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U–Pb geochronological constraints on the timing of Brioverian sedimentation and regional deformation in the St. Brieuc region of the Neoproterozoic Cadomian orogen, northern France

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Abstract

During the Neoproterozoic Cadomian orogeny in northern France, supracrustal rocks of the Brioverian Supergroup were deposited in marginal and back arc basins, and were subsequently variably deformed and metamorphosed. New U–Pb analyses of single, and small multigrain fractions of zircon from selected plutons from the Baie de St. Brieuc region provide robust geochronological constraints on the timing of these events. The Jospinet granodiorite forms part of the local basement directly overlain by Brioverian metasediments and basic volcanics, and yields a U–Pb zircon date of $625.9+3.6/-1.9$ (2σ) Ma. The pre-tectonic Port Moguer tonalite, which has been strongly sheared along with its amphibolite facies country rocks, has a crystallization age of 600.4 ± 0.9 Ma. Emplacement ages of $576.3+1.5/-1.2$ Ma for the syn-tectonic Fort La Latte quartz diorite and $574.6+1.8/-1.5$ Ma for the late-tectonic St. Quay quartz diorite place limits on termination of deposition and timing of subsequent regional deformation of the Brioverian sequence in the Baie de St. Brieuc region. The new dates constrain the age of Brioverian sedimentation to the interval 626–575 Ma, a range consistent with a previously published Pb–Pb zircon evaporation age of ca. 588 ± 22 Ma for Brioverian volcanic rocks (Lanvollen Formation). Deformation within this sector of the Cadomian belt is believed to have occurred shortly before 575 Ma, revising previously published estimates for the age of this major tectonothermal Cadomian event by 10–20 My. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: Cadomia; Brioverian; U–Pb geochronology; France

1. Introduction

The North Armorican Massif of northwest France and the British Channel Islands (Fig. 1) is

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one of a number of regions within the circum-Atlantic realm that preserve fragments of orogenic belts developed proximal to the western margin of Gondwana during the late Neoproterozoic (Auvray et al., 1980; Chantraine et al., 1988; Brun and Balé, 1990; D'Lemos et al., 1990; Strachan et al., 1996). The region exposes calc-alkaline magmatic-arc complexes and marginal basins formed during what has been termed the Cadomian orogeny (Bertrand, 1921). Unraveling the history of the Cadomian belt will lead to improved palaeogeographic reconstructions and correlations of circum-Atlantic Neoproterozoic accreted terranes, which is important for understanding the timing and geometry of subsequent supercontinental breakup and dispersal that occurred during the Precambrian–Cambrian transition (Nance and Murphy, 1994, 1996).

Geological units within the North Armorican Massif ascribed to Cadomian orogenesis (Fig. 1) range in age from ca. 740 to 540 Ma (Vidal et al., 1972, 1974; Graviou et al., 1988; Guerrot and Peucat, 1990; Egal et al., 1996). Previous work (Graviou et al., 1988; Strachan et al., 1989; Brun and Balé, 1990; Rabu et al., 1990) has identified contrasting lithologies and tectonothermal histories for different segments of the belt. In northernmost parts of the belt, well-dated ca. 615 Ma to 570 Ma plutons intrude ca. 2 Ga basement rocks (Calvez and Vidal 1978; Graviou et al., 1988; Samson and D'Lemos, 1998, 1999; Miller et al., 1999; D'Lemos et al., in press). Central and southern parts of the belt include thick late Neoproterozoic supracrustal units, collectively termed the Brioverian Supergroup, believed to have been deposited,

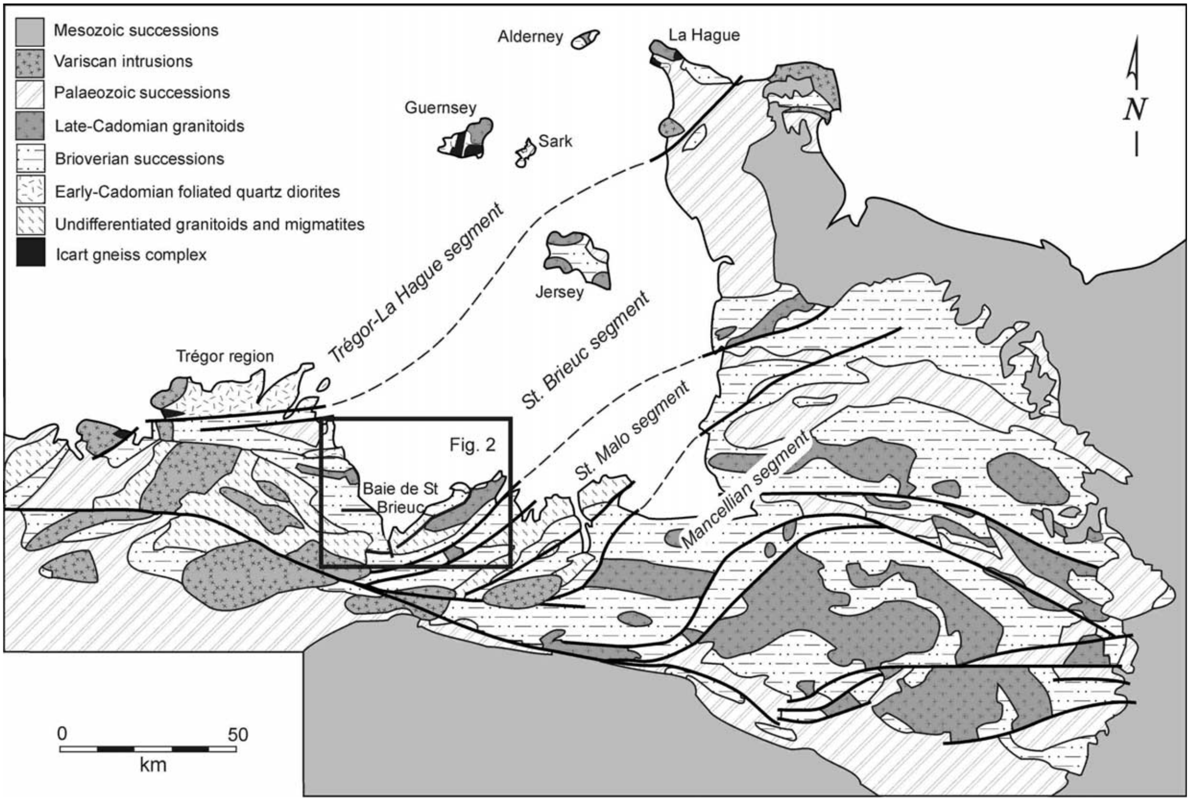

Fig. 1. Simplified geological map of the North Armorican Massif of northwestern France and the British Channel Islands, which consists of four fault-bounded segments. Location of Fig. 2 indicated.

deformed, metamorphosed and intruded by various plutons at several stages *during* Cadomian orogenesis (Rabu et al., 1983; Cabanis et al., 1987; Strachan and Roach, 1990). Available geochronological data for metamorphism and syn- to post-tectonic magmatism suggest that penetrative deformation had largely ceased by ca. 610 Ma in northernmost parts of the belt (e.g. Trégor La-Hague segment), but occured ca. 540 Ma in more southernly parts (e.g. St. Malo segment) (Peucat, 1986). The contrasting tectonothermal histories have been attributed alternatively to progressive migration of deformation from outboard to inboard parts of the orogenic system (Rabu et al., 1990), and to the juxtaposition of different tectonostratigraphic terranes with contrasting geological evolutions (Strachan et al., 1989, 1996).

The regionally extensive Brioverian Supergroup is a thick succession of metavolcanic and metasedimentary rocks that is generally interpreted to represent volcanism and clastic deposition in marginal and back arc basins during the Cadomian orogeny (Graindor, 1957; Cogné, 1962; Rabu et al., 1982, 1983; Chantraine et al., 1982). Basin development and subsequent inversion is considered to have occurred in response to major plate tectonic events within a subduction/collision setting (Strachan et al., 1996). Therefore, a key element to understanding the geological evolution of the region lies in placing precise and robust geochronological constraints on the possible time interval for Brioverian deposition and its subsequent regional deformation. Although past geochronological studies have been able to place constraints on these events, in some cases these are based upon techniques that may no longer be considered robust, or on data that lacks precision. This contribution documents U–Pb zircon dates based upon high precision single grain and small quantity multi-grain analyzes from four key plutonic units which occur in close geographic association with the Brioverian sequences in the Baie de St. Brieuc region. Two of the rock units were previously undated and includes the first direct dating of the sub-Brioverian basement. Two previously dated intrusive units have been redated with greatly improved precision.

2. Geological setting

The regional tectonic history of the Cadomian orogenic belt can be simplified into four episodes (Auvray et al., 1980; Graviou et al., 1988; D'Lemos et al. 1990 and references therein; Strachan et al., 1996; Samson and D'Lemos, 1998, 1999; Miller et al., 1999, 2001): (1) regional deformation, crustal thickening, amphibolite facies metamorphism, and intrusion of syn-tectonic calc-alkaline plutons at ca. 610 Ma; (2) development of marginal basin sequences (i.e. Brioverian deposition); (3) regional transpressive deformation, crustal thickening, and syn- to post-tectonic calc-alkaline intrusion; and (4) transition to transform plate boundary, amalgamation of crustal blocks, and syn-tectonic intracrustal magmatism by ca. 540 Ma. Various of the Neoproterozoic units are unconformably overlain by early Palaeozoic sedimentary rocks (Cogné, 1963; Doré, 1972; Went and Andrews, 1990) that do not exhibit pervasive late Palaeozoic (i.e. Variscan) reworking.

The Cadomian orogenic belt of North Armorica has been viewed as a composite terrane (Strachan et al., 1989, 1996). Using contrasting tectonothermal histories and lithological assemblages, these authors initially defined four terranes separated by steeply dipping ductile shear zones or brittle faults (Fig. 1). However, the degree to which such terranes represent dismembered and later juxtaposed parts of a single orogen or discrete crustal blocks which evolved independently of one another is unclear (Strachan et al., 1996). Hence, here we use the neutral term 'segment', to describe the contrasting parts of the belt to avoid the allochonous connotation of the terrane terminology. From north to south these are the Trégor-La Hague segment, St. Brieuc segment, St. Malo segment, and Mancellian segment. Only a brief summary of these segments is given here; see Strachan et al. (1996) and references therein for further details.

The Trégor-La Hague segment preserves the only known exposures of 2 billion year old Icartian basement gneisses, which are extensively intruded by, and tectonically interleaved with, syn-tectonic arc-related intrusions (ca. 615–610 Ma) and post-tectonic granitoids (ca. 580–560

Ma) (Adams, 1967; Vidal et al., 1974; Graviou et al., 1988; Samson and D'Lemos, 1998, 1999). The basement in the St. Brieuc segment is believed to be Neoproterozoic as demonstrated by the ca. 746 Ma metaigneous gneisses at Port Morvan (Egal et al., 1996) and tonalite boulders within the Brioverian Cesson conglomerate that yield ages of ca. 667 and 656 Ma (Guerrot and Peucat, 1990). The St. Brieuc segment preserves Brioverian metavolcanic and metasedimentary rocks (e.g. the Lanvollon and Binic Formations, respectively) that were inferred by Strachan et al. (1996), and references therein and Egal et al. (1996) to have accumulated in the approximate time interval ca. 600–575 Ma. The St. Malo segment consists of amphibolite facies migmatites and greenschist facies Brioverian metasedimentary rocks. Transitional contacts in some localities indicate that the migmatites formed from partial melting of Brioverian sediments. U–Pb and Rb–Sr studies (Peucat, 1986) indicate that anatexis occurred more recently (ca. 540 Ma) than the magmatism recorded in the Trégor-La Hague and St. Brieuc segments. The Mancellian segment consists of relatively low-grade Brioverian sedimentary rocks intruded by ca. 540 Ma granites (Pasteels and Doré, 1982). It has been suggested that the Mancellian segment represents a shallower crustal level counterpart of the St. Malo segment, and that the two segments have been tectonically juxtaposed by later sinistral transpression (D'Lemos et al., 1992). The Mancellian and St. Malo segments are dominated by intracrustal granite magmatism, in contrast to the dominantly calc-alkaline subduction-related magmatism of the St. Brieuc and Trégor-La Hague segments (Graviou and Auvray, 1990; Brown and D'Lemos, 1991; D'Lemos and Brown, 1993).

The deposition of Brioverian rocks in the Armorican Massif has been interpreted to have occurred principally in two geological settings: (1) volcanogenic deposition occurred within a volcanic arc and back-arc basin system on a rifted continental margin in an orogenic northern domain, which includes north Brittany and north Normandy; and (2) detrital Brioverian deposits dominate in a within-plate southern domain that represented a stable continental margin and shelf environment (Rabu et al., 1990). These deposits are primarily recorded in Normandy and central Brittany. The present-day boundary between these two domains corresponds very approximately to the boundary between the St. Brieuc and St. Malo segments in Fig. 1.

Facies changes, a lack of chronostratigraphic markers, and the tectonic interleaving of geological units have made it difficult to substantiate regional correlations and stratigraphic divisions within the Brioverian Supergroup. The original threefold division into Lower, Middle, and Upper Brioverian (Graindor, 1957; Cogné, 1962, 1970) was later revised to a twofold division (Cogné and Wright, 1980) that considered the primarily volcanogenic units (such as the Cesson and Lanvollon Formations) to be lower Brioverian deposits that were folded and metamorphosed prior to deposition of less deformed, primarily detrital, upper Brioverian deposits (e.g. Binic Formation). Subsequent studies (Rabu et al., 1982, 1983) found sedimentological continuity between the different units and could not identify any intra-Brioverian unconformity. However, more recent studies maintain that a weakly deformed upper sequence was derived from a deformed lower sequence (Dupret et al., 1990; Rabu et al., 1990). Guerrot and Peucat (1990) and Dupret et al. (1990) divided the Brioverian stratigraphic sequence into pre-585 (lower) and post-585 Ma (upper) members. This was based largely upon ca. 595–585 Ma ages for the St. Quay and Fort La Latte quartz diorites in Brittany and a 584 ± 4 Ma age for the Coutances diorite in Normandy, which were interpreted to have been intruded between the deposition of the two sequences.

However, only limited precise and reliable geochronological data have been available to constrain the age of Brioverian deposition and deformation around the Baie de St. Brieuc (Fig. 2) The Cesson conglomerate contains clasts and boulders that include meta-igneous material considered to have been derived from the sub-Brioverian basement. Guerrot and Peucat (1990) presented U–Pb zircon ages of 656 ± 5 and 667 ± 4 Ma from orthogneiss clasts that provided a maximum age for the deposition for the unit. A further constraint on the age of Brioverian sedimentary rocks comes from an estimate of the age of the Lanvol-

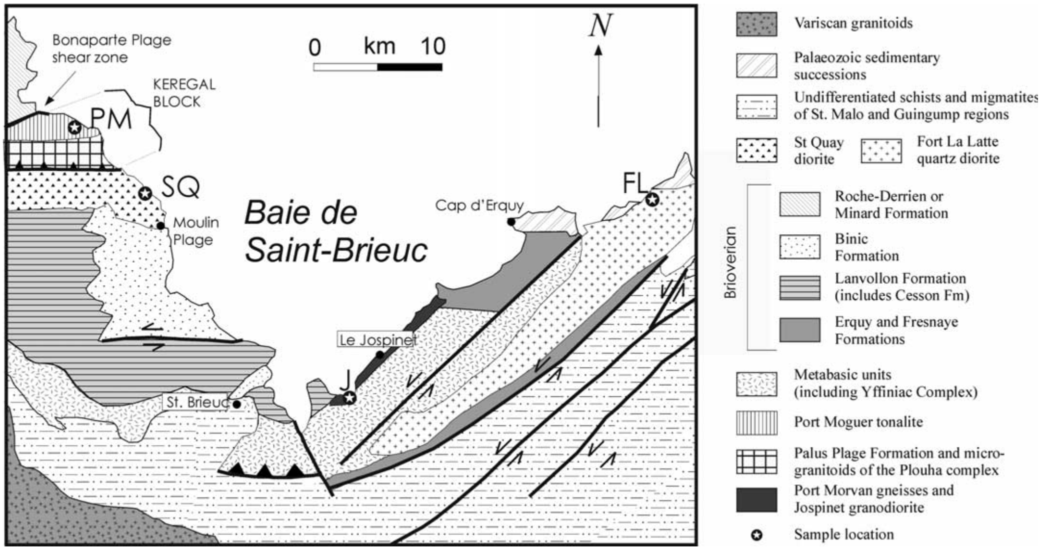

Fig. 2. Simplified geological map of Baie de St. Brieuc (modified from Strachan and Roach, 1990; Shufflebotham, 1990; Egal et al., 1996) showing sample localities (FL, Fort La Latte quartz diorite; J, Jospinet granodiorite; PM, Port Moguer tonalite; SQ, St. Quay diorite).

lon Formation, considered to stratigraphically underlie the Binic Formation (Egal et al., 1996). Single zircons from an acidic volcanic unit within the Lanvollon Formation were analyzed by Egal et al. (1996) using the zircon Pb-evaporation method (Kober, 1987). This technique allows a ^{207}Pb–^{206}Pb date to be calculated but because U and Pb contents are not determined, ^{238}U–^{206}Pb and ^{235}U–^{207}Pb dates cannot be derived, and hence the degree of discordancy cannot be assessed. The calculated ^{207}Pb–^{206}Pb date, which assumes the absence of a xenocrystic component, is taken as a minimum estimate of the age of crystallization of the zircon. Seven zircon crystals analyzed by Egal et al. (1996) from the Lanvollon Formation yielded a range of ^{207}Pb–^{206}Pb dates from 594 ± 8 to 579 ± 14 Ma (1σ uncertainties quoted). Combining all of the Pb isotopic ratios determined from the seven zircons yielded a *mean* ^{207}Pb–^{206}Pb date of 588 ± 11 Ma (1σ). This was used to argue that the sediments of the Binic Formation must have been deposited after ca. 588 Ma (Egal et al., 1996). However, if 2σ uncertainties are taken into account then a more realistic constraint on the oldest possible timing of deposition of the Binic Formation is between 610 and 566 Ma. Dallmeyer et al. (1991) provided ^{40}Ar/^{39}Ar mineral cooling ages from hornblende from metabasic units and plutons in the Baie de St. Brieuc region. These cooling ages (through ca. 500 °C) provided maximum post-metamorphic ages of ca. 575–565 Ma for deformation and amphibolite facies metamorphism.

Constraints on the timing of Brioverian deposition have also been based on the relationship between the Brioverian sequence and dated intrusive rocks, although we show here that some of these dates need revision. In particular, a ca. 593 ± 15 Ma age (U–Pb zircon, Vidal et al., 1974) for the Fort La Latte intrusion has been widely used to date regional metamorphism of the Brioverian sequences around the Baie de St. Brieuc at ca. 590 Ma (Guerrot and Peucat, 1990; Rabu et al., 1990; Brun, 1992). Outside of northern France, a very tight constraint on the youngest limit of deposition and timing of deformation of a part of the Brioverian succession has been established on the British Channel Island of Jersey. The U–Pb age of the youngest detrital zircon in the Brioverian Jersey Shale Formation is

587 ± 3 Ma, while zircon from an overlying Brioverian rhyolite unit yields an age of 583 ± 3 Ma. These units were subsequently folded then intruded by a post-tectonic granite dated at 580 ± 2 Ma (Miller et al., 2001).

A simplified geological map of the Baie de St. Brieuc region (Fig. 2) shows the key elements of the local geology pertinent to this contribution. The stratigraphically oldest rocks include the ca. 746 ± 17 Ma Port Morvan gneiss (Egal et al., 1996) and the Jospinet granodiorite (Hillion trondjemite of Egal et al., 1996), which form the local basement to Brioverian metasediments and basic volcanics (Erquy Formation) on the south and east side of the Baie de St. Brieuc (Cogné, 1959; Roach et al., 1990; Shufflebotham, 1990). On the south and west sides of the Baie de St. Brieuc, Brioverian deposits are locally divided into the volcanogenic Cesson and Lanvollon Formations, which are variably metamorphosed up to amphibolite facies, and the overlying, lower grade and mainly clastic Binic Formation (Rabu et al., 1983; Strachan and Roach, 1990; Egal et al., 1996). The Yffiniac Complex comprises various metagabbros and ultrabasic rocks metamorphosed to upper amphibolite facies conditions, and is considered to be the plutonic, co-magmatic equivalent of Brioverian volcanics. One component, the Le Croix-Gibat metagabbro, has yielded a precise U–Pb age of 587 ± 2 Ma, interpreted to date igneous crystallization (Guerrot and Peucat, 1990). The foliated Fort La Latte quartz diorite on the east side of the Baie de St. Brieuc is in fault contact with Brioverian metasediments and metavolcanics (Fresnaye Formation), but exhibits many features common to syn-tectonic plutons. The St. Quay quartz diorite on the northwest side of the Baie de St. Brieuc is largely undeformed, cross-cuts folds in Brioverian country rocks, and develops a late-tectonic metamorphic aureole. The Port Moguer tonalite, further to the north, has been strongly deformed along with its amphibolite facies country rocks and is considered to be a pre-tectonic intrusion (Strachan and Roach, 1990).

The St. Brieuc segment is in tectonic contact with schistose and migmatised Brioverian sequences to the south and east. This major orogenic boundary takes the local form of a southernly directed thrust (St. Brieuc thrust) and broad systems of sinistrally transpressive faults and ductile shear zones (Fresnaye and St. Cast shear zones) (Balé and Brun, 1983; Brun and Balé, 1990; Treloar and Strachan, 1990).

3. Sample descriptions and previous geochronology

We present U–Pb geochronological results for zircons from four plutonic rocks from the Baie de St. Brieuc region to derive emplacement ages. These are the Jospinet granodiorite, the Port Moguer tonalite, the St. Quay quartz diorite, and the Fort La Latte quartz diorite.

3.1. Jospinet granodiorite

The previously undated Jospinet granodiorite (Shufflebotham, 1990) consists of quartz, plagioclase, K-feldspar, relic mica and titanite (almost completely altered to low temperature replacement minerals), apatite, zircon, and opaque oxides. Chlorite fills late-stage brittle fractures. The Jospinet granodiorite exhibits widespread cataclasis and alteration but is only weakly penetratively deformed. Weak fabrics typically trend NE–SW and contrast with fabrics locally developed in the adjacent Brioverian sequences. One kilometer south of Le Jospinet, the granodiorite is unconformably overlain by pebbly psammite, which rapidly fines upwards into pelites (Cogné, 1959; Shufflebotham, 1990). Clasts within an undeformed matrix in the basal conglomerate horizon contain identical cataclastic features to the underlying basement, demonstrating local derivation from the weakly deformed basement. The basal units are overlain by interlayered metabasic units and pelites. To the north (around Cap d'Erquy) these units are believed to be succeeded stratigraphically by a thick sequence of weakly metamorphosed basic volcanic rocks including pillow lavas with local acid volcanics (Roach et al., 1990).

3.2. Port Moguer tonalite

The previously undated Port Moguer tonalite (Ryan, 1973) is a heterogeneously sheared medium to coarse-grained rock located in the Kérégal structural block (Strachan and Roach, 1990), ca. 20 km northwest of St. Brieuc (Fig. 2). It intrudes a mixed assemblage of meta-igneous rocks of the Plouha Complex (Ryan, 1973; Strachan and Roach, 1990). The Port Moguer tonalite and adjacent Plouha Complex are deformed in a major high strain zone, the Bonaparte Plage Shear Zone, marked by a heterogeneously developed, sub-vertical, east–west trending mylonitic L-S fabric that is present throughout the Kérégal block. Within central, low strain areas of the pluton, the Port Moguer tonalite comprises a coarse-grained assemblage of andesine, biotite and quartz and retains igneous textures. The Port Moguer tonalite is increasingly deformed and partially recrystallized towards its margins culminating in distinctive grey-coloured, finely banded mylonites. However, even in high strain zones, sheared tonalite is still distinguishable from meta-igneous members of the Plouha Complex by its coarser grain size and contrasting mineralogical and textural assemblage. Both the host microgranitoids and the Port Moguer tonalite include thin (<1 m) sub-vertical, concordant sheets of microgranite and amphibolite, the latter of which are variably retrogressed to actinolite–chlorite–biotite schist in high strain zones.

Field and petrographic data indicate a history of sequential intrusion and progressively focused down-temperature deformation within the following broad sequence of events. The Port Moguer tonalite was emplaced into the igneous host rocks of the Plouha Complex, and in turn intruded by microgranite and basic sheets. All of these units were deformed to give a widespread upright foliation and metamorphosed under regional upper greenschist to amphibolite facies conditions. Because of its mineralogy, the Port Moguer tonalite itself records little evidence of this metamorphism. However, the intruding basic units exhibit growth of amphibole at greenschist to low amphibolite grade (Ryan, 1973). Deformation was focused into narrow zones, such as the Bonaparte Plage Shear Zone, along the northern margin of the Port Moguer tonalite. Within the shear zone, deformation was initiated in the low amphibolite facies and continued under upper greenschist conditions to produce mylonites with dominantly upper greenschist facies parageneses and microtextures. At the same time, more-or-less static metamorphism and recovery took place in rocks not undergoing active deformation (e.g. the low strain core of the Port Moguer tonalite). Widespread cataclastic reworking took place at still lower temperatures. The intrusion is thus viewed as pre-tectonic with respect to the development of the Bonaparte Plage Shear Zone and the pervasive upright foliation throughout the Kérégal block.

The Kérégal block is in fault contact with weakly deformed, low-grade Brioverian turbidites and volcanics to the north, and has been thrust southwards over the St. Quay intrusion across a late structure known as the Port Goret thrust. The exact affinity and relationships of the units within the Kérégal block to those around St. Brieuc are thus unconstrained. However, because deformation within the Kérégal block has been regionally correlated with structures in other parts of the St. Brieuc region (Strachan and Roach, 1990), and is known to have occurred prior to ca.560 Ma (Strachan et al., 1996), an age for the emplacement of the pre-tectonic Port Moguer tonalite may help to constrain the timing of deformation in the St. Brieuc region as a whole.

3.3. St. Quay quartz diorite

The St. Quay quartz diorite contains quartz, plagioclase, K-feldspar, pyroxene, hornblende, biotite, zircon, and opaque oxides. Pressure estimates of 3–4 kbars suggest an emplacement depth of 10–12 km (Fabriès et al., 1984). A Rb–Sr date of 559 ± 54 Ma was originally reported by Vidal et al. (1972), later revised to 584 ± 56 Ma (Vidal 1980). Additional age determinations related to the St. Quay intrusion include ^{40}Ar/^{39}Ar cooling ages for magmatic amphiboles (563.1 ± 1.9 Ma) and for metamorphic muscovite from the aureole rocks (569.3 ± 0.6 Ma) (Dallmeyer et al., 1991).

The St. Quay quartz diorite carries a moderate to weakly developed fabric defined by the preferred alignment of zoned plagioclase prisms within an interstitial, undeformed quartz matrix. These features indicate fabric formation in a magmatic state, prior to complete crystallization. Mainly equant igneous hornblendes contain cores of orthopyroxene and clinopyroxene. The St. Quay quartz diorite develops a high-temperature inner aureole within Brioverian host rocks at Moulin Plage. Blocks of psammitic metasediment are surrounded by a matrix of biotite-rich granite, interpreted to be in-situ partial melt from pelitic portions of the host metasedimentary sequence. The matrix exhibits a granoblastic texture and does not carry a penetrative fabric, indicating a lack of deformation during and after cooling. Small, pipe-like diapiric bodies of diorite within the aureole indicate contemporaneous mobility of diorite and host (D'Lemos, 1992), and again, no subsequent deformation. The inner aureole is in fault contact with an extensive outer aureole at Moulin Plage. Approximately 200 m from the contact, and for a distance of up to 3 km, cordierite spotting is conspicuously developed in some horizons of the Brioverian Binic Formation. The cordierite has grown largely mimetically and is consequently prolate, but clearly overprints the cleavage. However, in some cases there is development of pressure shadows and of 'S' shaped inclusion trails in the outer parts of (replaced) cordierite. Quartz is largely recrystallized, as shown by only weak undulose extinction and a granoblastic texture. Together, we consider the relationships indicate that the country rocks had undergone significant deformation prior to emplacement of the pluton, and experienced only minor deformation during and following emplacement. As only one deformation is recorded in the host rocks this suggests that the pluton is late-tectonic. It is also possible, however, that the pluton was regionally syn-tectonic but that deformation was partitioned away from the pluton shortly following emplacement.

3.4. Fort La Latte quartz diorite

The Fort La Latte quartz diorite is a coarse-grained, containing quartz, plagioclase with white mica replacement, hornblende, chloritized and prehnitized biotite, zircon, and opaque oxides. Pressure estimates of 3–5 kbars suggest emplacement at depths of 9–15 km (Hallot, 1993). The Fort La Latte quartz diorite was previously considered to have been emplaced at 593 ± 15 Ma (U–Pb zircon, Vidal et al., 1974). Cooling ages include 579 ± 12 Ma determined by Rb–Sr analysis on biotite from a sample from the inner part of the pluton (Vidal et al., 1974) and 564.7 ± 1.6 Ma determined from ^{40}Ar/^{39}Ar analysis on amphibole (Dallmeyer et al., 1991).

The age of the Fort La Latte quartz diorite relative to deposition and deformation of the Brioverian sequence has been a matter of considerable debate, largely because all observable contacts are faults. Shufflebotham (1990) considered the quartz diorite to form part of the local 'Pentevrian' basement complex (which included the Jospinet granodiorite) and thus to predate Brioverian deposition. However, whereas many of the proven basement units are extensively intruded by a mafic dyke swarm (Dahouet dykes), interpreted as feeders to Brioverian volcanic sequences (Lees et al., 1987), they are conspicuously absent from the Fort La Latte intrusion indicating that it might be younger. Moreover, the Fort La Latte intrusion does not carry the extensive penetrative solid-state deformation and up to amphibolite facies metamorphism observed in many adjacent basement and Brioverian components. Balé and Brun (1983) interpreted the elongation of the Fort La Latte pluton and parallelism of a fabric within the pluton to the regional structural grain as evidence that the pluton was emplaced syn-tectonically. In turn, they used the existing U–Pb age of 593 ± 15 Ma (Vidal et al., 1974) to argue that Cadomian deformation in the region occurred at ca. 600–580 Ma. Strachan and Roach (1990) and Strachan et al. (in reply to Brun, 1992) pointed out that the structural relations could not prove unequivocally that the pluton was emplaced syn-tectonically with regional deformation (e.g. the fabrics might have formed by a variety of syn-emplacement processes not related to regional deformation). The syn-tectonic emplacement of the pluton at ca. 590 Ma is also brought into question because $^{40}Ar/^{39}Ar$ mineral ages from the pluton and metamorphic units in the region indicate post-tectonic cooling occurred considerably later, at around 575–565 Ma. In our field and petrographic analysis of the pluton, we concede that the evidence is equivocal. A well-developed fabric and flattening of mafic inclusions clearly parallels the regional structural grain. In large parts of the pluton, the fabric was demonstrably formed at a magmatic stage during down-temperature cooling and is only weakly overprinted by heterogeneously developed cataclastic zones. However, a low-grade metamorphic overprint, whereby biotite is replaced by chlorite, prehnite and epidote, is widely developed. The fault contacts mean that there are no reliable aureole porphyroblast–cleavage relationships to help evaluate the timing of emplacement relative to cleavage formation. Consequently, the evidence might be argued to be consistent with the fabric development simply as a result of ballooning within a regional stress field as opposed to emplacement during active regional deformation. However, we consider it most likely that the pluton *was* syn- to late-tectonic given: (1) the considerable size of the pluton and the consistent development of the fabric throughout the observable parts of the pluton; (2) the down-temperature nature of the fabric; and (3) the parallelism of deformation to regionally developed structures.

4. Analytical methods

U–Pb analyzes were performed using the isotope dilution method on grain-by-grain selected, abraded zircon fractions, and analyzed on a VG Sector 54 mass spectrometer at Syracuse University. We chose the clearest, crack-free grains with minor to no inclusions. See Samson and D'Lemos (1998, 1999) for details of the dissolution, chemistry, and mass spectrometric procedures. The total common Pb amounts (analytical laboratory blank plus initial zircon common Pb) in most samples were < 3 pg (in some cases < 1 pg) and U blanks were < 1 pg. Initial common Pb compositions were determined using the two-stage Pb evolution model of Stacey and Kramer (1975), and the data were reduced and regressed following the routines of Ludwig (1989, 1990). Data are summarized in Table 1. Analytical uncertainties throughout this paper are given at the 2σ level.

5. Results

5.1. Jospinet granodiorite

Seven multi-grain zircon fractions consisting of four to six crystals each form a linear array on a U–Pb concordia diagram (Fig. 3) with an upper intercept anchored by a concordant zircon fraction at 625 Ma. Regression of all seven analyzes gives an upper intercept value of $625.9 + 3.6/-1.9$ (2σ) Ma with an MSWD of 0.35 (lower intercept is 134 ± 157 Ma). The weighted average of the $^{207}Pb-^{206}Pb$ dates gives a similar value of 624.4 ± 0.9 Ma with an MSWD of 0.70.

5.2. Port Moguer tonalite

One single-grain and five multi-grain zircon fractions, consisting of five grains each, form a linear array on a U–Pb concordia diagram (Fig. 4) with an upper intercept well-anchored by two concordant fractions overlapping at 601 Ma. Regression of all six analyzes gives an upper intercept value of $601.6 + 6.0/-1.8$ Ma with an MSWD of 0.12 (lower intercept is 185 ± 241 Ma). The relatively large 'plus' error of 6.0 is simply an artifact of the regression algorithm's treatment of a tight grouping of data points at one end of the array. For this reason, a better estimate of the crystallization age is given by the weighted average of the $^{207}Pb-^{206}Pb$ dates, which gives a value of 600.4 ± 0.9 Ma (MSWD = 0.51).

5.3. St. Quay quartz diorite

Two single-grain zircon fractions, four two-

Table 1
U–Pb isotopic data for zircons from plutonic rocks in the Baie de St. Brieuc region

Fraction (grains)		Total U (ng)	Total Pb (pg)	Total Com. Pb (pg)	Atomic ratios								Ages (Ma)			ρ^{d}
					^{206}Pb^{a}/^{204}Pb	^{206}Pb^{b}/^{208}Pb	^{206}Pb^{b}/^{238}U	Error^{c} (%)	^{207}Pb^{b}/^{235}U	Error^{c} (%)	^{207}Pb^{b}/^{206}Pb	Error^{c} (%)	^{206}Pb/^{238}Pb	^{207}Pb/^{235}U	^{207}Pb/^{206}Pb	
Jospinet granodiorite																
1.	Five zircons	1.544	169.4	1.96	4663	4.68	0.09980	0.141	0.83364	0.175	0.06058	0.103	613.2	615.5	624.5	0.808
2.	Six zircons	1.149	126.6	2.30	2974	4.78	0.10043	0.174	0.83875	0.213	0.06057	0.121	616.9	618.5	624.0	0.824
3.	Five zircons	1.226	134.9	1.35	5292	4.68	0.10024	0.160	0.83686	0.189	0.06055	0.100	615.8	617.4	623.4	0.849
4.	Six zircons	1.062	118.7	1.26	4985	4.66	0.10185	0.220	0.85124	0.245	0.06062	0.107	625.2	625.3	625.7	0.899
5.	Six zircons	2.575	281.1	3.50	4510	5.02	0.10027	0.113	0.83719	0.151	0.06055	0.100	616.0	617.6	623.4	0.751
6.	Five zircons	2.057	225.2	1.73	7118	4.92	0.10054	0.128	0.83983	0.163	0.06058	0.101	617.6	619.1	624.4	0.787
7.	Four zircons	2.087	233.8	1.94	6490	4.45	0.10109	0.142	0.84478	0.172	0.06061	0.097	620.7	621.8	625.5	0.824
Port Moguer tonalite																
8.	Five zircons	1.536	152.7	2.30	3832	7.40	0.09647	0.143	0.79653	0.176	0.05988	0.102	593.7	594.9	599.4	0.816
9.	One zircon	0.590	58.6	2.20	1611	8.17	0.09730	0.291	0.80412	0.316	0.05994	0.122	598.5	599.2	601.5	0.923
10.	Five zircons	0.875	85.4	1.02	4776	9.33	0.09766	0.205	0.80712	0.232	0.05994	0.108	600.7	600.8	601.4	0.885
11.	Five zircons	1.392	139.7	4.50	1848	7.65	0.09695	0.157	0.80079	0.195	0.05991	0.114	596.5	597.3	600.3	0.810
12.	Five zircons	1.298	126.4	0.75	9097	9.12	0.09681	0.121	0.79950	0.154	0.05990	0.095	595.7	596.5	599.9	0.787
13.	Five zircons	1.670	166.4	4.00	2506	8.37	0.09756	0.139	0.80595	0.192	0.05992	0.129	600.1	600.2	600.6	0.739
St. Quay quartz diorite																
14.	Three zircons	8.400	767.3	1.80	24 622	8.64	0.09036	0.324	0.73765	0.339	0.05921	0.099	557.7	561.0	574.7	0.956
15.	Two zircons	6.404	590.6	1.34	24 647	7.96	0.09049	0.138	0.73892	0.176	0.05922	0.110	558.5	561.8	575.2	0.783
16.	Two zircons	3.446	317.1	1.47	12 188	7.44	0.08964	0.109	0.73169	0.149	0.05920	0.102	553.4	557.5	574.5	0.730
17.	Two zircons	2.720	255.0	1.16	12 448	8.43	0.09258	0.114	0.75572	0.148	0.05920	0.095	570.8	571.5	574.5	0.770
18.	One zircon	2.336	223.0	3.10	4219	6.52	0.09118	0.117	0.74418	0.153	0.05920	0.099	562.5	564.8	574.3	0.764
19.	Two zircons	7.076	636.1	1.07	33 374	7.92	0.08817	0.099	0.71967	0.136	0.05920	0.093	544.7	550.5	574.3	0.731
20.	One zircon	5.119	487.0	0.72	22 269	6.89	0.09183	0.101	0.74961	0.137	0.05920	0.092	566.4	568.0	574.5	0.738
Fort La Latte quartz diorite																
21.	Five zircons	1.396	132.1	1.00	7332	7.78	0.09272	0.135	0.75740	0.166	0.05924	0.097	571.6	572.5	576.0	0.812
22.	Four zircons	1.420	132.3	1.57	4956	8.83	0.09247	0.152	0.75581	0.185	0.05928	0.104	570.1	571.6	577.4	0.825
23.	Three zircons	2.183	201.1	1.58	7526	9.82	0.09228	0.122	0.75399	0.157	0.05926	0.099	569.0	570.5	576.5	0.777
24.	Two zircons	2.509	230.2	1.78	7729	10.04	0.09204	0.189	0.75249	0.212	0.05929	0.097	567.6	569.7	577.9	0.890
25	Four zircons	1.293	114.7	1.38	4775	8.75	0.08795	0.250	0.71946	0.270	0.05933	0.101	543.4	550.4	579.2	0.927

[a] Corrected for spike composition and fractionation/mass bias (Farady: 0.1 ± 0.05%/amu; Daly: 0.18 ± 0.07%/amu).

[b] Corrected for fractionation, blank, and initial common Pb.

[c] Errors quoted at 2 sigma.

[d] ^{207}Pb/^{235}U–^{206}Pb/^{238}U correlation coefficient of Ludwig (1989).

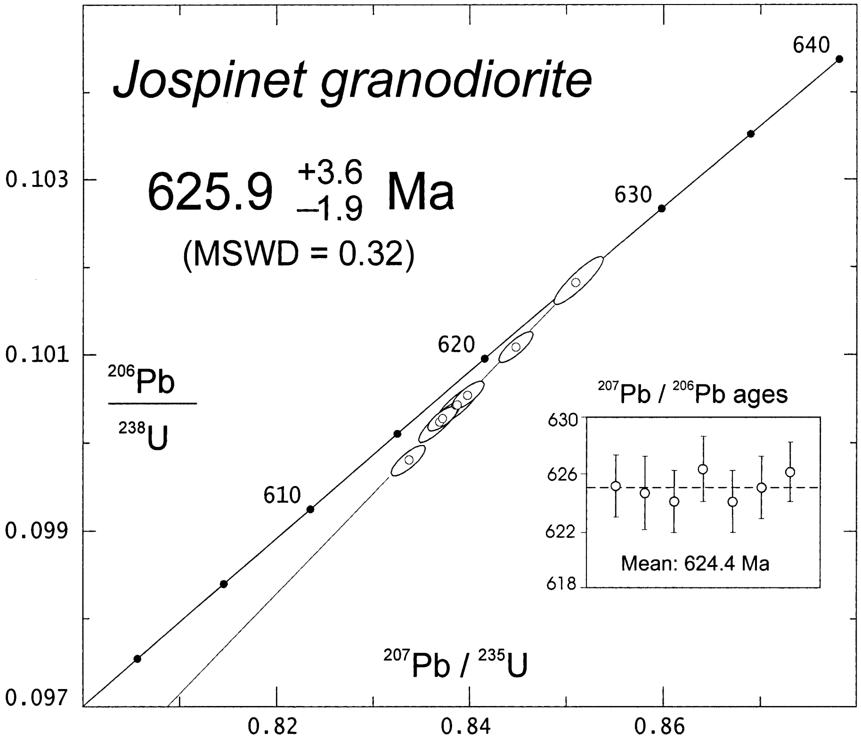

Fig. 3. U–Pb concordia diagram for zircons analyzed from the Jospinet granodiorite.

grain zircon fractions, and one three-grain zircon fraction form a very well-defined linear array on a U–Pb concordia diagram (Fig. 5) with an upper intercept of 574.6 + 1.8/ − 1.5 Ma (MSWD = 0.08). The weighted average of the ^{207}Pb–^{206}Pb ages gives an identical value of 574.6 ± 0.3 Ma with comparable MSWD of 0.07. The upper intercept date is considered the best estimate of the age of emplacement of this intrusion.

5.4. Fort La Latte quartz diorite

Five multi-grain zircon fractions, consisting of two to five grains each, form a linear array on a U–Pb concordia diagram (Fig. 6) with an upper intercept of 576.3 + 1.5/ − 1.2 Ma and an MSWD of 0.50. The weighted average of the ^{207}Pb–^{206}Pb ages gives a similar value of 577.4 ± 0.9 Ma with an MSWD of 1.34. The upper intercept date is considered the best estimate of the age of emplacement of this quartz diorite intrusion.

6. Discussion

6.1. Age of local basement to Brioverian

Our new geochronological results have determined that the Jospinet granodiorite, the local basement to the Brioverian, was emplaced at ca. 626 Ma. The age of this intrusion is about 10 m.y. older than the main phase of early Cadomian magmatism documented in the Trégor-La Hague segment to the north (Fig. 1), as constrained by the 611 Ma Perelle quartz diorite on Guernsey (Samson and D'Lemos, 1999), the 615 Ma Trégor batholith in Brittany (Graviou et al., 1988), and a 616 Ma orthogneiss from Sark (Samson and D'Lemos, 1998). Further studies are necessary to evaluate whether or not these early Cadomian magmatic events in the two segments are in any way related. The age is significantly younger than a ca. 740 Ma zircon evaporation age (Egal et al. 1996) for the Port Morvan Gneiss, the second major local basement component.

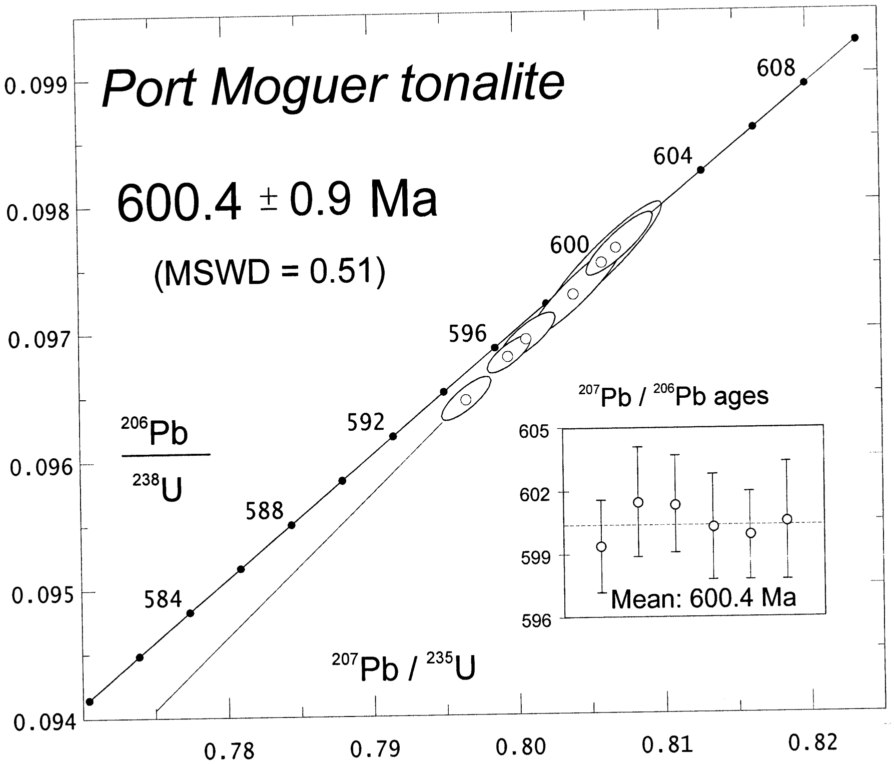

Fig. 4. U–Pb concordia diagram for zircons analyzed from the Port Moguer tonalite.

6.2. Brioverian deposition in the St. Brieuc segment

Local Brioverian sedimentation began after 626 Ma (age of the Jospinet granodiorite) and was completed by 575 Ma (age of the cross-cutting St. Quay quartz diorite). This allows a revision from the previous upper age limit of ca. 656 Ma based upon the age of the youngest clast within the Brioverian Cesson conglomerate. This timing is consistent with that deduced for Brioverian deposition on the island of Jersey, also located within the St. Brieuc segment of Cadomia (Fig. 1), where accumulation of the Jersey Shale Formation occurred prior to intrusion of a 580 ± 2 Ma (U–Pb) cross-cutting granite (Miller et al., 2001). The presence of detrital zircons as young as 587 ± 3 Ma (U–Pb) in one part of the Jersey Shale Formation implies that at least some Brioverian deposition occurred after 587 Ma (Miller et al., 2001).

The relative timing of emplacement of the Port Moguer tonalite and deposition of Brioverian sediments cannot be ascribed with any certainty due to the unclear affinity of the immediate host rocks, and to the fault-bounded nature of the Kérégal block. It is possible that the Plouha Complex comprises Brioverian volcanics and hyperbyssal rocks (Egal et al., 1996), but they could equally belong to an earlier (pre-Brioverian?) arc.

6.3. Timing of regional deformation and metamorphism of the Brioverian

If it is accepted that similarly orientated structures recorded in the Port Moguer tonalite and Brioverian units in adjacent blocks to the north and south all relate to the same regional deformation, then the pre-tectonic Port Moguer tonalite places an older age limit for deformation. Deformation of Brioverian deposits around the Baie de St. Brieuc was probably still occurring during the

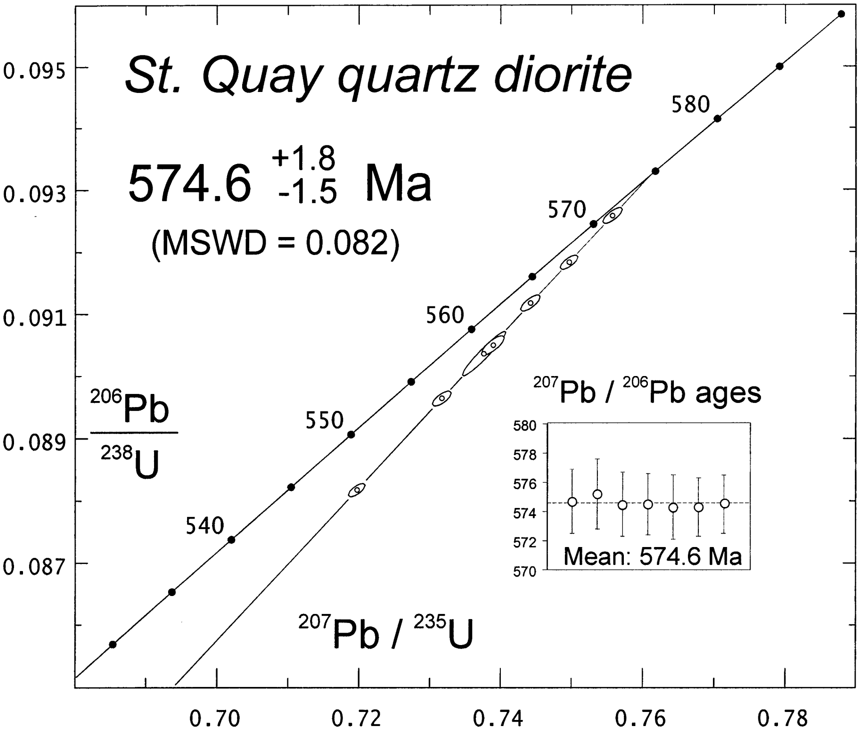

Fig. 5. U–Pb concordia diagram for zircons analyzed from the St. Quay quartz diorite.

emplacement of the Fort La Latte quartz diorite but was largely completed prior to emplacement of the late-tectonic St. Quay quartz diorite (at approximately 575 Ma). Our age data therefore constrains deformation to the interval between 600 and 575 Ma. This is consistent with the zircon evaporation age of for eruption of the Lanvollen Volcanic Formations provided by Egal et al. (1996) (588 ± 22 Ma, 2σ uncertainties). Age constraints on deformation on the island of Jersey lie within this age range, but indicate deformation was completed slightly earlier (583–580 Ma; Miller et al., 2001).

6.4. Timing of major magmatic events

The new ages for emplacement of the St. Quay quartz diorite ($574.6+1.8/-1.5$ Ma) and Fort La Latte quartz diorite ($576.3+1.5/-1.2$ Ma) are younger and significantly more precise than the commonly quoted ages of 584 ± 56 (Vidal et al., 1972; Vidal, 1980) and 593 ± 15 Ma (Vidal et al., 1974), respectively. The emplacement ages are consistent with ^{40}Ar/^{39}Ar hornblende cooling ages of 563.1 ± 1.9 (St. Quay quartz diorite) and 564.7 ± 1.6 Ma (Fort La Latte quartz diorite) (Dallmeyer et al., 1991). Protracted cooling to ca. 500 °C is consistent with the country rocks being at elevated temperatures at ca. 575 Ma, as demonstrated by the ^{40}Ar/^{39}Ar mineral ages from regionally metamorphosed amphibolites (Dallmeyer et al., 1991). This may also account for the partial recrystallization of mafic phases observed in the plutons.

The emplacement of the syn- to late-tectonic St. Quay and Fort La Latte intrusions at ca. 575 Ma occurred slightly after emplacement of 585–580 Ma post-tectonic magmatism in the neighboring Trégor-La Hague segment and the north-central part of the St. Brieuc segment (Jersey) (Vidal, 1980; Guerrot and Peucat, 1990; Dallmeyer et al., 1991, 1992; Egal et al., 1996; Miller et al., 2001; D'Lemos et al. 2001; Nagy and Samson, unpublished data). This is in agreement with re-

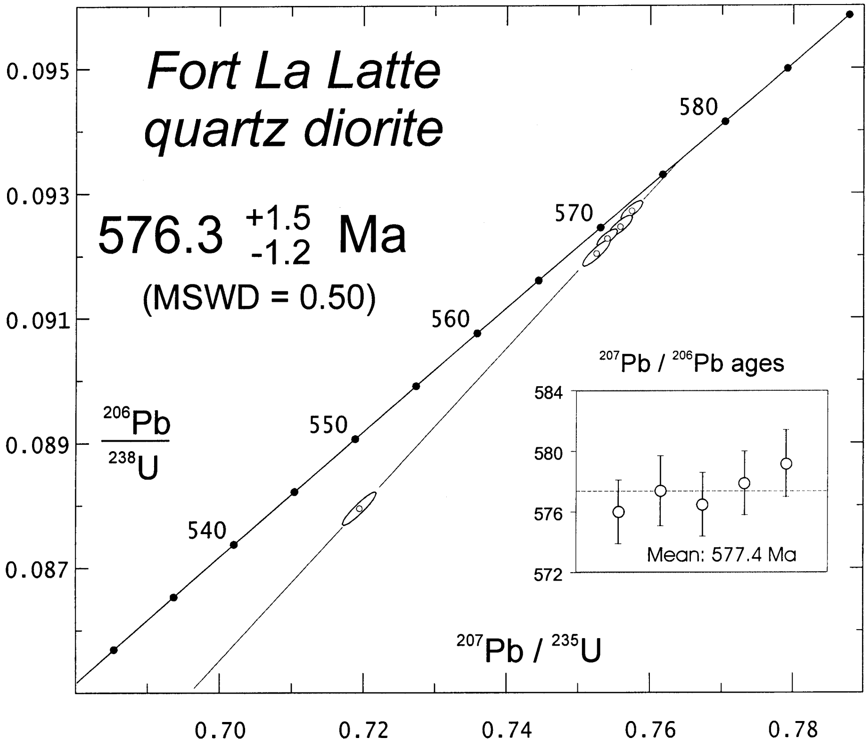

Fig. 6. U–Pb concordia diagram for zircons analyzed from the Fort La Latte quartz diorite.

gional models that suggest a migration of tectonothermal activity from outboard to inboard segments of the Cadomian orogenic system through time (Cogné and Wright, 1980; Rabu et al., 1983, 1990; Graviou and Auvray, 1985; Strachan et al., 1996). As discussed by Strachan et al. (1996) migrating magmatism and basin closure could be explained by variations in the dip of the subducting slab and the angle of plate convergence.

6.5. Revisions to previous tectonic models for the Baie de St. Brieuc

We have demonstrated that the St. Quay and Fort La Latte intrusions were emplaced 10–20 m.y. later than previously inferred, at ca. 575 Ma. These findings necessitate revision of tectonic models that have used the previously reported emplacement ages of these intrusions to constrain regional deformational and metamorphic events. Specifically, the inference that an early metamorphic event associated with back-arc closure in the Baie de St. Brieuc region occurred at ca. 590 Ma is largely based on the previously estimated age of the Fort La Latte intrusion (Rabu et al., 1990; Brun, 1992). Our new dates support instead the occurrence of a major tectonometamorphic Cadomian event closer to 575 Ma (Dallmeyer et al., 1991; Egal et al., 1996).

7. Conclusions

Our U–Pb geochronological data place tighter constraints on the timing of Brioverian deposition and subsequent deformation in the Baie de St. Brieuc region than has been previously possible and provide more precise and more robust ages for two key, syn- to late-tectonic plutons. In addition to providing the first direct age of the local basement, the date of 625.9 + 3.6/ − 1.9 Ma for the Jospinet granodiorite provides a maximum age for the accumulation of overlying Brioverian volcano-sedimentary sequence. The emplacement age of this local basement is slightly older than ages for the early phase of Cadomian magmatism (*c.* 615–610 Ma) recorded in the Trégor-La Hague segment to the north. The emplacement age of the Port Moguer tonalite is 600.4 ± 0.9 Ma. Although there are no direct geological constraints to determine whether intrusion of this unit pre- or post-dates the onset of deposition of the Brioverian succession, we infer that emplacement was prior to regional deformation of the Brioverian succession in the St. Brieuc region. The Fort La Latte quartz diorite ($576.3+1.5/-1.2$ Ma) is considered to be a regionally syn-tectonic intrusion which parallels the regional structural grain, while the St. Quay quartz diorite ($574.6+1.8/-1.5$ Ma) cross-cuts regional structures. The plutons thereby constrain deposition and deformation to pre- ca. 575 Ma. These emplacement ages (ca. 20 and 10 My younger, respectively, than previously quoted ages) necessitate revision of models which envisaged a major magmatic and metamorphic event in the St. Brieuc segment between ca. 600 and 580 Ma (Brun and Balé, 1990; Guerrot and Peucat, 1990). Our results offer a solution to the previous apparent inconsistency between such models and the argon geochronology of Dallmeyer et al. (1991). From a consideration of all the currently available, most robust geochronological data, we conclude that the main Cadomian deformation of the Brioverian sequence in the St. Brieuc segment occurred during the time interval 585–575 Ma.

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