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Special Issue Some applications of LA-ICP-MS U-Pb geochronology: A tribute to Jean Louis Paquette
Marc Poujol, Valérie Bosse and Dave Chew (Guest Editors)

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Zircon and apatite U-Pb geochronology of the Paleoproterozoic (Eburnean) basement and late Neoproterozoic (Pan-African) metamorphism and magmatism from Port-Béni, Armorican Massif (France)

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Abstract – By re-examining the historical outcrops of Port-Béni located in the Trégor unit of the North Armorican Cadomian belt, the present work delivers four new ages that provide additional constraints on the Proterozoic history of northern Brittany. It is established that granitic, porphyritic rocks crystallized at the end of the Rhyacian (Paleoproterozoic), 2038 ± 12 Ma ago, before being transformed into orthogneisses at a late Neoproterozoic (Ediacaran) age of 621 ± 2 Ma, which is a minimum age, given the retrograde alteration these rocks underwent. The age of ca. 1.8 Ga previously proposed for the protolith of the Port-Béni orthogneiss should be discarded, and these two new ages are consistent with most of those yielded so far by the other Icartian (*i.e.*, Eburnean) basement relics from the Armorican Massif. The gneissic basement was then intruded and disrupted into xenoliths by a granodioritic magma that crystallized 604.5 ± 2.0 Ma ago. This age, slightly younger than previously thought, corresponds to the emplacement age of one of the main units of the North Trégor batholith − the Pleubian-Talbert unit −, part of the Trégor volcano-plutonic complex, which may have built up over a longer period than that indicated by the uncertainty associated with this age. Caution should be exercised in extrapolating this age to that of the whole complex. Finally, doleritic dykes, possibly resulting in two swarms previously thought to be Paleozoic in age, have crosscut this complex. One of the latest yielded an age of 597 ± 15 Ma, indicating that the Trégor doleritic dyking episodes also occurred during the late Neoproterozoic, in between ca. 605 Ma and ca. 580 Ma. As the doleritic dykes are of tholeiitic composition, which distinguishes them from the earlier calc-alkaline magmas, they suggest that the intra-arc extension, documented in the southern, adjacent Saint-Brieuc unit of the belt, also affected the Trégor unit. They may likely have fed northern equivalents of the lava flows from the Paimpol Formation (exposed in between the Saint-Brieuc and the Trégor units), when magma production became moderately influenced by the Cadomian (*i.e.*, Pan-African) subduction and mostly dominated by extension, possibly as a result of a steepening of a north-dipping subduction slab. Indeed, a re-examination of the available geochemical and geochronological data in the light of our new results documents that arc-magma production moved progressively from north (Trégor unit) to south (Saint-Brieuc unit) over time, in the interval 605-580 Ma.

Keywords: Cadomian / Icartian / Trégor / dolerite dyke swarm / orthogneiss / granodiorite

Résumé – Géochronologie U-Pb sur zircon et apatite du socle paléoprotérozoïque (éburnéen), du métamorphisme et des évènements magmatiques néoprotérozoïques (panafricains) à Port-Béni, Massif armoricain (France). Réexaminés au cours de ce travail, les affleurements historiques de Port-Béni, qui sont localisés dans l'unité du Trégor de la chaine Cadomienne Nord-Armoricaine, ont permis d'obtenir quatre nouveaux âges fournissant des contraintes supplémentaires pour décrypter l'histoire du nord de la Bretagne au Protérozoïque. Les résultats établissement que des roches granitiques porphyriques ont cristallisées au Paléoprotérozoïque, à la fin du Rhyacien, il y a 2038 ± 12 Ma, avant de se transformer en orthogneiss tardivement au Néoprotérozoïque (à l'Édiacarien) à un âge de 621 ± 2 Ma, qui est minimum

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compte tenu de la rétromorphose subie par ces roches. L'âge proche de 1.8 Ga précédemment obtenu pour le protolithe des orthogneiss de Port-Béni devrait être écarté et nos deux nouveaux âges sont cohérents avec la plupart de ceux livrés à ce jour par les autres reliques de socle icartien (*i.e.*, éburnéen) du Massif armoricain. Ensuite, le socle gneissique a été fragmenté en xénolithes par des intrusions de magmas grandioritiques qui ont cristallisé il y a 604.5 ± 2.0 Ma. Cet âge, légèrement plus jeune que celui antérieurement admis, correspond à celui de la mise en place d'une des principales unités – celle de Pleubian-Talbert – du batholite nord-trégorrois, lui-même faisant partie du complexe volcano-plutonique du Trégor, dont la construction a pu durer plus longtemps que l'intervalle de temps donné par l'incertitude obtenue sur notre âge. La prudence est de mise en extrapolant cet âge à celui du complexe. Enfin, des filons doléritiques, possiblement mis en place en deux essaims distincts que l'on rattachait au Paléozoïque, ont recoupé ce complexe. Un des plus tardifs de ces filons a livré un âge de 597 ± 15 Ma indiquant que les épisodes filoniens doléritiques du Trégor sont également survenus tardivement au Néoprotérozoïque entre environ 605 Ma et 580 Ma. Ces filons étant de composition tholeiitique, ce qui les distingue des magmas calco-alcalins antérieurs, ils indiquent que l'extension intra-arc, documentée plus au sud dans l'unité adjacente de Saint-Brieuc, au sein de la chaîne, a aussi affecté l'unité du Trégor. Il est assez probable qu'ils aient alimenté des équivalents septentrionaux des coulées de lave de la formation de Paimpol (localisée entre les unités de Saint-Brieuc et du Trégor), lorsque la production magmatique est devenue modérément influencée par la subduction cadomienne (*i.e.*, panafricaine) et principalement dominée par l'extension, possiblement suite à un redressement d'un plan de subduction à pendage nord. En effet, le réexamen des données géochimiques et géochronologiques disponibles à la lueur de nos nouveaux résultats, montre que la production de magma d'arc s'est déplacée progressivement au fil du temps depuis le nord (unité du Trégor) vers le sud (unité de Saint-Brieuc) dans l'intervalle 605-580 Ma.

Mots clés : Cadomien / Icartien / Trégor / essaim de filons de dolérite / orthogneiss / granodiorite

1 Introduction

This work revisits a historical geological site, where Barrois (1898) first defined quite appropriately the "micaschistes et gneiss fondamentaux de Port-Béni" as a basement. These gneisses, exposed as enclaves within the local granitic rocks, were considered, at the time, to be the oldest rocks in the Trégor area of northern Brittany. Less than a century later, some of these rocks and their equivalents in the northern Armorican Massif (Fig. 1) were shown to be Paleoproterozoic in age, making them the oldest rocks to crop out in metropolitan France and a large part of westernmost Europe (*e.g.*, Calvez and Vidal, 1978; Auvray *et al.*, 1980). At the same time, together with the Trégor magmatic rocks, they inspired one of the very first geodynamical reconstructions incorporating the then-recent concepts of plate tectonics, and were used to propose that the local, late Neoproterozoic Pan-African mountain belt, known as the Cadomian belt, was an active continental margin (Auvray and Maillet, 1977; Auvray, 1979). As outlined below, the present work had four dating targets, each being related to one of the three main rock types in the Port-Béni outcrops, and provides new constraints on the Proterozoic history of the northern Armorican Massif.

A first target was to date the Port-Béni orthogneiss (Auvray, 1979), as it is a representative example of the disrupted Paleoproterozoic basement from the northern Armorican Massif (Fig. 1), now known as Icartian (Roach *et al.*, 1972; Calvez and Vidal, 1978), the local expression of the Eburnean orogeny. Wherever it has been identified so far, this basement consistently yielded protolith ages of around 2 Ga or older (Calvez and Vidal, 1978; Auvray *et al.*, 1980; Piton, 1985; Samson and D'Lemos, 1998; Inglis *et al.*, 2004; Martin *et al.,* 2018). The only exception is the Port-Béni locality where a younger age of about 1.8 Ga was proposed (Auvray *et al.*, 1980; Graviou *et al.*, 1988). One of the goals of this work is therefore to verify this age, in order to re-evaluate the age distribution of the Icartian relics of the Armorican Massif. In addition, getting the age of the metamorphism and deformation that affected the protolith of the orthogneiss was a second target. Indeed, if it has been demonstrated in Normandy that the ductile deformation and high-grade metamorphism that affected the Icartian protoliths took place during the Neoproterozoic (Inglis *et al.*, 2004; Martin *et al.*, 2018), although probable (Tribe *et al.*, 1996), this has yet to be confirmed in Brittany.

At Port-Béni, the Icartian xenoliths are enclosed within one of the intrusive units making up the composite North Trégor Batholith (Fig. 2 Barrois 1908; Auvray 1979), which is assumed to be ca. 615 Ma old (Graviou, 1984; Graviou *et al.*, 1988). As this age has been suggested to be overestimated by the authors themselves (Graviou *et al.*, 1988), the present work also provided an opportunity to date the microgranodiorites hosting the xenoliths.

Subsequently, several dykes, including some doleritic dykes exposed at Port-Béni and widespread elsewhere in the Trégor, crosscut the Trégor batholith (Fig. 2A; Barrois 1908; Auvray *et al.*, 1976; Auvray, 1979). These doleritic dykes, which may belong to two different swarms, were given a poorly constrained Paleozoic age (Lees *et al.*, 1991; Roach *et al.*, 1992; Ruffet *et al.*, 1992; Lahaye *et al.*, 1995). One additional goal of this study was therefore to verify the age of the Trégor dolerites.

After a brief review of the available geological, geochemical and geochronological data in the light of our results, this study explores the possibility that the Trégor doleritic dykes may have fed the mafic lava flows of the Paimpol volcanic formation exposed in southern Trégor (*e.g.*, Barrois 1908; Auvray *et al.*, 1976; Auvray, 1979) or its

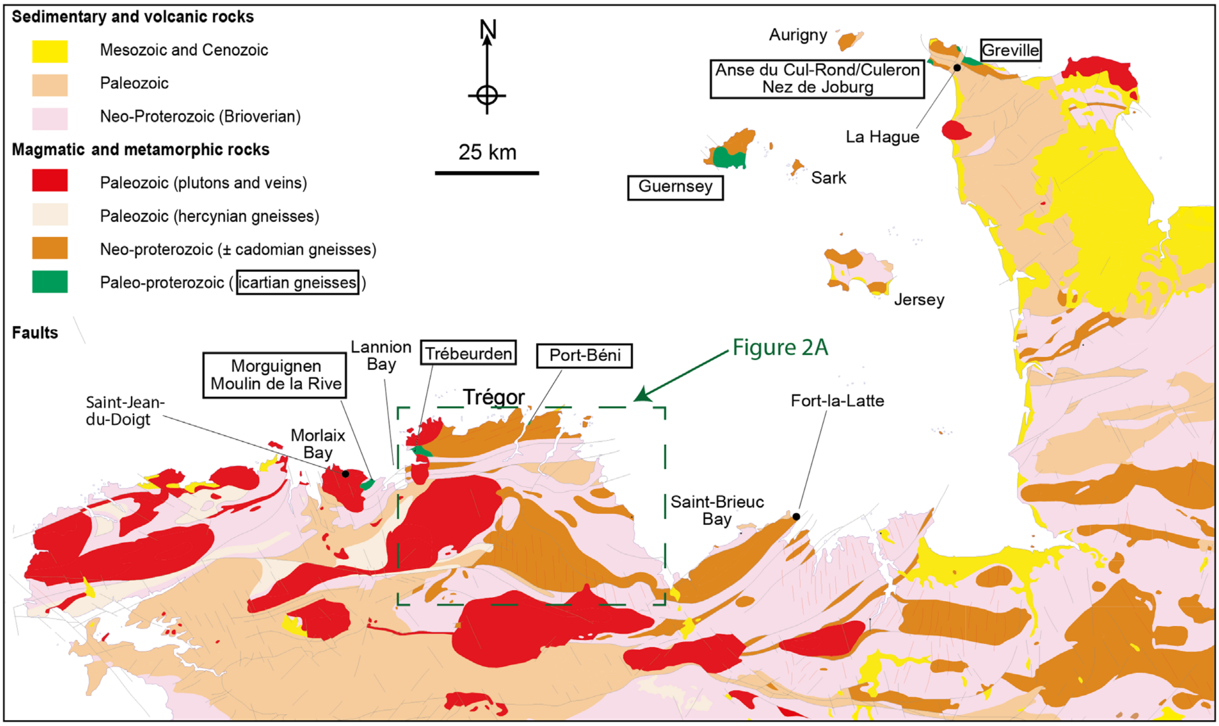

Fig. 1. Geological map of the northern part of the Armorican Massif (simplified from the 1/1 000 000 geological map of France from the BRGM).

northern, now eroded, lateral equivalents. Furthermore, both the spatial migration of the late Proterozoic magmatic activity and its temporal evolution are discussed in the framework of the last geodynamical model of the North Armorican Cadomian belt (Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001).

2 Geological setting and historical review

Today, several tectonically juxtaposed units or blocks that underwent distinct P-T-deformation paths, make up the North Armorican Cadomian belt (Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001). Among these, in northernmost Brittany, the Trégor unit (Fig. 2A) includes disrupted parts of the gneissic basement (Barrois, 1898), which exhibits Paleoproterozoic protolith ages ranging from about 2.2 to 1.8 Ga (Rhyacian-Orosirian) across the north of the Armorican Massif (Fig. 1; Calvez and Vidal, 1978; Auvray *et al.*, 1980; Vidal *et al.*, 1981; Piton, 1985; Samson and D'Lemos, 1998; Inglis *et al.*, 2004; Martin *et al.*, 2018). These gneisses witness an orogenic cycle defined locally under the name of Icartian (Roach *et al.*, 1972; Calvez and Vidal, 1978), corresponding to a world episode of crustal accretion and growth that is also known elsewhere as Eburnean (West Africa) or Transamazonian (South America). Both paragneisses and orthogneisses that experienced high-grade, sometimes migmatitic, metamorphic conditions are described among the Icartian rocks. They are found among Neoproterozoic or younger rocks from Northern Brittany (west and east of the Baie de Lannion and Port Béni in the Trégor area), Normandy (North Cotentin, near La Hague: Anse de Cul-Rond, Jobourg and Gréville) and the British Islands (Guernsey and possibly Aurigny; Fig. 1). Interbedded metasediments and metavolcanics of unknown ages that are crosscut by porphyroclastic granitic orthogneisses (gneiss “oeillés”) can be observed in most of these outcrops. Only some of the latter were dated and yielded protolith ages around 2 Ga and the age of the other gneiss types making up the Icartian remains unknown.

Evidence for the occurrence of an old gneissic basement in the Armorican Massif were recognized since the end of the 19th century, based first on relative chronology observations in the field and crude comparisons with other gneisses from elsewhere in Europe. After Barrois (1898, Barrois 1908), Cogné (1959) proposed to use “Pentévrien” to name the Armorican gneissic basement onto which rest unconformably the so-called Brioverian volcanic and sedimentary rocks that are nowadays assigned to the Neoproterozoic. Using also some observations from the island of Guernsey, Graindor (1960) renamed the gneissic basement “Sarnien” (from Sarnia, the latin name of the island). With the beginning of geochronology, using the Rb-Sr methods on samples from the Icart Point – the locality on Guernsey Island from which will derive the name Icartian (Roach *et al.*, 1972) –, Adams (1967) obtained a date of ca. 2 Ga, suggesting for the first time that Paleoproterozoic rocks do occur in the Armorican Massif. Then, additional K-Ar and Rb-Sr dates from other localities yielded older ages encompassing a time period between 2.7 and 0.9 Ga. This Precambrian Armorican gneissic basement was temporally thought to be widespread and polycyclic, with three successive orogenic periods, among which the oldest one was named Icartian (Roach *et al.*, 1972; Cogné, 1972; Leutwein *et al.*, 1973; Adams, 1976; Roach, 1977). However, as the K-Ar and Rb-Sr isotopic methods are very sensitive to thermal perturbations, most of the obtained dates at the time did not correspond to any geological event and did not confirm

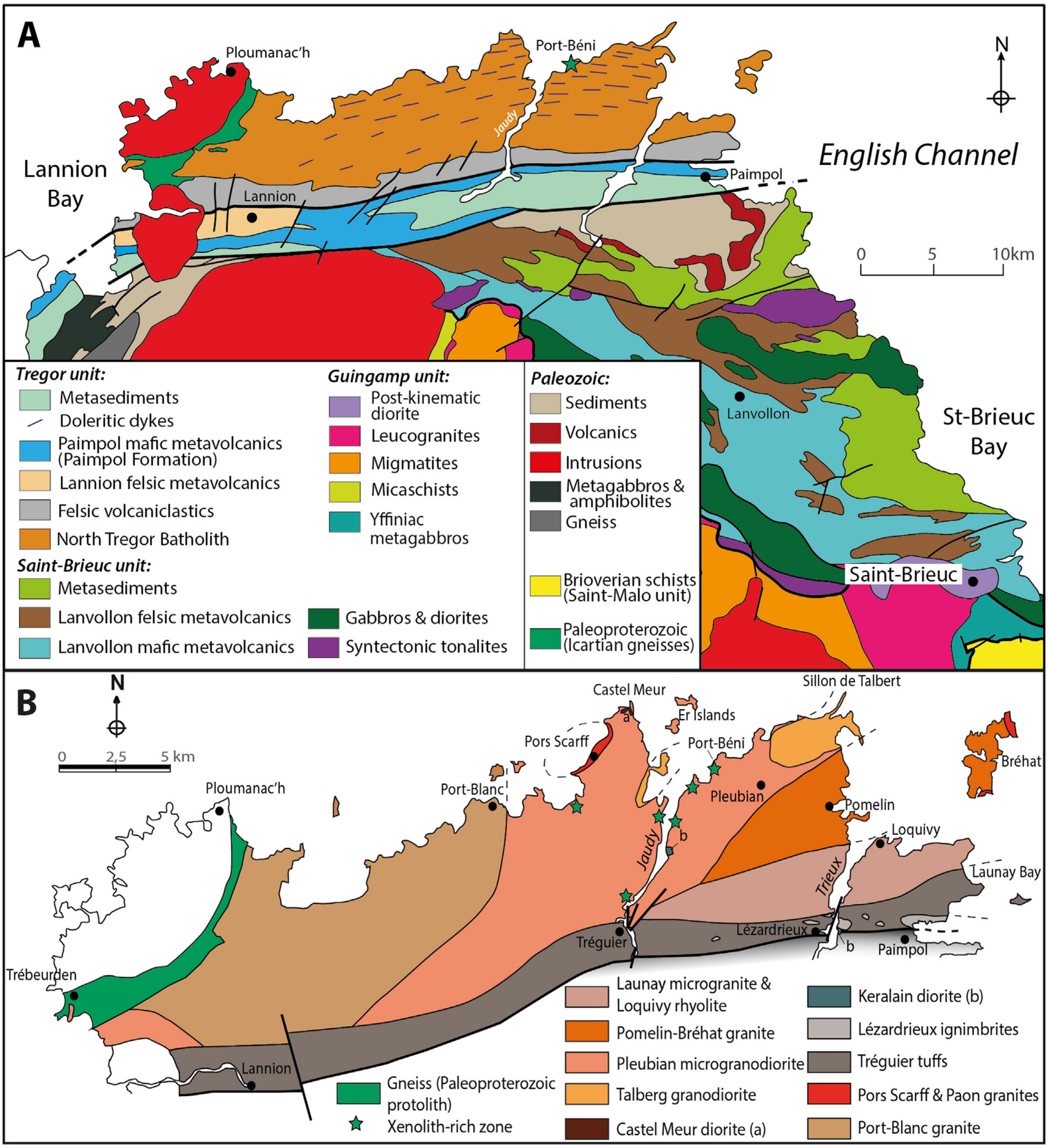

Fig. 2. (A) Sketch map identifying the Trégor, Saint-Brieuc and Guinguamp Cadomian tectonometamorphic units to the west of the Saint-Brieuc Bay (modified from Plaine, 2011, unpublished, and from the geological map of France 1/50 000 "vecteur harmonisée", BRGM, available at http://infoterre.brgm.fr; see also: Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001); note that some of the volcanic and sedimentary rocks from the southern Trégor unit (*e.g.*, Paimpol Formation) may alternatively be associated with the Saint-Brieuc unit (see text; Chantraine *et al.*, 2001); (B) Simplified geological map of the North Trégor volcano-plutonic complex (from Auvray, 1979; Graviou, 1984); dykes are not shown.

the polycyclic Precambrian history of the Armorican Massif (Vidal, 1980). Thus, as the geochronological methods improved, it was soon established that many of the Armorican gneisses were in fact Neoproterozic in age (*e.g.*, Vidal *et al.*, 1974; 1981; Égal *et al.*, 1996), with the exception of a few localities where Paleoproterozoic ages of about 2 Ga were finally obtained and confirmed (Fig. 3).

Calvez and Vidal (1978) published the first U-Pb data on zircon from the Icart Point orthogneisses in the Guernsey Island (Fig. 1). They obtained a discordia line yielding two dates at 2018 ± 15 Ma (2018 ± 68 recalculating the analytical error, Linnemann *et al.*, 2008) and 583 ± 90 Ma, interpreted respectively as the age of an Icartian magmatic protolith and the age of a Cadomian metamorphic event that affected this

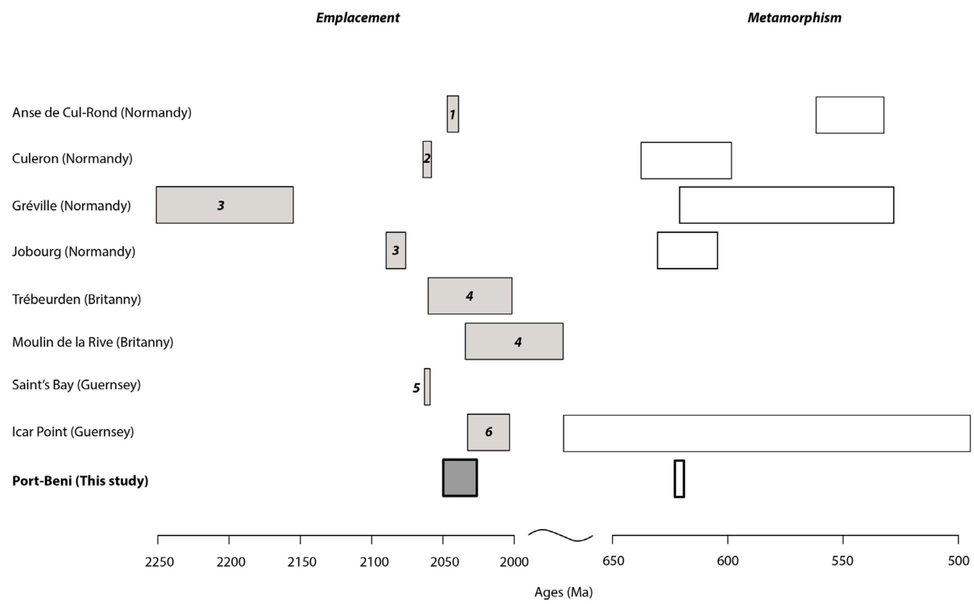

Fig. 3. Overview of the protolith and metamorphic ages obtained on Icartian orthogneisses from the Armorican Massif: (1) Martin *et al.* (2018); (2) Inglis *et al.* (2004); (3) Piton (1985); (4) Auvray *et al.* (1980); (5) Calvez and Vidal (1978); (6) Samson and D'Lemos (1998).

rock (Fig. 3). Samson and D'Lemos (1998) published a U-Pb age of 2061 ± 2 Ma for the Saint's Bay gneisses, close to the Icart Point (Fig. 3). In Britany, the Lannion's Bay – Morguignen-Moulin de la Rive to the West and Trébeurden to the East – and the Port-Béni orthogneisses (Figs. 1 and Fig 2) were also found to be Paleoproterozoic in age with U-Pb ages on zircon of, respectively, 2000 +35/-30 Ma, 2031 +36/-28 Ma and 1790 +19/-17 Ma (Fig. 3; Auvray, 1979; Auvray *et al.*, 1980; Graviou *et al.*, 1988). There again, the isotopic systems were re-opened by the Cadomian or Hercynian thermal events (*e.g.*, Vidal *et al.*, 1981). In Normandy, the orthogneisses from North-Cotentin yielded upper intercept ages of 2083 ± 7 Ma in Jobourg and 2203 ± 49 Ma Ga in Gréville (Fig. 3, Piton, 1985, unpublished data). The lower intercept dates where respectively of 618 ± 26 and 576 +45/-48 Ma (Fig. 3). It remained unclear for a while whether the granitic Icartian rocks from the British Islands, Britany and Normandy, underwent a metamorphic event at about 2 Ga or not (*e.g.*, Tribe *et al.*, 1996). Performing in-situ U-Pb analyses on zircon grains from the Cul-Rond orthogneisses, near Jobourg (Fig. 1), Inglis *et al.* (2004) found inherited cores that provided a protolith age of 2061.0 ± 2.7 Ma and growth rims, witnessing a Cadomian metamorphism at about 618 Ma, which supports the hypothesis that the first metamorphism affecting the Icartian magmatic protoliths of these orthogneisses occurred in fact during the Neoproterozoic (Fig. 3). This was recently confirmed by the study of Martin *et al.* (2018), who dated zircon crystals from an orthogneiss from the same locality (Anse du Cul-Rond, Fig. 1) and found a date of 2043.1 ± 3.8 Ma interpreted as the emplacement age of the protolith while one concordant analysis yielded a date of 547 ± 15 Ma that they interpreted as reflecting the Cadomian metamorphism in this area (Fig. 3).

In addition to these orthogneisses, detrital zircon grains extracted from Neoproterozoic (Ballouard *et al.*, 2018; Gougeon *et al.*, 2018; 2022) and Paleozoic sediments (Ducassou *et al.*, 2014), as well as xenocrystic zircon from Cadomian intrusions (*e.g.*, Guerrot 1989; Samson and D'Lemos 1998; Inglis *et al.*, 2004), provide other evidence for the presence of an Icartian basement aside and beneath the Armorican Massif. Most of these published ages are bracketed between 2.2 and 1.8 Ga. In addition, some rare Meso- to Neo-Archean, as well as a Mesoproterozoic zircon grains, have also been found, illustrating that moderate magmatic production and metamorphism were active during these periods, before the Neoproterozoic eo-Cadomian cycle began about 750 Ma ago (Égal *et al.*, 1996).

In the area of Port-Béni (Figs. 1, Fig 2 and Fig 4), the Icartian gneisses correspond to xenoliths dispersed within the Neoproterozoic Pleubian microgranodiorites (Auvray, 1979; Graviou, 1984). Together with the Talbert granodiorite, these microgranodiorites form one of the major intrusive units of the North Trégor calc-alkaline batholith (Fig. 2B; note that Talbert also spells Talberg, former name of the "Sillon de Talbert", the local coastal spit). The sources of these magmas were both the mantle wedge and the continental crust above a subduction zone along a Neoproterozoic continental margin according to the geodynamical scheme of the Cadomian orogeny (Auvray and Maillet, 1977; Auvray, 1979; Graviou and Auvray, 1985, 1990; Graviou *et al.*, 1988; Égal *et al.*, 1996). The Talbert granodiorite provided zircon crystal fractions that yielded an emplacement age of ca. 615 Ma for the batholith (Graviou, 1984; Graviou *et al.*, 1988). This age was also proposed for the volcanic rocks (Tréguier tuff, Lézardrieux ignimbrite, Loguivy rhyolite, Launay microgranite) from the south of the Trégor unit thus forming a volcano-plutonic complex (Fig. 2B; Ruffet *et al.*, 1991; Thiéblemont *et al.*, 1996; Chantraine *et al.*, 2001). Making up most of the Trégor unit, this complex corresponded then to a volcanic arc installed on a segment of old continental crust, along the subduction zone (Auvray, 1979; Graviou, 1984; Graviou *et al.*, 1988; Chantraine *et al.*, 2001). At the scale of the belt, whether the slab subducted to the north or to the south remains a puzzling question addressed in many papers and for which there is still no consensus. In a hypothesis that still holds (*e.g.*, Auvray, 1989; Graviou and Auvray, 1990; Tribe *et al.*, 1996), the slab was first thought to dip to the SE (Auvray and Maillet, 1977; Auvray, 1979), before additional arguments also support the hypothesis of a NW-dipping slab (Balé, 1986; Auvray, 1989; Brun and Balé, 1990; Graviou, 1992; Hébert, 1993; Chantraine *et al.*, 2001 and other references therein for both hypothesis).

Subsequently, the North Trégor volcano-plutonic complex was crosscut by numerous, dominantly mafic and E-W striking dykes, which, in the western Trégor, predate the intrusion of the Ploumanac'h granitic magmas at the end of the Carboniferous, ca. 300 Ma ago (Barrois 1908; Auvray, 1979; Vidal, 1980). Further west, in the Bay of Morlaix, other mafic intrusions that are Devonian to Carboniferous in age also crosscut the Trégor unit (Barnenez metadolerites and Saint-Jean-du-Doigt gabbro; Pochon *et al.*, 2023 and references therein). Among the dykes within the North Trégor volcano-plutonic complex, based on structural and petro-geochemical characteristics, as well as cross cutting relationships in the field, several swarms were characterized (Barrois 1908; Auvray *et al.*, 1976; Auvray, 1979; Lees *et al.*, 1991; Roach *et al.*, 1992; Lahaye *et al.*, 1995). The oldest dykes (spilitic dykes and "albitophyres d'Er"; Er being a group of islands fronting the estuary of the Jaudy river, near Port-Béni) named group 1 on Figure 4 form a first swarm that is basaltic to andesitic in composition and exhibits a calc-alkaline geochemical affinity, which links them to the very last episodes of the Trégor volcano-plutonic complex activity (Auvray, 1979). The Trieux dolerites, which are younger and more abundant across the whole Trégor volcano-plutonic system, show tholeiitic basaltic compositions and are therefore distinct from that of group 1 (Auvray, 1979; Lees *et al.*, 1991). Their compositions compare to that of continental tholeiites (Lahaye *et al.*, 1995). Among them, two groups of dykes that possibly represent distinct events, were distinguished, but in two different ways. In the first one, a group of dykes (defined as group 2) was observed to be crosscut at least locally (Port-Béni and neighboring area) by more abundant Fe-rich, often thicker dykes belonging to a group 3 (Lees *et al.*, 1991; Roach *et al.*, 1992). This group 3 also contains rare composite dykes displaying felsic centers and mafic margins (Lees *et al.*, 1991). In the second study, a widespread group of dolerites (defined as the Launay-type) with a high total FeO/MgO ratio was distinguished from a local group (the Talbert-type dolerites) that exhibits a lower total FeO/MgO ratio (Lahaye *et al.*, 1995). Two remarks are worth to note about these two studies. First, a careful examination of the available petrological and geochemical data clearly shows that groups 2 and 3 are comparable to the Talbert and the Launay dolerites, respectively (see Sup. App. 1). Second, Roach *et al.* (1992) suggested that the two groups of dykes emplaced sub-contemporaneously at the scale of the Trégor unit, as they mention to have observed that a few group 2 dolerites were younger than some group 3 dykes, calling into question the reality of distinct dyking episodes for these Trégor dolerites. Nonetheless, the ages of the Trégor dyking episodes are still poorly constrained, although these dolerites are considered to be Paleozoic in age (Ordovician to Carboniferous) by all authors (Auvray *et al.*, 1976; Lees *et al.*, 1991; Ruffet *et al.*, 1992; Roach *et al.*, 1992; Lahaye *et al.*, 1995). In addition, in the Trégor unit, rare younger lamprophyric dykes, also of unknown age, crosscut the dolerites (Auvray, 1979).

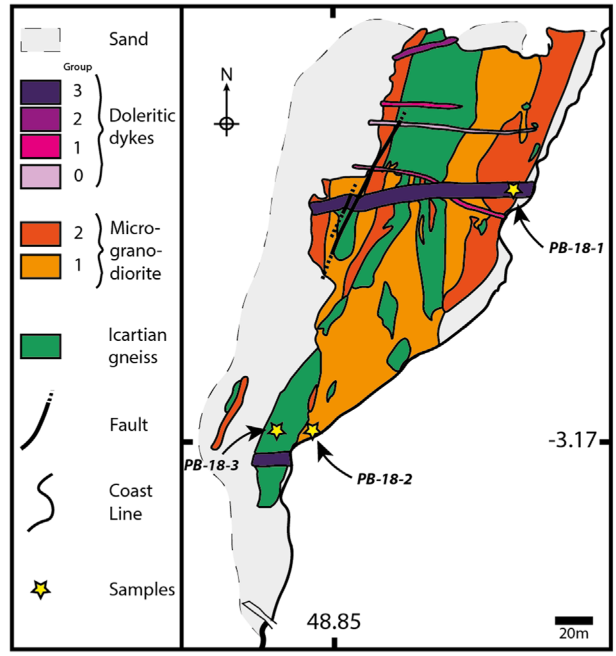

Fig. 4. Simplified geological map of the Port-Béni coastal foreshore and sampling locations. Microgranodiorites 1 and 2: porphyritic Pleubian microgranodiorite, phenocryst-rich and phenocryst-poorer facies, respectively. Doleritic dykes: 1 to 3, compositional groups from Lees *et al.* (1991) and Roach *et al.* (1992); 0, undifferentiated from any of these groups; 1, spilites and "albitophyres d'Er" (Auvray *et al.*, 1976; Auvray, 1979); 2 and 3, Trieux dolerites, Talbert and Launay dolerites (Lahaye *et al.*, 1995), respectively. Numbers also underline relative chronologies. Note that only two structural groups of doleritic dykes can be identified on the map (see text).

A distinctive feature of the Trégor unit is that it was weakly deformed during the Cadomian orogeny (*e.g.*, Chantraine *et al.*, 2001). By opposition, to the south, the contiguous, tectonically juxtaposed Saint-Brieuc unit (Fig. 2A) experienced stronger deformation at ca. 590-580 Ma (Ballèvre *et al.*, 2001). In addition, simultaneously the Saint-Brieuc unit was also increasingly metamorphosed towards the south (greenschist to amphibolite facies; Hébert, 1993; 1995; Hébert and Ballèvre, 1993; Hébert *et al.*, 1993; 1997; Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001). According to Chantraine *et al.* (2001), from north to south, the Saint-Brieuc unit comprises a volcano-sedimentary succession, a widespread bimodal volcanic suite (Lanvollon Formation) with a detrital cover, both crosscut by gabbro-dioritic to tonalitic synkinematic intrusions, and a composite orthogneiss complex forming a basement slice (Pentevrian from the eastern Saint-Brieuc Bay; Cogné, 1959) with ages as old as ca. 750 Ma (Égal *et al.*, 1996; Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001). Situated in between two major faults (Auvray, 1972; Chantraine *et al.*, 2001) and located in between the Saint-Brieuc and the Trégor units, the northern volcano-sedimentary succession can in fact be either associated with any of these tectonometamorphic units (see Fig. 3 in Chantraine *et al.*, 2001, *versus* Fig. 2 in

Ballèvre *et al.*, 2001 and Fig. 2A), or with none of them. This volcano-sedimentary succession includes basal subaqueous lava flows (pillow lava spilitic metabasalts) exhibiting typical arc-tholeiite compositions with felsic lava intercalations (Paimpol Formation) that yielded a Pb/Pb zircon age at 610 ± 9 Ma (Barrois, 1908; Auvray, 1979; Égal *et al.*, 1996). Together with the slightly younger Lanvollon metavolcanic rocks (588 ± 11 Ma; Pb/Pb zircon age) whose mafic terms have continental tholeiite compositions (Égal *et al.*, 1996), the Paimpol Formation and its sedimentary cover are assumed to be the witness of an intra-arc basin that emplaced on a thinned continental margin (Hébert, 1993; Chantraine *et al.*, 2001). It is also worth noting that some of the synkinematic intrusions of the Saint-Brieuc unit (*e.g.*, Coëtmieux-Fort la Latte intrusion, east of the Saint-Brieuc Bay) are composite and comprise mafic to intermediate rocks of mostly tholeiitic compositions and intermediate to felsic rocks that are mostly calc-alkaline (Égal *et al.*, 1996).

3 Samples and methods

3.1 Sampling

Three samples were collected on the Port-Béni rocky foreshore for dating (Fig. 4). Sample PB-18-1 was collected from one of the youngest E-W doleritic dyke (Fig. 4), identified as belonging to group 3 by Lees *et al.* (1991) and Roach *et al.* (1992). Sample PB-18-2 corresponds to a phenocryst-rich facies (microgranodiorite 1, Fig. 4) of the Pleubian microgranodiorite while sample PB-18-3 is a typical granitic orthogneiss with K-feldspar porphyroclasts (gneiss "oeillés") selected from one of the composite xenoliths found within the microgranodioritic body.

3.2 LA-ICP-MS U-Pb dating

A classical mineral separation procedure has been applied to concentrate zircon and apatite grains for U-Pb dating using the facilities available at the Géosciences Rennes laboratory (University of Rennes-CNRS, France). The samples were crushed and only the powder fraction with a diameter $< 250\,\mu m$ was kept. Heavy minerals were first concentrated by Wilfley table, then heavy minerals were separated with heavy liquids. Magnetic minerals were then removed with an isodynamic Frantz separator. Zircon and apatite grains were handpicked under a binocular microscope. The selected minerals were then embedded in epoxy mounts, which were grounded and polished on a lap wheel. Zircon and apatite grains were imaged by cathodoluminescence (CL) using a Reliotron CL system equipped with a digital color camera available at the GeOHeLiS platform (OSUR observatory, University of Rennes).

U-Pb geochronology of zircon (Zrn) and apatite (Ap) grains was conducted by in-situ laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at the GeOHeLiS analytical platform using an ESI NWR193UC Excimer laser coupled to an Agilent quadripole 7700x ICP-MS equipped with a dual pumping system to enhance sensitivity (Paquette *et al.*, 2014). The instrumental conditions are reported in the Supplementary Table 1 for apatite and Supplementary Table 2 for zircon.

Ablation spot diameters of $30\,\mu m$ and $25\,\mu m$ (depending on the size of the zircon grain) and rectangles of $30 \times 80\,\mu m$ (Ap), with repetition rates of 3 Hz (Zrn) and 5 Hz (Ap) and a fluence of $6\,J/cm^2$ were used. Data processing was carried out with the software Iolite 4 (Paton *et al.*, 2010; Chew *et al.*, 2014) with error propagation according to Horstwood *et al.* (2016). Data were corrected for U–Pb and Th–Pb fractionation and for mass bias by standard bracketing with repeated measurements of the GJ-1 zircon (Jackson *et al.*, 2004) and the Madagascar apatite (Cochrane *et al.*, 2014) reference materials, respectively. Along with the unknowns, zircon Plešovice (337.13 ± 0.37 Ma; Sláma *et al.*, 2008) and apatite reference materials McClure (523.51 ± 2.09 Ma; Schoene and Bowring, 2006) and Durango (31.44 ± 0.18 Ma; McDowell *et al.*, 2005) were measured to monitor precision and accuracy of the analyses, and produced ages of 337.7 ± 3.3 Ma (Plešovice, N = 18, MSWD = 3.4), 521.8 ± 5.6 Ma (McClure, N = 18, MSWD = 0.72) and 31.3 ± 1.7 Ma (Durango, N = 22, MSWD = 1.6) during the course of the analyses (Sup. Tab. 3). Further information on the dating protocol is given in in Nosenzo *et al.* (2022) for zircon and in Pochon *et al.* (2016) for apatite.

Concordia diagrams were generated with IsoplotR (Vermeesch, 2018). All errors given in the Supplementary Table 4 and Table 5 are listed at two standard errors (in percent), and where data are combined for concordia age or weighted mean calculations, the final results are provided with 95% confidence limits.

4 Results

4.1 Field observations and petrological results

At Port-Béni, the gneisses always show a strong ductile foliation that is crosscut by undeformed younger rocks, although a weak, late brittle deformation affects all the rocks. On the simplified geological map (Fig. 4), the gneissic xenoliths often take the form of large NNE-SSW trending elongated rafts mostly found within the phenocryst-rich microgranodiorite 1. These enclaves are usually smaller within the phenocryst-poorer microgranodiorite 2. In the field, the sharp, subparallel and mostly steeply dipping to subvertical contacts with the gneissic host-rocks, as well as the mostly NNE-SSW striking contacts between the two types of microgranodiorite, give the two porphyritic rocks the appearance of dykes. The contacts between the micro-granodiorites being diffuse and progressive, the observed difference in phenocryst contents further suggests that microgranodiorite 2 formed syn-plutonic dyke-like intrusions within microgranodiorite 1, which was then a more viscous, crystal-rich magma.

During field mapping of the area shown in Figure 4, we were not able to distinguish the three groups of doleritic dykes described by Lees *et al.* (1991) and Roach *et al.* (1992). Neither could we attribute the undifferentiated dyke quoted 0, which was incorrectly mapped by these authors, to one of their three groups. In addition, none of the local dolerites (Fig. 4) exhibits some of the distinctive macroscopic features that often characterize the group 1–spilites and “albitophyres d’Er” – dykes elsewhere in the Trégor area (see Auvray, 1979). In the absence of crosscutting relationships, thin sections and/or geochemical data are indeed essential to distinguish the Port-Béni groups of dykes, as defined by Lees *et al.* (1991) and Roach *et al.* (1992). At most, from our own observations only, a first order structural analysis of the map shows that a swarm of thinner and dominantly E-W striking dykes (identified as groups 1 and 0 on Fig. 4) might actually be older than the thicker and WSW-ENE trending dykes (groups 2 and 3 on Fig. 4).

The largest xenoliths are composite and made up of mostly two types of gneisses: migmatitic foliated and banded gneisses with garnet bearing melanosomes are found in direct, often progressive contacts with more abundant foliated granitic orthogneisses. The orthogneisses may contain centimetric K-feldspar porphyroclasts, or not, and often show scarce, more mafic gneissic enclaves and veins. In sections perpendicular to the foliation, the eye-like shape of these feldspars, due to recrystallization (at least partial and under stress) of former phenocrysts, underlines the rock fabric and indicates that metamorphism and deformation were synchronous. In thin section, the typical porphyroclastic facies selected for dating (sample PB-18-3) exhibits a granoblastic texture (Fig. 5A and Fig 5B). The perthitic K-feldspar porphyroclasts bear inclusions of plagioclase with albitic (Auvray, 1979) edgings, quartz, apatite and zircon. Plagioclase, perthitic to micro-perthitic microcline, quartz and chloritized biotite are the main phases in the matrix, reflecting the overall monzogranitic to granodioritic composition of the protolilth. Epidote is accessory. Late tiny veins, cross cutting the mineral fabric and intersecting each other, are made of feldspar, quartz and feldspar or chlorite, epidote and feldspar (Fig. 5A and Fig 5B). They may also include subhedral and euhedral zircon crystals, as well as apatite. All the observations (see also Auvray, 1979) indicate high-grade metamorphism and deformation of the xenoliths, up to partial melting of the least refractory lithologies, under amphibolite to granulite-facies conditions, with retrograde overprint in the greenschist facies.

In the porphyritic microgranodiorites hosting the xenoliths, the phenocrysts may exceed 70 vol % in microgranodiorite 1 and consist mostly of euhedral and altered (sericitised) plagioclase, euhedral hornblende and clinopyroxene pseudo-morphs or relics (Fig. 5C and Fig 5D). Fresh to chloritized biotite is also present. Plagioclase phenocrysts are zoned (labrador to oligoclase) and exhibit fresh albitic edgings (Auvray, 1979; Graviou 1984). Actinolite-chlorite clusters, including opaque minerals that sometimes show a skeletal habitus, often replace the clinopyroxene (bottom right of C & D, Fig. 5). The microgranular to fine-grained, granular mesostasis includes abundant quartz and feldspars (plagioclase and K-feldspar), but also hornblende and biotite, and opaque minerals (magnetite-ilmenite, Graviou, 1984), apatite, zircon, titanite and allanite, as accessory phases. In the phenocryst-poorer microgranodiorite 2, the phenocrysts remain abundant (up to about 40 vol %) and result in plagioclase and rare quartz. The mesostasis is granophyric, showing graphic-like quartz-feldspar associations, especially around the phenocrysts. Such textural and mineralogical-content variations are typical within the Pleubian microgranodiorite subunit of the Trégor batholith. They locally illustrate the overall composite nature of the Pleubian-Talbert unit (Auvray, 1979; Graviou, 1984), thus resulting from the accumulation of several granodioritic magma intrusions with diffuse and progressive contacts between them, which confirm that they were syn-plutonic.

In the sampled dyke, the aphyric dolerite exhibits a typical intergranular texture, mainly showing a framework of euhedral plagioclase laths with, mostly, anhedral augite and skeletal to subhedral opaque minerals (ilmenite and magnetite; Auvray, 1979) as interstitial phases (Fig. 5E and Fig 5F). Acicular apatite with large aspect ratios (length/width > 10) and quartz are the main accessory phases. Quartz-feldspar intergrowths and biotite are also present. Chlorite, epidote, actinolite, leucoxene, carbonate, sericite and euhedral pyrite are secondary. As expected from sampling, these characteristics are consistent with those of the group 3 dolerites defined by Lees *et al.* (1991) and Roach *et al.* (1992). As previously exposed (section 2), they are also similar to those of the Launay-type dolerites defined by Lahaye *et al.* (1995).

4.2 Geochronological results

For the orthogneiss sample PB-18-03, zircon grains were prismatic, stubby to elongated, with elongation ratios ranging from 1.5 to 3, and sizes from 50 to 250 microns. Some of the grains were yellow while others were pink in color. We therefore decided to mount them separately. Cathodoluminescence imaging of the yellow zircon grains (Sup. Fig. 1) reveals the presence of an inner core, usually very bright, and a darker outer rim for most of the grains. Some of the grains present growth zoning surrounding the core (*e.g.*, grain “ai” on Sup. Fig. 1), otherwise most of the grain outer rims are rather homogeneous (*e.g.*, grain “ae” on Sup. Fig. 1). For the pink crystals, the presence of bright cores is also evidenced (Sup. Fig. 2). Outer rims are either homogeneous (see for example rim1 on grain 1 or 2’ on Sup. Fig. 2) or zoned (*e.g.*, rim1 from grain 10 on Sup. Fig. 2).

For the pink zircon, seventy-five analyses out of 55 grains and, for the yellow one’s, forty analyses out of 27 grains were performed (Sup. Tab. 4). The inner cores yield variable U and Pb contents (1106-51 and 1942-12 ppm, respectively) while most of the Th/U ratios are above 0.1 (mean=0.23 ± 0.15). Their outer rims also yield variable U (3185-114 ppm) and Pb (682-3 ppm) contents while the Th/U are systematically lower than 0.1 (mean=0.05 ± 0.03). For the yellow grains, the U and Pb contents for the inner cores range in between 2156-59 ppm and 2659-27 ppm, respectively, while they are lower for the outer rims (1338-245 ppm for U and 173-34 ppm for Pb). The Th/U ratios are generally above 0.1 for the cores (mean=0.18 ± 0.13) and always below 0.1 for the rims (mean=0.05 ± 0.01).

Despite the color differences, all the grains have been plotted in the same concordia diagram (see Fig. 6A). With the exception of one analysis, they plot along a discordia yielding an upper intercept date of 2039.5 ± 23 Ma and a lower intercept date of 631 ± 6 Ma (MSWD=1.6). A first group of 30 concordant analyses acquired in the inner zone of the zircon grains allow to calculate a concordia date of 2038 ± 12 Ma (MSWD [Conc+Equiv]=2.4; Fig. 6B). With the exception of 5 analyses, they all yield Th/U ratios above 0.1. A second group of 68 concordant data, acquired in

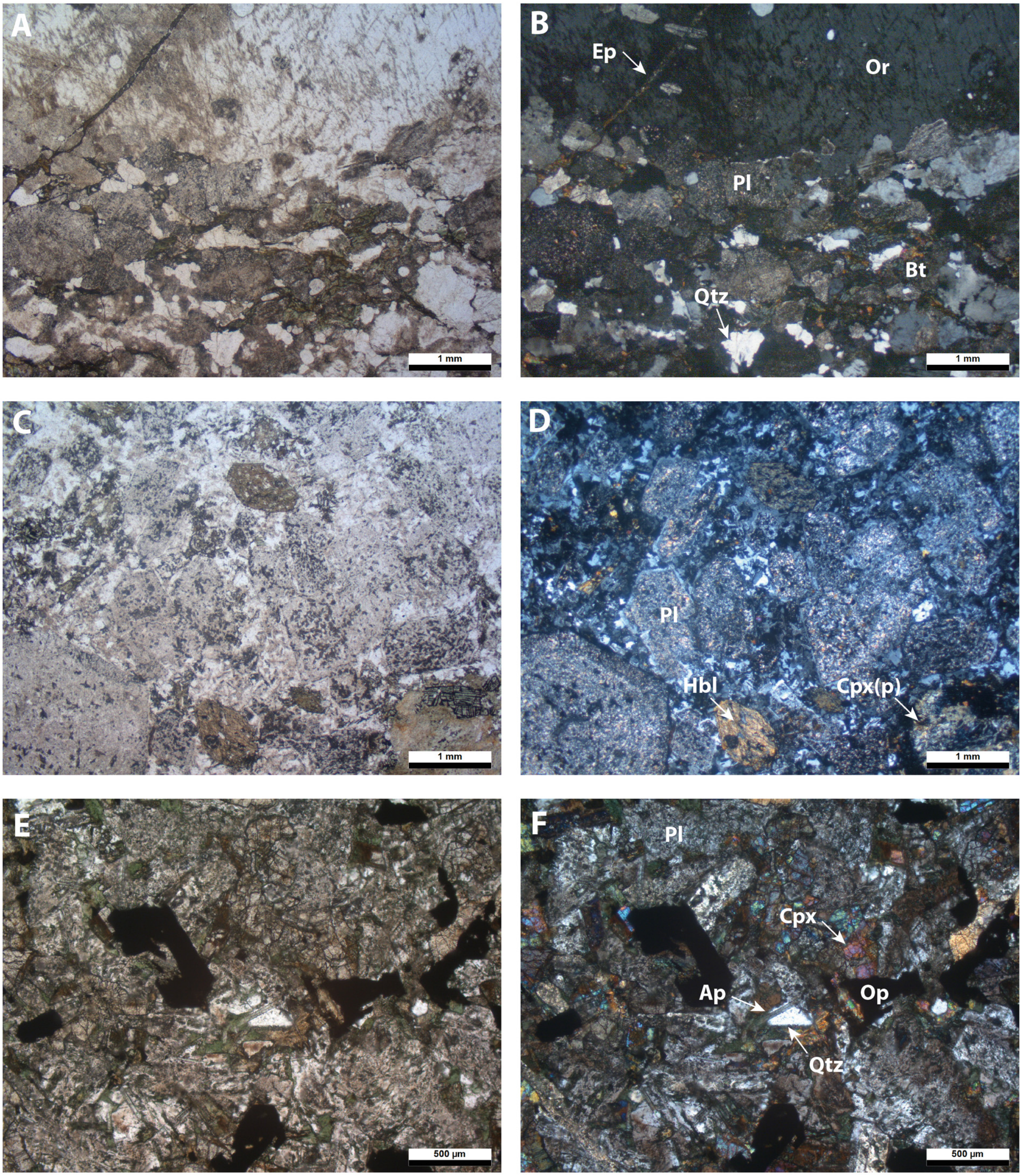

Fig. 5. Typical textures and mineralogical contents, in thin sections, of the dated orthogneiss (A & B, sample PB-18-3), microgranodiorite (C & D, sample PB-18-2) and dolerite (E & F, sample PB-18-1) from Port-Béni. Plan polarized light (A, C & E) and cross-polarized light (B, D & F). Ap: apatite; Bt: biotite; Cpx: clinopyroxene; Cpx(p): clinopyroxene pseudomorph; Ep: epidote (in a vein, A &B); Hbl: hornblende; Op: opaque mineral; Or: orthoclase; Pl: plagioclase; Qtz: quartz.

the outer parts of the zircon grains and/or outside the cores, yield a younger concordia date of 621 ± 2 Ma (MSWD [Conc+Equiv]=1.8; Fig. 6C). All these analyses but three present Th/U below 0.1. The remaining sixteen analyses plot in a discordant position between the two previously calculated concordant dates (Fig. 6A).

In addition, sixty-nine apatite grains extracted from the same orthogneiss sample have been analyzed (Sup. Tab. 5). Apatite appears as prismatic stubby to slightly elongated grains with lengths ranging from 60 to 200 microns. Cathodoluminescence imaging of the grains (Sup. Fig. 3) reveals that they luminesce homogeneously in the purple

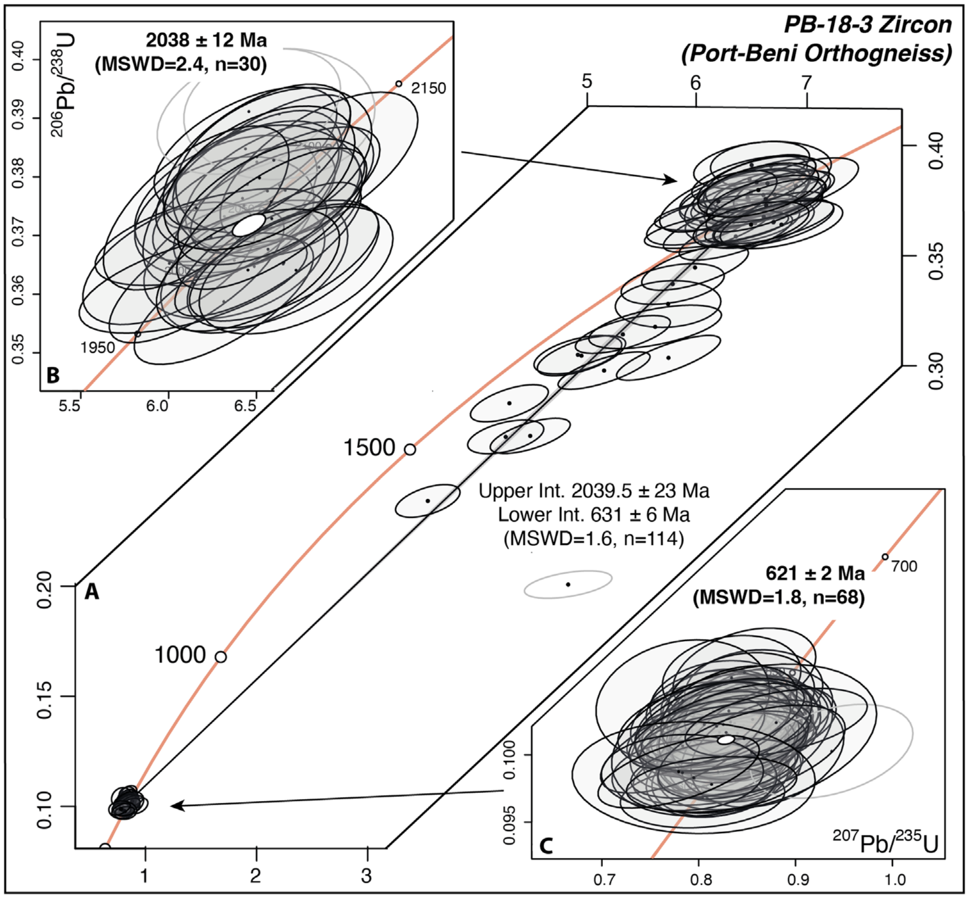

Fig. 6. U-Pb Zircon dating from the orthogneiss sample PB-18-3. (A) Wetherill concordia diagram for all the 114 different analyses and the resulting discordia line with a upper intercept date of 2039.5 ± 23 Ma and a lower intercept date of 631 ± 6 Ma; note the intermediate position of some analyses between the upper and the lower clusters; (B) Wetherill concordia diagram with a close up on the upper concordant cluster that yields a concordia date of 2038 ± 12 Ma; (C) Wetherill concordia diagram with a close up on the lower concordant cluster yielding a concordia date of 621 ± 2 Ma.

color (activation by REE, Rakovan and Reeder, 1994, usually Eu^{2+} and Ce^{3+}, Waychunas, 2002). Their compositions are fairly consistent in terms of U (21 ± 9 ppm) and Pb (6.1 ± 0.7 ppm) contents (Sup. Tab. 5). They also have consistent low Th/U ratios (0.2 ± 0.08). In a Tera-Wasserburg diagram (Fig. 7A), they plot in a discordant position due to the presence of initial common lead. All the analyses plot along a discordia that yields a lower intercept date of 614.5 ± 5.5 Ma (MSWD=0.45) and a $^{207}Pb/^{206}Pb$ initial ratio of 0.897 ± 0.029, a value consistent with that of common Pb at 620 Ma, according to Stacey and Kramers (1975).

In the microgranodiorite sample PB-18-2, zircon grains are prismatic and stubby with elongation ratios ranging from 1.3 to 2.2 and sizes between 40 and 200 microns. Cathodoluminescence imaging (Sup. Fig. 4) shows that the grains are rather simple, being either homogeneous or presenting growth zoning. Forty-six analyses were performed out of 39 grains (Sup. Tab. 4). Their average U (232 ± 143 ppm) and Pb (194 ± 123 ppm) contents are fairly consistent as their Th/U ratios (mean=0.33 ± 0.1). In a concordia diagram (Fig. 7B), thirty-seven analyses plot in a concordant position and yield a concordia date of 604.5 ± 2.0 Ma (MSWD=1.8). Four analyses yield an upper intercept date of 2025 ± 69 Ma (MSWD=0.62, not shown). The remaining five analyses (grey ellipses on Fig. 7B) are interpreted as the result of a very slight Pb loss.

Finally, no zircon grains were extracted from the doleritic dyke sample PB-18-01. Fortunately, apatite was quite abundant in this sample. The grains were elongated and, in cathodoluminescence, they are homogeneous and luminesce in the purple color, a frequent color in intermediate to mafic rocks (Waychnas, 2002; Sup. Fig. 5). Sixty-two apatite grains were analyzed (Sup. Tab. 5). They were all poor in U (1.8 ± 0.4 ppm, in average) and Pb (1.4 ± 0.3 ppm, in average) and present high Th/U ratios (mean 5.3 ± 0.3). In a Tera-Wasserburg diagram (Fig. 7C), they all plot in a discordant position owing for the presence of a variable proportion of initial common Pb. They define a discordia with a lower intercept date of 597 ± 15 Ma (MSWD=0.82) and a $^{207}Pb/^{206}Pb$ initial ratio of 0.827 ± 0.045, a value consistent with that of Stacey and Kramers (1975) for common Pb at this date.

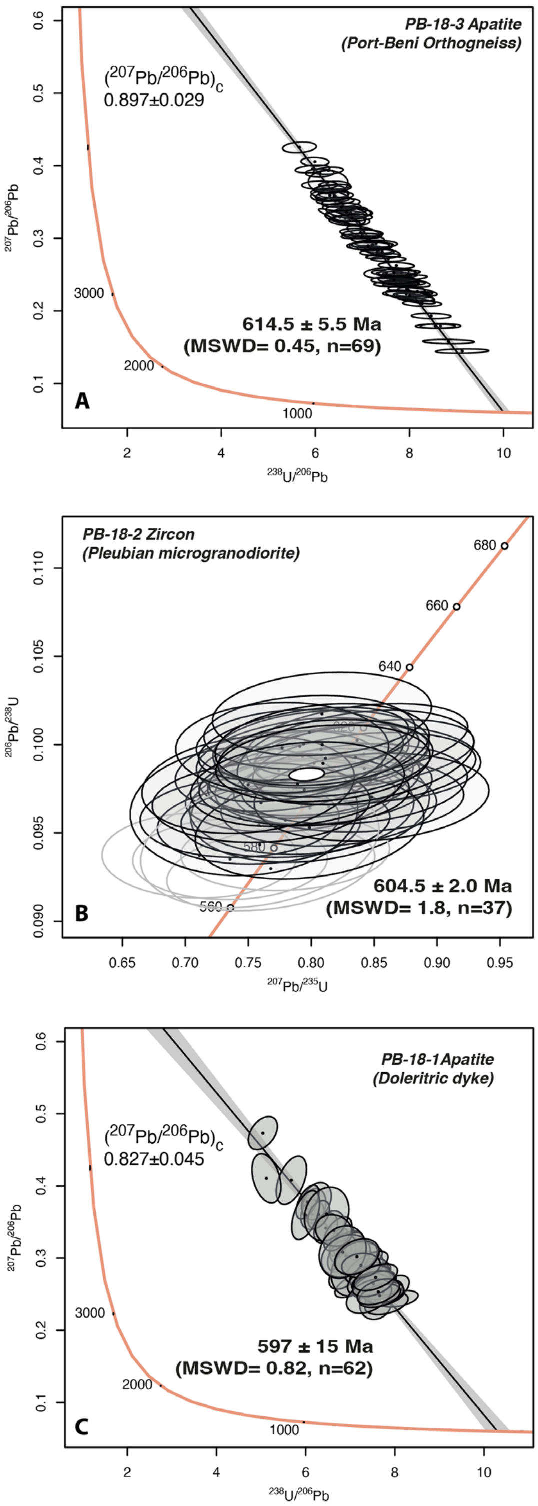

Fig. 7. (A) Tera-Wasserburg diagram for the apatite dating for the orthogneiss sample PB-18-3 yeilding a lower intercept date of 614.5 ± 5.5 Ma; (B) Wetherill concordia diagram for the zircon grains extracted from the granodiorite sample PB-18-2 with a concordia date of 604.5 ± 2.0 Ma; (C) Tera Wasserburg concordia diagraom for the apatite grains extracted from the doleritic dyke sample PB-18-1 yielding a lower intercept date of 597 ± 15 Ma.

5 Discussion

Four different new ages corresponding to four successive distinct events can be distinguished from the present study. They are consistent with geological observations and provide new insights into the history of the northern Armorican domain, especially during the late Neoproterozoic. Indeed, as expected, the obtained dates from the orthogneiss zircon and apatite grains can confidently be interpreted as corresponding to the emplacement and crystallization age of the magmatic protolith of the gneiss during the Paleoproterozoic (2038 ± 12 Ma) and to the age of the late Neoproteroic metamorphism and deformation (621 ± 2 Ma on zircon and 614.5 ± 5.5 Ma on apatite) that the Icartian granitic rock underwent. Then, two successive late Neoproterozoic intrusive events, whose respective ages can be deduced from the zircon data provided by the Pleubian microgranodiorite (604.5 ± 2.0 Ma) and from the apatite date given by the late doleritic dyke (597 ± 15 Ma), provide new information for the Cadomian history at Port-Béni, for the Trégor unit as a whole, and beyond.

In more details, the emplacement age at 2038 ± 12 Ma of the protolith of the Port-Béni orthogneiss is provided by the analyses of the zircon inner cores characterized by Th/U ratios above 0.1, a feature usually associated with a magmatic origin (Hoskin and Black, 2000). This age is confirmed by the upper intercept age obtained from all the measurements at 2039.5 ± 23 Ma. It is significantly older than the age of 1790 +19/-17 Ma yielded by zircon fractions from similar rocks of the same outcrop (Auvray, 1979; Auvray *et al.*, 1980; Graviou *et al.*, 1988). This age difference may either reflect the composite nature of the local Icartian xenoliths or the difficulty to obtain accurate ages from zircon grains that record a complex history, before the introduction of in-situ analyses. However, as the analyses defining the discordia obtained by Auvray *et al.* (1980) are grouped close to the lower intercept of the resulting discordia, the upper intercept date is poorly defined. Therefore, our new age very likely reflects the age of the granitic protolith of the orthogneiss at Port-Béni, fitting quite well with the other ages at ca. 2 Ga obtained elsewhere in the Trégor unit (Baie de Lannion; Auvray *et al.*, 1980), in the British Islands (Calvez and Vidal, 1978; Samson and D'Lemos, 1998), and for most of the samples dated in Normandy (Piton, 1985; Inglis *et al.*, 2004; Martin *et al.*, 2018), as one slightly older protolith age at ca. 2.2 was also found (Piton, 1985; Fig. 3).

The date of 621 ± 2 Ma obtained from the zircon rims from the orthogneiss yield Th/U ratios lower than 0.1, values that are typically found in metamorphic zircon (Corfu *et al.*, 2003). This date is interpreted as the age of the metamorphic and deformation event that affected the granitoid. This age is confirmed by the dating of the apatite grains from the same sample that yield an age of 614.5 ± 5.5 Ma. These apatite grains present rather low Th/U ratios (ca. 0.2) linked to high U and low Th contents (Sup. Tab. 5). This, together with the fact that these apatite crystals luminesce in the purple color (due to a high content in REE), could be the evidence that this mineral is metamorphic in origin (see Henrichs *et al.*, 2019 and references therein). As the retrograde epidote-bearing veins crosscutting the gneiss contain both zircon and apatite (see section 4), these new ages must probably be considered as minimum with respect to the peak metamorphic conditions. Nevertheless, these ages found on rocks from Brittany confirm what was already known for Normandy (Inglis *et al.*, 2004; Martin *et al.*, 2018), namely that the zircon grains did not record any Paleoproterozoic metamorphic event. Instead, in the Armorican Massif, the orthogneissic remnants of the Icartian basement underwent major deformation and metamorphism only during the late Neoprotezoic. In addition, this suggests that the protolith of these orthogneisses was emplaced after any metamorphic and/or deformation event that might have taken place during the Icartian orogenic cycle.

It follows from our results that, in the Trégor unit, the Cadomian deformation and metamorphism episode is older than the one that affected the adjacent southern Saint-Brieuc unit (Ballèvre *et al.*, 2001). It can also be noticed that the age of ca. 620 Ma for the metamorphism in the Trégor unit is comparable with the one obtained by Inglis *et al.*, (2004) in Normandy, near Jobourg (Fig. 3). It is however older than the date of ca. 550 Ma found with a single concordant zircon analysis from the same location by Martin *et al.* (2018). If we compare our results to the ages obtained so far for other Icartian gneisses from the Armorican Massif (Fig. 3), we can notice that most of these gneiss protoliths were emplaced between 2.08 and 2.02 Ga and were affected by a metamorphic event around 620 Ma. Therefore, we might question the meaning of the age of ca. 550 Ma published by Martin *et al.* (2018), based on a single measurement.

At 604.5 ± 2.0 Ma, the emplacement and crystallization age of the microgranodiorite that intruded the Icartian basement is slightly younger than the metamorphic and deformation event that affected the orthogneiss. It has to be noticed that some of the zircon grains from the microgranodiorite yielded a date of 2025 ± 69 Ma within error of the emplacement age of the Port-Béni orthogneiss protolith. They are interpreted as xenocrysts probably extracted from this orthogneiss or a similar one. Consistently, Graviou (1984) and Graviou *et al.* (1988) also evidenced heritage in the magmas of the Trégor batholith, and the source of the latter involved a significant crustal component (Égal *et al.*, 1996). Our new age of 605 ± 2 Ma is slightly younger, within the given uncertainty, than the age of 615 +13/-7 Ma proposed for the crystallization of the Talbert granodiorite, which together with the Pleubian microgranodiorite forms one the of the major intrusive unit of the batholith (Auvray, 1979; Graviou, 1984; Graviou *et al.*, 1988). As expected by Graviou *et al.* (1988), our result confirms that the age of ca. 615 Ma was overestimated due to the presence of inherited radiogenic lead in the zircon fractions. Given the syn-plutonic nature of the contacts between the granodioritic magmas of the composite unit (Graviou, 1984), it is expected that the ages of the subunits should be comparable. Therefore, one can rather confidently consider that the Talbert-Pleubian units of the Trégor bathotith were emplaced and crystallized ca. 605 Ma ago. However, extending the meaning of this new age to the whole batholith is more puzzling. Indeed, this calls for caution, firstly because the construction of the batholith itself is poorly documented (Graviou, 1984), and secondly because such magmatic systems are known to build up episodically over variable time periods that may be as long as a few tens of millions of years (*e.g.*, Saint-Blanquat *et al.*, 2011). Therefore, without more constraints, it remains possible that the duration of amalgamation of the units exceeded the rather small uncertainty (± 2 Ma) that we obtained on a single intrusion from these units. However, as the granodioritic rocks were not deformed under ductile conditions and crosscut the gneiss foliation, this work establishes that the whole batholith was emplaced between ca. 620 Ma and ca.597 Ma, *i.e.*, the age of the metamorphic event that affected the Icartian basement and that obtained for one of the latest doleritic intrusive event in the Trégor unit.

Indeed, the doleritic dyke that crosscuts all the previous lithologies as well as older dykes (Fig. 4) was emplaced 597 ± 15 Ma ago, age obtained on the apatite extracted from this sample. This age provides additional constraints on the history of the Trégor unit. First, it establishes that the Trégor doleritic dyke swarm is Ediacaran, and not Paleozoic in age. Thus, if there is any in the Trégor unit, only the lamprophyric dykes, which are younger (Auvray, 1979), could still be of Paleozoic age, and this remains to be determined. Second, as the Trégor dolerites also intrude volcanic rocks associated with the batholith (Auvray, 1979), it indicates that the Trégor volcano-plutonic complex (including the group 1 dykes, see section 2) as a whole was emplaced before ca. 597 Ma. This confirms that the Trégor volcanic rocks are older than previously thought, as it has already been suggested (Ruffet *et al.*, 1991; Thiéblemont *et al.*, 1996; Chantraine *et al.*, 2001). Third, considering the analytical uncertainties on the ages (604.5 ± 2.0 Ma and 597 ± 15 Ma), the age of ca. 597 Ma indicates that one of the latest doleritic dyking episode occurred within a maximum time lapse of ca. 25 My following the emplacements of the Pleubian-Talbert intrusive unit at ca. 605 Ma and that of the calc-alkaline feeder dykes (group 1, spilites and "albitophyres d'Er"). Of course, the shorter this time lapse, the more likely it is that the doleritic diking episodes corresponding to the group 2-Talbert and the group 3-Launay dyke swarms were sub-contemporaneous at the scale of the Trégor unit, as suggested by Roach *et al.* (1992) and by a structural analysis of the map (Fig. 4; see sections 2 and 4.1, respectively). It is plausible that this time lapse started with a period while magma production was reduced or even temporarily stopped, before starting again under new conditions. Indeed, the evolution of the compositions of the produced magmas from calc-alkaline to tholeiitic (Auvray, 1979; Lees *et al.*, 1991) demonstrates a significant change of the Cadomian subduction dynamics within this time interval of maximum 25 My. In the Trégor unit, this new geodynamical regime is fully compatible with the opening of a basin within the continental arc that was then recently in extension, as it has been proposed within the adjacent Saint-Brieuc unit to the south (Hébert, 1993; Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001).

Consequently, it is very likely that the doleritic dykes of the Trégor unit may have fed the subaqueous lava flows and volcanic deposits of the Paimpol Formation (Auvray, 1979; Égal *et al.*, 1996) or, more precisely, a northern lateral equivalent, now eroded, of this formation. Several observations support such a hypothesis. Indeed, the overall E-W direction of the doleritic dykes indicates a clear mostly N-S direction of extension. The ages of both the late dyke (597 ± 15 Ma, this work) and of the volcanic formation (610 ± 9 Ma; Égal *et al.*, 1996) are comparable within uncertainty. Both display subordinate felsic rocks and dominant mafic rocks exhibiting tholeiitic compositions (Égal *et al.*, 1996; Lees *et al.*, 1991; Lahaye *et al.*, 1995). Comparison of the available chemical compositions show that many Paimpol mafic rocks (Auvray, 1979; Égal *et al.*, 1996) have compositions that match those of the group 2-Talbert-type dyke or encompass, for some elements, the composition gaps between the two groups of dykes (*e.g.*, low to intermediate total FeO/MgO ratios and low TiO_2 contents; Lees *et al.*, 1991; Lahaye *et al.*, 1995 ; see Sup. App. 1). However, it is worth noting that the arc-like tholeiitic signature of the subaqueous Paimpol mafic lavas (Auvray, 1979; Égal *et al.*, 1996) is compatible with the combined effects of subduction and continental extension on magma production. In contrast, the continental tholeiitic signature of the dolerite dykes from the Trégor unit (Lahaye *et al.*, 1995) suggests a rather moderate influence of the subduction compared to that of extension. This is also the case for the continental tholeiites from the southern Lanvollon Formation in the Saint-Brieuc unit. These are slightly younger (588 ± 11 Ma) than the ones from the northern Paimpol Formation (Égal *et al.*, 1996), supporting the idea that the influence of subduction on magma production progressively diminished from north to south over time. In fact, it is as if the arc signature had disappeared to the north (Trégor doleritic dykes) but persisted further south (Paimpol Formation) before disappearing (Lanvollon Formation) as extension was developing in the Saint-Brieuc unit. However, this signature is still perceptible in the syntectonic intrusions of the Saint-Brieuc unit down to ca. 580 Ma (Égal *et al.*, 1996; Ballèvre *et al.*, 2001).

Before and during the tectonic juxtaposition of the units, the available geochronological data show that the main Cadomian volcanic arc migrated over time. First located during the early Cadomian (eo-Cadomian) within the Saint-Brieuc unit (Égal *et al.*, 1996; Chantraine *et al.*, 2001), it moved to the north within the present Trégor unit ca. 605 Ma ago, before moving back to the southern Saint-Brieuc unit at ca. 590-580 Ma (Chantraine *et al.*, 2001; Ballèvre *et al.*, 2001). Such a migration may likely relate to a change in the dip of the subduction slab. Assuming a northward subduction (Chantraine *et al.*, 2001), it follows that after an early period of decreasing dip, the subducting slab straightened with a steeper slope, making possible a southward retreat of an oceanic trench accommodating the onset of extension in the continental arc above the subduction (*e.g.*, Lallemand *et al.,* 2005; Heuret and Lallemand, 2005). This hypothesis requires further investigation but provides a plausible explanation for the opening of the intra-arc basin at the same time that arc magmas were still produced in the Saint-Brieuc unit. It may also explain the composite tholeiitic and calc-alcaline signatures of some of the magmatic mafic and felsic rock associations from the southern Trégor and the Saint-Brieuc units (*e.g.*, Paimpol Formation, Coëtmieux-Fort la Latte intrusion; Égal *et al.*, 1996). This hypothesis is also compatible with a significant decrease of the influence of the subduction on magma production to the north when the doleritic dykes emplaced in the Trégor unit after ca. 605 Ma.

6 Conclusion

Based on a geochronological study of representative samples from Port-Béni, this work sheds light on the geological history of the Trégor area and, more generally, of the North Armorican domain. It shows that the Paleoproterozoic granitic rocks from Port Béni were emplaced earlier than previously thought, 2038 ± 12 Ma ago, before being deformed and transformed into orthogneisses at 621 ± 2 Ma or just before, given the greenschist-facies alteration these rocks underwent. Not surprisingly, deformation and metamorphism preceded the emplacement of the calc-alkaline magmas of the North Trégor batholith, and of its associated dykes (spilites and “albitophyres d’Er”) and volcanic rocks. The Pleubian-Talbert granodioritic unit of this complex was emplaced and crystallized at 604.5 ± 2.0 Ma. Caution should be taken when considering that this is also the age of the other units making up the system, which may have been built up episodically over a period much longer than 4 My (the uncertainty on the age), between ca. 620 Ma and ca. 597 Ma. A late doleritic dyke from Port-Béni yields an age of 597 ± 15 Ma, which makes the Trégor dolerites a late Neoproterozoic intrusive event that must have occurred after ca. 605 Ma and before ca. 580 Ma. In contrast to the calc-alkaline nature of the older magmas, the tholeiitic signature of these magmas marks a significant evolution of the geodynamical environment of the Cadomian subduction zone at that time. Although it remains unclear whether a single or two successive episodes of doleritic dyking occurred at the scale of the Trégor unit, this change coincides with the opening of the intra-arc basin of the southern, adjacent Saint-Brieuc unit. Thus, the doleritic dykes from the Trégor unit probably fed northern equivalents of the subaqueous eruptions of the Paimpol Formation, and this, further north than the Saint-Brieuc unit itself, after ca. 605 Ma, but before the more recent episode of metamorphism and deformation that affected this unit. From ca. 605 Ma beneath the Trégor unit and down to ca. 580 Ma beneath the Saint-Brieuc unit, the steepening of a northward subducting Cadomian slab may explain both the formation of the intra-arc basin and the simultaneous production of magma with composite calc-alkaline and tholeiitic signatures, as well as the decreasing influence of subduction on magma production from north to south over time.

Supplementary material

Supplementary Appendix 1: Comparisons of the geochemical compositions of the Trégor dolerites and of the dolerites with mafic to intermediate volcanic rocks from the Paimpol Formation

Supplementary Figure 1: Cathodoluminescence imaging of the yellow zircon grains extracted from the orthogneiss sample PB-18-3. Letters refer to the grain numbers noted in Supplementary Table 4.

Supplementary Figure 2: Cathodoluminescence imaging of the pink zircon grains extracted from the orthogneiss sample PB-18-3. Numbers refer to the grain numbers noted in Supplementary Table 4.

Supplementary Figure 3:: Cathodoluminescence imaging of the apatite grains extracted from the orthogneiss sample PB-18-3.

Supplementary Figure 4:: Cathodoluminescence imaging of the zircon grains extracted from the granodiorite sample PB-18-2. Numbers refer to the grain numbers noted in Supplementary Table 4.

Supplementary Figure 5:: Cathodoluminescence imaging of some the apatite grains extracted from the doleritic dyke sample PB-18-1.

Supplementary Table 1:: Operating conditions for the apatite LA-ICP-MS dating in this study.

Supplementary Table 2: Operating conditions for the zircon LA-ICP-MS dating in this study

Supplementary Table 3: U-Pb analyses of the primary and secondary reference materials analyzed during this study.

Supplementary Table 4: U-Pb analyses of zircon grains from sample PB-18-3 and PB-18-2 analyzed during this study.

Supplementary Table 5: U-Pb analyses of apatite grains from sample PB-18-3 and PB-18-1 analyzed during this study.

The Supplementary Material is available at https://www.bsgf.fr/10.1051/bsgf/2024011/olm.

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References

Adams CJD. 1967. A geological and related isotopic study of rocks from north-western France and the Channel Islands (UK). PhD thesis dissertation, Oxford University (UK), unpublished.

Adams CJD. 1976. Geochronology of the Channel Islands and adjacent French mainland. *J Geol Soc London* 132 (3): 233–250. https://doi.org/10.1144/gsjgs.132.3.0233.

Auvray B. 1972. Les accidents tectoniques majeurs du Trégor. Age et signification. *C R Acad Sci Paris, série D*, 274: 2940–2943.

Auvray B. 1979. Genèse et évolution de la croÛte continentale dans le Nord du Massif Armoricain. Thèse d'État, université de Rennes (France), unpublished.

Auvray B. 1989. Pentevrian and Brioverian in the North Armorican Massif: characteristics and geo-dynamic evolution. Unpublished archive. https://insu.hal.science/insu-01576103.

Auvray B, Maillet P. 1977. Volcanisme et subduction au Protérozoïque supérieur dans le Massif armoricain (France). *Bull Soc géol France* 19 (5): 953–957. https://doi.org/10.2113/gssgfbull.S7-XIX.5.953.

Auvray B, Charlot R, Vidal P. 1980. Données nouvelles sur le Protérozoïque inférieur du domaine nord-armoricain (France) : âge et signification. *Can J Earth Sci* 17 (4): 532–538. https://doi.org/10.1139/e 80-050.

Auvray B, Lefort JP, Monnier JL. 1976. Feuille de Tréguier et notice. Carte géologique de France 1/50 000 N°171. Orléans (France): BRGM.

Balé P. 1986. Tectonique cadomienne en Bretagne Nord – interaction décrochement-chevauchement : champs de déformation et modélisations expérimentales. Thèse université de Rennes (France), unpublished.

Ballèvre M, Le Goff E, Hébert R. 2001. The tectonothermal evolution of the Cadomian belt of northern Brittany, France: a Neoproterozoic volcanic arc. *Tectonophysics* 331 (1-2): 19–43. https://doi.org/10.1016/S0040-1951(00) 00234-1.

Ballouard C, Poujol M, Zeh A. 2018. Multiple crust reworking in the French Armorican Variscan belt: implication for the genesis of uranium-fertile leucogranites. *Int J Earth Sci* 107 (7): 2317–2336. https://doi.org/10.1007/s00531-018- 1600-3.

Barrois C. 1898. Sur le gisement des roches cristallines anciennes du massif de Paimpol. 1re et 2e note. *Ann Soc géol Nord* 27: 22–29 et 265–267.

Barrois C. 1908. Légende de la feuille de Lannion. Carte géologique de France 1/80 000 N°41. *Ann Soc géol Nord* 37: 205–227.

Brun JP, Balé P. 1990. Cadomian tectonics in northern Brittany. In D'Lemos RS, Strachan RA, Topley CG, eds. *The Cadomian orogeny*. Geol Soc London, Spec Publ 51, pp. 95–114. https://doi.org/10. 1144/GSL. SP. ;1; 1990. 051. 01. 07.

Calvez JY, Vidal P. 1978. Two billion years old relicts in the Hercynian belt of Western Europe. *Contrib Mineral Petrol* 65 (4): 395–399. https://doi.org/10.1007/ BF00372286.

Chantraine J, Égal E, Thiéblemont D, Le Goff E, Guerrot C, Ballèvre M, *et al.* 2001. The Cadomian active margin (North Armorican Massif, France): a segment of the north Atlantic Panafrican belt. *Tectonophysics* 331 (1-2): 1–18. https://doi.org/10.1016/S0040-1951(00)00233-X.

Chew DM, Petrus JA, Kamber BS. 2014. U-Pb LA-ICPMS dating using accessory mineral standards with variable common Pb. *Chem Geol* 363: 185–199. https://doi.org/10.1016/j.chemgeo.2013.11.006.

Cochrane R, Spikings RA, Chew DM, Wotzlaw JF, Chiaradia M, Tyrrell S, *et al.* 2014. High temperature (>350 °C) thermochronology and mechanisms of Pb loss in apatite. *Geochim Cosmochim Acta* 127: 39–56. https://doi.org/10.1016/j.gca.2013.11.028.

Cogné J. 1959. Données nouvelles sur l'antécambrien dans l'Ouest de la France : Pentévrien et Briovérien en Baie de Saint-Brieuc. *Bull Soc géol France* S7-I (1): 112–118. https://doi.org/10.2113/gssgfbull.S7-I.1.112.

Cogné J. 1972. Le Briovérien et le cycle orogénique cadomien dans le cadre des orogènes fini-précambriens. In: *Actes du Colloque International sur les corrélations du Précambrien*, Agadir-Rabat, 3–23 mai 1970. Notes et Mémoire du Service géologique du Maroc (Rabat) 236, pp. 193–218.

Corfu F, Hanchar JM, Hoskin PWO, Kinny P. 2003. Atlas of zircon textures. *Rev Mineral Geochem* 53: 469–500. https://doi.org/10.2113/0530469.

Ducassou C, Poujol M, Ruffet G, Bruguier O, Ballèvre M. 2014. Relief variation and erosion of the Variscan belt: detrital geochronology of the Palaeozoic sediments from the Mauges Unit (Armorican Massif, France). In Schulmann K, Martínez Catalán JR, Lardeaux JM, Janoušek V, Oggiano G, eds. *The Variscan Orogeny: extent, timescale and the formation of the European crust*. Geol Soc London Spec Publ 405, pp. 137–167. https://doi.org/10. 1144/sp405. 6.

Égal E, Guerrot C, Le Goff E, Thiéblemont D, Chantraine J. 1996. The cadomian orogeny revisited in northern Brittany (France). In Nance RD, Thompson MD eds. *Avalonian and related peri-Gondwana terranes of the circum-north-Atlantic*. Geol Soc Am Spec Pap 304 pp. 281–318. https://doi. org/10. 1130/0-8137- 2304–3. 281.

Gougeon R, Néraudeau D, Dabard MP, Pierson-Wickmann AC, Polette F, Poujol M, *et al.* 2018. Trace Fossils from the Brioverian (Ediacaran-Fortunian) in Brittany (NW France). *Ichnos* 25 (1): 11–24. https://doi.org/10.1080/10420940.2017.1308865.

Gougeon R, Néraudeau D, Loi A, Poujol M. 2022. New insights into the early evolution of horizontal spiral trace fossils and the age of the Brioverian series (Ediacaran-Cambrian) in Brittany, NW France. *Geol Mag* 159 (7): 1284–1294. https://doi.org/10.1017/S0016756820001430.

Graindor MJ. 1960. Existence vraisemblable du Penthévrien dans les îles Anglo-Normandes et La Hague. *Bull Soc Linn Normandie* 10: 44–46.

Graviou P. 1984. Pétrogenèse des magmas calco-alcalins : exemple des granitoïdes cadomiens de la région trégorroise (Massif Armoricain). Thèse université de Rennes (France), unpublished.

Graviou P. 1992. Reconnaissance d'une suture majeure au sein de la chaîne cadomienne. C R Acad Sci Paris, s 2, 315 (13): 1799–1802.

Graviou P, Auvray B. 1985. Caractérisation pétrographique et géochimique des granitoïdes cadomiens du domaine nord-armoricain: implications géodynamiques. *C R Acad Sci Paris, série II*, 301 (5): 315–318.

Graviou P, Auvray B. 1990. Late Precambrian M-type granitoid genesis in the Cadomian belt of NW France. In D'Lemos RS, Strackan RA, Topley CG, eds. *The Cadomian Orogeny*. Geol Soc London Spec Publ 51, pp. 231–244. https://doi. org/10. 1144/GSL. SP. 1990. 051. 01. 14.

Graviou P, Peucat JJ, Auvray B, Vidal P. 1988. The Cadomian orogeny in the northern Armorican Massif. Petrological and geochronological constraints on a geodynamic model. *Hercynica: Bull Soc géol minér Bretagne* 4 (1): 1–13.

Guerrot C. 1989. Archéen et Protérozoïque dans la chaîne hercynienne ouest européenne. Géochimie isotopique (Sr-Nd-Pb) et géochronologie U-Pb sur zircons. Thèse université de Rennes (France). *Mém Doc Centre Arm Étude Struct Socles* 25, 180 p.

Hébert R. 1993. Évolution tectonométamorphique d'un arc insulaire au Protérozoïque supérieur : le domaine de Saint-Brieuc (Massif armoricain). Thèse université de Rennes (France). *Doc BRGM* 228, 356 p.

Hébert R. 1995. Evidence for multiple high-T metamorphism and regional migmatization within the Cadomian belt of northern Brittany, France. *J Geol Soc London* 152 (2): 213–216. https://doi.org/10.1144/gsjgs.152.2.0213.

Hébert R, Ballèvre M. 1993. Petrology of staurolite-bearing metapelites from the Cadomian belt, northern Brittany (France): constraints on low-pressure metamorphism. *Bull Soc géol France* 164 (2): 215–228.

Hébert R, Le Goff E, Ballèvre M. 1997. Évolution métamorphique des domaines de Saint-Brieux et de Guingamp (Massif armoricain) : implications géodynamiques pour l'orogène cadomien. *Géologie de la France* 1: 3–22.

Hébert R, Hallot E, Guerrot C, Chantraine J. 1993. New structural, petrological and radiometric constraints within the Cadomian belt: chronology of events in the baie de Saint-Brieuc, Northern Armorican Massif (France). *C R Acad Sci Paris, série 2,* 316 (3): 395–401.

Heuret A, Lallemand S. 2005.Plate motions, slab dynamics and back-arc deformation. *Phys Earth Planet Interiors* 149 (1-2): 31–51. https://doi.org/10.1016/j.pepi.2004.08.022.

Hoskin PWO, Black LP. 2000. Metamorphic zircon formation by solid-state recrystallization of protolith igneous zircon. *J Metamorph Geol* 18: 423–439. https://doi.org/10.1046/j. 1525-1314. 2000.00266.x.

Henrichs IA, Chew DM, O'Sullivan GJ, Mark C, McKenna C, Guyett P. 2019. Trace element (Mn-Sr-Y-Th-REE) and U-Pb isotope systematics of metapelitic apatite during progressive greenschist to amphibolite facies Barrovian metamorphism. *Geochem Geophys Geosyst* 20 (8):4103–4129. https://doi.org/10.1029/2019 GC008359.

Horstwood MSA, Košler J, Gehrels G, Jackson SE, Mc Lean NM, Paton C, *et al.* 2016. Community-derived standards for LA-ICP-MS U-(Th-)Pb geochronology – Uncertainty propagation, age interpretation and data reporting. *Geostand Geoanalyt Res* 40 (3): 311–332. https://doi., org/10.1111/j.1751-908X. 2016. 00379.x.

Inglis JD, Samson SD, D'Lemos RS, Hamilton M. 2004. U-Pb geochronological constraints on the tectonothermal evolution of the Paleoproterozoic basement of Cadomia, la Hague, NW France. *Precamb Res* 134: 293–315. https://doi.org/10.1016/j.precamres.2004.07.003.

Jackson SE, Pearson NJ, Griffin WL, Belousova EA. 2004. The application of laser ablation-inductively coupled plasma-mass spectrometry to in situ U-Pb zircon geochronology. *Chem Geol* 211 (1-2): 47–69. https://doi.org/10.1016/j.chemgeo.2004.06.017.

Lahaye Y, Blais S, Auvray B, Ruffet G. 1995. Le volcanisme fissural paléozoïque du domaine nord-armoricain. *Bull Soc géol France* 166 (5): 601–612. https://doi.org/10.2113/gssgfbull.166.5.601.

Lallemand S, Heuret A, Boutelier D. 2005. On the relationships between slab dip, back-arc stress, upper plate absolute motion, and crustal nature in subduction zones. *Geochem Geophys Geosyst* 6 (9), Q09006. https://doi.org/DOI doi:10.1029/2005 GC000917.

Lees GJ, Roach RA, Rowbotham G. 1991. The Trégor dyke Swarm at Port Béni, North Brittany: a preliminary study. *Proc Ussher Soc* 7 (4): 338–343.

Leutwein F, Power GM, Roach RA, Sonet J. 1973. Quelques résultats géochronologiques obtenus sur des roches d'âge précambrien du Cotentin. *C R Acad Sci Paris, série D*, 276: 2121–2124.

Linnemann U, D'Lemos RS, Drost K, Jeffries T, Gerdes A, Romer RL, *et al.* 2008. Cadomian tectonics. In: McCann T, ed.*Geology of Central Europe*, vol 1: Precambrian and Palaeozoic. Geol Soc London, pp. 103–154. https://doi. org/10. 1144/CEV1P. 3.

Martin E, François C, Paquette JL, Capdevila R, Lejeune AM. 2018. Petro-geochemistry and zircon U-Pb dating of the late Variscan Flamanville granodiorite and its Paleoproterozoic basement (Normandy, France). *Géologie de la France* 1: 34–48.

McDowell FW, McIntosh WC, Farley KA. 2005. A precise 40Ar-39Ar reference age for the Durango apatite (U-Th)/He and fission-track dating standard. *Chem Geol* 214 (3): 249–263. https://doi.org/10.1016/j.chemgeo.2004.10.002.

Nosenzo F, Manzotti P, Poujol M, Ballèvre M, Langlade J. 2022. A window into an older orogenic cycle: P-T conditions and timing of the pre-Alpine history of the Dora-Maira Massif (Western Alps). *J Metamorph Geol* 40 (4): 789–821. https://doi.org/10.1111/jmg.12646.

Paquette JL, Piro JL, Devidal JL, Bosse V, Didier A, Sanac S, *et al.* 2014. Sensitivity enhancement in LA-ICP-MS by N2 addition to carrier gas: application to radiometric dating of U-Th-bearing minerals. *Agilent ICP-MS J* 58: 4–5.

Paton C, Woodhead JD, Hellstrom JC, Hergt JM, Greig A, Maas R. 2010. Improved laser ablation U-Pb zircon geochronology through robust downhole fractionation correction. *Geochem Geophys Geosyst* 11 (3), Q0 AA06. https://doi.org/10.1029/2009 GC002618.

Piton P. 1985. Étude géochimique et géochronologique de la croÛte continentale protérozoïque (1,8-2,2 Ga) au Nord du Massif Armoricain. Mémoire DEA, université de Rennes (France), unpublished.

Pochon A, Poujol M, Gloaguen E, Branquet Y, Cagnard F, Gumiaux C, *et al.* 2016. U-Pb LA-ICP-MS dating of apatite in mafic rocks: Evidence for a major magmatic event at the Devonian-Carboniferous boundary in the Armorican Massif (France). *Am Mineral* 101 (11): 2430–2442. https://doi.org/10.2138/am- 2016-5736.

Pochon A, Iacono-Marziano G, Gloaguen E, Tuduri J, Erdmann S. 2023. High-temperature alteration during cooling of mafic intrusions: Insights from the Saint-Jean-du-Doigt intrusive complex (Armorican Massif, France). *Lithos*, 436-437: 106977. https://doi.org/10.1016/j.lithos.2022.106977.

Rakovan J, Reeder RJ. 1994. Differential incorporation of trace elements and dissymmetrization in apatite; the role of surface structure during growth. *Am Mineral* 79 (9-10): 892–903.

Roach RA. 1977. A review of the Precambrian rocks of the British variscides and their relationships with the precambrian of N. W. France. Colloque international du CNRS n°243 sur la chaine Varisque d'Europe moyenne et occidentale et Programme international de corrélation géologique (Projet UNESCO - PICG n°22), *Rennes*, 25 septembre – 6 octobre 1974. Actes pp. 61–79.

Roach RA, Lees GJ, Rowbotham G. 1992. Le champ filonien paléozoïque du Trégor, Nord du massif armoricain. *C R Acad Sci Paris, série II*, 315 (7): 813–820.

Roach RA, Adams CJD, Brown M, Power GM, Ryan P. 1972. The precambrian stratigraphy of the Armorican Massif, NW France. *Proc Int Geol Congr* 24 (1): 246–252.

Ruffet G, Féraud G, Amouric M. 1991. Comparison of 40Ar-39Ar conventional and laser dating of biotites from the North Trégor Batholith. *Geochim Cosmochim Acta* 55 (6): 1675–1688. https://doi.org/10.1016/0016-7037(91)90138-U.

Ruffet G, Perroud H, Féraud G. 1992. Palaeomagnetism and 40Ar/39Ar dating of the Trégor dolerites (Armorican Massif, France). *Tectonophysics* 201 (1-2): 121–140. https://doi.org/10.1016/0040-1951(92)90179-A.

Saint-Blanquat M (de), Horsman E, Habert G, Morgan S, Vanderhaeghe O, Law R *et al.* 2011. Multiscale magmatic cyclicity, duration of pluton construction, and the paradoxical relationship between tectonism and plutonism in continental arcs. *Tectonophysics* 500 (1-4): 20–33. https://doi.org/10.1016/j.tecto.2009.12.009.

Samson SD, D'Lemos RS. 1998. U-Pb geochronology and Sm-Nd isotopic composition of Proterozoic gneisses, Channel Islands, UK. *J Geol Soc London* 155 (4): 609–618. https://doi.org/10.1144/gsjgs.155.4.0609.

Schoene B, Bowring S. 2006. U-Pb systematics of the McClure Mountain syenite: thermochronological constraints on the age of the 40Ar/39Ar standard MMhb. *Contrib Mineral Petrol* 151 (5): 615–630. https://doi.org/10.1007/s00410-006-0077-4.

Sláma J, Košler J, Condon DJ, Crowley JL, Gerdes A, Hanchar JM *et al.* 2008. Plesovice zircon – A new natural reference material for U-Pb and Hf isotopic microanalysis. *Chem Geol* 249 (1-2): 1–35. https://doi.org/10.1016/j.chemgeo.2007.11.005.

Stacey JS, Kramers JD. 1975. Approximation of terrestrial lead isotope evolution by a two stage model. *Earth Planet Sci Lett* 26: 207–221.

Thiéblemont D, Égal E, Chantraine J. 1996. Révision du volcanisme cadomien dans l'Est du Trégor (Bretagne nord). *Géologie de la France* 4: 3–21.

Tribe IR, Strachan RA, D'Lemos RS. 1996. Neoproterozoic shear zone tectonics within the Icartian basement of Guernesey and Sark, Channel Islands. *Geol Mag* 133 (2): 177–192.

Vermeesch P. 2018. IsoplotR: A free and open tool box for geochronology. *Geosci Front* 9: 1479–1493. https://doi.org/10.1016/j.gsf.2018.04.001.

Vidal P. 1980. L'évolution poly-orogénique du Massif Armoricain : apport de la géochronologie et de la géochimie isotopique du strontium. *Mém Soc géol minéral Bretagne* 21, 162 p.

Vidal P, Deutsch S, Martineau F, Cogné J. 1974. Nouvelles données radiométriques en Baie de Saint-Brieuc. Le problème d'un socle antécadomien nord-armoricain. *C R Acad Sci Paris, série D*, 279: 631–634.

Vidal P, Auvray B, Charlot R, Cogné J. 1981. Precadomian relicts in the Armorican Massif: their age and role in the evolution of the westhern and central european cadomian-hercynian belt. *Precamb Res* 14: 10–20. https://doi.org/10.1016/0301-9268(81)90032-2.

Waychunas GA. 2002. Apatite luminescence. *Rev Mineral Geochem* 48(1): 701–742. https://doi.org/10.2138/rmg.2002.48.19.

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