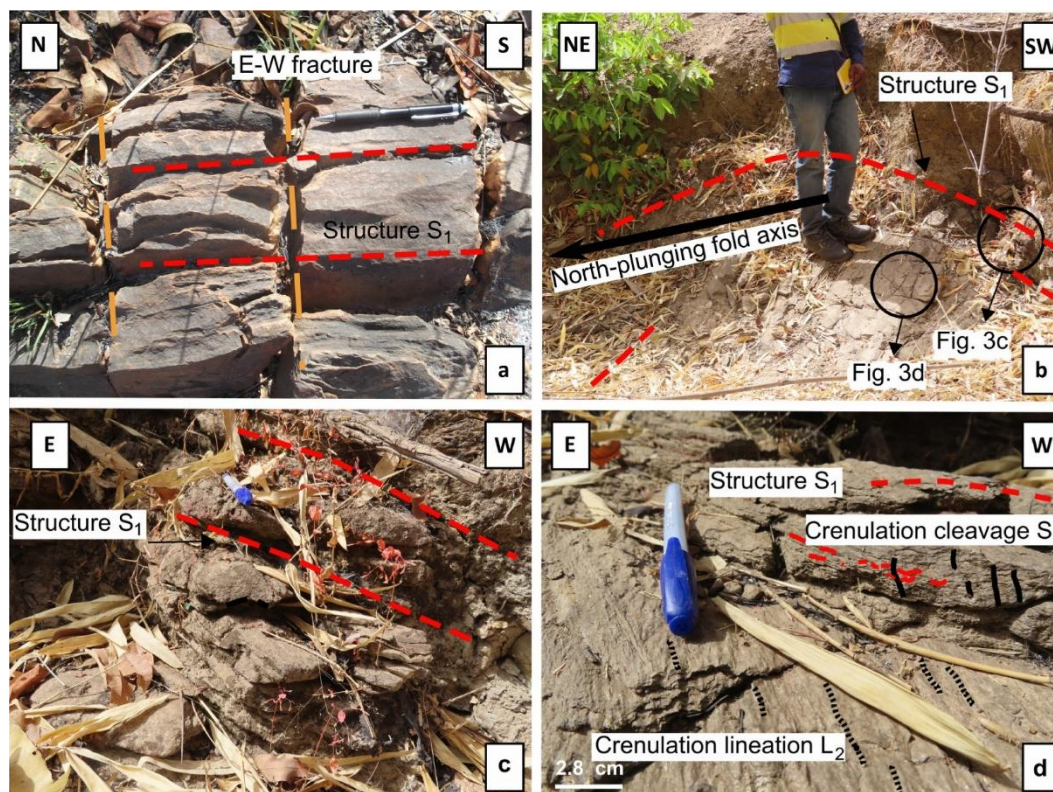
The leucocratic and mesocratic granitic facies are both mylonitized near the SW border of the pluton. In this high-strain zone, hereafter called the Mylonitic Zone, S/C fabrics are common (Figs. 2 and 4), with (1) a mylonitic foliation (or schistosity, S) delineated by the SPO of feldspar porphyroclasts and, locally, flattened MMEs and (2) shear bands (C) defined by clusters of biotite (Fig. 4). The S and C planes are steeply-dipping and ca. N-S-striking, with mean orientations of N013-70ESE and N172-77E, respectively (Fig. 2). On flat-lying outcrops, the S and C planes are systematically better defined than on steep surfaces (Fig. 4a), suggesting they bear a subhorizontal stretching lineation and the relationships between the two planes, as well as the shape of asymmetric objects (e.g. deformed veins) indicate an apparent sinistral sense of movement (Fig. 4b, c-d and e-f). The Mylonitic Zone is located to the south of and has the same orientation as the FSZ, and it materializes therefore most certainly the southern extension of this high-strain zone which hosts the Fekola gold deposit ~1.5 km to the west of the Gamaye pluton (Fig. 2). However, the Mylonitic Zone forms two segments which are offset from each other and which are also not aligned with the FSZ. These offsets suggest displacements along late E-W or ESE- WNW faults with an apparent sinistral strike-slip component. An array of such faults is indeed present in the area, as mapped on the Senegalese territory just to the west of the Falémé river (Delor et al., 2010a) and has been added on Fig. 2.

Fig. 2.



Geological map of the Gamaye pluton and surrounding area (western part, after Diene et al., 2015; Fekola gold deposit and associated structures, from Garagan et al., 2019; and B2Gold, unpublished data; eastern part, after Feybesse et al., 2006b; Gamaye pluton and structural measurements in the metasediments close to the northern tip of the granite, after observations of the authors; limits of the pluton apex under the metasediments, after Feybesse et al., 2006b; array of ESE-WNW faults with an apparent sinistral strike-slip component of movement to the south-west of the pluton, from Delor et al., 2010a). Note that the Senegal-Mali shear zone, partly visible to the west of the Gamaye pluton and to which the Fekola shear zone possibly belongs (see text), consists of a network of shears which cut across or follows the contact between the Falemé and Kofi series. Localization of the field photographs displayed in Figs. 3 and 4 is shown. Color codes for the structural elements: red for D₁-, black for D₂- and orange for D₃-related structures (application of this color code to the shear zones and faults to the west of the Gamaye pluton is highly hypothetical). Equal-area plots (lower hemispheres) show projections of various structures, with average orientations (weighted means) (data from the authors and/or from literature when stated, compiled in Supplementary Material 2): S₁ cleavage poles (mainly from Feybesse et al. 2006e); S₂ cleavage poles (Feybesse et al., 2006e) and one L₂ crenulation lineation; all in metasediments of the Kofi series, mostly to the north of the Gamaye pluton; S and C planes defining S/C fabrics in the Mylonitic Zone.

Fig. 3.



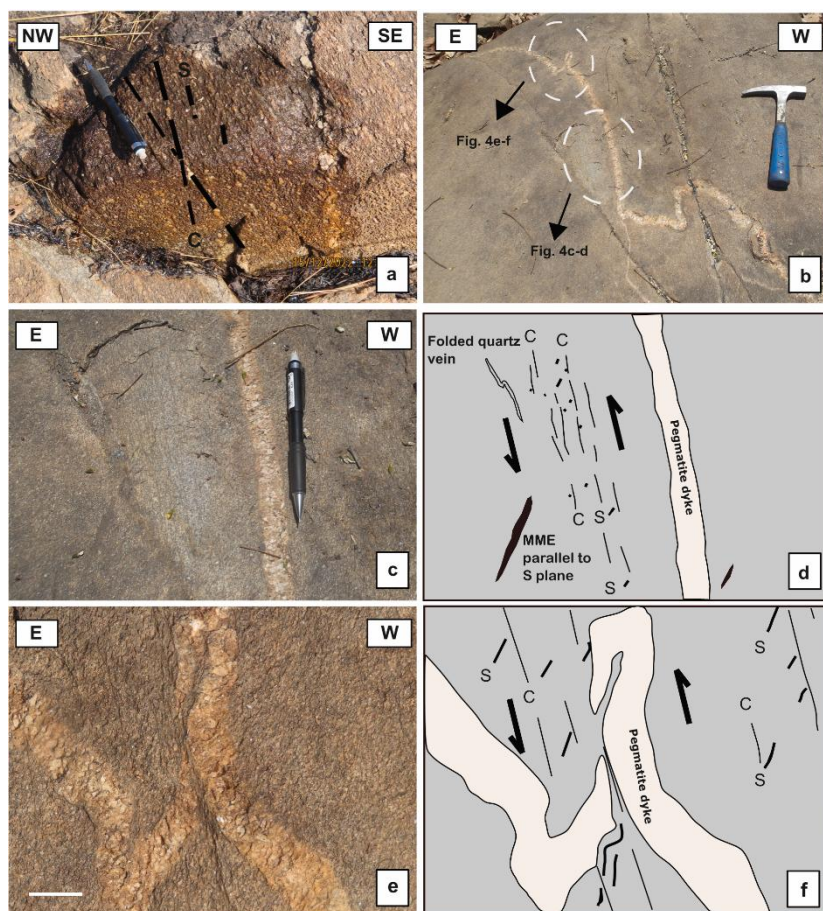
Field photographs of metasediments from the Kofi series to the north of the Gamaye pluton, near its margin. (a) Sandstone bed cut by E-W-striking late fractures (orange dashed lines); (b) S₁ cleavage in an argillite, deformed into an up-right antiformal anticline (F₂ fold) (looking south; insets show the location of Fig. 3c and d); (c) western limb of the F₂ fold shown in Fig. 3b; (d) crenulation cleavage (S₂) and lineation (L₂) in the western limb of the F₂ fold shown in Fig. 3b. Coordinates of the photographs (WGS 84 / UTM zone 29N): (a) 1396038N, 244244E; (b, c, d) 1400134N, 246266E.

4.2. PETROGRAPHY AND MICROSTRUCTURES

4.2.1. ROCK FACIES

The dominant, leucocratic granitic facies is composed mainly of quartz (grain size 0.4–2 mm), alkali feldspar (either microcline with its typical tartan twinning or orthoclase with Carlsbad twinning, 0.2–4 mm up to 14 mm), plagioclase (generally oligoclase, 0.8–2.8 mm) and biotite, with apatite, zircon, rutile, allanite, hornblende and opaque phases (mostly ilmenite; Supplementary Material 1) as accessory minerals. The texture ranges from heterogranular to locally porphyritic in the southern part of the pluton. Myrmekites are frequent. The western mesocratic facies has a similar mineralogy, however, it contains more biotite and it has titanite as an additional accessory phase. Its grain size is also smaller, 0.8–1 mm for quartz, ~0.5 mm for alkali feldspar and ~0.8 mm for plagioclase. The modal composition determined by point counting in two selected thin sections, one for each of the main facies, plot along the boundary between (syeno)granite and alkali feldspar granite in a QAP diagram (Streckeisen, 1976). There is no significant alteration that could influence the AMS and thermomagnetic measurements: saussuritization of plagioclase cores and chloritization of biotite grains are observed only locally and are both of weak intensity (Supplementary Material 1).

Fig. 4.



The Mylonitic Zone (S and C planes are in black, with thick dashed lines and thin dashed lines, respectively). (a) S/C fabric in mylonitized leucocratic granite (the outcrop is horizontal in the upper part of the photograph and vertical in the lower part; the S plane is oriented at N010-60E and the C plane at N170-70E; (b) mylonitized mesocratic granite with a S/C fabric and a pegmatite dyke affected by ptygmatic folds (the outcrop is subhorizontal); (c) is a detail of part (b) of the figure and (d) its interpretation, where it can be seen that orientation of the S plane with respect to the C plane and the asymmetry of a small folded quartz vein are both consistent with apparent sinistral shearing (note also a flattened mafic microgranular enclave, MME, parallel to the S plane –another enclave of this type is visible in part (b) of the figure, to the right of the hammer); (e) is another detail of part (b) and (f) its interpretation, showing an apparent sinistral offset of the folded pegmatite dyke. Coordinates of the photographs (WGS 84 / UTM zone 29N): (a) 1379284N, 246550E; (b, c-d, e-f) 1383337N, 244228E

4.2.2. MICROSTRUCTURES

We have classified microstructures in the Gamaye pluton into three main types, using microscale evidence of solid-state deformation, recovery and recrystallization, as observed in quartz, feldspars and biotite (Fig. 5). The main criteria used in this classification are detailed in Table 1. Distribution of the microstructural types, as seen in map view, is presented in Fig. 6.

Microstructures Ma (low-strain-high-T solid-state deformation). This microstructure type is characterized by a random distribution of the main rock-forming minerals, feldspars, quartz and biotite, and by their weak SPO. Undulous extinction in quartz grains (Fig. 5a), as well as the occurrence of some gently bent grains of biotite (Fig. 5b) and plagioclase (Fig. 5c) reveal a ductile deformation of very weak intensity. A chessboard pattern may occur in quartz (Fig. 5d). This

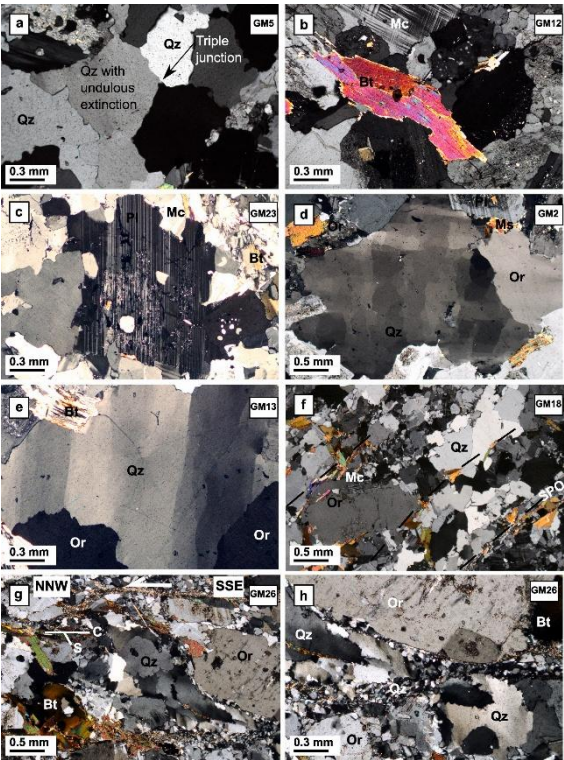
recovery microstructure, which results from the association between prismatic and basal subgrain boundaries due to combined basal $\langle a \rangle$ and prism $[c]$ intracrystalline slip, points to deformation at high T of ~650–750°C (Mainprice et al., 1986; Stipp et al., 2002), i.e. close to the granite solidus. Elongate quartz subgrains, indicative of $\langle a \rangle$ slip at low to moderate T (Kruhl, 1996), are also locally present (Fig. 5e), which shows that deformation may have lasted down to lower T conditions at some places.

Microstructures Mb (high-strain-high-T solid-state deformation). Compared to Ma, the Mb microstructural type is characterized by more widespread microscale evidence of ductile deformation. Moreover, the rock-forming minerals have a significant SPO and quartz shows signs of dynamic recrystallization (Fig. 5f). This microstructural type reflects higher-strain, but still relatively high-T conditions (de Saint Blanquat and Tikoff, 1997; Gleizes et al., 1998; Nédélec and Bouchez, 2011).

Microstructures Mc1-c2 (high-strain-low-T solid-state deformation). The main characteristic of the Mc microstructural type is the systematic occurrence of a S/C structure (Fig. 5g), a typical high-strain and also medium to low-T deformational feature (Trouw et al., 2010). Accordingly, elongate subgrains in quartz are more abundant. Also, dynamic recrystallization of quartz is significant, in particular in high stress and strain areas such as along C-bands and grain boundaries (Fig. 5g, h). The existence of S/C structures and the quartz microstructural features collectively suggest medium-T conditions (probably ~400–500°C; Stipp et al., 2002; Tullis, 2002; Passchier and Trouw, 2005), here classified as low-T to make a clear distinction with the higher temperature conditions associated with microstructures Ma and Mb. We have distinguished two subtypes in the microstructural type Mc: Mc1, with an incipient S/C structure, only visible at the microscale and Mc2, with a well-defined S/C mylonitic fabric.

The Ma microstructural type corresponds to a magmatic texture, almost unmodified by solid-state deformation. It is found only in the north-east of the pluton and its east-central part (Fig. 6), in agreement with the magmatic aspect of the fabric, as seen in the field in these areas. The Mc2 type is found solely in the Mylonitic Zone and the Mc1 type occurs both in this high-strain zone and adjacent to it. The Mb type, which attests to deformation conditions intermediate between these of the Ma and Mc types, is found only in the south and west of the pluton. In these areas, outside the Mylonitic Zone, the microstructural analysis definitely helps to constrain a higher degree of solid-state deformation, a feature difficult to decipher from the field observations alone. The distribution in map view of the microstructures highlights therefore a strain gradient (Fig. 6): intensity of the solid-state deformation is negligible in the NE part of the pluton and increases towards the Mylonitic Zone in the south-west where it reaches a maximum. The T-conditions of the deformation, as revealed also by the microstructures, decrease in the same direction.

Fig. 5.



Microstructures in quartz, feldspars and biotite (all microphotographs in cross-polarized light and all samples from the leucocratic facies). (a) Quartz grains with undulous extinction (note boundaries showing triple junction at $\sim 120^\circ$, as a result of static recrystallization; sample GM5, microstructural type Ma); (b) slightly bent biotite (GM12, type Ma); (c) slight bending of polysynthetic twins in a plagioclase (GM23, type Ma); (d) chessboard pattern in quartz (sample GM2, type Ma); (e) elongate subgrains in quartz (GM13, type Ma); (f) dynamic recrystallisation of quartz and significant SPO of individual minerals and grain aggregates (GM18, type Mb); (g) S/C structure, as seen in a thin section cut perpendicular to the S and C planes, and parallel to the lineation observed on an oriented sample (note the asymmetry, indicating a sinistral sense of shear; GM26, type Mc2); (h) detail of part (g), showing elongate aggregates of fine-grained recrystallized quartz (note the slightly irregular, serrated contacts between the larger quartz grains, pointing to limited grain boundary displacement. Bt, biotite; Mc, microcline; Ms, muscovite; Or, orthoclase; Pl, plagioclase; Qz, quartz.

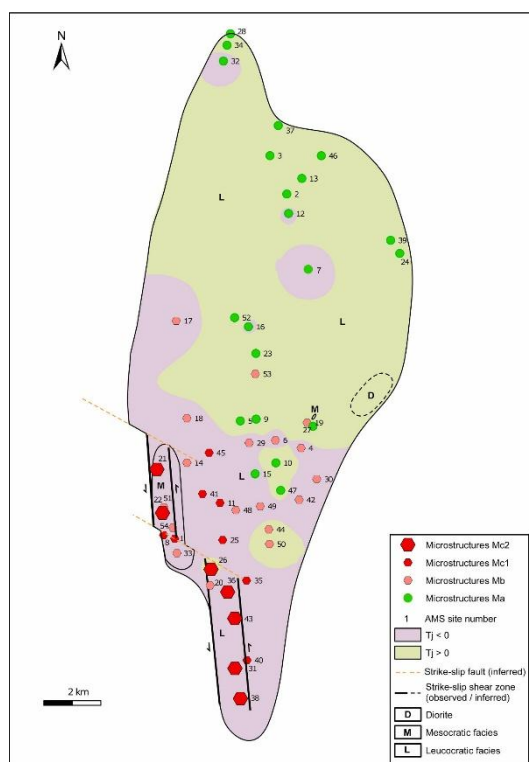
Table 1

Classification of the microstructures.

	Microstructure Ma	Microstructure Mb	Microstructure Mc1	Microstructure Mc2
Undulous extinction in quartz	X	X	X	X
Chessboard pattern in quartz	X	X	X	X
Elongate subgrains in quartz	X	X	XX	XX
Significant SPO of minerals and dynamic recrystallisation of quartz		X	XX	XXX
Bending of biotite cleavage	X	X	XX	XX
Bending of plagioclase twins	X	X	X	X
S/C structure			X	XXX

X, XX and XXX: increasing frequency of the microstructure.

Fig. 6.



Distribution map of the microstructure types observed in thin sections (Ma, Mb and Mc1-c2; see text for details). Areas with positive or negative values of the shape parameter (T_j), i.e. oblate or prolate magnetic fabric, were delineated from the AMS data (Table 2).

4.3. ROCK MAGNETIC STUDY

4.3.1. THERMOMAGNETIC ANALYSIS

The low- and high-T heating curves of samples GM29 (Fig. 7a), GM34 (Fig. 7b) and GM54 (Fig. 7e) may be approximated by a single concave upward (paramagnetic) hyperbola tending towards a low-slope (GM29, GM34) or horizontal (GM54) asymptote. At $\sim 560^\circ\text{C}$, there is a small susceptibility drop down to maximum $\sim 600^\circ\text{C}$ (GM29, GM54) or simply a slope break in the curve (GM34), which corresponds to the ill-defined Curie transition of a nearly pure (Ti-poor) (titano)magnetite (the Curie T of magnetite is $\sim 580^\circ\text{C}$; Clark, 1997; Dunlop and Özdemir, 2015). After this point, susceptibility decreases more rapidly than in the paramagnetic part of the curve until the final T (700°C) is reached, a behavior we attribute to the occurrence of hematite even if there is no obvious susceptibility drop at around 675°C (the Curie or Néel T of hematite; Dunlop and Özdemir, 2015). The heating curves of sample GM36 (Fig. 7c) are similar to these of GM34, except for an increase in susceptibility from $\sim 330^\circ\text{C}$ to $\sim 560^\circ\text{C}$. This increase, not to be confused with a Hopkinson peak, probably reflects the early formation of secondary magnetite from oxidation of the Fe-Mg silicates (mostly biotite) during heating (Deng et al., 2001; Trindade et al., 2001; Izquierdo-Llavall et al., 2014).

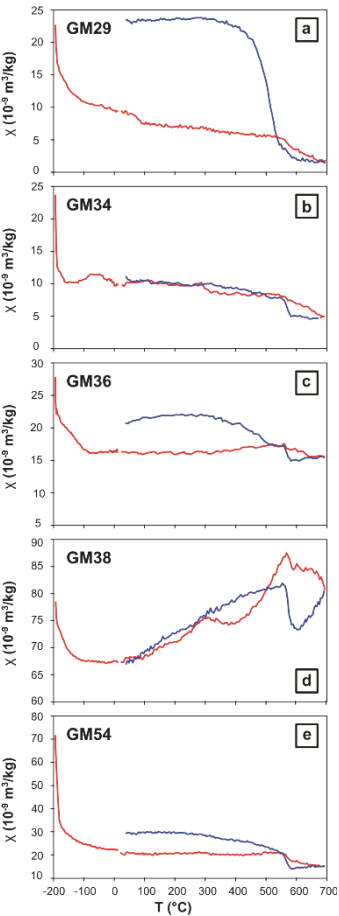
For sample GM38 (Fig. 7d), the evolution of susceptibility with increasing temperature is peculiar: after a hyperbolic decrease for most of the low-T curve, susceptibility increases from $\sim 0^\circ\text{C}$ to $\sim 310^\circ\text{C}$,

then drops down to a minimum at ~375°C (likely evolution following Curie transition of monoclinic pyrrhotite known to occur at 320°C; Clark, 1997; Dunlop and Özdemir, 2015), before increasing again up to ~550–570°C and finally decreasing down to the end (700°C) with a slightly more marked drop at ~580°C (Curie T of magnetite) and ~675°C (Néel T of hematite). This evolution is best interpreted as reflecting the neoformation of ferromagnetic (*sensu lato*) phases through early destabilization of the primary minerals during heating, at a larger scale than in GM36.

We consider the heating curves of GM29, GM34 and GM54 as the best representatives of the primary magnetic mineralogy of the Gamaye pluton. Their shape indicates that the magnetic mineralogy is controlled by paramagnetic minerals, mostly biotite which is ubiquitous and by far the most abundant paramagnetic mineral in the Gamaye pluton according to the petrographic observations, with a probably minor contribution of magnetite and hematite, two ferromagnetic carriers not detected in the thin sections. Due to the simultaneous occurrence of magnetite and hematite in the ferromagnetic fraction, the respective contribution to susceptibility of the ferro- and paramagnetic minerals cannot be quantified using methods such as those of Hrouda (1994) and Hrouda et al. (1997) or Richter and van der Pluijm (1994) which are magnetite-only based. A preliminary calculation using these methods give aberrant values, especially with regard to petrographic observations and the shape of the thermomagnetic curves (the calculated ferromagnetic contribution generally falls in the interval ~70–100%, some values being even >100%).

The *K-T* curves are irreversible (Fig. 7), with high-T cooling curves showing a more marked Curie point for magnetite, as well as negligible to strong increase in susceptibility. Such a susceptibility increase reflects oxidation of the samples during heating and the neoformation of magnetite (Henry et al., 2003; Izquierdo-Llavall et al., 2014).

Fig. 7.



Magnetic susceptibility versus temperature curves. Susceptibility measurements (χ) are normalized based on the mass of the samples. (a) Sample GM29 (leucocratic facies, microstructural type Mb); (b) sample GM34 (leucocratic facies, type Ma); (c) sample GM36 (leucocratic facies, type Mc2); (d) sample GM38 (leucocratic facies, type Mc2); (e) sample GM54 (mesocratic facies, type Mb). Red and blue lines represent heating and cooling curves, respectively.

Table 2

Anisotropy of magnetic susceptibility data.

Site	X(m)	Y(m)	Rock facies	n	Km (10 ⁻⁶ SI)	P _j	T _j	K1			K3		
								Dec/Inc	Conf	σ1/σ2	Dec/Inc	Conf	σ1/σ2
GM1	244,759	1,381,085	Mesocratic facies	6	134.0	1.057	0.164	22/2	0.97	17/8	113/17	0.98	15/9
GM2	248,561	1,392,788	Leucocratic facies	4	68.4	1.038	0.294	12/9	0.99	13/4	266/61	0.99	7/6
GM3	247,990	1,394,093	Leucocratic facies	4	80.6	1.034	0.413	199/18	0.98	14/4	304/38	1.00	7/4
GM4	249,046	1,384,160	Leucocratic facies	5	94.2	1.107	-0.072	181/1	0.99	9/4	281/72	0.97	18/5
GM5	246,984	1,385,083	Leucocratic facies	4	89.5	1.063	0.383	76/15	0.88	40/4	266/75	0.96	21/6
GM6	248,175	1,384,425	Leucocratic facies	4	78.0	1.072	-0.140	234/3	0.9	37/3	334/72	0.86	44/8
GM7	249,261	1,390,142	Leucocratic facies	4	83.8	1.018	-0.284	23/7	0.97	21/4	244/80	0.96	22/7
GM8	244,379	1,381,204	Mesocratic facies	4	150.7	1.088	-0.356	202/7	0.9	35/13	111/12	0.84	40/25
GM9	247,522	1,385,145	Leucocratic facies	7	76.3	1.097	0.262	34/6	0.9	34/11	227/84	0.98	15/8
GM10	248,191	1,383,657	Leucocratic facies	5	65.1	1.170	0.707	22/2	0.98	17/5	282/76	0.99	10/3
GM11	246,296	1,382,296	Leucocratic facies	4	78.1	1.080	-0.250	148/1	0.96	22/7	56/68	0.98	16/6
GM12	248,620	1,392,127	Leucocratic facies	4	81.5	1.029	-0.035	201/2	0.91	35/2	109/52	0.95	25/3
GM13	249,070	1,393,319	Leucocratic facies	7	85.7	1.036	0.632	236/4	0.79	53/12	327/18	0.97	18/10

Site	X(m)	Y(m)	Rock facies	n	Km (10 ⁻⁶ SI)	P _j	T _j	K1 Dec/Inc	Conf	σ1/σ2	K3 Dec/Inc	Conf	σ1/σ2
GM14	245,171	1,383,662	Leucocratic facies	4	76.6	1.133	0.047	180/1	0.94	27/11	270/5	0.92	27/18
GM15	247,487	1,383,283	Leucocratic facies	4	74.1	1.051	-0.147	58/3	0.93	27/14	168/82	0.92	27/17
GM16	247,261	1,388,283	Leucocratic facies	4	138.0	1.032	-0.087	66/17	0.88	38/11	249/73	0.89	39/7
GM17	244,808	1,388,484	Leucocratic facies	4	80.9	1.138	-0.324	171/8	0.97	19/4	263/16	0.95	25/8
GM18	245,170	1,385,171	Leucocratic facies	4	81.2	1.073	0.000	220/16	1.00	7/1	129/3	0.99	13/3
GM19	249,248	1,385,022	Mesocratic facies	6	100.7	1.120	-0.035	169/4	0.91	33/6	17/86	0.98	17/4
GM20	245,932	1,379,486	Leucocratic facies	4	80.8	1.051	-0.740	183/12	1.00	6/3	280/30	0.89	39/2
GM21	244,152	1,383,435	Mesocratic facies	5	131.9	1.069	-0.135	198/6	0.96	20/12	106/18	0.87	39/17
GM22	244,342	1,382,056	Mesocratic facies	5	124.0	1.153	0.130	186/7	0.97	21/3	93/22	0.96	21/5
GM23	247,511	1,387,374	Leucocratic facies	3	93.0	1.327	0.807	18/0	0.89	38/8	115/88	0.99	9/1
GM24	252,394	1,390,772	Leucocratic facies	5	67.7	1.128	0.857	167/2	0.65	72/4	284/85	1.00	4/3
GM25	246,386	1,381,035	Leucocratic facies	4	117.4	1.112	-0.143	186/12	0.99	8/2	95/1	0.98	13/8
GM26	245,987	1,380,040	Leucocratic facies	4	118.3	1.228	0.414	2/12	0.92	32/3	93/3	0.92	34/2
GM27	249,248	1,385,022	Leucocratic facies	4	75.7	1.142	0.547	158/6	0.79	53/9	279/79	0.97	21/3
GM28	246,628	1,398,321	Leucocratic facies	4	75.0	1.202	0.691	36/0	0.97	17/7	305/83	0.98	15/4
GM29	247,283	1,384,333	Leucocratic facies	4	129.7	1.065	-0.190	64/13	0.99	8/2	157/15	1.00	5/1
GM30	249,575	1,383,105	Leucocratic facies	4	75.1	1.097	-0.259	7/13	1.00	5/1	267/37	0.99	9/2
GM31	246,799	1,376,674	Leucocratic facies	4	109.0	1.072	-0.327	7/7	1.00	3/3	100/23	1.00	7/3
GM32	246,411	1,397,306	Leucocratic facies	4	60.0	1.065	-0.346	211/7	1.00	5/2	104/66	1.00	5/2
GM33	244,832	1,380,590	Mesocratic facies	4	120.0	1.059	-0.310	21/6	1.00	6/1	112/15	1.00	7/3
GM34	246,534	1,397,847	Leucocratic facies	4	65.0	1.079	0.366	185/28	1.00	6/4	30/60	1.00	5/3
GM35	247,193	1,379,659	Leucocratic facies	4	73.0	1.080	-0.058	190/4	0.99	8/3	289/65	0.99	8/5
GM36	246,552	1,379,272	Leucocratic facies	4	49.9	1.077	-0.593	169/7	0.98	16/1	78/7	0.66	72/5
GM37	248,269	1,395,115	Leucocratic facies	3	65.5	1.032	0.504	240/69	0.97	19/5	342/4	1.00	7/1
GM38	246,987	1,375,648	Leucocratic facies	4	105.7	1.046	-0.276	192/13	0.97	19/3	100/8	0.97	20/4
GM39	252,082	1,391,217	Leucocratic facies	4	83.9	1.020	0.161	29/9	0.94	28/8	298/6	0.96	20/11
GM40	247,208	1,376,955	Leucocratic facies	4	90.0	1.061	-0.236	30/11	1.00	4/2	263/72	1.00	5/2
GM41	245,699	1,382,605	Leucocratic facies	4	92.4	1.068	-0.344	182/11	1.00	6/2	276/22	0.99	10/2
GM42	248,979	1,382,404	Leucocratic facies	4	93.1	1.072	-0.507	199/8	1.00	6/1	305/62	0.99	10/5
GM43	246,791	1,378,372	Leucocratic facies	4	84.9	1.076	-0.764	356/3	1.00	2/2	265/24	0.95	26/2
GM44	247,952	1,381,396	Leucocratic facies	4	64.3	1.078	0.132	199/9	1.00	4/1	302/55	1.00	5/3
GM45	245,924	1,383,999	Leucocratic facies	4	86.9	1.033	-0.459	14/17	0.98	12/9	106/4	0.91	34/6
GM46	249,731	1,394,091	Leucocratic facies	4	56.4	1.032	0.179	174/22	0.98	14/10	359/81	0.97	19/9
GM47	248,353	1,382,722	Leucocratic facies	6	96.0	1.116	0.252	186/1	0.99	9/3	12/15	0.99	9/4
GM48	246,830	1,382,051	Leucocratic facies	5	64.9	1.130	-0.527	204/11	0.99	9/1	113/4	0.76	59/6
GM49	247,659	1,382,183	Leucocratic facies	5	92.0	1.079	-0.263	45/9	0.99	10/4	307/39	0.90	37/4
GM50	247,972	1,380,900	Leucocratic facies	7	80.5	1.182	0.586	205/1	0.98	16/2	307/83	0.99	11/3
GM51	244,342	1,381,955	Mesocratic facies	5	127.0	1.100	0.000	181/4	0.98	16/3	88/45	0.95	21/13
GM52	246,789	1,388,591	Leucocratic facies	4	97.8	1.131	0.420	176/2	0.99	10/4	82/69	0.98	16/6
GM53	247,494	1,386,678	Leucocratic facies	4	80.7	1.149	0.640	201/3	0.96	23/2	45/87	0.99	10/2
GM54	244,691	1,381,451	Mesocratic facies	4	130.6	1.107	-0.327	190/5	0.99	8/2	9/54	0.98	17/2

X, Y: coordinates in the WGS84 kilometric UTM grid (zone 29 N); n: number of specimens used in the AMS measurements; K_m: magntitude of the bulk magnetic susceptibility; P_j and T_j: anisotropy degree and shape parameters of Jelinek (1981); Conf: 2σ confidence level of the axis measurements; σ₁/σ₂: dimensions (in degrees) of the 2σ confidence cone expressing the scattering of individual axis measurements; K₁ and K₃: long and short principal axes of the AMS ellipsoid (K₁ is the magnetic lineation and K₃ the pole to the magnetic foliation); Dec/Inc: declination/inclination of the axes.

4.3.2. AMS SCALAR PARAMETERS

The mean bulk magnetic susceptibility (K_m) of the 54 AMS samples ranges from 49.9 to 150.7 $\times 10^{-6}$ SI (Table 2). Such low susceptibility values are typical of paramagnetic granites (e.g. Bouchez, 2000; Nédélec and Bouchez, 2011) and agree with the dominantly paramagnetic signature deduced from the thermomagnetic analysis. Bulk susceptibility has an unimodal frequency distribution, for both the leucocratic and mesocratic facies (Fig. 8a). This indicates that susceptibility is dominated by a single magnetic carrier (Borradaile and Henry, 1997), namely biotite in the present case. Accordingly, K_m is generally lower in the leucogranite facies than in the mesocratic facies which contains more biotite (Table 2; Fig. 8a). Higher K_m values in a few samples from the former facies probably reflect local enrichments in mafic minerals, mostly biotite.

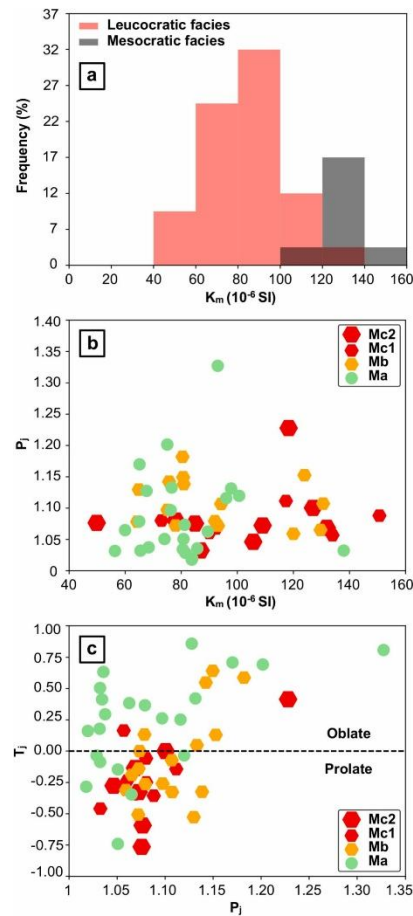
The mean values of the corrected anisotropy degree (P_j) vary from 1.02 to 1.33 and those of the shape parameter (T_j) range from - 0.76 to 0.86 (Table 2). The strength of the magnetic fabric is therefore (very) low and its shape is variable, from strongly prolate to strongly oblate. There is no obvious positive correlation between P_j and K_m (Fig. 8b), contrary to what is commonly observed in granitoids, both for para- and ferromagnetic lithologies (e.g. Bouchez, 2000; Ferré et al., 2014; Bolle et al., 2018). It can also be seen that the P_j value does not vary according to the type of microstructures (Fig. 8b, c), which suggests that P_j cannot be used as a proxy for strain intensity. Conversely, T_j is related to the microstructural types (Fig. 8c): samples with Ma microstructures tend to have $T_j > 0$, whereas many samples with Mb and Mc1-2 microstructures have $T_j < 0$. This means that the magnetic fabric is mostly oblate in the almost undeformed NE part of the pluton and becomes predominantly prolate in the west and south, i.e. close to and within the Mylonitic Zone (Fig. 6). Such a variation of the magnetic fabric shape must mimic a somewhat similar modification of the petrofabric shape.

4.3.3. AMS DIRECTIONAL DATA

First of all, it must be reminded that the AMS sampling grid in the Gamaye pluton is denser in the southern and central parts of the intrusion than in the northern part (see above, section “Material and methods”). Consequently, the magnetic foliation and lineation patterns are better constrained in the southern and central portions of the pluton.

Magnetic foliations are predominantly striking parallel to the N-S elongation of the intrusion, with highly variable dips, from 2° to 89° (Fig. 9a). Their average orientation (weighted mean) is N022-23ESE and on an equal-area projection, most foliation poles are roughly organized along a great circle, with a best fit π -axis (or magnetic zone axis; Henry, 1997) of 196/02. When the magnetic foliation poles are plotted on equal-area projections separately for samples with Ma, Mb and Mc microstructural types, compared to the whole data set we find quite similar mean orientations, as well as distributions along great circles, with two notable exceptions (Fig. 9a): a mean foliation at N006-79W for the Mc type, which reflects the predominantly steep dips of magnetic foliations in the Mylonitic Zone and close to it, i.e. where the Mc microstructures are found, and an ill-defined zone axis for the Ma type (052/03), which expresses likely the weaker SPO of the rock-forming minerals in the samples with this kind of microstructure.

Fig. 8.



AMS scalar parameters. (a) Frequency histogram of the bulk magnetic susceptibility (K_m) (the dark color indicates an overlap in the K_m values between the leucocratic and mesocratic facies); (b) corrected anisotropy degree (P_i) versus K_m diagram; and (c) P_i versus shape parameter (T_i) diagram.

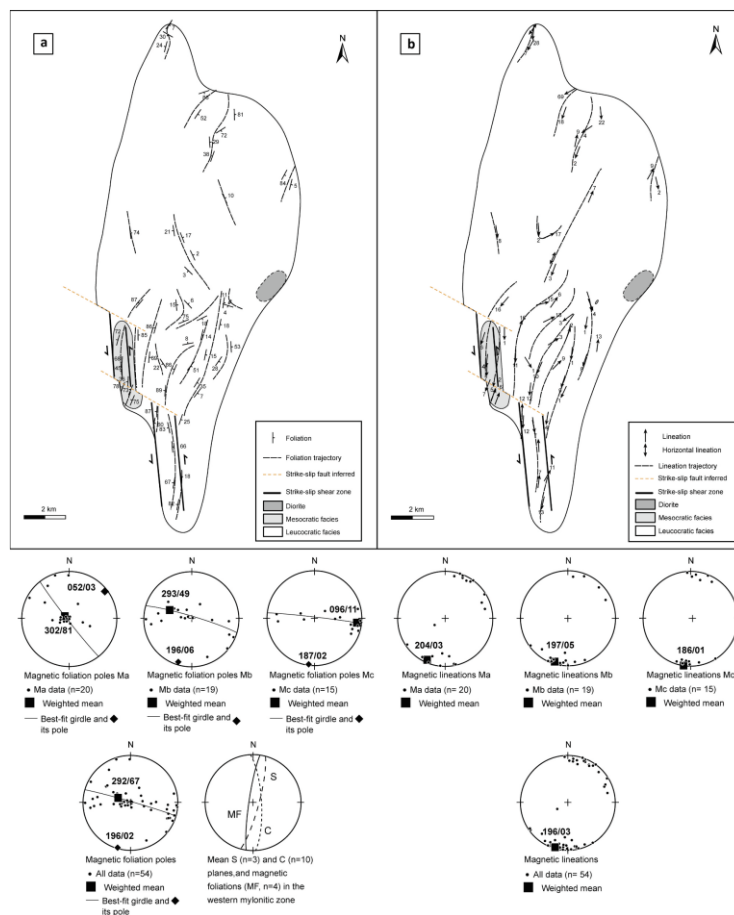
Outside the Mylonitic Zone, a comparison between field structural data in the pluton and magnetic foliation is difficult, mostly since measurements of the mesoscopic foliation are scarce and concern only the strike (Fig. 2). Within the shear zone, orientations of the S/C fabric and magnetic foliation can be more directly and accurately compared (Fig. 9a), at least using mean orientations (we have no paired measurements of both strike and dip for S plane, C plane and magnetic foliation on the same outcrop). We have seen above (Fig. 2) that the well-defined and regularly oriented C plane, and the S plane are both dominantly dipping to the east, with mean orientations of N172-77E and N013-70ESE, respectively. The associated magnetic foliation is also steeply dipping, to the east or to the west, with an average orientation (weighted mean) of N018-88WNW (Fig. 9a), making a low angle with the mean S and C planes (23° and 30°, respectively).

As for the magnetic foliations, many magnetic lineations have a submeridian orientation, roughly parallel to the N-S pluton elongation (Fig. 9b). These lineations are gently-plunging (maximum plunge of 28°) or even horizontal, with only one exception (plunge of 69°). Their average orientation (weighted mean) is 196/03, which perfectly fits, to within 1°, the zone axis of the magnetic foliations (Fig. 9a). Samples with Ma and Mb microstructures have mean magnetic lineations at 204/ 03 and

197/05, respectively, hence not significantly different from each other and also very close to the overall mean lineation. Magnetic lineations for the Mc type have also a very similar mean orientation, which is however a bit closer to the N-S direction (186/01). There is thus some deflection of the lineation in the Mylonitic Zone and its vicinity, however less marked than for magnetic foliation. We note also slight rotations of the magnetic lineations in the central part of the pluton, leading to sigmoidal lineation trajectories whose shape is in agreement with a deflection caused by a sinistral strike-slip movement (Fig. 9b). Sigmoidal arrangement of the magnetic foliations in the same area, as seen in map view, is less obvious, probably because of the variable dip values (Fig. 9a).

Magnetic lineations at the northern tip of the pluton (Fig. 9b) are subparallel to the axis of the small F_2 anticline we have measured in that area in the Kofi series (010/20; Fig. 2). At a larger scale, the mean magnetic lineation and the zone axis of the magnetic foliation in the Gamaye pluton (Figs. 9a, b) are parallel, to within 1° , to the zone axis defined by S_1 cleavages in the Kofi series (195/02; Fig. 2).

Fig. 9.



Maps of the magnetic foliation (plane perpendicular to the K_3 axis) (a) and magnetic lineation (K_1 axis) (b) in the Gamaye pluton. Below the maps, from left to right, equal-area plots (lower hemispheres) with: poles of the magnetic foliation and their cylindrical best-fit, three mean planes (S, C and magnetic foliation, MF, as measured in the Mylonitic Zone; number of measurements used to calculate the mean orientations: 3, 10 and 4, respectively) and the magnetic lineations.

5. Discussion

5.1. THE FEKOLA SHEAR ZONE

The use of AMS as a petrofabric tool is based in particular on the coaxiality between magnetic and rock fabrics (Borradaile and Henry, 1997). This principle admits exceptions (Rochette et al., 1992), but is verified in most granitoids, in particular in biotite-bearing plutons (Bouchez, 1997, 2000; Nedélec and Bouchez, 2011). In mylonitized granitoids, where the pre-deformational fabric is weak, the magnetic lineation is a direct equivalent of the stretching lineation (in a rock with a strong pre-existing fabric, mylonitization may produce a magnetic lineation perpendicular or oblique to the stretching lineation; Parés and van der Pluijm, 2002). In mylonitic rocks with a S/C structure, in particular granitic S/C mylonites, magnetic lineation is also parallel to the stretching lineation in many cases, not to the intersection of the S and C planes as might be expected (Ferré et al., 2014; and references therein). Attitude of the magnetic lineations in the Mylonitic Zone (Fig. 9b) thus confirms the field assumptions: the stretching lineation is subhorizontal there or at least gently-plunging and, consequently, the observed kinematics (Fig. 4) is indeed that of a sinistral strike-slip shear zone (this kinematics is also coherent with the microscale asymmetries illustrated in Fig. 5g). According to Ferré et al. (2014), the magnetic foliation in S/C mylonites lies in between the C and S planes, and will rotate towards the shear plane as finite strain increases. Such an orientation has been well documented in two mylonitized biotite granitoids by Aranguren et al. (1996) and Ferré and Améglio (2000), in the Variscan belt of northern Spain and the late Neoproterozoic–early Cambrian Saldanian basement of South Africa, respectively. However, the magnetic foliation associated with S/C structures may also approximate the S plane, as found by Ono et al. (2010) in a mylonitized granite from the Late Cretaceous Ryoke belt (SW Japan), where biotite and hornblende are assumed to control AMS. Here, we do not have enough information to compare accurately the orientation of the magnetic foliation with that of the S/C structure in the Mylonitic Zone (Fig. 9a): we can only say that, on average, the magnetic foliation approximates the S and C planes.

As stated above, there is little doubt that the Mylonitic Zone materializes the southern extension of the FSZ. Accordingly, shear bands mapped in the host rocks of the Fekola gold deposit (B2Gold, unpublished data; Supplementary Material 2) have the same orientation as the C planes measured in the Mylonitic Zone (Fig. 2), and lineations in the FSZ are predominantly N-S-trending and gently-plunging (Rhys, 2015; Supplementary Material 2). Also, kinematic indicators in the FSZ (A. Toure, personal communication 29th November 2025) demonstrate a strike-slip sinistral shearing, largely dominating over a possible reverse movement evoked by Rhys (2015). In short, the high-strain zone hosting the Fekola gold deposit and that affecting the SW part of the Gamaye pluton correspond to a single sinistral strike-slip shear zone, the FSZ, which has a ca. N170° orientation and extends over more than 15 km (Fig. 2). The present study thus confirms the occurrence of a roughly N-S-trending, sinistral strike-slip shear zone affecting the western margin of the Gamaye pluton, as previously suggested by Ledru et al. (1989, 1991), and it further specifies its orientation and minimal extension.

The gently-plunging attitude of the lineations, as well as the prolateness of the fabric revealed by the AMS data in the Mylonitic Zone (Figs. 6, 9b), indicate that the FSZ is not necessarily a “true” strike-slip (transcurrent) shear zone (only simple shear deformation), but could equally correspond to a transpressional (or transtensional) simple-shear- dominated shear zone with a relatively moderate magnitude of strain (Tikoff and Greene, 1997; Fossen and Tikoff, 1998). If a reverse component of movement is indeed present (Rhys, 2015), this would make the FSZ a transpressional structure with a small component of coaxial deformation.

The FSZ is maybe a segment of the SMSZ. In studies conducted to the west and north-west of the Gamaye pluton (Dabo and Aïfa, 2010, 2011; Diene et al., 2015), and further north in the Sadiola mining district (Masurel et al., 2017c), the SMSZ has been described as a large brittle- ductile corridor made of a network of shear zones of varying importance, cutting across or following lithological contacts and related to the D₂ sinistral transpressional event. Relating the FSZ to the SMSZ system deserves, however, further field investigations. We will simply retain here that according to its ca. N-S orientation and sinistral kinematics, the FSZ may be considered as a D₂-related structure, like the SMSZ.

5.2. SYN-D₂ EMPLACEMENT AND DEFORMATION OF THE GAMAYE PLUTON

Available U-Pb and Pb-Pb dates on zircon ($\pm$ monazite) constrain emplacement of the late-Eburnean granitoids in the KKI at ca. 2095–2060 Ma, with two large *syn*-D₂ igneous complexes, the Saraya (Pons et al., 1992) and Fal’em’è (Diene et al., 2015) batholiths, dated at ca. 2080–2060 Ma and ca. 2090–2075 Ma, respectively (see age compilation above). The Gamaye pluton forms with the *syn*-D₂ Saraya batholith and some other late-Eburnean intrusions a well-defined group of biotite-bearing, high-K and peraluminous granitoids with high (La/Yb)_N and (Dy/Yb)_N ratios (Dioh et al., 2006; Lambert-Smith et al., 2016a; Masurel et al., 2017c). However, the emplacement age of the Gamaye pluton is currently not well constrained. The Rb-Sr date obtained by Bassot and Caen-Vachette (1984), 2045 ± 27 Ma, has a quite large uncertainty and, moreover, it must be younger than the emplacement age of the granite. Indeed, the Rb⁸⁷ decay constant of $1.42 \cdot 10^{-11} \text{ yr}^{-1}$, as recommended by Steiger and Jäger (1977) and used by Bassot and Caen-Vachette (1984), is slightly higher than the actual value, leading to too young calculated ages (e.g. Nebel et al., 2011; Rotenberg et al., 2012).

In absence of any robust age constraint, it would be conceivable that the Gamaye pluton was emplaced before the ca. 2085-(?)2060 Ma D₂ event responsible for its mylonitization in the FSZ, either in a last stage of the ca. 2160–2085 Ma D₁ event or during the D₁-D₂ transition. However, our microstructural study, coupled with field observations, reveals a strain and temperature gradient from an almost pristine magmatic fabric in the north-east of the pluton to a mylonitic fabric in the south-west, in the FSZ (Fig. 6). Such a continuum of deformation, as well as the concordance between magnetic fabric in the pluton and tectonic structures in the country rocks (Figs. 2, and 9) are typical of a granitic intrusion emplaced in a regional strain field and deformed in the course of its crystallization, down to subsolidus conditions (e.g. Bouchez, 1997; Paterson et al., 1998; Bouchez, 2000; Vernon, 2000). Since this regional strain field is that of the D₂ event, as detailed below,

emplacement of the Gamaye pluton is therefore *syn-D₂*, as for most other late-Eburnean granitoids in the KKI.

Compared to the magnetic foliations (Fig. 9a), orientation of the magnetic lineation in the Gamaye pluton is homogeneous, with only small variations when approaching the FSZ (Fig. 9b). Also, this orientation coincides with that of the direction of *D₂* folding recorded in the metasediments of the Kofi series (Figs. 2, and 9a), as shown by the local concordance between magnetic lineations and a *F₂* fold axis, as well as the parallelism between mean magnetic lineation and a *S₁* zone axis (organization of the *S₁* cleavages around a common NNE-SSW, subhorizontal zone axis reflects roughly cylindrical folding during *D₂*, with the predominantly NNE-SSW, subvertical *S₂* cleavages as axial-planar fabrics).

Parallelism between a mineral/stretching lineation, materialized here by the magnetic lineation, and a fold axis indicates that the latter is parallel to the X-axis of the finite strain ellipsoid, classical explanations of this feature including sheath fold geometry, strong deformation in a shear zone and fold formation in a constriction flow (e.g. Twiss and Moores, 2007). However, parallelism between X-axis and fold axis is also typical of folds produced by oblique tectonics, whether transpressional (Tikoff and Peterson, 1998; Frehner, 2016) or transtensional (Fossen et al., 2013). Thus, we propose that in the present case, the concordance between stretching and folding directions reflects the transpressional nature of the *D₂* event. Large-scale examples of such a concordance in transpressional systems are found in the northern Appalachian orogen, in western Main, USA (Solar and Brown, 2001) and the East Variscan shear zone system of the Southern European Variscan belt (Carosi and Oggiano, 2002; Corsini and Rolland, 2009; Petroccia et al., 2024) where transpression is partitioned between strike-slip corridors and less deformed folded zones, parallelism between mineral/ stretching lineations and fold axes being observed, at least at the regional scale, even in the less deformed zones.

Geometry of the *D₂* folding in the host rocks of the Gamaye pluton and orientation of the mean lineation in the latter thus collectively indicate stretching parallel to the *D₂* fold axes, i.e. in a NNE-SSW direction. The contraction direction is perpendicular, i.e. ESE-WNW. Also, the pluton fabric is chiefly planar outside the influence area of the FSZ, as revealed by the predominantly oblate shape of the AMS ellipsoid away from the Mylonitic Zone (Fig. 6). The *D₂* contraction was therefore stronger than the stretching acting parallel to the folding direction, as expected for transpressional folding (Fossen et al., 2013).

Magnetic foliations defining a zone axis parallel to the mean magnetic lineation, as observed in the Gamaye pluton (Fig. 9a, b), are quite frequent in granitoids (see case studies in Henry, 1997; and Nedelec and Bouchez, 2011). In particular, there are well-documented examples of syntectonic granitoids emplaced in partitioned transpressional systems, either along strike-slip shear zones or in the intervening domains, where a gently-plunging magnetic foliation zone axis, parallel to the mean magnetic lineation, is also concordant with both the directions of stretching, materialized by mineral/stretching lineations, and folding in the country rocks, locally with deformation and fabric deflection along the transcurrent shear zones (e.g. several plutons in the Meguma terrane of the northern Appalachians in Nova Scotia, Canada; Benn et al., 1997; Pignotta and Benn, 1999; and in

the Neoproterozoic Serido belt of the Borborema province in NE Brazil; Archanjo et al., 2002). These granitoids display striking fabric similarities with the Gamaye pluton. In the Meguma terrane, the peculiar disposition of the magnetic foliations in the granitoids has been interpreted as the consequence of a syn-crystallization folding of early foliations related to emplacement kinematics, whether (Benn et al., 1997) or not (Pignotta and Benn, 1999) the regional fold axes cut across pluton margins. In the Serido belt, no specific signification has been given to that disposition by Archanjo et al. (2002), other than a zonal distribution around the mean magnetic lineation which materializes the regional stretching direction recorded by the syntectonic granitoids. In the Gamaye pluton, it cannot be ruled out that “winding” of the magnetic foliations around the NNE-SSW mean magnetic lineation would reflect syn-emplacement folding in response to the ESE-WNW contraction. However, in absence of any other evidence of folding in the granite, this remains hypothetical.

To summarize, during its syntectonic emplacement, the Gamaye pluton has recorded an ESE-WNW contraction and a NNE-SSW stretching of weaker intensity. It has also accommodated a N-S sinistral strike-slip shearing along the FSZ, i.e. at a slight angle to the stretching direction, and this shearing has lasted down to lower temperatures (probably ~400–500°C) giving rise to mylonitization (the sigmoidal lineation trajectories in the central part of the pluton suggest also sinistral shearing in that area, however at higher temperatures; Figs. 6, and 9b). Since these synmagmatic deformations are related to the D_2 transpressional event, we propose here a model of progressive deformation (*sensu* Fossen et al., 2019) into which the contraction and the weaker associated stretching on the one hand, and the shearing on the other hand represent two components of the same partitioned transpression (e.g. Tikoff and Teyssier, 1994; Dewey et al., 1998; Fossen and Tikoff, 1998), rather than being two consecutive or slightly overlapping discrete phases (D_{2a} and D_{2b} phases, respectively). In this model, the Gamaye pluton was emplaced in a contraction-dominated domain, at the contact with a relatively steep sinistral strike-slip shear zone roughly perpendicular to the contraction direction, the FSZ. During crystallization the pluton was deformed in the contractional domain, concomitantly with folding of the Kofi series, and sheared along the FSZ, with localization of the shearing after solidification leading to mylonitization of the granite. The ESE-WNW faults with an apparent sinistral strike-slip component which cut across the FSZ and part of the Gamaye pluton are probably the witness of a late brittle deformation phase that we refer here as D_3 .

The close association between the Gamaye pluton and the FSZ further suggests that the ascent and emplacement of the granitic magma could have been tectonically controlled by this weakness zone. Indeed, it is now widely admitted that diapirism is not a viable mechanism of upward transport and emplacement of granitic magmas, at least in the upper crust (Clemens and Mawer, 1992; Petford, 1996). Diapirism fails, in particular, to account for the rapid ascent rates of these magmas (approximately 10^{-2} m/s) and the sheet-like geometry commonly observed for large granitoid plutons. A fracture-driven transport mechanism of the granitic magmas, i.e. dyking, has therefore been proposed as an alternative to diapirism (Clemens and Mawer, 1992; Petford, 1996). However, the ascent and emplacement of granitic magmas at different levels in the crust may also be controlled by active tectonic structures, in particular faults and shear zones, both in compressional/transpressional and extensional/transensional settings (e.g. Brown and Solar, 1999; Petford and Clemens, 2000; Papeschi et al., 2025). There are for instance many examples of major strike-slip

shear zones or faults that have guided the upward migration of granitic magmas, as well as their emplacement possibly in voids related to the strike-slip tectonics such as large-scale extension fractures or dilational jogs (e.g. Castro, 1986; Hutton and Reavy, 1992; Ferré et al., 1997; Gébelin et al., 2006; Bolle et al., 2010). Nevertheless, adding the Gamaye pluton to that list would require further investigations again focusing on the FSZ.

The *syn*-D₂ emplacement model proposed here for the Gamaye pluton fits in some respects with these of Pons et al. (1992) and Diene et al. (2015) for the Saraya and Fal'ém'e batholiths, respectively. Following a preliminary model proposed by Ledru et al. (1989, 1991), a structural analysis of the Saraya batholith led Pons et al. (1992) to propose segregation, collection and ascent of magmas triggered by a major D₂ strike-slip structure, as well as interference between pluton "ballooning" and D₂ transpressional tectonics during emplacement, as supported in particular by the subhorizontal to gently-plunging lineations, parallel to the batholith long axis and concordant with L₂ stretching lineations in the country rocks. To complement this model, it must be reminded that the Saraya batholith crops out in the core of a F₂ mega-antiform (Bassot and Dommanget, 1986; Ledru et al., 1989, 1991), which may be taken as a further evidence of synemplacement D₂ stretching, parallel to the fold axis. In their model, based on integrated geophysical and satellite data, field observations, structural analysis and geochronology, Diene et al. (2015) proposed that emplacement of the Fal'ém'e batholith was coeval with the D₂ sinistral transpressional deformation, with intrusion of the plutons controlled by the SMSZ system and occurring in dilational jogs between lower-order strike-slip shears.

5.3. THE GAMAYE PLUTON: ANOTHER EVIDENCE FOR WIDESPREAD, LATE-EBURNEAN TRANSPRESSION AND MAGMATISM IN THE SOUTHERN WAC

In addition to the Gamaye pluton (this study), the Saraya batholith (Pons et al., 1992) and the Falémé batholith (Diene et al., 2015) in the KKI, a strong link between late-Eburnean transpressional tectonics and plutonism has been established in the Baoul'e-Mossi domain. For example, a ca. 2090–2070 Ma plutonic belt was emplaced in the NW part of this domain, in Guinea, near the northern margin of the Archean Kénéma-Man domain, along the NW branch of the Sassandra shear zone (Egal et al., 2002) and ca. 2100–2075 Ma granitoids from the Massigui batholith were emplaced along the Banifing shear zone in Southern Mali (Liégeois et al., 1991; Wane et al., 2018). In Ivory Coast, the Bonikro porphyritic granodiorite, dated at 2086 ± 4 Ma (U-Pb zircon age), was emplaced along the Bonikro shear zone during the late stage of a compressive event, prior to late-Eburnean transcurrent deformation, but it released residual, mineralized fluids at the onset of the transcurrent phase (Masurel et al., 2022b).

A spatio-temporal (and commonly genetic) relationship between late-Eburnean magmatism and structurally-controlled gold mineralization, as observed here between the Gamaye pluton and the FSZ which hosts the Fekola gold deposit, is also quite widespread in the southern WAC. In the Kofi series, north of the Gamaye pluton, the gold deposits of Alamoutala and Sadiola (Sadiola mining district; Masurel et al., 2017a, b), as well as Yal'ea and Gounkoto (Loulo mining district; Lawrence et al., 2013; Lambert-Smith et al., 2016b) are all located along shear zones, the Loulo district being

close to the Falémé batholith (which has led to various models invariably proposing a more or less direct genetic link between this batholith and the neighboring gold deposits; see Allibone et al., 2020 for the most recent contribution). In the Mako volcanic belt, the Sabodala gold deposit consists of two successive hydrothermal Au mineralizations associated with a shear zone, the first one being related to felsic lavas (Sylla and Ngom, 1997). Also in the Mako volcanic belt, the Massawa gold deposit is a shear-hosted mineralization formed from magmatic fluids released from a series of felsic sheets coeval with the mineralization (Treloar et al., 2015). The Bonikro gold deposit in Ivory Coast, dated at 2074 ± 16 (Re-Os molybdenite age), is hosted by late- Eburnean, strike-slip brittle-ductile shear zones and overprints an earlier intrusion-related gold system linked to the emplacement of the Bonikro porphyritic granodiorite (Masurel et al., 2022b).

6. Conclusions

The FSZ, which hosts the world-class Fekola gold deposit, is a ca. N-S sinistral-strike slip structure of first order (extension >15 km) related to the late-Eburnean transpressional D₂ event dated elsewhere in the KKI at ca. 2085-(?)2060 Ma. Analysis of field structural data, microstructures and magnetic fabric support *syn*-D₂ emplacement of the Gamaye pluton at the contact with the FSZ in a partitioned transpressional system: throughout its emplacement down to subsolidus conditions, the pluton has recorded an ESE-WNW contraction and a weaker NNE-SSW stretching, away from the FSZ and concomitantly with folding of the Kofi series, and it has accommodated as well a N-S sinistral strike-slip shearing along the FSZ, which has lasted down to lower temperatures (probably ~400–500°C) leading to mylonitization of the granite. The present study confirms the close association between magmatism, tectonics and also metallogeny during the late-Eburnean, transpressional evolution of the southern WAC. Furthermore, the emplacement model proposed for the Gamaye pluton highlights the importance of regional strain partitioning in shaping parts of this crustal domain during the last stages of its orogenic evolution.

CREDIT AUTHORSHIP CONTRIBUTION STATEMENT

Aliou Dembele: Writing – review & editing, Writing – original draft, Resources, Project administration, Methodology, Investigation, Funding acquisition, Formal analysis, Conceptualization. **Olivier Bolle:** Writing – review & editing, Visualization, Validation, Supervision, Resources, Methodology, Investigation, Formal analysis, Conceptualization. **Sarah A. Friedman:** Writing – review & editing, Validation, Resources. **Moussa Dabo:** Writing – review & editing, Validation, Supervision, Resources, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **Mamadou Lamine Bouare:** Writing – review & editing, Validation, Investigation.

DECLARATION OF COMPETING INTEREST

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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APPENDIX A. SUPPLEMENTARY DATA

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.precamres.2026.108094>.

DATA AVAILABILITY

Data will be made available on request.

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