Paleozoic sedimentation and Caledonian terrane architecture in NW Svalbard: indications from U–Pb geochronology and structural analysis

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Abstract: Svalbard’s Northwestern Basement Province is traditionally divided into the Albert I Land and the Biscayarhalvøya terranes. New U–Pb age data on zircon and monazite and structural and geochemical data provide first evidence of early Paleozoic deposits south of the Biscayarhalvøya Terrane indicating the possible existence of a third terrane: the Germaniahalvøya Terrane. This area is represented by a Cambro-Ordovician succession of mica schist and marble (Lernerygane Group) and its higher-grade metamorphic equivalent (Liefdefjorden Migmatite Complex), which were affected by the Taconian phase (migmatization at c. 469 Ma) and the Scandian phase (c. 422–415 Ma) of the Caledonian Orogeny. During the Scandian phase, the ductile Lerner Deformation Zone was formed. New isotopic data from the eclogite-bearing Richarddalen Complex of the Biscayarhalvøya Terrane imply the formation as an Ordovician–Silurian collision-related mélangé dominantly composed of c. 730 to 600 Ma Timanian island-arc-derived detritus and igneous rocks, partly eclogite-facies metamorphosed at c. 656 Ma, and Tonian meta-igneous rocks. After amphibolite-facies metamorphism of the mélangé matrix at c. 423 Ma, the Richarddalen Complex and the Stenian–Tonian Biscayarfonna Group were juxtaposed and mylonitized by the dextral Biscayarhalvøya Deformation Zone.

Keywords: Albert I Land Terrane; Biscayarhalvøya Terrane; Richarddalen Complex; Spitsbergen; Haakon VII Land; eclogite; Lerner Deformation Zone; Liefdefjorden

Supplementary material: The complete geochemical and U-Pb isotope geochemical dataset as well as additional figures are available at https://doi.org/10.6084/m9.figshare.c.5778735

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The Arctic Ocean and the northern North Atlantic Ocean are surrounded by continents with cratonic nuclei (Laurentia, Baltica, Siberia) and orogenic belts, which were repeatedly amalgamated and disrupted during Earth’s history. Svalbard, today an archipelago at the northwestern edge of the Barents Shelf (Fig. 1), was located close to NE Greenland after the Caledonian Orogeny and prior to the opening of the northern North Atlantic and the Eurasian Basin in the Paleogene. Thus, Svalbard represents a key area for studying Caledonian processes, particularly the transition from the North Atlantic Caledonides to the Arctic region. The study of pre- and post-Caledonian events recorded on Svalbard is important for circum-Arctic plate-tectonic reconstructions and correlations (Fig. 1).

The pre-Devonian basement of Svalbard is divided into the Northwestern, Southwestern, and Northeastern basement provinces (e.g. Gee and Page 1994; Gasser 2014; Dallmann 2015) (Fig. 2a). The three provinces consist of several crustal blocks or terranes of partly different (and partly unclear) age, tectonic history and stratigraphic relationship. They were juxtaposed along major, north-south-trending shear or fault zones in late Caledonian times (Harland et al. 1974; Manby 1990; Gee and Page 1994; Harland 1997a, b; Gee and Teben’kov 2004; Labrousse et al. 2008).

Between the Northwestern and the Northeastern basement provinces, several kilometres thick Old Red Sandstone deposits of the Devonian Andrée Land Basin cover an unknown basement (Harland et al. 1974; Dallmann and Piepjohn 2020). The Northwestern Basement Province of Svalbard is dominated by Stenian–Tonian metasedimentary rocks (mica schists, marbles) intruded by Tonian granitoids, which were overprinted by up to high-grade metamorphism during the Caledonian Orogeny and intruded by Silurian–Early Devonian granitoids (Myhre et al. 2008, and references therein; Pettersson et al. 2009a, b). An exception within the Northwestern Basement Province is the tectonically bounded, eclogite-bearing Richarddalen Complex on Biscayarhalvøya (Fig. 2b). The protoliths of the eclogites, for which Cryogenian ages were reported (Peucat al. 1989; Gromet and Gee 1998), have been interpreted as mafic dykes and layered gabbro intrusions into the Stenian–Tonian basement (e.g. Elvevold et al. 2014).

In contrast to the geochronologically studied western part of the Northwestern Basement Province at Albert I Land, the ages of the lithologically similar rock units east of the Raudfjorden Fault in Haakon VII Land are still poorly constrained. This paper is focused on this eastern part with study areas on Biscayarhalvøya.
We present new U–Pb zircon and monazite ages from the major rock units of the study areas (13 samples), completed by whole-rock geochemical and Sr–Nd isotope data, and a description and structural analysis of major deformation zones that have affected the basement. Our new geochronological data reveal that both study areas east of the Raudfjorden Fault show a pre-Caledonian and early Caledonian geological history that is different from previous interpretations. We propose a new approach for the chronological and tectonic interpretation of the basement between the Raudfjorden and Breibogen faults.

**Geological setting and results of previous isotopic dating**

Our study areas are located in the northern part of the Northwestern Basement Province, which is subdivided into the Albert I Land Terrane and the Biscayarhalvøya Terrane (e.g. Harland 1997a; Dallmann 2015). The Albert I Land Terrane constitutes the basement areas of Albert I Land and main parts of Haakon VII Land west of the Rabotdalen–Hannabreen and Breibogen faults (Fig. 2a, b). The Biscayarhalvøya Terrane is situated NE of the Albert I Land Terrane between the Rabotdalen–Hannabreen Fault Zone in the west and the Breibogen Fault Zone in the east (e.g. Harland 1997a; Dallmann 2015).

**Albert I Land Terrane**

The pre-Devonian basement of the Albert I Land Terrane consists of two major structural units: a metasedimentary unit and a migmatite complex (Fig. 3). The metasedimentary Krossfjorden Group is dominated by 6500 to 8000 m thick mica schists and subordinate amphibolites of the Nissenfjella and Signehamna formations and marbles of the Generalfjella Formation (Gee and Hjelle 1966; Abakumov 1979; Hjelle 1979). The high-grade Smeerenburgfjorden Migmatite Complex consists of heterogeneous migmatites and...
gneisses, metatexites, diatexites and granitoids (Holtedahl 1914; Schenk 1937; Gee and Hjelle 1966, 1979; Hjelle and Ohta 1974; Abakumov 1979; Gjelsvik 1979; Dallmann et al. 2002).

Isotopic age determinations have shown that the Smeerenburgfjorden Complex represents a high-grade metamorphosed equivalent of the Krossfjorden Group, strongly affected by Caledonian deformation, migmatization and granite genesis between c. 434 and 417 Ma (Myhre et al. 2008; Pettersson et al. 2009a, b). The sedimentary protolith of both units was deposited in late Stenian to early Tonian times between 1021 ± 11 Ma (maximum depositional age) and c. 995 Ma (intrusion age of a dyke, interpreted as the minimum depositional age) (Pettersson et al. 2009b). The succession was intruded by Tonian granitoids at c. 970 to 960 Ma and Silurian anatectic grey granites at c. 435 to 420 Ma (Ohta et al. 2002; Pettersson et al. 2009a). Inherited zircon grains in granitoids and migmatites have an age range between 1820 and 1030 Ma, comparable to the detrital zircon age span of the sedimentary protoliths (Pettersson et al. 2009a, b). During the final stage of the Caledonian Orogeny, the Smeerenburgfjorden Complex was intruded by the post-tectonic Hornemantoppen Batholith that yielded Rb–Sr ages of 414 ± 10 Ma (Hjelle 1979) and 413 ± 5 Ma (Balašov et al. 1996), and a U–Pb age of 418 ± 1 Ma on zircon and titanite (Myhre et al. 2008).

**Biscayarhalvøya Terrane**

The basement in Biscayarhalvøya is subdivided into the Richarddalen Complex and the Biscayarfonna Group with the metasedimentary Biscayarhuken and Montblanc formations (e.g. Peucat et al. 1989; Harland 1997a, b; Gromet and Gee 1998; Ohta et al. 2003; Dallmann 2015) (Figs 3 and 4). Both tectonostratigraphic units are separated by a tectonic boundary (e.g. Peucat et al. 1989; Gromet and Gee 1998; Ohta et al. 2003). The Biscayarfonna Group has been correlated with the Krossfjorden Group by Harland (1997a, b).

The Biscayarfonna Formation of the Biscayarfonna Group is a >3.5 km thick succession of metapelitic rocks (garnet–mica schists, phyllites) with some quartzite, marble and interlayered amphibolite. A granitic intrusion into phyllites of the Biscayarfonna Formation at its southernmost outcrop on an island at the northern margin of Liefdefjorden yielded an age of 961 ± 4 Ma, interpreted as...
Northwestern Basement Province

| Albert I Land | Biscayarhalvøya | Germaniahalvøya |
|---------------|-----------------|-----------------|
| General-fjella Fm. | Sneerenburgfjorden Migmatite Complex (Ordovician–Silurian) | Lernerøyane Group (Cambro-Ordovician protolith) |
| Marble | with Cryogenian and Tonian meta-igneous rocks | Marble Unit |
| Signehamna Fm. | | Mica schist Unit |
| Mica schist | | FHK 11+12 ★ FHK53 |
| Nissen-fjella Fm. | Biscayarhalvøya | Lerner Deformation Zone |
| Gneiss, amphibolite, (passes into migmatites) | Deformation Zone | ★ FHK06 |

Fig. 3. Schematic tectono-stratigraphic overview chart of the Northwestern Basement Province (modified from Gee and Hjelle 1966; Gromet and Gee 1998; Myhre et al. 2008; Pettersson et al. 2009a, b; Dallmann 2015). For Biscayarhalvøya and Germaniahalvøya, the new results of this study are already considered. See the text for further explanations. Stars with numbers: dated samples of this study; white stars: leucogranitic dykes and veins (see Table 1).

minimum depositional age for the sedimentary protolith (Pb evaporation method on four zircon grains; Ohta and Larionov 1998).

The Montblanc Formation comprises c. 1 km of banded garnetiferous mica schist, gneiss, foliated amphibolite and quartzite. A sample of the Montblanc Formation from NW of Richardvatnet yielded detrital zircon ages of c. 2670 to 940 Ma (Pb evaporation method; Ohta et al. 2003). The youngest ages from this sample (four analyses between 938 ± 4 Ma and 952 ± 2 Ma) are younger than the age of the above-mentioned granitic intrusion and the orthogneisses that intruded the Albert I Land Terrane (970 to 960 Ma). Thus, it cannot be excluded that the Montblanc Formation (or a part of it) was deposited after the Krossfjorden Group.

40Ar/39Ar amphibole ages of c. 429 and c. 437 Ma as well as muscovite and biotite 40Ar/39Ar and Rb–Sr ages of 430 to 400 Ma were interpreted as cooling ages after Caledonian amphibolite-facies metamorphism of the Biscayarfonna Group (Dallmeyer et al. 1990).

The NNW–SSE-trending Richarddalen Complex, the tectonically highest unit of the terrane, comprises lenses of eclogite partly retrogressed to amphibolite, orthogneiss, metagabbro and marble (Gee 1966; Peucat et al. 1989; Elvevold et al. 2014). The eclogites were formed under pressure–temperature conditions of 1.9 to 2.5 GPa and 720 to 740°C and experienced amphibolite-facies retrogression to c. 1.2 GPa and 650°C along a steep exhumation path (Elvevold et al. 2014). Previous zircon dating yielded Tonian ages of c. 980–940 Ma for granitic orthogneiss and metagabbro (Peucat et al. 1989; Ohta et al. 2003; Pettersson et al. 2009a) and Cryogenian ages of 670–620 Ma for eclogite, mafic and felsic gneiss (Peucat et al. 1989; Gromet and Gee 1998) either indicating rift-related mafic and felsic magmatism at this time (e.g. Gromet and Gee 1998; Elvevold et al. 2014) or representing mixed ages (e.g. Pettersson et al. 2009a). Cambro-Ordovician 40Ar/39Ar ages (553 to 505 Ma on amphibole, 481 to 443 Ma on phengite; Dallmeyer et al. 1990) and U–Pb titanite ages (483 to 431 Ma; Gromet and Gee 1998) probably post-date the eclogite-facies metamorphism and represent the age of amphibolite-facies retrogression (Dallmeyer et al. 1990; Gromet and Gee 1998). Late Caledonian cooling ages of 430–410 Ma as recorded in the Biscayarfonna Group were obtained in the Richarddalen Complex only by Rb–Sr dating on phengitic muscovite and biotite (Dallmeyer et al. 1990).

The rock units of the Biscayarhalvøya Terrane with their inverse metamorphic tectonostratigraphic structure were interpreted as a collisional nappe stack that has experienced northward thrusting and north–south constrictional stretching (Labrousse et al. 2008).

Lernerøyane and Germaniahalvøya

The basement rocks in the study area south of Liefdefjorden on Lernerøyane and Germaniahalvøya (Fig. 5a) are dominated by gneisses, migmatites, granitoids and tectonically overlying mica schists and marbles, which have been correlated with the Sneerenburgfjorden Migmatite Complex and the metasedimentary Krossfjorden Group, respectively, of the Albert I Land Terrane (Abakumov 1979; Gjelsvik 1979; Ohta et al. 2003). The tectonic boundary between these differently metamorphosed rocks on Lernerøyane and Germaniahalvøya is the ductile Lerner Deformation Zone (Piepjohn and Thiedig 1995, 1997) (Figs 2b and 5). The most frequent rock types in the Lerner Deformation Zone are marble mylonites and cale-silicate rocks. Additionally, mica schists, gneisses, migmatitic rocks, grey granites and dykes are involved and ductilely deformed. The ductile Lerner Deformation Zone is accompanied by several parallel mylonite zones in the higher-grade metamorphosed complex below and in the lower-grade metamorphosed unit above (Piepjohn and Thiedig 1995).

The migmatites are exposed in the central parts of the NNW–SSE-trending Bockfjorden Anticline (Gee and Moody-Stuart 1966; Gee 1972) (Fig. 5). They are locally intruded by homogeneous...
granitic rocks, similar to the grey granite of the Albert I Land Terrane, which contains numerous xenoliths of marble, mica schist, fine-grained gneiss and calc-silicate rocks (Gjelsvik 1979; Piepjohn and Thiedig 1995; Ohta et al. 2003). The migmatites contain 100 m-scale domains of fine-grained and dark-coloured foliated mica schists which, in some parts, are characterized by feldspar porphyroblasts on a centimetre-scale. The structures and layering of the pre-migmatitic deformations are still preserved in these domains. Where exposed, the contacts with the surrounding metatexites and diatexites are sharp (Wyss et al. 1998).

The migmatites are tectonically overlain by a monotonous sequence of mica schists, which were correlated with the Signehamna Formation of the Krossfjorden Group in the Albert I Land Terrane (e.g. Gjelsvik 1979; Piepjohn and Thiedig 1995; Harland 1997a, b). The assemblage garnet–biotite–muscovite–chlorite indicates amphibolite-facies metamorphism. The mica

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**Fig. 4.** Geological map of Biscayarhalvøya (based on Gromet and Gee 1998; Ohta et al. 2003, 2007; Dallmann et al. 2005a, b; Labrousse et al. 2008; and own field data) with the sample sites (yellow stars). Additionally shown are the lower hemisphere, equal-area stereoplots of fabric elements of the sheared and mylonitized rocks. In the Hoeppener plots (Hoeppener 1955), the lineation is projected on the respective pole of foliation plane; arrows depict the relative sense of shear of the hangingwall. Symbols without arrowheads show the lineation only without clear sense of shear.
Schists are overlain by a marble unit, which can be divided into a lower and an upper variety (Gjelsvik 1979). This marble unit was correlated with the Generalfjella Formation of the Krossfjorden Group (Gjelsvik 1979; Piepjohn and Thiedig 1995; Harland 1997a, b). No isotopic age data from the basement rocks of Lernerøyane and Germaniahalvøya exist so far.

Late and post-Caledonian deformation events

The Northwestern Basement Province is overlain by deposits of the late Silurian to Devonian Old Red Sandstone (Siktefjellet, Red Bay and Andrée Land groups; see Dallmann and Piepjohn 2020, and references therein), which are exposed to the east and in a...
north–south-trending graben along the Raudfjorden Fault (Fig. 2). The three stratigraphic units of the Old Red Sandstone are separated by strike-slip faulting and regional uplift during the Haakonian (Gee 1972) and Monacobreen events (McCann 2000). In the early Late Devonian (Frasnian) to earliest Carboniferous (Tournaisian) interval, the Old Red deposits and the older basement on the west coast of NW Spitsbergen were affected by the contractional Ellesmerian (Svalbardian) orogenic event (Piepjohn et al. 2015; Dallmann and Piepjohn 2020; Berry and Marshall 2015).

Analytical methods

Fieldwork, including mapping, structural studies and rock sampling were carried out on Biscayarhalvøya and on Lernerøyane and Germaniahalvøya (Fig. 2b) during the expedition CASE 10 (Circum–Arctic Structural Events) of the German Federal Institute for Geosciences and Natural Resources (BGR) in the summer of 2007. Rock samples were collected and documented by several members of the CASE 10 field team. Altogether 13 samples from the major rock units of the study areas (Figs 3–5) were selected for geochronological studies after examination by thin-section petrography and whole-rock geochemistry detected by X-ray fluorescence (XRF) and inductively coupled plasma-mass spectrometry (ICP-MS). Thin-section preparation and XRF analysis were carried out at BGR Hannover, and ICP-MS analysis by Activation Laboratories Ltd. (Actlabs), Ancaster, Canada. Additionally, Rb–Sr and Sm–Nd whole-rock isotope analyses were performed at BGR, Hannover. A description of the analytical methods and results from XRF, ICP-MS and isotope measurements are available in Supplementary material 1. For zircon dating, in some cases two lithological identical rock pieces from the same outcrop but with own sample number were united after careful thin-section and XRF check.

Separation of zircon grains, mount preparation, cathodoluminescence (CL) imaging, and analysis of U–Th–Pb isotopes by laser ablation combined with inductively coupled plasma-mass spectrometry (LA-ICP-MS) techniques were carried out at the geochronology laboratories at the Goethe University Frankfurt, Germany, the University of Toronto, Canada, and the Senckenberg Naturhistorische Sammlungen Dresden, Germany (see Table 1). For a better definition of igneous and metamorphic events, U–Pb analyses on zircon and monazite using isotope dilution-thermal ionization mass spectrometry (ID-TIMS) were done on five (meta-) igneous rock samples (two of them were also dated by the LA-ICP-MS method) at the University of Toronto. Additionally, sensitive high-resolution ion microprobe zircon dating (SHRIMP II) at the Geological Survey of Canada in Ottawa, Canada, was performed on two granitic samples. Analytical details and the complete U–Pb LA-ICP-MS, ID-TIMS and SHRIMP II data are presented in Supplementary material 2. Only analyses with a concordance of 100 ± 10% are considered for our interpretation below.

We present our structural and geochronological results separately for the study areas. The samples are sorted according to structural and genetic aspects, resulting in the following order: (i) Biscayarhalvøya: meta-igneous and meta-sedimentary rocks of the Richarddalen Complex, meta-sedimentary rocks of the Biscayarfonna Group; (ii) Lerneroyne and Germaniahalvøya: meta-sedimentary rock, anatexic granite and xenoliths, leucogranitic dykes and veins (Table 1, Fig. 3). The stratigraphic terms suggested in Figure 3 are used for the description of the structural events and the sample sites.

Structural observations

Biscayarhalvøya

Regional structural analyses were performed in the metamorphic basement in the central part of Richarddalen on Biscayarhalvøya (Fig. 4). The results are described in the following paragraphs from west to east.

The Biscayarhalvøya basement complex comprises a zone of strong deformation that is sandwiched between major shear zones following Raudfjorden and Breibogen (Fig. 4). In general, foliation planes of the three main lithological units, i.e. Richarddalen Complex and Biscayarfonna Group with Montblanc and Biscayarhuken formations, dip steeply to the west and east; main fold axes are approximately north–south-oriented and shallowly plunging (Figs 4, 6a and b). This is in line with a general syn- and antiform geometry, tentatively related to imbrication of the three units during east–west contraction (Labrousse et al. 2008).

Metapsammopelitic rocks of the Richarddalen Complex show dominant subvertical to steeply west-dipping foliation planes (Fig. 6b), which are at least S2, with a pronounced subhorizontal lineation L2 indicated by aligned mica and elongated quartz and feldspar aggregates (Fig. 6c). Isoclinal intrafolial folds F2 with variably oriented axes are related to this fabric. The rocks are also affected by centimetre- to decimetre-scale open folding with axes strictly parallel to L2 and differing vergence. Folding is more intense where the foliation becomes mylonitic. Amphibolite bodies with penetrative shearing along the margins and distinct shear planes in the interior form σ-like shapes indicating dextral shear parallel to L2, as do widely distributed shear bands and σ-clasts in the metasedimentary country rocks that interlayer with banded (garnet) amphibolites. All intensely foliated rocks are interpreted as platy mylonites with Smy-foliation and Lmy-lineation.

The metapsammopelitic rocks of the Richarddalen Complex continuously grade into mica schist with quartzites and amphibolites of the Montblanc Formation in the east. There, the main foliation S2 is mylonitic and carries a pronounced subhorizontal lineation L2 recorded by aligned mica and elongated quartz and feldspar aggregates. All structural elements are comparable to those in the Richarddalen Complex, including isoclinal intrafolial F2-folds and open folds on a decimetre-scale with axes strictly parallel to L2 and differing vergence. Flattened rootless and boudinaged isoclinally folded quartz segregates and isolated fold hinges within S2 could indicate a relit earlier foliation S1 (transposed to S2) and deformation event D1, or they may alternatively relate to strong progressive deformation. However, we did not observe east–west trending down-dip lineations in any of the three units. Mafic metamagmatic bodies on a several metres scale often show σ-shapes. Like widely distributed shear bands, S–C fabrics and σ-clasts of quartz and feldspar (Fig. 6e, f), they show a dextral sense of shear parallel to lineation L2/Lmy recorded by long axes of aligned amphibole on the again steeply west-dipping foliation planes S2/Smy. In some intensely mylonitized parts, C planes of S–C fabrics become the dominant fabric.

The mylonites of the Montblanc Formation continuously grade into mylonites and phyllonites of the Biscayarhuken Formation in the east. There, the rocks derived from former mica schists (presently phyllonites) with intercalated mylonitized quartzites and feldspar-rich layers. The latter might represent former felsic aggregate clasts in metapsammopelitic rocks, and long axes of oval and up to centimetre-long mica plates in phyllonites (formerly coarser-grained mica schist), quartz–feldspar aggregate clasts in metapsammopelitic rocks, and long axes of aligned amphibole in greenschist lenses. A dextral sense of shear parallel to the lineation L2/Lmy is shown by σ-shapes of quartz and feldspar aggregates and amphibolite/greenschist bodies and widely distributed shear bands and S–C fabrics. The structural elements, therefore, are comparable to those of the other two units. These include open folds with axes parallel to the lineation and differing...
### Table 1. Sample overview and dating results

| Sample*          | Unit                        | Lithology                                | Latitude N | Longitude E | U–Pb geochronology                                                                 |
|------------------|-----------------------------|------------------------------------------|------------|-------------|-----------------------------------------------------------------------------------|
| **Biscayarhalvøya** |                             |                                          |            |             |                                                                                   |
| FHK07-45         | Richarddalen Complex        | Garnet amphibolite (meta-eclogite)       | 79°45'59.4″ | 12°18'06.6″ | LA-ICP-MS zircon (T), ID-TIMS zircon (T) a) 670 ± 5 b) 656 ± 4 a) Crystallization |
|                  |                             |                                          |            |             |                                                                                   |
| C6 -1            | Richarddalen Complex        | Mylonite orthogneiss                     | 79°46'41.3″ | 12°22'14.3″ | LA-ICP-MS zircon (T) 687 ± 5 a) Metamorphism                                      |
| FHK07-43 + 44    | Richarddalen Complex        | Banded clinopyroxene-garnet gneiss        | 79°45'56.4″ | 12°18'02.4″ | LA-ICP-MS zircon (F) 423 ± 5 b) Maximum depositional age                          |
| C9 - 1           | Montblanc Fm.               | Garnet-mica schist                       | 79°46'39.3″ | 12°23'22.7″ | LA-ICP-MS zircon (D, T) 1056 ± 9 Maximum depositional age                        |
| C20 - 1          | Biskayarhuken Fm.           | Garnet-mica schist                       | 79°46'38.0″ | 12°26'52.5″ | LA-ICP-MS zircon (D) 1150 ± 10 Maximum depositional age                          |
| **Germaniahalvøya** |                             |                                          |            |             |                                                                                   |
| FHK07-11 + 12    | Mica schist Unit (Lernerøyane Group) | Cordierite-andalusite-sillimanite-   | 79°34'03.6″ | 12°42'15.6″ | LA-ICP-MS zircon (F) c. 490 Maximum depositional age                             |
| FHK07-13 + 14    | Liefdefjorden Migmatite Complex | Grey granite                          | 79°34'21.6″ | 12°52'40.8″ | LA-ICP-MS zircon (F), SHRIMP zircon (O), ID-TIMS monazite, zircon (T) 421 ± 7 (SHRIMP zircon); 421 ± 1 Ma (abraded monazite) |
| FHK07-15         | Liefdefjorden Migmatite Complex | Leucosome of migmatite schist            | 79°34'24.0″ | 12°52'10.2″ | LA-ICP-MS zircon (F) 469 ± 2 Metamorphism                                       |
| FHK07-17         | Liefdefjorden Migmatite Complex | Amphibole gneiss                       | 79°34'24.6″ | 12°52'30.0″ | LA-ICP-MS zircon (D, T) c. 880 Maximum depositional age                           |
| FHK07-23         | Liefdefjorden Migmatite Complex | Leucogranitic segregation              | 79°34'06.0″ | 12°49'11.4″ | LA-ICP-MS zircon (D) 459–408 Metamorphism                                       |
| FHK07-53         | Mica schist Unit (Lernerøyane Group) | Leucogranitic dyke (boudin)           | 79°34'01.8″ | 12°45'42.0″ | ID-TIMS monazite, zircon (T) 963 ± 4 Inheritance                               |
| FHK07-06         | Lerner Deformation Zone     | Leucogranitic vein cross-cutting        | 79°34'08.4″ | 12°46'37.2″ | ID-TIMS monazite, zircon (T), SHRIMP zircon (O) 421–415 (monazite)               |
| FHK07-60         | Liefdefjorden Migmatite Complex | Leucogranitic dyke (boudin)            | 79°34'01.2″ | 12°46'37.8″ | ID-TIMS monazite, zircon (T) 417 ± 1 (monazite); 404 ± 9 (zircon Concordia lower intercept) |

* Short numbers for FHK07-xx samples are used in graphic presentations (without ‘07’, e.g. FHKxx).

†Lab: F, Goethe University Frankfurt/Main (Germany); D, Senckenberg Naturhistorische Sammlungen Dresden (Germany); T, University Toronto (Canada); O, GSC Ottawa (Canada).
vergence as well as relicts of isoclinal intrafolial folds. Furthermore, (kink) folds on a centimetre- to decimetre-scale with axes at a high angle to the lineation occur in all three units. These asymmetries suggest that they accommodated the final stages of dextral shearing during exhumation of the rocks.

All basement units along the investigated area in central Richarddalen record structures of the same deformational history during the same deformation event D2 (Dmy). However, in contrast to the Richarddalen Complex and the Montblanc Formation, the Biscayarhalvøya Formation is characterized by the growth of chlorite, micro-biotite and partly small garnet within thin layers in the mylonitic rocks. This points to an increase in retrograde greenschist-facies metamorphism during dextral shearing. This retrogression also affected the eastern parts of the Montblanc Formation to a variable extent where, however, there also occur layers with garnet growth syn- to post-S2/Smy. In contrast, the western parts of the Montblanc Formation and the Richarddalen Complex display late syn- to post-S2/Smy garnet growth (up to 1 cm) and post-S2/Smy static growth of quartz and feldspar. These are indicative of amphibolite-facies metamorphic overprint during late-stage and/or after dextral shearing. This variable metamorphism in the three units also affected fold structures with axes parallel to the lineation (L2 or Lmy) whereas folds with axes at a high angle to the lineation formed under decreasing temperatures mostly in the Biscayarhuken Formation in the east.

**Lernerynane and Germaniahalvøya**

Structural analyses were performed in the Liefdefjorden area (Lernerynane, Germaniahalvøya, Hornbækpollen) considering previous work by Piepjohn and Thiedig (1994, 1995). Several stages of deformation (D1–D5) were identified. Their cross-cutting relationship is well exposed on Lernerynane and gives relative timing.

**Events D1–D3**

After deposition of the protoliths of the Lernerynane Group, the first deformational event D1 led to the formation of the first detectable metamorphic foliation S1 parallel to the primary compositional bedding S0. During this deformation, numerous quartz segregations were generated in the mica schist parallel to the S1 foliation (Fig. 7a, b).

During the second deformation D2, quartz segregations and the S1 foliation were folded by isoclinal F2-folds on a centimetre-
metre-scale (Fig. 7a, c). The poorly developed corresponding S2 cleavage is only preserved in very fine-grained, quartzitic mica schist or is sometimes indicated by thin biotite layers between long limbs of isoclinally folded quartz segregations. The strike of B2-fold axes and δ2-intersection lineations shown by the intersection of the foliation S1 and cleavage S2 parallel to the B2-fold axes is predominantly NW–SE (Fig. 8a).

The D3-event represents the main deformation in the metasedimentary rocks on Germaniahalvøya and Lernerøyane. It is characterized by a penetrative S3 foliation (Fig. 7b, d) and corresponding, mostly isoclinal, F3-folds in the mica schist and marble units (Fig. 7c, d). Within the fine-grained, quartzitic parts of the mica schists, isoclinal F2-folds and quartz segregations are refolded by isoclinal F3-folds (Fig. 7b, d). In contrast to the NW–SE-striking F2-folds, the strike of B3-fold axes is NNE–SSW (Fig. 8). No clear evidence for tectonic transport directions or relative sense of shear of the hangingwall could be found in the Liefdefjorden area. Numerous unmigmatized domains of mica schist of the Liefdefjorden Migmatite Complex display isoclinal F2- and F3-folds, suggesting that deformations D1–D3 took place prior to migmatization.

High-grade metamorphism up to crustal anatexis led to the formation of anatectic granitic melts, which intruded the migmatites of the Liefdefjorden Migmatite Complex (and the D1–D3 structures, which are equivalent to those in the mica schist of the Lernerøyane Group) and crystallized as grey granites, dykes and
sills. Granitic intrusions were not observed in the overlying marbles. The anatectic event was followed by the intrusion of late-stage leucogranitic dykes, one of which cuts through the mylonitic foliation but is still slightly boudinaged. This event might mark the cessation of activity along the Lerner Deformation Zone.

Lerner Deformation Zone (D4)

A mylonitic foliation on Lernerøyane mostly dips towards WSW with a NNW–SSE-striking lineation (Fig. 8). At Hornbækpollen at the north coast of Liefdefjorden (Fig. 2b), a mylonitic foliation dips towards SW with SSW-plunging lineations (Fig. 8b). This shows that most basement rock types, which are now exposed in the Germaniahalvøya area, are included in, and affected by, the Lerner Deformation Zone. The deformation zone can be divided into two structural units (Fig. 7e). The upper unit (thrust zone) consists of mylonitic and boudinaged calc-silicate rocks, marble mylonites, marbles, porphyroblastic mica schists and some gneisses (Fig. 7f). The lower unit (fold zone) is characterized by intensely and ductilely folded and boudinaged marbles, granitic dykes and sills, mica schists and gneissose rocks (Fig. 7g). In some parts, continuous layers of granitic rocks and mica schists are affected by monoclinal, east-vergent folds on a several metres scale. In more intensely deformed parts, reliefs of isoclinal intrafolial folds with boudinaged limbs occur. Biotite in mylonitic layers and quartz mylonites in the hangingwall are disrupted but not recrystallized, indicating that the mylonitization occurred below biotite stability conditions. Feldspar is rotated and only cataclastically deformed. This indicates that the deformation took place under lower greenschist-facies conditions.

The kinematics of the Lerner Deformation Zone are still unclear. On the one hand, the rotation of porphyroclasts and boudins within the mylonites and boudinage zones as well as the east-vergent folds in the lower part of the Lerner Deformation Zone indicate transport direction of the hangingwall towards the east. This is supported by rotated feldspar porphyroclasts in the mylonitized mica schists and rotated calc-silicate boudins within the mylonites of the Wilhelmtinden Mylonite Zone (Fig. 5a). On the other hand, more frequent NNE–SSW- to NNW–SSE-striking lineations on the mylonites of the Lerner Deformation Zone (Fig. 8a, b) indicate transport direction to the north or to the south. The structural analyses indicate that the higher-grade Liefdefjorden Migmatite Complex and the lower-grade Lernerøyane Group were juxtaposed during the formation of the Lerner Deformation D4. However, it is still a question if the Lernerøyane Group was thrusted on top of or moved adjacent to the Liefdefjorden Migmatite Complex.

Event D5

The last deformation event in the study areas is marked by a contractional and brittle deformation D5, which is dominated by the major, NNW–SSE-trending, Bockfjorden Anticline and secondary parallel folds (Figs 5 and 7e). The folds are west-vergent on the eastern limb and east-vergent on the western limb of the Bockfjorden Anticline and are accompanied by a fracture cleavage. The fold axes and intersection lineations gently plunge to the south.
Sample description, petrography and geochemistry

For geochemical and isotopic data of the studied samples, we refer to Supplementary material 1. Photomicrographs are shown in Supplementary material 3-1.

Biscayarhalvøya

Garnet amphibolite FHK45 (Richarddalen Complex)

Sample FHK45 was collected from a dark lens of foliated garnet-rich rock NW of Richardvatnet (Fig. 4). The rock consists of poikiloblastic garnet (up to 1 mm in diameter), hornblende, clinzoisite, plagioclase, quartz, clinoxyroxene and biotite (partly chloritized). Accessories are carbonates, apatite, titanite and rutile. Some garnet contains quartz inclusions, which are surrounded by radial cracks. This has been explained by Chopin (1984) as result of inversion from coesite to quartz and as indication of ultra-high-pressure metamorphism.

The geochemical composition of the sample (c. 45 wt% SiO2, 15 wt% total FeO, 5 wt% MgO, 11 wt% CaO, 291 ppm V) points to a mafic igneous protolith. The chondrite-normalized rare-earth element (REE) pattern of the sample is similar to enriched mid-ocean ridge basalt (E-MORB) but more enriched. The flat pattern in the range of heavy REE excludes melt formation in the presence of garnet at greater depth. High Al2O3 (18 wt%) as well as negative Nb and Ta anomalies relative to Th and La (in a primitive mantle-normalized pattern) indicate magma formation influenced by subduction-zone-related fluids. That is supported by a εNd(t) value of 0.4, inconsistent with values of primitive MORB. Thus, a formation of the basaltic protolith in a subduction-related setting can be inferred.

Mylonitic orthogneiss C6-1 (Richarddalen Complex)

Sample C6-1 is a light-coloured, mylonitic, felsic orthogneiss from the Richarddalen Complex NE of Richardvatnet (Fig. 4). The foliated rock is formed of fine-grained (<1 mm) quartz and feldspar/plagioclase (slightly altered) as well as dark bands of biotite (partly chloritized) and titanite. Accessories are apatite, metamict orthite, pumpellyite, epidote/clinozoisite, carbonate, zircon and opaque grains.

Banded gneiss FHK43 + 44 (Richarddalen Complex)

The sample was collected next to the garnet amphibolite FHK45 NW of Richardvatnet (Figs 4 and 9a). It represents a mylonitic, felsic gneiss (72 to 74 wt% SiO2). Darker, greenish-grey bands consist of a fine-grained matrix of quartz, plagioclase, biotite (partly altered to chlorite) and hornblende, as well as elongated aggregates (parallel to the foliation) of muscovite, clinzoisite and accessory titanite. Light-grey bands also have a fine-grained matrix of plagioclase, quartz, clinzoisite and some muscovite flakes. Garnet (partly chloritized) is present in both bands. The protolith is probably a siliciclastic sediment rich in felsic igneous material, which is supported by relatively low 87Sr/86Sr(t) values of 0.7076 to 0.7094 and the negative εNd(t) values of −4.8 and −6.7.

Garnet–mica schist C9-1 (Monthblanc Formation, Biscayarfonna Group)

Sample C9-1 is a mica schist from the Monthblanc Formation NE of Richardvatnet (Fig. 4). The schist is formed of long, partly irregularly folded aggregates of biotite (partly chloritized) and muscovite in a fine-grained matrix of quartz and plagioclase. Accessories are garnet, apatite, tourmaline, titanite aggregates, zircon and opaque lenses and grains.

Garnet–mica schist C20-1 (Biscayarhukene Formation, Biscayarfonna Group)

Sample C20-1 comes from the Biscayarhukene Formation east of Richardvatnet (Fig. 4). The strongly foliated schist is formed of muscovite, biotite, fine-grained granoblastic quartz, garnet (c. 1.5 mm in diameter) and some staurolite grains (up to 0.8 mm in length). Accessories are tourmaline, apatite, zircon and small opaque grains and lenses. Mica forms bands parallel to the foliation, but biotite (partly chloritized) also forms large lenses (up to c. 1.5 mm in diameter) speared by small muscovite flakes.

Lerneryøane and Germanialhalvøya

Mica schist FHK11 + 12 (Lerneryøane Group)

The mica schist sample was collected on the western Lerneryøane, east of the Wilhelmtinden Mylonite Zone, from a dark layer of a banded and deformed schist (Figs 5 and 9b). The amphibolite-facies-metamorphosed, fine-grained, foliated metasedimentary rock consists of reddish-brown biotite, muscovite, quartz, feldspar/plagioclase, minor andalusite, sillimanite, cordierite and garnet. Accessories are tourmaline, apatite and zircon. Andalusite forms large crystals parallel to the foliation, which are cut by fine orthogon al muscovite veins. In hydrothermally affected parts of the rock, biotite is altered to chlorite and fine-dispersed pyrite is present.

From the mineral paragenesis, the metamorphic P–T conditions can be estimated to c. >500°C and <0.4 GPa (corresponding to a depth of c. 15 km). The two geochemically analysed rock pieces of the sample show only little variation in the composition (63–67 wt% SiO2, 15.7–17.2 wt% Al2O3, 6.4–7.4 wt% total FeO, 2.1–2.4 wt% MgO, 2.0–2.2 wt% CaO, 2.9–3.2 wt% Na2O, 2.7–3.2 wt% K2O). Their chondrite-normalized REE patterns are almost equal and typical for pelitic sediments. This is supported by 87Sr/86Sr(t) values of 0.72898 and 0.72712 and the εNd(t) of −8.5 and −9.0 for the time of deposition, which is typical for crustal components.

Grey granite FHK13 + 14 (Liefdefjorden Migmatite Complex)

The sample was collected from an anatectic grey granitoid rich in xenoliths at the southern margin of Liefdefjorden, at the eastern side of the Bockfjorden Anticline (Figs 5 and 9c). The medium-grained granitoid consists of quartz, plagioclase and perthitic feldspar. Nests of fine biotite flakes (partly chloritized), apatite, as well as accessory orthite, zircon and opaque phases are also present. The rock is of granodioritic to granitic composition (69–74 wt% SiO2, 2.6–2.7 wt% Na2O, 4.1–4.3 wt% K2O). High 87Sr/86Sr(t) values of 0.72120–0.72083 and negative εNd(t) of −8.4 and −9.1 indicate the formation of this granitoid from crustal material.

Leucosome of migmatic schist FHK15 (Liefdefjorden Migmatite Complex)

The sample was collected from a large (c. 2 m × 4 m) xenolith of migmatic schist within anatectic grey granite at the south coast of Liefdefjorden (Figs 5 and 9d). Metamorphic foliation of the xenolith is cut at the contact to the country rock. The xenolith is a
dark grey, strongly foliated rock penetrated by light grey leucosome veins. The studied sample represents the leucosome. It consists of medium- to coarse-grained alkali feldspar (up to 5 mm in length), plagioclase, quartz, and bands and lenses of biotite (partly chloritized). Minor to accessory amounts of apatite, myrmekite, muscovite flakes and garnet (in alkali feldspar) are present.

The major element composition of the leucosome sample is very similar to that of the mica schist and grey granite samples (FHK11 + 12, FHK13 + 14) from the study area; however, the concentrations of Ba (1470 ppm), Sr (569 ppm), Th (120 ppm), U (8.5 ppm) and light REEs are clearly higher. The REEs are more strongly fractionated, showing decreasing chondrite-normalized values from La to Yb (except a negative Eu anomaly. The εNd(t) value (~1.7) is less enriched and the Nd model age TDM of 1.07 Ga is lower compared with the other samples.

**Amphibole gneiss FHK17 (Liefdefjorden Migmatite Complex)**

Sample FHK17 is from a large body in grey granite at the south coast of Liefdefjorden (Figs 5 and 9e). This body consists of leucocratic and melanocratic banded, in parts migmatitic gneiss (Fig. 9e) and is in sheared contact to the hangingwall host rock. The footwall contact is not exposed. The sample was taken from the melanocratic part. The fine-grained, foliated gneiss consists of plagioclase (partly sericitized), quartz and amphibole (hornblende;
up to 1 mm in length). Some biotite aggregates (partly chloritized) are also present. Accessories are apatite, zircon and opaque grains (Fe oxides and sulfides).

The geochemical composition of the amphibole gneiss is mostly similar to mica schist FHKK1 + 12, including the chondrite-normalized REE pattern, but is richer in CaO (9 wt%), MnO and Sr, and very poor in alkalis (Na2O+K2O = 1.5 wt%). The protolith was probably a carbonate-bearing psammopelitic sediment.

**Leucogranitic segregation FHKK23 (Liefdefjorden Migmatic Complex)**

This leucogranitic segregation is exposed on the eastern island of Lernøyane (Fig. 5). It is part of a banded metasedimentary rock and cross-cut by a late generation of leucocratic veins (Fig. 9f). The leucocratic medium-grained rock consists mainly of plagioclase, perthite (c. 2 mm), and quartz that forms interstitial anhedral grains with undulose extinction, indicative of metamorphic overprint, as well as dispersed large (2 mm) reddish-brown biotite flakes (partly chloritized). Accessories are apatite inclusions in biotite and zircon inclusions in feldspar, and scarce myrmekitic textures.

The leucogranitic segregation (76 wt% SiO2, 3.25 wt% CaO, 3.0 wt% Na2O, 2.6 wt% K2O) is weakly peraluminous and represents an I-type granitic melt with an aluminum saturation index (ASI) of 1.03. Most trace elements have very low concentrations with the exception of Ba, Sr and Rb. Except for a positive Eu anomaly, the REEs are normalized REE pattern, but is richer in CaO (9 wt%), MnO and Sr, and very poor in alkalis (Na2O+K 2O = 1.5 wt%). The protolith normalized REE pattern, but is richer in CaO (9 wt%), MnO and Sr, and very poor in alkalis (Na2O+K 2O = 1.5 wt%). The protolith was probably a carbonate-bearing psammopelitic sediment.

**Leucogranitic boudin FHKK53 (Lernøyane Group)**

The sample is from a small leucogranitic boudin sheared from a larger boudin within dark mica schist west of the Lerner Deformation Zone (Fig. 5). Both the granitic rock and the country rock (biotite–muscovite schist) are foliated. The granitic rock consists of undulose quartz, alkali feldspar (partly sericitized), lenticular aggregates of muscovite, as well as accessory zircon (in feldspar) and apatite.

**Granitic vein FHKK06 (Lerner Deformation Zone)**

The leucogranitic sample stems from a small and partly disrupted vein cross-cutting mylonitic rocks of the Lerner Deformation Zone at the contact between the Lernøyane Group and the Liefdefjorden Migmatic Complex (Figs 5 and 9h).

The leucogranite (76 wt% SiO2, 0.7 wt% CaO, 2.2 wt% Na2O, 5.9 wt% K2O) is peraluminous and of S-type character (ASI = 1.28). In contrast to FHKK23, the heavy REEs are not fractionated. Some trace elements such as Sb (17 ppm), Ta (4.4 ppm), Pb (44 ppm), U (10.6 ppm) are enriched and show the highest values of the whole sample set. 87Sr/86Sr(t) values of 0.71430 and 0.714 are present. Accessories are apatite, zircon and opaque grains (Fe oxides and sulfides).

The geochemical composition of the amphibole gneiss is mostly similar to mica schist FHKK1 + 12, including the chondrite-normalized REE pattern, but is richer in CaO (9 wt%), MnO and Sr, and very poor in alkalis (Na2O+K 2O = 1.5 wt%). The protolith was probably a carbonate-bearing psammopelitic sediment.

**Leucogranitic boudin FHKK60 (Liefdefjorden Migmatic Complex)**

The sample derives from a boudin of a leucogranitic dyke cutting through migmatic east of the Lerner Deformation Zone (Figs 5 and 9g).

**Geochronological results and interpretations**

For the detailed method description and zircon and monazite isotope analyses we refer to Supplementary material 2.

**Biscayarylavoya**

**Garnet amphibolite FHKK45 (Richarddalen Complex)**

The sample yielded abundant colourless, stubby and prismatic subrounded zircon grains with sizes between 100 and 250 µm. A total of 60 spots were measured on 20 zircon grains by the LA-ICP-MS method with 48 of them yielding 90 to 110% concordant 206Pb/238U ages. Eight of 48 spots showed a high 88Sr signal and were rejected due to possible intersection of the beam with altered zones or inclusions. The remaining 40 spots yielded 206Pb/238U ages in the range of 754 to 602 Ma (Figs 10a–c and 11a, b).

There are two groups of zircon grains with respect to the internal structures: zircon grains with oscillatory-zoned core and thick U-poor rim and zircon grains with sector zoning and very thin U-poor rim (Fig. 10c). While the oscillatory-zoned cores yield a Concordia age of 670 ± 5 Ma, the sector-zoned zircon grains show a 14 Ma younger Concordia age of 656 ± 4 Ma (Fig. 10a, b). The rims of the oscillatory-zoned grains often show discordant ages. Concordant rim ages are somewhat younger than the ages of the corresponding core. One zircon grain with oscillatory-zoned core and thick U-poor rim shows older ages for both core and rim (754 to 671 Ma; five spots) and is interpreted to represent an inherited grain. The lack of older inherited continental-sourced zircon grains additionally supports formation of this rock in an island-arc setting.

The Th/U ratios for all analyses, including the U-poor rims, are high (0.28 to 1.95) and fall within the range for magmatically formed zircon. These high values would be unexpected for metamorphic zircon grains that have formed under amphibolite-facies metamorphic conditions. However, high Th/U ratios are reported from high-grade, granulite-facies and eclogite-facies metamorphic zircon grains (Ashwal et al. 1999; Vavra et al. 1999; Li et al. 2015). Additional ID-TIMS analysis of three zircon grains with 206Pb/238U ages of 658 ± 1 Ma, 654 ± 1 Ma, and 648 ± 1 Ma are well within the LA-ICP-MS age distribution of the sector-zoned zircon grains.

The LA-ICP-MS U–Pb data suggest that the zircon grains of FHKK45 underwent a three-stage growth/recrystallization. (1) Ages of oscillatory-zoned cores are interpreted to represent the magmatic zircon growth at 670 Ma. (2) The Concordia age of the sector-zoned grains of 656 Ma is interpreted as the age of eclogite-facies metamorphism. Such zircon domains usually form as result of textural disruption during high-pressure events (Corfu et al. 2003). A subduction within 14 Ma into depths of 70 to 100 km followed by eclogite-facies metamorphism would agree with calculated subduction velocities (e.g. Rubatto et al. 1999). (3) The rims show slightly younger ages than the corresponding cores; however, the spread of ages for the zircon rims does not allow for defining a Discordia.

Although we prefer the interpretation given above, another interpretation cannot be excluded. Sector-zoned grains are also common in mafic rocks and all zircon grains could have recorded magmatic ages. In this case the rock formed over a longer time span in an island arc and has not experienced eclogite-facies metamorphism but prograde amphibolite-facies metamorphism (not recorded by the zircon data).

**Mylonitic orthogneiss C6-1 (Richarddalen Complex)**

The sample yielded abundant colourless, stubby and prismatic subrounded zircon grains consisting mostly of euhedral to subrounded fresh prisms with oscillatory to sector-zoned internal structures mantled by secondary low-U rims. Almost all LA-ICP-MS U–Pb analyses (n = 69, including two on rims) yielded concordant to slightly discordant 206Pb/238U ages in a range of 708 to 619 Ma, similar to sample FHKK45 (Figs 10d and 11c) and...
Th/U ratios of 0.26 to 1.33 (average: 0.59 ± 0.39; Fig. 11d) typical for igneous zircon growth. A Discordia calculation resulted in a well-defined upper intercept $^{207}\text{Pb}^{206}\text{U}$ age of 687 ± 4 Ma (MSWD 1.1) when forcing the Discordia through zero (Fig. 10d). This age is interpreted to represent the formation age of the felsic igneous protolith of the gneiss.

Fig. 10. Concordia diagrams for analysed zircon grains of samples from Biscayarhalvøya. (a)–(c) Garnet–amphibolite FHK45 from the Richarddalen Complex. Concordia ages are given for zircon grains with oscillatory-zoned zircon cores (grey error ellipses in (a); $n = 15$) and with sector zoning (green error ellipses in (b); $n = 28$). Red error ellipses in (a) are rim measurements ($n = 15$). Zircon cathodoluminescence images (c) depict sector-zoned and oscillatory-zoned zircon grains with core (dark) and rim (bright) structures. Yellow circles show spots analysed. Numbers next to the zircon grains refer to the measurement spot (first line) and the age in Ma (second line). (d) Mylonitic orthogneiss C6-1 from the Richarddalen Complex. Upper intercept age for C6-1 is calculated by forcing the Discordia through zero ($n = 64$). (e) Banded gneiss FHK43 + 44 from the Richarddalen Complex ($n = 76$). Youngest Concordia zircon ages for measurements with Th/U < 0.1 are given. (f) Garnet–mica schist C9-1 from the Montblanc Formation ($n = 85$). (g) Garnet–mica schist C20-1 from the Biscayarhuken Formation ($n = 86$). For both samples, C9-1 and C20-1, the youngest zircon (Concordia) age is given. Data are in Supplementary material 2. Abbreviations: disc – discordant measurements (level of concordance <90% and >110%); Sr – high Sr signal; C + E – concordance and equivalence.
Banded gneiss FHK43 + 44 (Richarddalen Complex)
The sample yielded abundant zircon grains, which are stubby to short-prismatic and mostly rounded. CL images show that sector- to oscillatory-zoned areas are often overgrown or to variable degrees replaced by unzoned bright CL domains. LA-ICP-MS U–Pb dating of zircon core and rim areas yielded concordant to subconcordant ages (100 ± 10%) between c. 2300 and 414 Ma (n = 78 of 91 spots) (Figs 10e and 11e, f) with Th/U ratios of 0.01–1.38. The most dominant group of zircon ages ranges from c. 730 to 600 Ma (n = 40) with a peak at around 680 Ma.

Mesoprotrozoic data show a broad range from c. 1560 to 994 Ma (n = 22). Only one Paleoproterozoic age (c. 2300 Ma) was found. A small group of rim analyses (n = 5) with very low Th/U of <0.1, typical for metamorphic zircon, and 206Pb/238U ages between 425 and 414 Ma, yield a Concordia age of 423 ± 5 Ma (MSWDc+e= 0.8; n = 5). The youngest grains with Th/U > 0.1 comprise five concordant 206Pb/238U ages between 527 ± 9 and 448 ± 9 Ma. Although these ages do not give a coherent group, their presence points to an early Paleozoic (probably Late Ordovician) deposition of the sedimentary protolith.

Garnet–mica schist C9–1 (Monteblanc Formation, Biscayarfonna Group)
The sample yielded only tiny (<50 µm), mostly well-rounded zircon grains. LA-ICP-MS analysis was carried out at two laboratories (Toronto and Dresden). The number of concordant to slightly discordant (+10%) analyses was 57 (out of 76 spots) at Toronto and 28 (out of 43 spots) at Dresden. The analytical results show a good match and are combined for graphic presentations and interpretation (Figs 10f and 11g, h).

The combined dataset yields ages in the range of c. 2850 to 1050 Ma with Th/U ratios mostly >0.1 to 1.7 typical for igneous zircon formation. Most of the ages are in the interval between c. 2000 and 1050 Ma, with a main cluster between c. 1800 and 1400 Ma (n = 49). Nine ages are older than 2000 Ma. A metamorphic event at around 2000 Ma is recorded by six analyses with Th/U < 0.1. The maximum depositional age is defined by a group of three concordant zircon grains with ages between 1045 and 1058 Ma and Th/U ratios of 0.12 to 0.18 (Fig. 11b). A single age of 508 Ma with a very low Th/U of 0.02 may have been partially reset by later overprint.

Garnet–mica schist C20–1 (Biscayarhuken Formation, Biscayarfonna Group)
Zircon grains of sample C20–1 are tiny fragments (c. 50 µm) and larger, subrounded grains. They mostly show oscillatory or sector zoning in CL images. LA-ICP-MS dating yielded concordant to slightly discordant U–Pb ages between c. 2500 and 1140 Ma with Th/U ratios of 0.13–1.21 (n = 86 of 114 spots). Most prominent are 206Pb/238U ages in the range of c. 1760 to 1450 Ma (n = 58) (Figs 10g and 11i, j). The maximum depositional age is defined by the seven youngest analyses to be 1150 ± 10 Ma (MSWDc+e = 0.51) (Fig. 10g).

Lerneroyane and Germaniahalvoyen
Mica schist FHK11 + 12 (Lerneroyane Group)
Zircon grains of sample FHK11 + 12 are small (mostly c. 50 to 100 µm), subrounded or fragmented relics, sometimes with oscillatory or sector-zoned internal structure. Some grains are mantled by a small inhomogeneous low-U rim mostly too small to be dated. Grain margins are strongly ragged, indicative of corrosion processes. LA-ICP-MS U–Pb dating yielded concordant to subconcordant ages (+10%) in the range of c. 1890 to 490 Ma (n = 64) (Figs 12a and 13a, b). Most of the dated grains are characterized by Th/U of 0.1 to 1.4 typical of magmatic zircon growth. The dominant group of ages is in the range of c. 1770 to 1350 Ma (n = 48), with peaks at around 1650 and 1500 Ma. The youngest grains comprise five ages between c. 571 and 492 Ma (Fig. 12a) with Th/U ratios of 0.14 to 0.92, which do not form a coherent group. These youngest ages point to a Cambrian (or later) age of deposition of the sedimentary protolith of the Mica-schist Unit.

Grey granite FHK13 + 14 (Liefdefjorden Migmatite Complex)
U–Pb dating was performed on zircon and monazite. Zircon was investigated by LA-ICP-MS, SHRIMP and ID-TIMS techniques, monazite by ID-TIMS. Zircon grains are hydridomorphically well rounded. Internal structures are very heterogeneous. Oscillatory to sector-zoned domains predominate. Many grains show cores partially replaced by optically homogeneous domains. Both are overgrown by weakly zoned thin rims. LA-ICP-MS U–Pb dating (Frankfurt) yielded 100 concordant to subconcordant ages (out of 112 spots) in the range of 2811 to 426 Ma (Fig. 13c, d). The dominant age population is between c. 1000 and 900 Ma (n = 49) with a weighted mean age of 947 ± 8 Ma (MSWD = 13). A smaller age group of c. 1170 to 1110 Ma (n = 9) and a less well-defined group between c. 1790 and 1500 Ma (n = 19) are present. Only two ages of 579 and 536 Ma give a hint to a possible involvement of latest Neoproterozoic to Cambrian sediments (Fig. 12b). Th/U ratios of mostly 0.1 to 1.0 for all spots with ages >536 Ma are indicative of a predominantly magmatic origin. The youngest population is represented by five ages between 442 and 426 Ma with a weighted mean of 436 ± 9 Ma, all with low Th/U < 0.1, indicative of metamorphic zircon growth.

SHRIMP II dating (Ottawa) yielded 19 discordant to slightly discordant zircon ages (out of 30 spots) in the range of c. 1800 to 390 Ma. The dominant age population was of c. 1800 to 1550 Ma (n = 11), whereas no ages were found in the range of the most dominant population of the LA–ICP–MS dataset (1000 to 900 Ma). The 10 youngest, mostly stronger discordant ages with Th/U < 0.1 were measured at rims and are in the range of 439 to 390 Ma. Eight of them give an average 206Pb/238U age of 421 ± 7 Ma (Fig. 12b). They overlap with the five youngest ages from LA-ICP-MS measurement (Fig. 12b). Six abraded zircon grains dated by ID-TIMS (Toronto) gave mostly discordant ages in the range of the LA–ICP–MS data. Only a single, long prismatic grain with Th/U = 0.5 yielded a nearly concordant (+1.9%) 206Pb/238U age of 422 ± 1 Ma. Six monazite grains (abraded and unabraded) dated by ID-TIMS yielded nearly concordant 206Pb/238U ages between 422 and 418 Ma (Fig. 12c).

The LA-ICP-MS and SHRIMP data (obtained on zircon grains separated from different pieces of the same sample) show that most zircon grains of FHK13 + 14 are inherited from the metasedimentary rocks of the basement from which the anatectic granitoid was generated. The zircon data give information about the sources of the sedimentary protolith and suggest an early Paleozoic deposition age. Differences in the frequency of age populations between the LA-ICP-MS and SHRIMP datasets are probably due to heterogeneities within the S-type granite.

The 206Pb/238U ID-TIMS ages of c. 422 to 421 Ma on three abraded monazite grains give a mean age of 421 ± 1 Ma (95% conf.; MSWD = 1.5, probability = 0.23). Together with the age of 422 Ma of one zircon grain, this is interpreted as the best estimate of the emplacement age of the anatectic grey granite. The slightly older 442 to 426 Ma age group from the LA-ICP-MS analysis is probably either related to reset of inherited zircon grains during metamorphism or to the onset of metamorphic zircon growth during progressing high-grade metamorphism and migmatization.
The latter seems to be more likely since it is supported by the youngest group of measured SHRIMP ages on zircon rims.

**Leucosome of migmatitic schist FHK15 (Liefdefjorden Migmatite Complex)**

The sample yielded abundant zircon grains, which are mostly subrounded prismatic to stubby. The internal structures revealed by CL imaging are variable and complex with oscillatory to sector-zoned central areas mantled and/or replaced by either high-U or low-U zones partly showing weak zoning. LA-ICP-MS U–Pb analyses yielded mostly concordant ages \( n = 93 \) of 101) in the range of c. 1560 to 460 Ma (Figs 12d and 13e, f). Most prominent is a population of late Stenian to Tonian ages \( (c. \ 1015 \ \text{to} \ 899 \ \text{Ma}; \ n = 47) \). Ages \( >1015 \ \text{Ma} \) are represented only by three analyses. A smaller population is of Cryogenian age \( (c. \ 833 \ \text{to} \ 651 \ \text{Ma}; \ n = 15) \). A Cambro-Ordovician population varies between 542 and 465 Ma \( (n = 8) \). All measurements revealed Th/U ratios of 0.2 to 1.4, typical of a magmatic origin.

The Cambro-Ordovician age group with high Th/U ratios overlaps with a population of 539 to 463 Ma \( (n = 19) \), which is characterized by replacement/overgrowth domains with Th/U < 0.1

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Fig. 11. Combined frequency and probability plots (a), (c), (e), (g), (i) and Th/U v. age plots (b), (d), (f), (h), (j) of LA-ICP-MS dated samples from Biscayhalvoya. All U–Pb data are available in Supplementary material 2.

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The latter seems to be more likely since it is supported by the youngest group of measured SHRIMP ages on zircon rims.
typical for metamorphic zircon growth. The 11 youngest ages of this metamorphic population form a coherent group with a Concordia age of 469 ± 2 Ma (Fig. 12d). This age is interpreted to date high-grade metamorphism and migmatization during the Middle Ordovician.

LA-ICP-MS U–Pb ages indicate that zircon grains of the leucosome FHK15 originated from their metasedimentary host rock. The protolith of the latter was mainly sourced from a 1000 to 900 Ma island arc as suggested by the almost complete lack of ages >1010 Ma. The maximum depositional age of the sedimentary protolith is difficult to define due to the metamorphic overgrowth of many zircon grains. A Cambrian age is suggested from the youngest ages of 511 and 525 Ma combined with high Th/U > 0.8 measured on zircon cores (Fig. 13c, f).
Amphibole gneiss FHK17 (Liefdefjorden Migmatite Complex)

Zircon grains separated from two different pieces of sample FHK17 were analysed by means of LA-ICP-MS at two laboratories (Toronto and Dresden). Due to differences in zircon preparation, CL images are only available from the Dresden dataset. The sample yielded mostly moderately to well-rounded zircon grains of small size. Altogether, LA-ICP-MS dating yielded 93 concordant to slightly discordant (±10%) ages (out of 173 spots) with a range of c. 3100 to 400 Ma (Figs 12e and 13g, h). Most ages are in the range of c. 1900 to 950 Ma (n = 71) with a cluster between c. 1650 and 1400 Ma (n = 32). The Th/U ratios (mostly >0.1 to 1.3) are typical for magmatic zircon. Only four ages are older than 1900 Ma, and two are younger than 950 Ma (893 and 878 Ma). The youngest age of 878 ± 27 Ma with high Th/U of 0.6 constrains the maximum depositional age of the sedimentary protolith. The youngest ages with Th/U < 0.1 (n = 16) vary between 459 and 408 Ma, but do not form a coherent group. They are most likely related to metamorphic overprint of older zircon during the Caledonian Orogeny. The c. 470 Ma high-grade metamorphic event (see sample FHK15) is not recorded in this sample. A comparison with the age spectrum of...
the grey granite (sample FHK13 + 14; Fig. 13c, d) indicates that this type of rock was involved in the generation of the granite and might have contributed mainly zircon ages >1000 Ma.

**Leucogranitic segregation FHK23 (Liefdefjorden Migmatite Complex)**

Zircon grains for LA-ICP-MS dating are mostly euhedral and in CL images they often show faint oscillatory or sector zoning along the rims but also show domains which are featureless. Some grains have irregular patchy internal textures. Fine cracks are common. Most grains have very high U concentration (mostly >1000 ppm) and low Th/U < 0.1. Only 55 out of 114 analyses yielded concordant to subconcordant ages (±10%) between 503 and 339 Ma (Figs 12f and 13i, j). One grain has an older age of 1650 Ma and Th/U = 0.35 measured on a bright U-poor domain and is considered to be from inheritance.

The euhedral shape of most grains combined with very low Th/U ratios and irregular patchy domains indicates the influence of later metamorphic or metasomatic processes on original magmatic crystallized zircon grains. Such a combination of textures and Th/U ratios is described by Hoskin and Black (2000) as recrystallization in the solid state. During this process, primary features such as zoning patterns and external morphology may be retained but certain elements, e.g. Th, are removed from the lattice due to structural strain resulting in low Th/U ratios. This is also supported by the large spread of concordant U–Pb ages (Fig. 12f), which results from differently intense memory of the protolith age (Ashwal et al. 1999; Hoskin and Black 2000).

Zircon grains with irregular internal structures such as cloudy or featureless dark domains (n = 17) show U–Pb ages from 468 to 339 Ma. The patchy internal texture of the zircon grains is probably caused by metasomatic replacement of originally low-U domains by zircon rich in U during late- to postmagmatic recrystallization (Corfu et al. 2003). The youngest age group of c. 390 Ma is interpreted as the final stage of metasomatism during late Caledonian fluid migration. However, the majority of zircon grains (n = 38) show faint oscillatory or sector zoning with U–Pb ages from 503 to 405 Ma and variable, low Th/U ratios from 0.01–0.19, which we interpret as a result of variously intense solid-state recrystallization during metamorphism (Figs 12f and 13j). A significant subset of the data (n = 20) clusters within error, giving a Concordia age of c. 460 Ma (Fig. 12f), which might approximate the time of magmatic crystallization.

**Leucogranitic boudin FHK53 (Lernerøyane Group)**

Six abraded zircon grains and two monazite grains were analysed by ID-TIMS. The zircon data form a Discordia with upper and lower intercepts of 963 ± 4 Ma and 410 ± 4 Ma, respectively (MSWD = 1.15) (Fig. 14b). The monazite grains yielded concordant 206Pb/238U ages of 418 ± 1 and 415 ± 1 Ma (Fig. 14d). The young ages and geochemistry suggest that this sample might represent a late magmatic, hydrothermally enriched dyke.

**Granitic vein FHK06 (Lerner Deformation Zone)**

The sample yielded only a small number of zircon grains. Monazite is abundant as small crystals. ID-TIMS analyses on zircon and

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**Fig. 14.** Concordia diagrams for analysed zircon and monazite grains of leucogranite dykes and veins from Germaniahalvøya. (a) SHRIMP data for zircon overgrowths of samples FHK06. (b, c) ID-TIMS upper intercept age of zircon grains from samples FHK53 and FHK60. (d) ID-TIMS monazite ages of samples FHK06, FHK53 and FHK60. All U–Pb data are available in Supplementary material 2.
monazite grains as well as SHRIMP analyses on polished zircon were carried out. ID-TIMS analyses of six abraded zircon grains gave scattered and mostly discordant data between 2643 and 752 Ma, indicating the presence of widely different age domains. Two grains gave nearly concordant 207Pb/206Pb ages of around 1960 Ma. Analyses of two fresh monazite grains produced concordant data with 206Pb/238U ages of 420.5 ± 1 Ma and 415 ± 1 Ma (Fig. 14d). SHRIMP analyses (13 spots) on 10 zircon grains yielded eight near-concordant (+10%) data. Ages of core areas scatter from c. 2700 to 1600 Ma. Overgrowths show very high U concentrations (3000 to 12 000 ppm) and low Th/U (<0.1) with an age range from 447 to 397 Ma (Fig. 14a), indicative of a metamorphic overprint. Therefore, the two monazite ages of c. 421 and 415 Ma might be the best estimate for the emplacement of the vein during formation of the high-strain Lerner Deformation Zone.

**Leucogranitic boudin FHK60 (Liefdefjorden Migmatite Complex)**

The sample contains only a small amount of zircon; monazite is very rare. Six abraded zircon grains and one monazite grain were analysed by ID-TIMS. Five of the zircon data form a Discordia with intercepts of 950 ± 11 Ma and 404 ± 9 Ma (MSWD = 0.65) (Fig. 14c). The monazite grain gave a concordant 207Pb/206Pb, 238U age of 417 ± 1 Ma (Fig. 14d). The very similar ID-TIMS dating results of the leucogranitic boudin samples FHK53 and FHK60 from west and east of the Lerner Deformation Zone indicate a similar formation as late magmatic rocks.

**Summary of geochronological results and interpretation**

**Biscayarhalvøya**

LA-ICP-MS zircon data of two mica schist samples from the Biscayarfonna Group show that the sedimentary protolith was deposited after 1050 Ma (maximum depositional age of C9-1) and was dominantly sourced from c. 1700 to 1500 Ma igneous rocks and subordinately from rocks of Grenvillian age, whereas ages >2000 Ma are very scarce. Thus, the age spectra of the Biscayarfonna Group are comparable with such of the metasedimentary Krossfjorden Group of Albert I Land (Pettersson et al. 2009b). The minimum depositional age is only indicated by the age of 961 Ma of a granite intrusion from the southernmost outcrop of the Biscayarhukene Formation at the north coast of Liefdefjorden (Ohta and Larionov 1998).

LA-ICP-MS zircon dating of a mylonitic gneiss from the Richarddalen Complex (sample FHK43 + 44) indicates a sedimentary protolith that was deposited not earlier than Late Ordovician and was amphibolite-facies metamorphosed at c. 423 ± 5 Ma. The sediment was dominantly sourced from Cryogenian–Ediacaran-aged (700 to 600 Ma) igneous material. Meta-igneous rock fragments with Cryogenian ages in the Richarddalen Complex are represented by garnet amphibolite (FHK45) and felsic orthogneiss (C6-1). Felsic magmatism occurred at c. 687 Ma. The mafic igneous protolith of sample FHK45 crystallized at c. 670 Ma in an island-arc setting, probably followed by eclogite-facies metamorphism at c. 656 Ma.

**Lernerøyane and Germanialhalvøya**

The Mica-schist Unit, formerly correlated with the Stenian–Tonian Signehamna Formation of the Krossfjorden Group of the Albert I Land Terrane, is represented by sample FHK11 + 12. The youngest detrital zircon ages between 571 and 492 Ma indicate a late Cambrian maximum depositional age of the pelitic protolith in contrast to the Stenian–Tonian metasedimentary rocks in the Krossfjorden Group. We assume that this result is representative for all metasedimentary rocks on Lernerøyane and Germanialhalvøya (Lernerøyane Group in Fig. 3); however, confirmation by further work is needed.

The Lernerøyane Group is tectonically juxtaposed with a migmatite complex in the core of the Bockfjorden Anticline (Fig. 5), formerly correlated with the Smeerenburgfjorden Complex, the high-grade metamorphic equivalent of the Stenian–Tonian Krossfjorden Group of the Albert I Land Terrane. LA-ICP-MS zircon dating on leucogranitic dykes and veins from the Lernerøyane Group shows that the sedimentary protolith was amphibolite-facies metamorphosed at 700 to 600 Ma (Fig. 14d). The very similar ID-TIMS dating results of the leucogranitic boudin samples FHK53 and FHK60 from west and east of the Lerner Deformation Zone indicate a similar formation as late magmatic rocks.

**Discussion**

**A new terrane interpretation for the Northwestern Basement Province**

Until now, two terranes were described in Svalbard’s Northwestern Basement Province: the Albert I Land Terrane and the Biscayarhalvøya Terrane (e.g. Dallmann 2015). Our geochronological data require a revision and new interpretation for the basement east of the Raufjorden Fault.

**The newly postulated Germanialhalvøya Terrane**

Due to the early Paleozoic deposition ages found for the protoliths of the metasedimentary rocks on Lernerøyane and Germanialhalvøya, we have decoupled these rock units from the previous correlation with Albert I Land and defined a separate group: the Cambro–Ordovician Lernerøyane Group and its higher-grade metamorphic equivalent, the Liefdefjorden Migmatite Complex (Fig. 3). Furthermore, two Caledonian metamorphic events are recorded in that area: a Taconian event at c. 469 to 415 Ma with crustal anatexis and formation of anatectic grey granite followed by late-magmatic leucogranitic dykes and veins.

The new geochronological data give a more precise time frame for the deformation events on Lernerøyane and Germanialhalvøya described above. The migmatite xenolith (sample FHK15) shows the same deformation features D2–D3 as the unmigmatized domains in the Liefdefjorden Migmatite Complex, which can be...
correlated with deformation events D1–D3 in the Lernerøyane Group. Therefore, the deformations D1 to D3 are suggested to be older than the Middle Ordovician (Fig. 15).

These geochronological data and structural events show that the pre-Scandian geological evolution of Lernerøyane and Germaniahalvøya differs from that of the Albert I Land Terrane (Supplementary material 3-2). Therefore, we suggest that the area between the Raudfjorden Fault in the west, the Breibogen Fault in the east, and the Rabotdalen–Hannabreen Fault Zone in the NE should be considered as a new terrane in Svalbard’s Northwestern Basement Province: the Germaniahalvøya Terrane. The previous Albert I Land Terrane should be limited to the area west of the Raudfjorden Fault (Fig. 2b).

The Germaniahalvøya Terrane represents a former Cambro-Ordovician basin. The sedimentation probably did not start before c. 500 Ma, estimated based on the scarcity of youngest, non-metamorphic zircon ages older than this. It has a minimum depositional age of c. 469 Ma, the age of the Taconian migmatization, but may be earlier as indicated by the penetrative deformation in the unmigmatized domains of the migmatite. The metamorphic rocks of the Germaniahalvøya Terrane are characterized by two different zircon age patterns in the range of the Precambrian ages:

![Diagram](http://jgs.lyellcollection.org/)

**Fig. 15.** Simplified chart of late Cambrian to Carboniferous structural events in the Biscayarhalvøya Terrane (left column) and in the Germaniahalvøya Terrane (right columns). Time scale after Walker *et al.* (2018). For Precambrian events and comparison with the Albert I Land Terrane see also Supplementary material 3-2.
including subduction, island-arc formation and eclogite-facies populations in the Arctic realm (e.g. Gee and Pease 2004; Kuznetsov et al. 2010). Similar zircon ages are known from the Midtkap igneous suite of North Greenland (U–Pb zircon ages of 650 to 570 Ma) that represents fragments of an oceanic island arc (Rosa et al. 2016; Estrada et al. 2018b). In the composite Pearya Terrane on northern Ellesmere Island, Cambro-Ordovician volcanoclastic sediments of Succession 3 show a dominant zircon age population of c. 620 to 540 Ma peaking at 573 Ma, which is interpreted to be sourced almost completely from a Timanian island arc (Estrada et al. 2018a). The Timanian ages in Pearya are younger than much of the Richarddalen Complex but overlap with the Midtkap igneous suite of North Greenland.

The mélangé comprising the Richarddalen Complex is interpreted to have formed during the collision between a terrane that was affected by late Grenville–Sveconorwegian magmatism and a terrane with an early Timanian volcanic arc. The Timanian arc itself is no longer exposed. Perhaps it was transported far from the Biscayarfonna Terrane during continuous shearing along the deformation zone. The origin of the Timanian arc-related orthogneisses within the Richarddalen Complex is also unclear. Their origin from intrusions into deeper parts of the Biscayarfonna Group is not likely. The complete absence of detrital zircon ages between 1800 and 1500 Ma in the melange matrix (FHK43 + 44; Fig. 11e–f), which represents the main grouping of the Biscayarfonna Group, is also consistent with the idea that the latter was not involved in the formation of the melange matrix. However, the mélangé is very heterogeneous and cannot be well characterized by a single sample.

**Revised genetic and structural interpretation for the Biscayarfonna Terrane**

**A mélangé origin for the Richarddalen Complex**

In the Richarddalen Complex of the Biscayarfonna Terrane, we have found Cryogenian–Ediacaran-aged mafic and felsic metagranitic rocks, including a mafic rock formed at c. 670 Ma and probably eclogite-facies metamorphosed at c. 656 Ma (garnet amphibolite FHK45), and a felsic rock formed at c. 678 Ma (sample C6-1), as well as Late Ordovician to early Silurian metasedimentary rocks with a dominant detrital zircon age population of 730 to 600 Ma (sample FHK43 + 44). Our data are in the range of former zircon evaporation \(^{207}\text{Pb}/^{206}\text{Pb}\) ages of garnet–amphibole gneiss (655 ± 10 and 653 ± 9 Ma) and felsic gneiss (667 ± 4 and 647 ± 4 Ma) of the Richarddalen Complex (Gromet and Gee 1998). Their garnet–amphibole gneiss sample was collected close to our sample C6-1 and stems from a ‘25 cm-thick layer intermixed with felsic and calc-silicate gneiss between boudins and lenses of eclogite and retro-eclogite’ and represents a ‘unit of mixed protoliths’. Petrographically, this sample is very similar to our sample FHK43 + 44 and according to the description of Gromet and Gee (1998) it may represent a part of a melange unit. Their felsic gneiss sample is from a ‘zone of foliated, garnet and epidote-bearing felsic and amphibolite gneisses’ in the southwestern part of the Richarddalen Complex. It is petrographically similar to our mylonitic felsic orthogneiss sample C6-1. However, Timanian U–Pb zircon ages (c. 966 to 950 Ma) were also reported from meta-granite and meta-gabbro (Peucat et al. 1989; Ohta et al. 2003; Pettersson et al. 2009a). Such a very heterogeneous composition of the Richarddalen Complex suggests a block-in-matrix fabric, in which the Late Ordovician to early Silurian metasedimentary rocks would represent the matrix while the Cryogenian meta-igneous rocks (including eclogites) and the Timanian meta-granites and -gabbros are exotic components, as would be typical for a mélangé (e.g. Festa et al. 2010, 2019).

Due to the island-arc geochemical character of sample FHK45 and comparisons with geochemical studies of the last few years that have shown a wide distribution of Timanian zircon age populations in the Arctic realm (e.g. Gee and Pease 2004; Kuznetsov et al. 2010; Anfinson et al. 2012a, b), we relate the Cryogenian–Ediacaran zircon ages to Timanian orogenic events, including subduction, island-arc formation and eclogite-facies metamorphism. In the type locality of the Timanides in NE Baltica, Upper Neoproterozoic sandstone in the Pechora Basin (Fig. 1) contains a unimodal population of detrital zircon ages of c. 750 to 600 Ma sourced from the active margin of a continental block that collided with the passive margin of Baltica at around 550 Ma (Kuznetsov et al. 2010). Similar zircon ages are known from the Midtkap igneous suite of North Greenland (U–Pb zircon ages of 650 to 570 Ma) that represents fragments of an oceanic island arc (Rosa et al. 2016; Estrada et al. 2018b). In the composite Pearya Terrane on northern Ellesmere Island, Cambro-Ordovician volcanoclastic sediments of Succession 3 show a dominant zircon age population of c. 620 to 540 Ma peaking at 573 Ma, which is interpreted to be sourced almost completely from a Timanian island arc (Estrada et al. 2018a). The Timanian ages in Pearya are younger than much of the Richarddalen Complex but overlap with the Midtkap igneous suite of North Greenland.

**Dating of metamorphic events in the Biscayarfonna Terrane**

No other metamorphic event between eclogite-facies metamorphism at c. 656 Ma and Scandian metamorphism at c. 423 Ma is recorded by our zircon data. However, \(^{40}\text{Ar}/^{39}\text{Ar}\) plateau ages of 553 to 505 Ma on hornblende from foliated amphibolite and retrogressed eclogite (Dallmeyer et al. 1990), U–Pb titanite ages on Richarddalen garnet–amphibole gneiss (618 to 431 Ma; 7 fractions) and felsic gneiss (458 to 452 Ma; 4 fractions) with a weighted average of 452 ± 20 Ma (Gromet and Gee 1998), and similar \(^{40}\text{Ar}/^{39}\text{Ar}\) plateau ages of 481 to 443 Ma on phengitic muscovite concentrates from Richarddalen meta-granite and amphibolite (Dallmeyer et al. 1990) were previously obtained in this time interval. They could point to additional metamorphic processes, which have not affected the U–Pb isotopic system of zircon. The U–Pb titanite ages with a grouping between 460 and 450 Ma (Gromet and Gee 1998) were obtained on the same two samples used by these authors for zircon dating (667–647 Ma) and were interpreted to indicate an amphibolite-facies metamorphic event during the Ordovician that has affected the Richarddalen Cryogenian rocks. However, without microchemical and microtextural analysis, which are not reported for the dated minerals, it is impossible to distinguish if these ages represent different events or if they are the result of partial resetting of titanite, amphibole and phengitic muscovite chromotometers during a late-stage single event accompanied or followed by a ductile shearing tectonic phase with local fluid circulation during the Scandian phase (e.g. Bosse and Villa 2019 and references therein). The latter possibility could be supported by U–Pb dating results on titanite from c. 1750 Ma granitoids, which was partially reset to c. 410 Ma by Scandinavian ductile deformation related to shearing along the Billefjorden Fault Zone in Ny Friesland (Johansson et al. 1995), which is parallel to the Raudfjorden and...
Breibogen faults (Fig. 2a). Therefore, a similar effect for the titanite U–Pb ages of the Richaldallen rocks can be assumed. This is supported by the youngest titanite ages (c. 431 and 435 Ma; Gromet and Gee 1998) found in the amphibolite–gneiss matrix, which are similar to a titanite age of c. 430 Ma for the Montblanc Formation. The youngest metamorphic overprint of the mélangé matrix of the Richaldallen Complex at c. 423 Ma took place under amphibolite-facies conditions in a deep setting. Such zircon ages are not reported minimum age of deformation (e.g. Labrousse et al. 2008) for the Richaldallen Complex that can be related to nappe stacking. They differ from the Richaldallen Complex only by their primary lithological compositions and the lack of high-pressure relics. This calls into question the interpretation of at least the Biscayarhalvøya and Montblanc formations as two distinct lithotectonic units of an inverse regional-metamorphic nappe stack. The consistent distribution of apparently coeval sinistral and dextral shearing in strongly folded zones and on both limbs of syn- or antiforms as well as top-to-the-north-directed shear planes in their cores can alternatively be interpreted as the result of tight to isoclinal east–west folding of lateral shear zones during continuous east–west contraction in Late Caledonian times. On a regional scale, the Svalbard Caledonian basement is affected by large-scale sinistral shear zones that transferred crustal blocks in late Silurian–Early Devonian times (e.g. Harland et al. 1974, 1992; Manby 1990; Lyberis and Manby 1999; Faehnrich et al. 2020). The observed dextral shear kinematics in Richaldallen, however, seem to represent a more local rather than regional scenario that may relate either to (i) more or less coeval sinistral and dextral shearing and/or reversal within a relatively short time span or to (ii) isoclinal folding of pristine left-lateral shear zones during ongoing east–west contraction in Late Caledonian times. Faehnrich et al. (2020) recently dated dextral motion along the Vinsodden–Kosipasset Shear Zone in SW Svalbard at c. 424 Ma contemporaneous with Scandian deformation in the Scottish Caledonian Orogen and the early collision between Laurentia and Baltica. These authors propose a similar timing for sinistral strike-slip displacement in both NE and SW Svalbard and interpret c. 450 Ma Ar–Ar muscovite ages of the Billefjorden Fault Zone (Michalski et al. 2012) to result from partial resetting by Ordovician metamorphism or cooling rather than strike-slip deformation.

**Timing of terrane amalgamation**

**Correlation of structural events in Biscayarhalvøya and Germaniahalvøya terranes**

The examination of the pre-Devonian basement in the Biscayarhalvøya and Germaniahalvøya terranes shows that their structural evolution differs (Fig. 15). The Taconian contraction (D1–D3) and the migmatization event are characteristic of the Germaniahalvøya Terrane. The possibly oldest deformation D1 in the basement rocks in the Biscayarhalvøya Terrane is represented by flattened rootless and boudinaged isoclinal folded quartz segregates and isolated fold hinges within S2. However, a clear correlation with either the Taconian or Scandian phase is not possible. The Scandian phase of the Caledonian Orogeny is the first phase that evidently affected both terranes, although there are differences in the way it affected each. While the Germaniahalvøya Terrane is characterized by migmatization and the subsequent intrusion of late Caledonian granites and leucogranites, an amphibolite-facies metamorphic overprint was found for the rocks of the Biscayarhalvøya Terrane. These events were followed by the formation of the Biscayarhalvøya Deformation Zone in the Biscayarhalvøya Terrane (D2) and the Lerner Deformation Zone in the Germaniahalvøya Terrane (D4). The intense ductile shearing suggests that they belong to the same stage of late Caledonian shearing, which is supported by similar orientation of lineations (compare Figs 4 and 8). Furthermore, an approximately coeval relationship of the Lerner Deformation Zone and the Biscayarhalvøya Deformation Zone is confirmed by the age data. While the Lerner Deformation Zone post-dates the anatectic event (422–421 Ma) and syn- to post-dates the late-tectonic intrusions (c. 421–415 Ma), the Biscayarhalvøya Deformation Zone is coeval with the amphibolite-facies metamorphic event (423 ± 5 Ma). Therefore, both belong to a late stage of the Scandian phase.
The differences in structural development suggest that the Germanialahvalvya and Biscayarhalvoya terranes were separated during the Taconian phase of the Caledonian Orogeny in the Lower and Middle Ordovician. During the Scandian phase in the late Silurian, both terranes were affected in different ways by crustal shortening with formation of a mélangé and dextral shearing in the Biscayarhalvoya Terrane and thrusting and heating up to anatexis in the Germanialahvalvya Terrane. Although the kinematic relationship along the Lerner Deformation Zone is still unclear, it can be concluded that late Caledonian ductile shearing along the Biscayarhalvoya Deformation Zone in the Biscayarhalvoya Terrane and the Lerner Deformation Zone in the Germanialahvalvya Terrane probably juxtaposed the Biscayarhalvoya and Germanialahvalvya terranes.

Indications from the basal Old Red sediments

The new zircon data in this study indicate that the formation of the Lerner Deformation Zone and the start of deposition of the Old Red sedimentary rocks overlap in time. This is supported by Dallmann and Piepjohn (2020) who describe an angular unconformity between subvertical late Silurian Siktefjellet and the horizontal Early Devonian Red Bay groups of the Old Red sedimentary rocks north of Liefdefjorden. The deformation of the Siktefjellet Group is ascribed to the late Silurian–Early Devonian tectonic Haakonian phase, which corresponds in time with the late Scandian phase (Fig. 15). Assuming that the Haakonian phase might be the latest stage of the Scandian phase, this suggests that the tectonic activity of the Scandian phase continued during early deposition of the late Silurian–Devonian Old Red sediments.

Detrital zircon studies on basal Old Red sediments deposited before (Siktefjella Group) and shortly after the Haakonian phase (lower Red Bay Group), which overlie the Biscayarhalvoya and Germanialahvalvya terranes east of Raudfjorden, show similar age spectra (Pettersson et al. 2010; Beranek et al. 2020). Apart from abundant ages between c. 2000–1000 Ma and minor ages between 3000 and 2500 Ma, these samples contain varying amounts of Tonian ages (1000–900 Ma), Cryogenian–Ediacaran (Tianman) ages (c. 680 to 570 Ma) and Cambrian–Ordovician ages (c. 530–450 Ma with peaks at 498, 490–480 and 475 Ma in individual samples) (Pettersson et al. 2010; Beranek et al. 2020). These data represent a mixture of the age signatures recorded by our samples from both the Biscayarhalvoya and the Germanialahvalvya terranes (Figs 11 and 13). The common presence of Tianman and Tonian zircon ages in the basal Old Red sediments demonstrates that the amalgamation of the Biscayarhalvoya and the Germanialahvalvya terranes must have taken place before the deposition of these sediments in the latest Silurian.

A sample from the eastern margin of the Albert I Land Terrane (Konglomeratodden), collected from sandstone of the lowermost Red Bay Group, yielded mainly ages between 1800 and 990 Ma, some older ages, and a small Silurian population with a Concordia age of 426 ± 12 Ma (Pettersson et al. 2010). This age spectrum indicates that this sandstone was most likely exclusively sourced from the Albert I Land Terrane (Pettersson et al. 2009a, b) and, additionally, it implies the absence of Cambrian–Ordovician sediments in Albert I Land. Furthermore, it indicates that the Albert I Land Terrane was still divided from the other terranes during the deposition of the lowermost Red Bay Group in the earliest Lochkovian.

It can be concluded that the basal Old Red sediments were largely, if not completely, sourced from the uplifted, local basement rocks (e.g. Gee and Moody-Stuart 1966; Gjelsvik and Ilyes 1991; Friend et al. 1997). The Cambrian to Ordovician zircon grains, which are only scarcely present in our dataset but are more abundant in the basal Old Red sediments east of Raudfjorden, were probably more frequent in the upper, meanwhile eroded strata of the Germanialahvalvya Terrane and the Richarddalen Complex.

The change in age signatures in samples from stratigraphically younger Old Red sediments (Beranek et al. 2020) records a change in source areas, which might be explained by the approach of Albert I Land Terrane to the Biscayarhalvoya and Germanialahvalvya terranes in Early Devonian times.

Conclusions

As a result of our study, we propose to subdivide the Northwestern Basement Province into three terranes (Fig. 2):

- Albert I Land Terrane (now limited to west of Raudfjorden Fault);
- Biscayarhalvoya Terrane between Rabotdalen–Hannabreen Fault Zone in the SW and the Breibogen Fault in the east;
- the newly defined Germanialahvalvya Terrane (informal name) between the Raudfjorden Fault in the west, the Rabotdalen–Hannabreen Fault Zone in the NE and the Breibogen Fault in the east.

The Biscayarhalvoya Terrane is composed of Stenian–Tianman metasedimentary rocks of the Biscayarfonna Group, comparable with the Krossfjorden Group of the Albert I Land Terrane, and the eclogite-bearing Richarddalen Complex. The latter is now interpreted as a Late Ordovician to Silurian tectonic mélangé composed predominantly of Tianman clastic material (730 to 600 Ma), Tianman igneous rock fragments, which formed at c. 690–670 Ma in an island arc setting and were partly metamorphosed to eclogite facies at c. 656 Ma, as well as Tianman igneous rocks known from previous studies. The mélangé matrix was metamorphosed under amphibolite-facies conditions at 423 ± 5 Ma, which we interpret as the maximum age of collision and dextral shearing during the Scandian Orogeny. Final juxtaposition and mylonitization of the Biscayarfonna Group and the Richarddalen Complex occurred after the c. 423 Ma metamorphic event. The process of Caledonian arc–terrane collision is not yet fully understood and needs further work (e.g. the origin of the Tianman orthogneisses involved in the Richarddalen mélangé).

The Germanialahvalvya Terrane comprises the Cambro–Ordovician amphibolite-facies metamorphic Lernerenyane Group and its higher-grade equivalent, the Liefdefjorden Migmatisit Complex. The terrane has experienced multiple deformation and two metamorphic events during the Caledonian Orogeny: (i) Taconian orogenetic events D1–D3 followed by migmatization at c. 469 Ma (recorded by a xenolith in anatexic grey granite), and (ii) Scandian migmatization and anatexis at c. 422–421 Ma and intrusion of late-tectonic leucogranitic dykes and veins between 421 to 415 Ma. During the Scandian phase, the Lernerenyane Group and the Liefdefjorden Migmatisit Complex were juxtaposed along the ductile Lerner Deformation Zone. The age of this event is best estimated from U–Pb ages of 421 and 415 Ma on two monazite grains from a syntectonic leucogranitic vein cutting through the Lerner Deformation Zone.

Detrital zircon age populations in the overlying basal Old Red sedimentary rocks of local provenance (Pettersson et al. 2010; Beranek et al. 2020) give indications of the time of amalgamation of these terranes. The Biscayarhalvoya and Germanialahvalvya terranes were probably amalgamated in the latest Silurian, whereas the approach of the Albert I Land Terrane took place during the Early Devonian. The Tianman igneous rocks and detritus within the Richarddalen mélangé, the Tianman Midtikap Igneous Suite of North Greenland, and the late Tianman detritus in Peary Succession 3 were probably related to the same Tianman island-arc system that was split and displaced by large-scale sinistral slip-slip movements during the Caledonian Orogeny. However, a
detailed regional comparative discussion is outside the scope of this study.

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