Multistage Cretaceous magmatism in the northern coastal region of Ellesmere Island and its relation to the formation of Alpha Ridge – evidence from aeromagnetic, geochemical and geochronological data

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The poorly studied Alpha–Mendeleev Ridge of the Amerasia Basin of the Arctic Ocean is covered by a large magnetic high and interpreted to be part of the Cretaceous High Arctic Large Igneous Province (HALIP). At the north coast of Ellesmere Island, in the Yelverton Bay area, this characteristic magnetic high overlaps the shelf and the coastal area. Results of aeromagnetic surveys, geological fieldwork, geochronological, geochemical and Sr–Nd isotope analysis were combined to study the tectonic history of this part of the Canadian Polar margin. The basement, consisting of metamorphic rocks of the Pearya Terrane, was strongly affected by HALIP-related magmatism that is responsible for characteristic positive magnetic anomaly patterns. A broad magnetic high that can be traced to the Alpha Ridge covers the shelf and coastal areas of Yelverton Bay. It is caused by a voluminous intrusion into the continental crust at a depth of ca. 6–15 km. A chain of small positive magnetic anomalies crosses Wootton Peninsula from SW to NE and then curves into a NNE–SSW trend. It follows the coastline onshore and offshore and superimposes the southern and eastern margins of the broad magnetic high. On land, the anomaly chain covers fault-bounded outcrops of the Wootton Intrusive Complex (WIC) and the Hansen Point Volcanic Complex (HPVC), indicating that the entire anomaly chain is associated with a volcanogenic sequence. Rhyodacite that forms large parts of the HPVC east and west of Yelverton Bay extruded between c. 104 and 97 Ma (LA–ICP–MS U–Pb zircon ages). It originates from a tholeiitic suite and is probably genetically related to the deep intrusion. The HPVC together with the rhyodacite has an age range of c. 104–74 Ma and includes the mildly alkaline WIC (93–92 Ma). The northern coast of Ellesmere Island shows some characteristic features of a magma-rich rifted margin indicating an extensional evolution parallel to the North American margin between the openings of the Canada Basin and the Eurasian Basin.

Keywords: HALIP, Canada, Arctic, magma-rich rifted margin, aeromagnetic survey, magnetic susceptibility, U–Pb zircon age, Sr–Nd isotope data

Electronic Supplement 1. U–Pb zircon data of sample SE 27/01.

Electronic Supplement 2. U–Pb zircon data of sample SE 9/01.

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Introduction

The Arctic Ocean includes two main basins, the Cenozoic Eurasia Basin and the Mesozoic Amerasia Basin, that are separated by the Lomonosov Ridge (Fig.

1), which is interpreted to be a fragment of continental crust that split off the Eurasian Shelf during the opening of the Eurasia Basin (e.g., Ostenso & Wold, 1977; Sweeney et al., 1982; Jokat et al., 1992). The geological history of the Eurasia Basin is well recorded by parallel

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seafloor spreading magnetic anomalies, which indicate that opening started at c. 55 Ma (Vogt et al., 1979) or already at c. 56 Ma (Brozena et al., 2003). The Amerasia Basin is crossed by an over 1500 km-long and 250–400 km-wide topographic feature, the Alpha–Mendelev Ridge, which divides it into the Canada Basin and the Makarov Basin. The Mendelev Ridge strikes towards the East Siberian Shelf approximately along 180 degrees of longitude and the Alpha Ridge strikes towards the Canadian Ellesmere Island approximately parallel to the Lomonosov Ridge. The age and tectonic development of the Amerasia Basin are still under debate. Typical seafloor spreading magnetic anomaly patterns are poorly or not developed, probably because the opening of the Amerasia Basin overlapped the interval of the Cretaceous Normal Superchron between c. 124 and 84 Ma (Walker et al., 2013). In the recently compiled Circum-Arctic magnetic anomaly map (Gaina et al.,

2011), the magnetic domain over the Alpha–Mendelev Ridge and neighbouring areas of the Makarov and Canada basins is characterised by an irregular anomaly pattern with dense high-amplitude and long- to mid-wavelength highs and lows in multiple orientations (Saltus et al., 2011). This magnetic pattern is interpreted to represent a Large Igneous Province (LIP), comparable with other LIPs, such as the Greenland–Iceland–Faroe Ridge and the Siberian traps (Saltus et al., 2011). Some linear, parallel, magnetic and gravity anomaly patterns identified between the Alpha Ridge and the Lomonosov Ridge, trending approximately orthogonal to the ridges, probably represent spreading anomalies (Døssing et al., 2013a, b). The interpretation of the Alpha–Mendelev-Ridge basement as a LIP of oceanic origin is supported by seismic investigations (Forsyth et al., 1986; Jackson et al., 1986; Jokat, 2003; Dove et al., 2010; Funck et al., 2011; Jokat et al., 2013; Jokat & Ickrath, 2015). The

Figure 1. Overview map of the Arctic (based on Jakobsson et al., 2012). Onshore and offshore igneous provinces interpreted to be part of the High Arctic Large Igneous Province (HALIP) including the Alpha–Mendelev magnetic high are encircled in yellow (based on Saltus et al., 2011; Tegner et al., 2011; Shipilov, 2015). The outline of the Canadian portion of the HALIP coincides approximately with the extent of the Sverdrup Basin. The outlines of aeromagnetic surveys and the ARTA refraction seismic profile (Funck et al., 2011) are shown. The white dashed line is the outline of Fig. 2. Abbreviations: AHI – Axel Heiberg Island, CBL – Chukchi Borderland, EI – Ellesmere Island, ERI – Ellef Ringnes Island, LR – Lomonosov Ridge, LS – Lincoln Sea, MB – Makarov Basin, MJ – Morris Jesup Rise, R. – Ridge.

interpretation of the Funck et al. (2011) 'ARTA' refraction line (see Fig. 1) characterised crustal velocities to define continental, (underplated) transitional and 'igneous' crustal types. Although the 6.8 km/s to 7.2 km/s velocities for the 'igneous' lower crust are typical of oceanic crustal layer 3, the Moho depths (28 to 38 km) are not (Forsyth et al., 1986; Lebedeva-Ivanovna et al., 2006; Funck et al., 2011). Funck et al. (2011) suggested that the Alpha Ridge crust resulted from the "interaction of a Cretaceous plume and a seafloor spreading centre parallel to the Canadian Polar margin". This and other tectonic models for the formation of the Alpha–Mendeleev Ridge are still discussed controversially (summarised by Dove et al., 2010; Funck et al., 2011; Petrov et al., 2016). Alternative models include (i) interaction between a mantle plume and seafloor spreading parallel to the Alpha–Mendeleev Ridge, (ii) a hotspot track on oceanic crust, (iii) a magma-rich continental margin rifted off the Barents Shelf, and (iv) a submerged block of intruded and underplated continental crust, favoured by Petrov et al. (2016).

To date, only a very few rock samples provide any information about the composition and/or the age of the Alpha–Mendeleev Ridge basement (Van Wagoner & Robinson, 1985; Jokat, 2003; Andronikov et al., 2008; Jokat et al., 2013, 2015; Morozov et al., 2013; Sergeev et al., 2013; Mukasa et al., 2015; Petrov et al., 2016). More information about the age and duration of the magmatism can be provided by the better accessible Cretaceous igneous rocks from coastal areas around the Arctic Ocean, which together with the submarine Alpha–Mendeleev Ridge are interpreted to be part of the High Arctic Large Igneous Province (HALIP; Tarduno et al., 1998; Maher, 2001; Drachev & Saunders, 2006). The HALIP includes continental flood basalts, mafic dykes and sills emplaced at the Barents Shelf region, De Long Islands, North Greenland, and the Canadian Queen Elizabeth Islands between c. 130 and 80 Ma (recently summarised by Buchan & Ernst, 2006; Nejberr et al., 2011; Tegner et al., 2011; Senger et al., 2014; Polteau et al., 2016; Thórarinnsson et al., 2015). Late Cretaceous to Palaeogene, alkaline, mafic to felsic suites are known in North Greenland and in the Canadian Arctic (e.g., Dawes & Soper, 1970; Batten et al., 1981; Brown et al., 1987; Trettin & Parrish, 1987; Embry & Osadetz, 1988; Estrada et al., 2010). Their genetic relationship to the older, dominantly tholeiitic suites is still under debate (Tegner et al., 2011; Estrada, 2015).

Northern Ellesmere Island in the Canadian Arctic was affected by intense intrusive and extrusive Cretaceous magmatism (e.g., Trettin & Parrish, 1987; Embry & Osadetz, 1988; Estrada et al., 2006; Estrada & Henjes-Kunst, 2004, 2013). It is the only place in the Arctic where both tholeiitic and alkaline igneous suites are in close proximity. At the north coast of Ellesmere Island, in the Yelverton Bay area (Fig. 2), the characteristic magnetic domain over the Alpha–Mendeleev Ridge overlaps the shelf and the coastal area. Geological fieldwork and an

aeromagnetic survey were carried out during CASE expeditions (Circum-Arctic Structural Events) of the German Federal Institute for Geosciences and Natural Resources (BGR) in cooperation with the Geological Survey of Canada (GSC) to understand the structural architecture of the North American passive continental margin. Geological fieldwork in the Yelverton Bay area including sampling for petrographical, geochemical and geochronological analyses was performed in 2000, 2001 and 2008 (Estrada et al., 2006; Piepjohn et al., 2013). The joint helicopter-borne magnetic survey in 2008 covered the coastal area of Ellesmere Island to the east of the previous Canadian survey over Axel Heiberg Island and adjacent marine areas in 1991 (Fig. 1; Geological Survey of Canada, 1995; Teskey et al., 1995).

We present:

- results of the 2008 BGR–GSC aeromagnetic survey combined with the eastern part of the 1991 GSC Axel Heiberg survey for the Yelverton Bay area (within the map area between 78–86°W and 82°00'–83°15'N);
- magnetic susceptibility data of typical lithologies of the Yelverton Bay area;
- results of U–Pb zircon age dating on three volcanic samples from west of Yelverton Bay;
- geochemical and Sr–Nd isotope data on the Cretaceous igneous rocks of the Yelverton Bay area.

We compare the magnetic anomalies with the geology on land and show the relationship between Cretaceous igneous rocks and positive magnetic anomalies over the land and marine areas. Implications of the crustal structure beneath the Yelverton Bay area and the depth of the source of the magnetic anomalies are obtained by magnetic modelling by consideration of seismic refraction data from the 2008 ARTA profile, that extends farther west from the continental shelf between Axel Heiberg and Ellesmere islands across the slope onto the southern Alpha Ridge (Funck et al., 2011). Furthermore, our data provide new constraints on the age, duration and geochemical variation of the HALIP-related magmatism in the Canadian Arctic and on the tectonic history of the Amerasia Basin.

Geological setting

Geological overview of the northeast Canadian Arctic

Northern Ellesmere Island and Axel Heiberg Island are characterised by three major structural units (e.g., Trettin, 1991a):

- the Precambrian to Early Palaeozoic Pearya Terrane,
- the Palaeozoic Franklinian Basin, and
- the Late Palaeozoic to Early Palaeogene Sverdrup Basin (Fig. 2).

Figure 2. Geological overview map of northern Ellesmere Island and Axel Heiberg Island with the Cretaceous igneous suites (after Embry & Osadetz, 1988; Trettin, 1991a). The outline of Fig. 3 (blue dashed line) and the transects of magnetic modelling (grey dashed lines; see Figs. 15, 16) are shown. Abbreviations: EFFZ – Emma Fiord Fault Zone, HPVC/EF – Hansen Point Volcanic Complex/Emma Fiord, HPVC/PI – Hansen Point Volcanic Complex/Phillips Inlet South, HPVC/YB – Hansen Point Volcanic Complex/Yelverton Bay, MPFZ – Mitchell Point Fault Zone, PBF – Petersen Bay Fault, PI – Phillips Inlet, WIC – Wootton Intrusive Complex, YB – Yelverton Bay.

The Pearya Terrane at the northern coast of Ellesmere Island is a microcontinental fragment and can be divided into five major units (Trettin, 1998): the Late Mesoproterozoic to Early Neoproterozoic crystalline basement (succession 1), Neoproterozoic to Lower Ordovician metasedimentary and metavolcanic rocks (succession 2), Lower–Middle Ordovician arc-type metavolcanic and metasedimentary rocks including ultramafic-mafic complexes (succession 3), and Middle Ordovician to Upper Silurian sedimentary and volcanic rocks (successions 4 and 5). All successions are separated either by unconformities or by major faults (Trettin, 1998). During the Mid Ordovician (Caledonian) M’Clintock Orogeny, successions 1 to 3 were metamorphosed under lower greenschist to lower amphibolite facies conditions. Additionally, granitic plutons were emplaced during this event (Trettin, 1991b). Pearya was thrust over deep-water sediments of

the Franklinian Basin during its approach and docking against the passive margin of Laurentia after the Late Silurian (McClelland et al., 2012) and before the late Early Carboniferous (Piepjohn et al., 2013).

The Late Neoproterozoic to Devonian Franklinian Basin overlies the crystalline basement rocks of the Greenland–Canadian Shield and consists of up to 8 km-thick clastic and carbonate shelf and deep-water sediments deposited on the passive margin of Laurentia (Trettin, 1991a). The development of the Franklinian Basin was terminated by the Late Devonian to Early Carboniferous Ellesmerian Orogeny which caused intense crustal shortening and kilometre-scale folding of the entire basin succession (e.g., Trettin, 1991a; Piepjohn et al., 2008, 2013).

The folded Franklinian Basin is unconformably overlain by up to 15 km-thick carbonate and clastic sediments of

the 1300 km-long and 400 km-wide, Carboniferous to Palaeogene, Sverdrup Basin that extends over large areas of the Queen Elizabeth Islands parallel to the coast (Fig. 1; Trettin, 1991a). Between the Mid Jurassic and the Early Cretaceous, the northern Canadian margin became part of a rift shoulder, the Sverdrup Rim, that separated the Sverdrup Basin in the south from the Amerasia Basin in the north (e.g., Balkwill, 1978; Embry & Dixon, 1990; Embry & Beauchamp, 2008).

An extensive phase of igneous activity in the Sverdrup Basin took place during the Cretaceous (see next section). The last pulse of clastic sedimentation in the Sverdrup Basin is represented by the locally more than 3 km-thick, latest Cretaceous to Palaeogene Eureka Sound Group (Miall, 1991; Trettin, 1991a).

The Palaeogene Eureka deformation affected parts of both the Franklinian Basin and the Sverdrup Basin and led to the formation of a deformation belt that extends from the Canadian Arctic in the west over North Greenland towards Svalbard in the east (e.g., Thorsteinsson & Tozer, 1970; Soper et al., 1982). It was caused by the plate tectonic reorganisation during the simultaneous seafloor spreading in the Northeast Atlantic and Labrador Sea–Baffin Bay between c. 56 and 34 Ma, the related northward motion of the Greenland plate, and the opening of the Eurasia Basin starting c. 56 Ma ago (e.g., Menzies, 1982; Srivastava & Tapscott, 1986; Tessensohn & Piepjohn, 2000; Piepjohn et al., 2013).

The HALIP-related Cretaceous magmatism of the Canadian Arctic

At the end of the development of the Sverdrup Basin and prior to the Eureka deformation, the Canadian Arctic was affected by igneous activity related to the HALIP (Figs. 1 & 2). The oldest basalt flows and volcanoclastic deposits occur within the lower Isachsen Formation at Axel Heiberg Island (Embry & Osadetz, 1988) and Ellef Ringnes Island at c. 130 Ma (Evenchick et al., 2015). The main phase of basalt volcanism took place in the Late Barremian to Early Aptian, upper Isachsen Formation (ca. 127–115 Ma) mainly on Axel Heiberg Island (Embry & Osadetz, 1988). The overlying Late Aptian to latest Albian Christopher Formation contains pyroclastic deposits with 105 Ma old detrital zircons and c. 103 Ma old basalt pillows and hydroclastic breccias on Ellef Ringnes Island (Evenchick et al., 2015), as well as volcanoclastic sandstones (Embry & Osadetz, 1988), and bentonite layers (Schröder-Adams et al., 2014) on Axel Heiberg Island.

The volcanogenic Late Albian to Cenomanian Strand Fiord Formation of Axel Heiberg Island consists mainly of subaerial basalt flows that reach a thickness of almost 800 m (Ricketts et al., 1985; Embry & Osadetz, 1988; Embry, 1991). Ar–Ar ages of 100 ± 2 Ma (whole-rock

and sanidine data, MacRae et al., 1996), of 102.5 ± 2.6 Ma (whole-rock age, Estrada & Henjes-Kunst, 2013) and of 95.3 ± 0.4 Ma (whole-rock age, Tarduno et al., 1998) were reported for the Strand Fiord Formation. Basalts within the stratigraphically equivalent Hassel Formation are exposed on northern Amund Ringnes Island (Balkwill, 1983), Ellef Ringnes Island (Evenchick et al., 2015), and northeast Ellesmere Island (Embry & Osadetz, 1988; Osadetz & Moore 1988). The mildly alkaline Hassel Formation basalts from NE Ellesmere Island have an Ar–Ar whole-rock age of 96.4 ± 1.6 Ma (Estrada, 2015).

Mafic tholeiitic dykes and sills are emplaced approximately through the whole extent of the Sverdrup Basin (e.g., Balkwill, 1978; Embry, 1991; Buchan & Ernst, 2006). NE–SW-trending dykes dominate in the western part of the Sverdrup Basin, whereas dykes with N–S to NW–SE trends are present on northern Axel Heiberg Island and on northwest Ellesmere Island (Buchan & Ernst, 2013). These dyke trends are interpreted to be directed to a possible mantle plume centre beneath the Canadian terminus of the Alpha Ridge (Embry & Osadetz, 1988; Buchan & Ernst, 2006). Sills occur in Carboniferous to lower Upper Cretaceous strata. Ar–Ar age data for sills and dykes on Axel Heiberg and northern Ellesmere islands correspond roughly to the time spans of the main volcanic activities in the Barremian/Aptian and Albian/Cenomanian (Villeneuve & Williamson, 2006, and references therein; Estrada & Henjes-Kunst, 2013). Mafic intrusions at Ellef Ringnes Island with U–Pb ages of 126.6 ± 1.2 Ma and 120.8 ± 0.8 Ma, respectively, correlate with the older interval (Evenchick et al., 2015).

Late Cretaceous to Palaeocene alkaline, mafic to felsic igneous suites are known in the Canadian Arctic and North Greenland. Isolated outcrops of the 93–92 Ma old Wootton Intrusive Complex (WIC) and of the c. 80 Ma old Hansen Point Volcanic Complex (HPVC) are exposed in the northwestern coastal area of Ellesmere Island (Fig. 2; Trettin & Parrish, 1987; Embry & Osadetz, 1988; Estrada et al., 2006; Estrada & Henjes-Kunst, 2004, 2013). They form two, parallel, NE–SW-striking chains, which are tectonically controlled by the Mitchell Point Fault Zone in the NE (the study area) and the Emma Fiord Fault Zone in the SW (Estrada & Henjes-Kunst, 2013). The southwestern HPVC outcrops (HPVC/EF and HPVC/PI; see Fig. 2) are formed by a c. 1000 m-thick succession of volcanic and pyroclastic rocks, as well as minor amounts of sandstone, siltstone, shale and coal, and rest unconformably on Carboniferous to Permian strata of the Sverdrup Basin (Trettin & Parrish, 1987; Embry & Osadetz, 1988). Biostratigraphically, a Maastrichtian age is possible, whereas Ar–Ar ages vary between 82 and 65 Ma, and a Rb–Sr isochron age of 80 ± 2 Ma has also been obtained (Estrada & Henjes-Kunst, 2004, 2013, and references therein). The northeastern outcrops around Yelverton Bay are part of the study area and are described below.

Geology and magmatism of the Yelverton Bay area

A large portion of the northern coastal margin of Ellesmere Island consists of rocks of the Pearya Terrane (Figs. 2, 3B, 4). The study area around Yelverton Bay is dominated by the Pearya successions 1 and 2, which are separated by the NE–SW-trending Mitchell Point Fault Zone (Estrada et al., 2006; Piepjohn et al., 2013) and the Pearya Shear Zone (Fig. 3B; von Gosen et al., 2012). Rocks of Pearya successions 3 and 4 crop out only in the northeastern part of the map area south and northeast of Ayles Fiord (Fig. 3B). Intrusive rocks related to the Mid Ordovician M'Clintock orogeny (Trettin et al., 1982) are represented by the Cape Fanshaw Martin pluton north of Ayles Fiord with a preliminary LA–ICP–MS U–Pb zircon age of c. 458 Ma (unpublished BGR data) and the Cape Richards Intrusive Complex farther east dated at 463 ± 5 Ma (U–Pb titanite age, Trettin et al., 1987). The Pearya Terrane is bounded by faults, among others the Peterson Bay Fault and the M'Clintock Glacier Fault (Trettin, 1987; Trettin & Frisch, 1987; Piepjohn et al., 2013), against Early Palaeozoic metasedimentary rocks of the Franklinian Basin in the south (Fig. 3B). A post-tectonic intrusion related to the Ellesmerian orogeny, the Cape Woods pluton with a U–Pb titanite age of 390 ± 10 Ma (e.g., Trettin et al., 1987; Trettin, 1991a), is present on western Wootton Peninsula.

Remnants of the Sverdrup Basin, Carboniferous to Permian carbonates, evaporites and clastic sediments, are limited to the southeastern and eastern parts of the map area (Fig. 3B). On northern Wootton Peninsula, west of the Yelverton Bay, an outlier of folded and thrust-faulted, unconsolidated sandstones, assigned to the Eureka Sound Group, unconformably overlies the Pearya succession 2 (Trettin & Frisch, 1996; Piepjohn et al., 2013). The Eureka deformation led to the reactivation

of the older, mainly SW–NE-striking fault zones as brittle strike-slip faults (Piepjohn et al., 2013).

Several outcrops of Cretaceous igneous rocks are present in the study area (Table 1; Figs. 3B, 4). They are represented by

- (i) basaltic dykes cutting rocks of Pearya successions 1 and 2;
- (ii) a NE–SW-trending chain of plutonic rocks, the WIC, crossing the Wootton Peninsula between the master faults of the Mitchell Point Fault Zone; and
- (iii) the HPVC west and east of Yelverton Bay, which are fault-bounded against metamorphic rocks of Pearya successions 1 and 2.

The outcrops are described in more detail by Estrada et al. (2006). Samples of basaltic dykes cutting Pearya rocks east and west of Yelverton Bay provided Ar–Ar whole-rock ages between 122 ± 2 and 94 ± 3 Ma (Estrada & Henjes-Kunst, 2013). The several lithologies of the WIC (gabbro, syenite, granites, hypabyssal microgranite) have uniform U–Pb zircon (and titanite) ages between 92.7 ± 0.3 and 92.2 ± 0.2 Ma (Trettin & Parrish, 1987; Estrada & Henjes-Kunst, 2013). Large parts of the HPVC outcrops west and east of Yelverton Bay are formed by rhyodacite and related volcanoclastic rocks. Former U–Pb zircon dating yielded an imprecise lower intercept age of $88 +20/-21$ Ma (Trettin & Parrish, 1987). In this study, we report new precise U–Pb zircon ages of these rocks (see below). East of Yelverton Bay, the rhyodacites are truncated by intersecting basaltic dykes (Fig. 5) with Ar–Ar whole-rock ages of 94 ± 1 and 83 ± 2 Ma (Estrada & Henjes-Kunst, 2013). West of Yelverton Bay, the rhyodacitic rocks are overlain by an alkali-olivine basalt flow with an Ar–Ar whole-rock age of 82 ± 2 Ma and a trachyandesite with an Ar–Ar feldspar age of 74 ± 2 Ma. Farther south, separated from the outcrop described before by an E–W-trending glacier, another outcrop of

Table 1. Overview of Cretaceous igneous rocks of the Yelverton Bay area, northern Ellesmere Island.

Unit	Lithologies	Magma series	Age (Ma)	Method	Reference
Dykes cutting Pearya rocks	dolerite	tholeiitic	122.0 ± 2.4 to 94.3 ± 2.8	Ar–Ar (whole rock)	Estrada & Henjes-Kunst, 2013
Hansen Point Volcanic Complex (east and west of Yelverton Bay)	rhyodacite	tholeiitic	103.6 ± 1.0 99.5 ± 1.1 96.9 ± 0.9	U–Pb zircon	this study
Hansen Point Volcanic Complex (east of Yelver-ton Bay)	dolerite (dykes cutting rhyodacite), minor basalt	tholeiitic / transitional	93.9 ± 1.3 83.0 ± 1.8	Ar–Ar (whole rock)	Estrada & Henjes-Kunst, 2013
Wootton Intrusive Complex	ultramafics, gabbro, dolerite, monzonite, syenite, granite, microgranite	mildly alkaline	92.7 ± 0.3 to 92.2 ± 0.2	U–Pb zircon, titanite	Trettin & Parrish, 1987; Estrada & Henjes-Kunst, 2013
Hansen Point Volcanic Complex (west of Yelverton Bay)	basalt, trachyandesite, rhyolite	mildly alkaline	82.3 ± 2.0 73.5 ± 2.4	Ar–Ar (whole rock, feldspar)	Estrada & Henjes-Kunst, 2013

Figure 3. Map of the magnetic anomalies (A) and geological map (B) of the Yelverton Bay area at the same scale. The map of the magnetic anomalies (A) is compiled from the 1991 GSC and the 2008 BGR-GSC aeromagnetic surveys (see text for further explanations). The light-grey coastline is from the geological map (B). Magnetic anomalies are labelled as H (magnetic high), L (magnetic low) and C (chain of positive magnetic anomalies). The geological map (B) is based on Trettin & Frisch (1996), Trettin & Mayr (1996), Estrada et al. (2006), von Gosen et al. (2012) and Piepjohn et al. (2013). Abbreviations: CFM – Cape Fanshawe Martin, HP – Hansen Point.

volcanic rocks is formed by thick rhyolite and ignimbrite units (Fig. 4). The rhyolites provided very young Ar–Ar whole-rock ages of c. 49 and 46 Ma, which are interpreted to be reset due to a hydrothermal overprint during a reactivation of the Mitchell Point Fault Zone in the Eocene (Estrada & Henjes-Kunst, 2013).

Methods

Aeromagnetic survey

The joint German (BGR)–Canadian (GSC) helicopter-borne onshore/offshore magnetic survey in 2008 covered the coastal area of Ellesmere Island to the east of the previous Canadian survey over Axel Heiberg Island and adjacent marine areas in 1991 (Geological Survey of Canada, 1995; Teskey et al., 1995). The map of the anomalies of the magnetic field for the Yelverton Bay area (Fig. 3A) was compiled using the data of the 1991 and 2008 aeromagnetic surveys.

The BGR–GSC aeromagnetic survey was designed in a way to map magnetic sources close to the surface as well as to cover a sufficiently large area. Considering the physiography of the survey area with water depths along the margin generally less than 250 m, a profile spacing of 2 km and a flight elevation of 450 m was chosen. Over the coastal mountains survey lines were flown at terrain clearances of 300 m to 500 m, which allowed the survey lines to be continued from the marine area across the coastal cliffs without extreme changes in flight elevations. Survey lines were flown perpendicular to the coast (roughly N–S). E–W tie lines were flown every 10 km (except over the offshore part due to logistic constraints). The surveying platform was a single-engine helicopter (type Bell 106 – Jet Ranger). A Scintrex CS3 (Cesium) magnetometer was slung on a 25 m cable below the helicopter as a ‘bird’ installation (see Reeves, 2005). Flight-line navigation was done using the helicopter’s Global Positioning Satellite (GPS) system. High-amplitude diurnal variations in the magnetic field produced by high-energy solar radiation in the Arctic Auroral Zone made it necessary to monitor the geomagnetic activity in order to avoid surveying during times of high magnetic activity and apply corrections in later data processing. The diurnal variations were recorded with a Geometrics G856 proton precession base-station at 1 minute intervals at the Taconite Inlet camp. Over the entire survey period no magnetic storms (with high-frequency oscillations in the diurnals) occurred. During the field program, the maximum daily variation at the base camp in Taconite Inlet ranged from 60 nT to 220 nT. There were only two times that the magnetic field varied for more than 50 nT during a 2 ½ hour survey flight.

The parameters of the 1991 Axel Heiberg survey are briefly described in Teskey et al. (1995): Flight line spacing was 4 km on average at a flight altitude of 305 m mean terrain clearance. Control lines were flown at an average spacing of 40 km. Navigation and positioning of the Convair 580 survey aircraft was primarily by GPS backed up by Inertial Guidance. Regarding the geomagnetic activity, Teskey et al. (1995) simply noted that the survey was “*not restricted by diurnal activity*”.

Data processing of the BGR–GSC survey included a correction using magnetic base-station data which were smoothed with a 30-minute low-pass filter to exclude short-period variations, thus avoiding introduction of ‘artificial’ anomalies in the survey data. The internal (Earth main-field) component was removed by calculating the IGRF (International Geomagnetic Reference Field) for 01 June 2008 (as the central date for the survey) using the IGRF model 2005 (IAGA, 2005) and subtracting it from the magnetic field values measured for each survey location.

For both surveys, a levelling procedure was applied to adjust the differences of the magnetic values at the intersection of profile and control lines. While for the 1991 survey the adjustment was done manually (Teskey et al., 1995), the BGR–GSC survey used the Geosoft OasisMontaj levelling package including microlevelling (see e.g., Ferraccioli et al., 1998).

Magnetic susceptibility

To aid interpretation of the aeromagnetic results, magnetic susceptibility readings were performed on rock samples collected in the Yelverton Bay area (see below). Magnetic susceptibility was measured using a KT-6 kappameter of SatisGeo. The sensitivity is 1×10^{-5} SI units. The mean value was taken mostly from 7 single readings for each sample, omitting the lowest and highest value. Results are presented in Table 2.

U–Pb zircon dating

Zircons from three rock samples were separated and analysed by Laser Ablation combined with Inductively-Coupled-Plasma Mass-Spectrometry (LA–ICP–MS) techniques at the Senckenberg Naturhistorische Sammlungen Dresden, Germany. The zircon concentrate was separated from approx. 1–2 kg sample material. Samples were crushed in a jaw crusher (Retsch) and sieved by a sieving machine (Retsch). The fraction of 400 to 40 µm was used for heavy-mineral separation by making use of a heavy liquid (LST, lithium heteropolytungstate), which had a density of 2.85 g/cm³. Sample powder and heavy liquid was mixed in a separatory funnel. After ~1 hour heavy minerals were separated, washed with distilled water, and dried in a

Figure 4. Enlarged section of the geological map from Fig. 3B with the sites of dated samples (redrawn from Estrada et al., 2006; Piepjohn et al., 2013).

drying cabinet at a temperature of 50°C. Final selection of the zircon grains for U–Pb dating was achieved by hand-picking under a binocular microscope. Zircon grains of all grain sizes and morphological types were selected, mounted in resin blocks and polished to half their thickness. Zircons were analysed for U, Th and Pb isotopes by LA–ICP–MS techniques, using a Thermo-Scientific Element 2 XR sector field ICP–MS coupled

to a New Wave UP–193 Excimer Laser System. A teardrop-shaped laser cell was used to enable sequential sampling of heterogeneous grains (e.g., growth zones) during time-resolved data acquisition. Each analysis consisted of approximately 15 s background acquisition followed by 30 s data acquisition, using a laser spot size of 25 and 35 µm, respectively. A common-Pb correction based on the interference-corrected and background-

Table 2. Magnetic susceptibility of typical lithologies in the Yelverton Bay area, northern Ellesmere Island.

Sample	Site	GPS coordinates		Lithology	n	Susceptibility x 10 ⁻³ SI units (mean)
		N	W			
Pearya Terrane						
Pearya succession 1						
SE 14/08	W of Ayles Fiord	82°47'20.8"	80°40'48.0"	orthogneiss, mylonitic	7	0.1
SE 13/08	W of Ayles Fiord	82°47'20.8"	80°40'48.0"	orthogneiss	7	0.1
SE 11/08	W of Ayles Fiord	82°47'20.8"	80°40'48.0"	orthogneiss	5	0.1
SE 09/08	W of Milne Fiord	82°38'15.2"	81°41'45.6"	orthogneiss	5	0.0
SE 48/08	SW of Petersen Bay	82°27'27.6"	81°50'15.0"	orthogneiss	5	0.1
SE 53/08	SW of Petersen Bay	82°25'45.6"	81°48'06.0"	orthogneiss	5	6.1
SE 84/08	island in Petersen Bay	82°30'11.4"	81°51'36.6"	orthogneiss	5	0.1
SE 121/08	S of Milne Glacier	82°21'15.7"	80°54'03.6"	orthogneiss	7	0.1
SE 12/08	W of Ayles Fiord	82°47'20.8"	80°40'48.0"	amphibolite	5	0.5
SE 07/08	W of Milne Fiord	82°38'15.2"	81°41'45.6"	amphibolite	7	0.7
SE 33/08A	SW of Petersen Bay	82°27'04.4"	81°50'55.3"	amphibolite	7	0.6
SE 38/08	SW of Petersen Bay	82°27'55.4"	81°53'12.5"	amphibolite	5	0.7
SE 54/08	SW of Petersen Bay	82°25'37.8"	81°48'25.8"	amphibolite	7	0.5
SE 51/08	SW of Petersen Bay	82°27'29.4"	81°49'22.2"	amphibolite	5	1.2
Pearya succession 2						
SE 134/01 A	Wootton Peninsula	82°13'22.2"	84°00'17.9"	marble	7	0.0
SE 151/01	Wootton Peninsula	82°16'40.7"	84°20'30.6"	carbonat. sericite slate	7	0.4
SE 155/01	Wootton Peninsula	82°14'58.3"	83°53'18.5"	silty slate	7	0.2
SE 111/01 D	Wootton Peninsula	82°19'35.7"	83°35'29.8"	chert	8	0.3
SE 108/01 A	Wootton Peninsula	82°17'06.2"	83°31'49.1"	metabreccia	7	1.0
SE 108/01 D	Wootton Peninsula	82°17'06.2"	83°31'49.1"	metabreccia	7	1.3
SE 110/01 B	Wootton Peninsula	82°19'31.3"	83°35'49.2"	psammitic-psephitic metapelite	7	3.8
SE 180/01	Wootton Peninsula	82°17'06.3"	83°31'48.3"	metabreccia	6	0.8
SE 112/01	Wootton Peninsula	82°19'33.0"	83°35'39.8"	metavolcanic rock, mylonitic	7	114.3
SE 173/01	Wootton Peninsula	82°17'18.8"	83°30'32.3"	metabasalt	7	63.4
Cretaceous igneous rocks						
dykes in Pearya succession 1						
SE 49/08	SW of Petersen Bay	82°27'30.6"	81°49'10.2"	dolerite	7	16.8
SE 42/08	SW of Petersen Bay	82°28'26.7"	82°00'47.8"	dolerite, altered	5	0.5
SE 43/08	SW of Petersen Bay	82°28'26.7"	82°00'47.8"	dolerite	5	8.5
SE 44/08	SW of Petersen Bay	82°28'26.7"	82°00'47.8"	dolerite	5	10.0
SE 45/08	SW of Petersen Bay	82°28'26.7"	82°00'47.8"	dolerite	5	19.2
SE 56/08	SW of Petersen Bay	82°28'30.6"	81°00'15.6"	dolerite	5	30.8
dykes in Pearya succession 2						
SE 148/01	Wootton Peninsula	82°16'39.6"	84°21'15.8"	dolerite	7	20.3
SE 150/01	Wootton Peninsula	82°16'40.7"	84°20'30.6"	dolerite	7	20.2
SE 159/01	Wootton Peninsula	82°14'58.3"	83°53'18.5"	dolerite	7	59.9
Wootton Intrusive Complex (WIC)						
SE 36/01	Wootton Peninsula	82°09'48.3"	84°39'01.5"	olivine hornblendite (ultramafite)	6	29.2

Table 2. Continued

Sample	Site	GPS coordinates		Lithology	n	Susceptibility $\times 10^{-3}$ SI units (mean)
		N	W			
SE 81/01	Wootton Peninsula	82°12'18.4"	84°47'57.4"	olivine hornblendite (ultramafite)	7	29.4
SE 28/01	Wootton Peninsula	82°03'59.8"	85°47'04.8"	hbl dolerite	7	0.7
SE 29/01	Wootton Peninsula	82°03'59.8"	85°47'04.8"	hbl dolerite	5	1.5
SE 33/01	Wootton Peninsula	82°03'59.8"	85°47'04.8"	hbl gabbro	5	44.5
SE 34/01	Wootton Peninsula	82°03'59.8"	85°47'04.8"	hbl gabbro	5	0.9
SE 43/01	Wootton Peninsula	82°09'48.3"	84°39'01.5"	hbl dolerite, porphy.	5	1.8
SE 85/01	Wootton Peninsula	82°13'04.5"	84°23'51.6"	hbl dolerite, porphy.	8	8.9
SE 86/01	Wootton Peninsula	82°13'04.5"	84°23'51.6"	hbl dolerite	5	0.7
SE 80/01	Wootton Peninsula	82°12'18.4"	84°47'57.4"	monzonite	8	29.7
SE 87/01	Wootton Peninsula	82°13'04.5"	84°23'51.6"	monzonite	7	22.0
SE 30/01	Wootton Peninsula	82°03'59.8"	85°47'04.8"	syenite	5	64.1
SE 32/01	Wootton Peninsula	82°03'59.8"	85°47'04.8"	syenite	6	68.5
SE 76/01	Wootton Peninsula	82°12'18.4"	84°47'57.4"	syenite	7	5.8
SE 38/01	Wootton Peninsula	82°09'48.3"	84°39'01.5"	granite	7	13.1
SE 78/01	Wootton Peninsula	82°12'18.4"	84°47'57.4"	granite	8	17.7
SE 91/01	Wootton Peninsula	82°13'04.5"	84°23'51.6"	granite (aplite)	7	5.6
SE 1/00	Wootton Peninsula	82°17'32.3"	83°27'13.8"	microgranite	6	15.3
SE 3/00	Wootton Peninsula	82°17'32.3"	83°27'13.8"	microgranite	7	15.9
SE 5/1/00 A	Wootton Peninsula	82°17'39.8"	83°28'32.5"	microgranite	5	11.1
SE 5/1/00 B	Wootton Peninsula	82°17'39.8"	83°28'32.5"	microgranite	6	10.3
Hansen Point Volcanic Complex (HPVC)						
SE 130/01	Wootton Peninsula	82°13'22.2"	84°00'17.9"	olivine basalt, altered	7	77.8
SE 16/01	Wootton Peninsula	82°16'27.6"	83°32'58.6"	alkali basalt	10	49.3
SE 18/01	Wootton Peninsula	82°16'27.6"	83°32'58.6"	alkali basalt	6	48.7
SE 46/01	Wootton Peninsula	82°16'30.0"	83°29'12.4"	basalt	5	26.5
SE 52/01	Hansen Point South	82°25'52.5"	82°29'00.6"	basalt	5	26.3
SE 53/01	Hansen Point South	82°25'51.1"	82°29'11.7"	dolerite	7	10.4
SE 58/01	Hansen Point	82°27'10.3"	82°26'56.8"	dolerite	5	40.5
SE 66/01	Hansen Point	82°27'10.3"	82°26'56.8"	dolerite	8	10.9
SE 67/01	Hansen Point	82°29'52.9"	82°25'33.8"	dolerite	5	34.0
SE 136/01	W of Phillips Inlet	82°00'57.7"	84°42'45.7"	basalt	5	38.5
SE 10/01	Wootton Peninsula	82°17'56.6"	83°30'58.6"	trachyandesite	5	25.1
SE 163/01	Wootton Peninsula	82°18'53.6"	83°34'26.2"	trachyandesite	7	60.8
SE 164/01	Wootton Peninsula	82°18'53.6"	83°34'26.2"	trachyandesite	5	57.3
SE 6/01	Wootton Peninsula	82°17'51.1"	83°28'43.3"	rhyodacite	5	23.0
SE 9/01	Wootton Peninsula	82°17'55.2"	83°30'00.9"	rhyodacite	5	24.3
SE 70/01	Hansen Point	82°29'52.9"	82°25'33.8"	rhyodacite	5	38.8
SE 113/01	Wootton Peninsula	82°14'54.9"	83°38'25.2"	rhyolite, hydrotherm. alt.	5	0.5
SE 114/01	Wootton Peninsula	82°14'54.9"	83°38'25.2"	rhyolite, hydrotherm. alt.	7	0.4
SE 4/01 B	Wootton Peninsula	82°17'48.4"	83°28'28.6"	volcaniclastic arkose	5	50.9
HPVC/Pearya succ. 2						
SE 169/01	Wootton Peninsula	82°18'58.1"	83°34'48.8"	vent breccia	7	1.3

corrected ^{204}Pb signal, and a model Pb composition (Stacey & Kramers, 1975), was carried out, if necessary. The necessity of the correction is judged on whether the corrected $^{207}\text{Pb}/^{206}\text{Pb}$ lies outside of the internal errors of the measured ratios. Discordant analyses were generally interpreted with caution. Raw data were corrected for background signal, common Pb, laser-induced elemental fractionation, instrumental mass discrimination, and time-dependent elemental fractionation of Pb/Th and Pb/U using an Excel[®] spreadsheet program developed by Axel Gerdes (Institute of Geosciences, Goethe University Frankfurt/M., Germany). Reported uncertainties were propagated by quadratic addition of the external reproducibility obtained from the standard zircon GJ-1 (~0.6% and 0.5–1% for the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$, respectively) during individual analytical sessions, and the within-run precision of each analysis. Concordia diagrams (2 σ error ellipses) and concordia ages (95% confidence level) were produced using Isoplot/Ex 2.49 (Ludwig, 2001), and frequency and relative probability plots using AgeDisplay (Sircombe, 2004). The $^{207}\text{Pb}/^{206}\text{Pb}$ age was taken for interpretation for all zircons >1.0 Ga, and the $^{206}\text{Pb}/^{238}\text{U}$ ages for younger grains. For further details on analytical protocol and data processing, see Gerdes & Zeh (2006) as well as Frei & Gerdes (2009). The complete LA-ICP-MS zircon data are presented as Electronic Supplements 1–3.

Geochemistry

Major and trace element whole-rock chemistry of the Cretaceous igneous rock samples was measured by X-ray fluorescence (XRF) at BGR, Hannover, Germany. The sample powders were mixed with lithium metaborate at a ratio of 1:5, fused at 1200°C for 20 min, and analysed using two different X-ray fluorescence spectrometers: a Philips PW 2400 (Rh tube) for the elements Si, Al, Fe, Mn, Mg, Na, P, S, F, As, Bi, Ce, Co, Cr, Cu, Ga, Hf, La, Mo; and a PW 1480 (Cr tube) for the elements Ti, Ca, K, Cl, Ba, Cs, Sb, Sc, and Sn. The trace element detection limits are as follows: 20 ppm for Ce and La, 10 ppm for Cu, 5 ppm for Ba, Cs, Hf, Ta, Th, and V, 4 ppm for Pb, 3 ppm for Co, Cr, Ga, Ni, U, Y, Zn, and Zr, and 2 ppm for Nb, Rb, Sc and Sr. The calibrations are validated by reference material analysis. 'Monitor' samples and 130 certified reference materials (CRM) are used for the correction procedures. Loss-on-ignition (LOI) was determined by heating the sample powder to 1,030°C for 10 minutes.

Additional analyses of trace elements, including rare earth elements (REE), by inductively coupled plasma mass spectrometry (ICP-MS) were performed on selected samples by Activation Laboratories Ltd. (Actlabs), Ancaster, Canada, using the lithium metaborate/tetraborate fusion ICP-MS method (code 4B2 Research). For the fusion process, the sample powder was mixed with lithium metaborate and lithium tetraborate in graphite crucibles and fused in induction

furnaces at 1,150°C. The fused and diluted samples were analysed using a Perkin Elmer Sciex ELAN Inductively Coupled Mass Spectrometer. Quality assurance is warranted by repeated analysis of samples and certified reference materials. Further analytical details and detection limits are available from the Actlabs website (<http://www.actlabs.com>).

Rb–Sr and Sm–Nd isotope analysis was carried out at BGR, Hannover, Germany. The isotope data were obtained on the unleached whole-rock powders using a Finnigan MAT 261 mass spectrometer for Rb measurements, and a ThermoFinnigan TRITON mass spectrometer for Sr, Sm and Nd measurements. Element concentrations were determined by isotope-dilution techniques. All elements were loaded on Re filaments and run on a double-filament assembly in static mode. Isotopic ratios were normalised to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ for Nd, to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ for Sr and to $^{152}\text{Sm}/^{149}\text{Sm} = 1.93476$ for Sm. Uncertainties at the 95% confidence level are 0.0025% for $^{143}\text{Nd}/^{144}\text{Nd}$, 0.005% for $^{87}\text{Sr}/^{86}\text{Sr}$ and 0.75% for $^{87}\text{Rb}/^{86}\text{Sr}$ and 0.5% for $^{147}\text{Sm}/^{144}\text{Nd}$. The $^{87}\text{Sr}/^{86}\text{Sr}$ values of the samples were corrected to $^{87}\text{Sr}/^{86}\text{Sr} = 0.710245$ for the NIST SRM987 Sr standard. The $^{143}\text{Nd}/^{144}\text{Nd}$ values of the samples were corrected to $^{143}\text{Nd}/^{144}\text{Nd} = 0.511845$ for the LaJolla Nd standard. Procedural blanks are less than 0.1% of the relevant sample concentration and are therefore negligible. Calculation of the Sm–Nd model parameter ϵ_{Nd} and the single-stage depleted-mantle model ages (t_{DM}) is based on DePaolo (1988) with $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ for a CHUR reference ('chondritic uniform reservoir', Jacobsen & Wasserburg, 1980).

Results and interpretation

The magnetic anomalies and relationship to the geology on land

The map of the anomalies of the total magnetic field (Fig. 3A) shows distinct units of different magnetic patterns:

- (i) a strong, long-wavelength, magnetic high (H1) mainly over the shelf and neighbouring coastal areas around Yelverton Bay;
- (ii) a broad magnetic low (L) mainly over the inland area;
- (iii) a chain of small positive anomalies (C1–C3) crossing Wootton Peninsula from SW to NE (C1), then crossing Yelverton Bay from SSW to NNE between Yelverton Inlet and north of Milne Fiord (C2), and turning to a renewed SW–NE trend offshore north of Ayles Fiord (C3);
- (iv) a broader magnetic high (H2) and several smaller circular positive anomalies within the inland area of the magnetic low (L).

The broad magnetic high (H1) over Yelverton Bay likely originates from large and deep-seated magnetic bodies. The southwestern margin of the anomaly follows the coastal area of northeast Wootton Peninsula that is covered by ice, Quaternary and Tertiary sediments, rocks of Pearya succession 2, as well as Cretaceous volcanogenic rocks of the HPVC (Fig. 3B). The southern and southeastern margin of anomaly H1 covers rocks of the non-magnetic (see below) Pearya succession 1 and outcrops of the Cretaceous HPVC. Thus, the source for the magnetic high H1 should be an intrusion within or below Pearya succession 1. Northeast of Milne Fiord, the anomaly covers only coastal areas. The anomaly H1 can be traced offshore to the large Alpha–Mendeleev Ridge magnetic domain (Saltus et al., 2011). The southeastern and northeastern margins of the anomaly H1 are superimposed by the chain of small positive anomalies (C2 and C3).

The chain of small, short-wavelength, positive anomalies is related to sources at shallow depth or at the surface. On land, these anomalies can be identified as outcrops of the Cretaceous WIC and HPVC. The WIC outcrops (and the anomaly chain C1) cross the Wootton Peninsula from SW to NE, following the Mitchell Point Fault Zone (Fig. 3B). The segment C2 of the anomaly chain forms a wide, c. 50 km-long, NNE–SSW-striking lineament that includes on land fault-bounded outcrops of the HPVC southwest and northeast of Yelverton Bay. We interpret the segment C2 and its SW–NE striking offshore elongation C3 as a fault-controlled structure that places HPVC-type volcanogenic material against Pearya rocks. The fault structure runs approximately parallel to a pre-existing structure on land, the Pearya Shear Zone, and coincides with segments of the Mitchell Point Fault Zone.

Compared with the geological map (Fig. 3B), the magnetic low (L) comprises the Mesoproterozoic to Neoproterozoic crystalline basement rocks of Pearya succession 1, the Neoproterozoic to Ordovician metamorphic rocks of Pearya succession 2 and, in the southwest, Lower Palaeozoic metasedimentary rocks of the Franklinian Basin. Small positive anomalies within

the inland region correlate with Palaeozoic intrusive rocks, such as the Cape Fanshaw Martin and the Cape Richards plutons (northeast in the geological map Fig. 3B). In contrast, the large Devonian Cape Wood pluton of western Wootton Peninsula, that is marginally touched by the magnetic anomaly map (Fig. 3A), does not significantly differ from the surrounding magnetic low. A broader magnetic high (H2) in the northeastern onshore area (east of c. 79°W) comprises Ordovician units of Pearya successions 3 and 4, which are dominated by volcanogenic rocks, and includes the Cape Richards Intrusive Complex. A short SW–NE trending chain of small anomalies between Ayles Fiord and Milne Fiord may be linked with a mafic dyke of unknown age within the Pearya Shear Zone (mapped by Trettin & Frisch, 1996, but not visited during our fieldwork). Other small positive anomalies within the area of the magnetic low do not correlate with any of the mapped geological features. They probably represent dykes or smaller plutons of unknown age.

The magnetic susceptibility of the rocks

The main lithologies of Pearya succession 1, orthogneisses and minor amphibolites, are characterised by very low values of magnetic susceptibility (Table 2). Most orthogneiss samples have values between 0 and 0.2×10^{-3} SI units, and only a single sample shows a higher value of 6.1×10^{-3} SI units. The range of the amphibolites is $0.5\text{--}1.2 \times 10^{-3}$ SI units. The low-grade metasedimentary rocks of Pearya succession 2 on Wootton Peninsula (phyllites, quartzitic slates, marble, chert) also show low values between 0 and 0.4×10^{-3} SI units. Metasedimentary breccias, a minor rock type, have higher values up to 3.8×10^{-3} SI units due to the presence of small amounts of volcanogenic lithoclasts. Very high values (up to 114×10^{-3} SI units) are given by metavolcanic rocks within succession 2. However, these rocks form small, decimetre-scale lenses within the metasedimentary rocks (Estrada et al., 2006) and cannot markedly influence the magnetic field.

Figure 5. Cross-cutting dykes truncate rhyodacitic rocks east of Yelverton Bay. (A) Southern margin of the HPVC outcrop near Hansen Point. Helicopter (encircled in red) for scale. (B) View to the snow-covered coastal area at Hansen Point from the helicopter.

The Cretaceous igneous rocks of the WIC and HPVC show a wide range of magnetic susceptibility. The values vary between 0.7 and 68×10^{-3} SI units in the WIC rocks and between 0.4 and 61×10^{-3} SI units in the HPVC rocks (Table 2). The highest values are related to rocks of intermediate to felsic composition (WIC syenite, HPVC rhyodacite, trachyandesite). Hydrothermally altered rhyolites from SW of Yelverton Bay are characterised by low values of 0.5×10^{-3} SI units. Basalts and dolerites of the HPVC show values between 10 and 49×10^{-3} SI units. The magnetic susceptibility of WIC mafic rocks (gabbros and dolerites) is relatively low, mostly $<10 \times 10^{-3}$ SI units. Only one gabbro sample has a higher value of 44.5×10^{-3} SI units. WIC ultramafic rocks have values of c. 30×10^{-3} SI units.

Cretaceous dolerite dykes cutting Pearya succession 2 on Wootton Peninsula have high values between 20 and 60×10^{-3} SI units. South of Petersen Bay (southeast of Hansen Point), the dykes are intruded into the Pearya succession 1 (Fig. 4) following pre-existing structures of the older amphibolite dykes. Relatively fresh Cretaceous dyke rocks show values between 10 and 31×10^{-3} SI units and thus differ clearly from the low values of the amphibolites. Some dyke rocks are strongly hydrothermally altered

and clinopyroxene is completely replaced by fine-grained green amphibole. Those dyke rocks show lower values of magnetic susceptibility. However, all these dykes are relatively small in width (decimetre to metre) and do not influence the magnetic field at the scale of the survey.

In summary, there is a notable difference in the magnetic susceptibilities between the Cretaceous WIC and HPVC igneous rocks and the Pearya metamorphic rocks, which supports the interpretation that Cretaceous igneous rocks are the source of the positive anomaly chain (C1–C3; Fig. 3A).

U–Pb zircon geochronology

Volcanic rocks of rhyodacitic composition are a large component of the HPVC outcrops west and east of Yelverton Bay. These rocks are too altered for dating by the Ar–Ar method. Three samples from west of Yelverton Bay (SE 9/01, SE 27/01, SE 162/01) yielded a sufficient amount of zircon grains for LA–ICP–MS U–Pb analysis (Fig. 4). No zircons were found in a sample (SE 65/01) from the east side, at Hansen Point.

Figure 6. Photomicrographs of the dated samples. (A) Sample SE 27/01, rhyodacite with devitrified matix (plane-polarised light). (B) Sample SE 9/01, porphyritic rhyodacite (crossed polars). (C–D) Sample SE 162/01, microbreccia with fragments of crystals, and volcanic and metasedimentary rocks. (C) Overview photo of the plain rock surface. (D) Photomicrograph, crossed polars.

Figure 7. U–Pb data of zircon grains from sample SE 27/01. (A) Concordia diagram for all ages. (B) Combined frequency and probability density distribution plots for the concordant Cretaceous zircon ages showing two age populations. (C) Concordia diagram for the two Cretaceous age populations.

Sample SE 27/01

Sample 27/01 is an altered reddish-grey, porphyritic rhyodacite. The rock is tectonically deformed and cut by fine fissures. It contains phenocrysts of alkali feldspar and plagioclase. The matrix is devitrified (Fig. 6A).

Forty-five out of 87 analysed zircon grains gave concordant ages (Electronic Supplement 1; Fig. 7A). The concordant Cretaceous ages (40 grains) are in the range of 98 ± 5 Ma (2σ) to 112 ± 3 Ma. Within this interval, two age populations with mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 106.3 ± 0.9 Ma (26 grains) and 99.8 ± 1 Ma (14 grains) as well as concordia ages of 105.0 ± 0.9 Ma and 99.5 ± 1.1 Ma are present (Fig. 7B, C). Concordia ages and mean $^{206}\text{Pb}/^{238}\text{U}$ ages are within error identical. Five grains yielded Neoproterozoic ages between 915 ± 22 Ma and 988 ± 28 Ma.

Sample SE 9/01

Sample SE 9/01 is a dark porphyritic rhyodacite (Fig. 6B). Phenocrysts are strongly saussuritised alkali feldspar and plagioclase up to 2 mm in size within a spherulitic matrix. Accessory minerals are apatite, zircon and opaque Fe oxides.

From 106 analysed zircon grains 101 grains yielded discordant ages (Electronic Supplement 2; Fig. 8). Only 5 grains gave concordant ages between 935 ± 25 Ma and 1015 ± 20 Ma (Neoproterozoic). The 98 Cretaceous $^{206}\text{Pb}/^{238}\text{U}$ ages are in the interval of 88 ± 3 Ma to 111 ± 5 Ma with a mean age of 96.9 ± 0.9 Ma. This mean age corresponds approximately with the youngest age peak from sample SE 27/01.

Figure 8. U–Pb data of zircon grains from sample SE 9/01. (A) Concordia diagram for all ages. (B) All Cretaceous $^{206}\text{Pb}/^{238}\text{U}$ ages with the mean.

Sample SE 162/01

Sample SE 162/01 was taken near the tectonic boundary between the HPVC and Pearya succession 2 rocks at the central part of a mountain crest that strikes parallel to the west coast of Yelverton Bay ($82^{\circ}18'53.6''\text{N}$; $83^{\circ}34'26.2''\text{W}$). The rock is a microbreccia, formed in a volcanic pipe. It contains several variegated angular rock and crystal fragments up to 1 cm in size (Fig. 6C). The rock fragments comprise carbonate and calc-silicate rocks, feldspar and plagioclase porphyritic volcanic rocks, microgranite, metasedimentary rocks, quartz breccia and undulose polycrystalline quartz aggregates. Crystal fragments are plagioclase, alkali feldspar and undulose quartz. Accessories are chlorite, sericite, zircon and apatite, as well as opaque Fe–Ti oxides, sulphides, and leucoxene.

118 zircon grains were analysed, and 63 grains of these gave concordant ages (Electronic Supplement 3; Fig. 9). Concordant Cretaceous $^{206}\text{Pb}/^{238}\text{U}$ ages (16 grains) are between 101 ± 3 Ma and 120 ± 4 Ma. The 12 youngest

Figure 9. U–Pb data of zircon grains from sample SE 162/01. (A) Concordia diagram for all ages. (B) Concordia diagram for the concordant Cretaceous ages.

grains give a concordia age of 103.6 ± 1.0 Ma. This age can be interpreted as the time for the onset of explosive silicic volcanism in the Yelverton Bay area. It corresponds within the error with the older age peak from sample SE 27/01. The other 47 concordant ages are in the range from 897 ± 21 Ma to 1558 ± 41 Ma with a peak at 993 ± 5 Ma. These Neoproterozoic and Mesoproterozoic ages are related to the Pearya country rocks.

The Cretaceous zircons of all samples are characterised by unusual low U concentrations and relatively high Th/U ratios (sample SE 27/01: 2–77 ppm U, Th/U = 0.58–1.24; sample SE 9/01: 9–116 ppm U, Th/U = 0.60–1.33; sample SE 162/01: 4–137 ppm U, Th/U = 0.65–1.78) indicating a common origin from an igneous source. The Cretaceous zircons were formed in magma chambers before and during the onset of the extrusion of the rhyodacitic magma at c. 104 Ma. An upper crustal level for the magma chambers is supported by the presence of Neoproterozoic and Mesoproterozoic zircons inherited from Pearya country rocks. Thus, the Hansen Point

Figure 10. Classification of the Cretaceous igneous rocks of the Yelverton Bay area in the $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 diagram, modified from Cox et al. (1979) with the dividing line between alkaline and subalkaline magma series from Miyashiro (1978). (A) Volcanic lithologies of the Hansen Point Volcanic Complex (HPVC) and subvolcanic mafic dykes of different ages. (B) Intrusive lithologies of the Wootton Intrusive Complex (WIC) with an age of 93–92 Ma.

volcanism was active over a much longer time interval than indicated by previous age determinations (80 ± 2 Ma, Rb–Sr isochron age, Estrada & Henjes-Kunst, 2004; 83 ± 2 Ma to 77 ± 4 Ma, Ar–Ar ages, Estrada & Henjes-Kunst, 2013).

Geochemical characterisation of the Cretaceous igneous rocks

Major and trace element compositions of the Cretaceous igneous rocks of the Yelverton Bay area are presented in the Electronic Supplement 4. For graphic presentation, the dataset was recalculated to a loss-on-ignition (LOI) free basis. In the total alkalis vs. silica (TAS) diagram (Fig. 10A), the igneous lithologies of the HPVC can be classified as basalts, rhyodacites, rhyolites and minor trachyandesites, mugearites and hawaiites. They represent a transitional to mildly alkaline magma series, with exception of the c. 100 Ma old rhyodacites and the dolerites from east of Yelverton Bay that tend to the subalkaline (tholeiitic) field. The mafic dykes cutting Pearya rocks east and west of Yelverton Bay are tholeiitic basalts to basaltic andesites (Fig. 10A). The samples of the 93–92 Ma old WIC comprise ultramafic rocks, hornblende gabbro and hornblende dolerite, monzonite, syenite, granite and alkali feldspar microgranite, which represent a transitional to mildly alkaline magma series (Fig. 10B). The same result is obtained using classification diagrams based on immobile elements, e.g., the Zr/TiO_2 vs. Nb/Y diagram from Winchester & Floyd (1977) (see Estrada et al., 2006; Estrada & Henjes-Kunst, 2013).

Two discrimination diagrams were selected to characterise the geotectonic setting of the Cretaceous igneous rocks. In the Ti–V diagram (Shervais, 1982), most of the mafic igneous samples including the tholeiitic varieties plot in the field for ocean-island

Figure 11. The Cretaceous igneous rocks of the Yelverton Bay area in discrimination diagrams for tectonic setting. (A) The mafic samples in the V–Ti diagram (Shervais, 1982). (B) The felsic samples in the Nb–Y diagram (Pearce et al., 1984). Abbreviations: HPVC – Hansen Point Volcanic Complex, MORB – mid-ocean ridge basalt, OIB – ocean-island basalt, ORG – ocean-ridge granites, VAG – volcanic arc granites, syn-COLG – syncollision granites, WIC – Wootton Intrusive Complex, WPG – within-plate granites.

Figure 12. Chondrite-normalised REE patterns of the Cretaceous igneous rocks of the Yelverton Bay area. (A) Mafic dykes cutting Pearya rocks. (B–F) Volcanic and subvolcanic lithologies of the Hansen Point Volcanic Complex (HPVC). (G–J) Intrusive lithologies of the Wootton Intrusive Complex (WIC). The field of the Isachsen and Strand Fiord flood basalts (shaded grey; data from Estrada, 2015) and the line of ocean-island basalt (OIB) are shown for comparison. Chondrite and OIB values are from Sun & McDonough (1989).

Figure 13. Primitive mantle-normalised multi-element patterns of the Cretaceous igneous rocks of the Yelverton Bay area. (A) Mafic dykes cutting Pearya rocks. (B–F) Volcanic and subvolcanic lithologies of the Hansen Point Volcanic Complex (HPVC). (G–J) Intrusive lithologies of the Wootton Intrusive Complex (WIC). The field of the Isachsen and Strand Fiord flood basalts (shaded grey; data from Estrada, 2015) and the line of ocean-island basalt (OIB) are shown for comparison. Primitive mantle and OIB values are from Sun & McDonough (1989).

basalts (OIB) and alkali basalts or along the division line between this field and the field for continental flood basalts (Fig. 11A). Only a few samples from the tholeiitic dykes cutting Pearya rocks plot in the field of the Isachsen and Strand Fiord basalts from Axel Heiberg Island, which are interpreted to be tholeiitic continental flood basalts (e.g., Estrada & Henjes-Kunst, 2004; Jowitt et al., 2014). The tholeiitic samples are characterised by high V contents (>270 ppm) like the Isachsen and Strand Fiord basalts. Most of the mafic WIC samples and of the HPVC basalts from west of Yelverton Bay are in good accordance with the field of the alkaline basalts from the southwestern HPVC outcrops (Emma Fiord and Phillips Inlet South; see Fig. 2). OIB is geochemically very similar to continental rift basalts (e.g., Fitton, 2007). Therefore, the mafic Cretaceous igneous rocks of the Yelverton Bay area are interpreted to have formed in a continental rift zone. The geotectonic setting of the silicic WIC and HPVC rocks is demonstrated by the Nb–Y diagram (Pearce et al., 1984). Nearly all samples fall in the field for within-plate granites (Fig. 11B). A relationship to subduction-zone magmatism can be excluded for all Cretaceous igneous rocks.

To obtain more information about the relationship between the tholeiitic and alkaline rocks of the study area, their chondrite-normalised REE patterns and primitive mantle-normalised multi-element patterns are compared with the patterns of the Isachsen and Strand Fiord basalts and OIB (Figs. 12, 13). The mafic dykes cutting Pearya rocks (Fig. 12A) and the HPVC dolerites from east of Yelverton Bay (Fig. 12B) show very similar chondrite-normalised REE patterns ($\text{La}_n/\text{Lu}_n = 3.4\text{--}5.4$ and $\text{La}_n/\text{Lu}_n = 4.9\text{--}6.6$, respectively), which run parallel to the field of the Isachsen and Strand Fiord basalts ($\text{La}_n/\text{Lu}_n = 2.5\text{--}5.4$), but are slightly more enriched for all REE. The heavy REE (HREE) are barely fractionated, indicating magma generation at lower pressure and the absence of garnet. The HPVC basalts from west of Yelverton Bay ($\text{La}_n/\text{Lu}_n = 8.2\text{--}28.1$) and the mafic WIC samples ($\text{La}_n/\text{Lu}_n = 6.7\text{--}23.7$) follow the OIB trend ($\text{La}_n/\text{Lu}_n = 13.2$) with a greater variation of the light REE (LREE) contents, probably due to variable degrees of partial melting (Fig. 12C, G). Some WIC samples with less fractionated HREE patterns can be interpreted as differentiates from a tholeiitic magma (hornblende gabbro SE 79/01, monzonite SE 84/01). The silicic rocks of the HPVC (Fig. 12D–F) and WIC (Fig. 12H–J) show very similar patterns, which are characterised by enrichment of the LREE over HREE ($\text{La}_n/\text{Lu}_n = 24.3\text{--}68.4$ and $\text{La}_n/\text{Lu}_n = 7.1\text{--}34.1$, resp.), little fractionation of the HREE and stronger fractionation of the LREE, a negative Eu peak (apart from the WIC syenite samples), and enrichment of the REE concentrations (apart from Eu) with increasing silica content.

Similar to the REE plots, the primitive-mantle-normalised multi-element patterns of the mafic dykes cutting Pearya rocks (Fig. 13A) and the HPVC dolerites

Table 3. Rb–Sr and Sm–Nd isotope data of Cretaceous igneous rocks of the Yelverton Bay area.

Unit, lithology	Sample	Age (Ma)	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}}(0)$	$\epsilon_{\text{Nd}}(t)$	t_{DM} (Ga)
Hansen Point Volcanic Complex (HPVC)														
basalt, HPVC west	SE 16/01	80	26.13	655.80	0.1153	0.705490	0.70536	5.34	28.62	0.1123	0.512774	2.7	3.5	0.47
basalt, HPVC west	SE 45/01	80	7.89	437.30	0.0522	0.704208	0.70415	5.28	23.35	0.1361	0.512881	4.7	5.4	0.40
basalt, HPVC west	SE 49/01	80	9.59	415.50	0.0668	0.704282	0.70421	4.99	22.33	0.1345	0.512860	4.3	5.0	0.43
basalt, HPVC west	SE 136/01	80	21.57	803.20	0.0777	0.703775	0.70369	11.63	69.94	0.1005	0.512837	3.9	4.9	0.33
basalt, HPVC west	SE 137/01	80	137.86	791.57	0.5037	0.704113	0.70354	11.02	70.91	0.0939	0.512804	3.2	4.3	0.36
trachyandesite, HPVC west	SE 10/01	80	140.29	269.75	1.5055	0.713692	0.71198	14.09	67.71	0.1258	0.512497	-2.8	-2.0	1.04
trachyandesite, HPVC west	SE 44/01	80	126.41	594.93	0.6147	0.706597	0.70590	12.49	79.45	0.0950	0.512634	-0.1	1.0	0.58
trachyandesite, HPVC west	SE 163/01	80	105.72	380.41	0.8045	0.712419	0.71150	9.96	47.36	0.1271	0.512484	-3.0	-2.3	1.08
rhyolite, HPVC west	SE 114/01	80	192.30	50.38	11.0690	0.720338	0.70776	14.77	74.20	0.1198	0.512621	-0.3	0.5	0.76
rhyolite, HPVC west	SE 119/01	80	115.70	96.80	3.4627	0.713496	0.70956	16.64	80.73	0.1241	0.512624	-0.3	0.5	0.79
dolerite, HPVC east (dyke cutting rhyodacite)	SE 74/01	90	42.48	253.00	0.4862	0.706747	0.70613	9.64	43.12	0.1346	0.512636	0.0	0.7	0.88
dolerite, HPVC east (dyke cutting rhyodacite)	SE 59/01	90	26.56	275.00	0.2797	0.705508	0.70515	9.07	37.84	0.1443	0.512768	2.5	3.1	0.71

Table 3. Continued

Unit, lithology	Sample	Age (Ma)	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}}(0)$	$\epsilon_{\text{Nd}}(t)$	$t_{\text{DM}}^{\text{Ga}}$
dolerite, HPVC east (dyke cutting rhyo-dacite)	SE 67/01	90	34.50	265.20	0.3763	0.706053	0.70557	9.35	39.73	0.1423	0.512746	2.1	2.7	0.73
rhyodacite	SE 9/1/00	100	159.70	222.40	2.0818	0.719247	0.71629	11.25	52.08	0.1301	0.512362	-5.4	-4.5	1.34
rhyodacite	SE 8/00A	100	255.00	65.35	11.3290	0.734621	0.71852	11.30	52.93	0.1286	0.512338	-5.9	-5.0	1.36
rhyodacite	SE 7/01	100	200.18	152.34	3.8094	0.727578	0.72216	11.97	57.48	0.1258	0.512282	-6.9	-6.0	1.42
rhyodacite	SE 27/01	100	199.85	86.57	6.6962	0.733132	0.72362	13.05	62.53	0.1261	0.512281	-7.0	-6.1	1.43
rhyodacite	SE 65/01	100	166.34	123.85	3.8891	0.715979	0.71045	12.72	59.83	0.1285	0.512553	-1.7	-0.8	0.97
rhyodacite	SE 69/01	100	167.20	117.97	4.1067	0.722893	0.71706	11.33	53.08	0.1290	0.512382	-5.0	-4.1	1.29
rhyodacite	SE 9/01	100	258.09	175.00	4.2744	0.725317	0.71924	14.38	68.13	0.1276	0.512299	-6.6	-5.7	1.42
Wootton Intrusive Complex														
hbl gabbro	SE 29/01	92	58.83	482.40	0.3530	0.704015	0.70355	5.41	24.53	0.1328	0.512883	4.8	5.5	0.38
hbl gabbro	SE 34/01	92	50.49	739.80	0.1976	0.703937	0.70368	9.86	58.40	0.1017	0.512834	3.8	4.9	0.34
hbl gabbro	SE 79/01	92	47.45	316.80	0.4337	0.707243	0.70668	8.49	39.26	0.1223	0.512522	-2.3	-1.4	0.95
monzonite	SE 84/01	92	100.30	292.80	0.9916	0.706626	0.70533	11.44	54.00	0.1276	0.512666	0.5	1.4	0.75
	SE 84/01, repl.							11.41	53.88	0.1275	0.512663	0.5	1.3	0.76
syenite	SE 76/01	92	92.42	194.60	1.3755	0.708425	0.70663	5.24	27.20	0.1159	0.512570	-1.3	-0.4	0.81
	SE 76/01, repl.							5.23	27.16	0.1159	0.512565	-1.4	-0.5	0.82
syenite	SE 32/01	92	57.72	785.50	0.2127	0.703939	0.70366	6.53	47.59	0.0826	0.512816	3.5	4.8	0.31
granite	SE 78/01	92	152.90	117.60	3.7651	0.711103	0.70618	7.65	38.60	0.1193	0.512620	-0.4	0.6	0.76
granite	SE 90/01	92	123.20	89.90	3.9695	0.710456	0.70527	4.07	24.65	0.0994	0.512566	-1.4	-1.4	0.70
microgranite	SE 100/01A	92	163.10	35.05	13.5030	0.725073	0.70742	21.82	111.51	0.1184	0.512626	-0.2	0.7	0.74
microgranite	SE 171/01	92	178.80	52.81	9.8139	0.720807	0.70798	14.49	71.55	0.1219	0.512632	-0.1	0.8	0.76
microgranite	SE 1/00	92	207.90	80.10	7.5252	0.718386	0.70855	9.82	53.71	0.1100	0.512586	-1.0	0.0	0.74
microgranite	SE 5/2/00	92	227.60	37.92	17.4210	0.731677	0.70890	11.75	60.70	0.1165	0.512601	-0.7	0.2	0.77
dykes cutting Pearya rocks														
dolerite	SE 154/01	109	33.40	216.98	0.4454	0.707485	0.70680	8.89	35.08	0.1532	0.512648	0.2	0.8	1.13
dolerite	SE 158/01	122	35.09	227.03	0.4472	0.708040	0.70726	8.93	35.18	0.1534	0.512655	0.3	1.0	1.11
dolerite	SE 7/2/00	95	50.76	260.87	0.5632	0.710852	0.71009	7.20	30.00	0.1452	0.512564	-1.4	-0.8	1.18
dolerite	SE 9/3/00	95	17.94	237.90	0.2184	0.705813	0.70552	8.27	34.39	0.1449	0.512780	2.8	3.4	0.68
dolerite	SE 9/4/00	95	30.07	275.58	0.3157	0.707590	0.70716	8.43	35.78	0.1425	0.512677	0.8	1.4	0.89

from east of Yelverton Bay (Fig. 13B) run parallel to the pattern of the Isachsen and Strand Fiord basalts. The mildly alkaline HPVC basalts from west of Yelverton Bay (Fig. 13C) and most of the mafic WIC samples (Fig. 13G) follow the OIB trend. Most of the silicic HPVC and WIC lithologies show very similar patterns, with typical Ta–Nb and Sr–P troughs as well as negative Ti peaks (Fig. 13D–F, H–J). Low Ba, Sr, P, Eu and Ti concentrations are the result of fractional crystallisation of plagioclase, apatite and Ti minerals. However, fractional crystallisation of a tholeiitic or OIB-like magma alone cannot explain the formation of the silicic rocks.

The Sr and Nd isotope ratios of the Cretaceous igneous rocks of the Yelverton Bay area vary over a wide range (Table 3; Fig. 14A). Two major trends are apparent in the Sr–Nd isotope variation diagram (Fig. 14A). The first, ‘steep’ trend with relatively unradiogenic Sr isotope ratios includes the mafic and felsic WIC rocks (apart from the subvolcanic microgranite), the mildly alkaline HPVC basalts from west of Yelverton Bay, and the mafic and felsic HPVC lavas from the outcrops farther southwest on northern Ellesmere Island (Phillips Inlet South and Emma Fiord; see Fig. 2; data from Estrada & Henjes-Kunst, 2004). The Sr and Nd isotope ratios of the mildly alkaline, low-phosphorous (LP) and high-phosphorous (HP) Hassel basalts from northeast Ellesmere Island also correspond with the ‘steep’ trend 1 (Estrada, 2015). The c. 100 Ma old HPVC rhyodacite, the HPVC rhyolite and two trachyandesite samples, the samples from the dykes cutting Pearya rocks, the WIC microgranite as well as the Isachsen and Strand Fiord basalts (data from Estrada & Henjes-Kunst, 2004) follow a second, ‘flat’ trend towards high-Sr isotope ratios. The isotope ratios of the tholeiitic to transitional dolerites from east of Yelverton Bay vary between the two trends.

The isotopically most primitive samples related to trend 1 are characterised by $\epsilon_{\text{Nd}}(t)$ of c. +4 to +6 and $(^{87}\text{Sr}/^{86}\text{Sr})_t$ of c. 0.703 to 0.704 typically for an OIB-like mantle source. Isotopically more evolved samples of trend 1 point to a mixing with a lithospheric component of likely lower continental crustal origin having experienced large ion lithophile (LIL) element depletion during an earlier granulite-facies metamorphism. This will have resulted in a time-integrated low Rb/Sr evolutionary path and unradiogenic Sr isotope ratios (DePaolo & Wasserburg, 1979). Alternatively, one may speculate about mixing with a low- ϵ_{Nd} subcontinental lithospheric mantle component (e.g., DePaolo, 1988). The isotopically primitive end member of trend 2 is characterised by Nd isotopic compositions similar to those of trend 1, but by slightly higher Sr isotope ratios. The enriched end member of trend 2 is a lithospheric component with very high Sr isotope ratios, probably upper continental crust of the Pearya Terrane as indicated by the Sr–Nd isotope composition of the Late Mesoproterozoic to Early Neoproterozoic orthogneisses of Pearya succession 1 (unpublished BGR data; see inset in Fig. 14A).

In the $(^{87}\text{Sr}/^{86}\text{Sr})_t$ vs. SiO_2 diagram (Fig. 14B), the mafic and evolved rocks of the mildly alkaline suites (WIC, HPVC west of Yelverton Bay, HPVC of the southwestern outcrops Phillips Inlet South and Emma Fiord) show an almost horizontal trend indicative of fractional crystallisation. Only the elevated $(^{87}\text{Sr}/^{86}\text{Sr})_t$ values of the WIC microgranites and the HPVC rhyolites indicate a limited assimilation of crustal rocks. The upward trend of the tholeiitic basalts (dykes cutting Pearya rocks, dolerites from east of Yelverton Bay, Isachsen and Strand Fiord basalts) is characteristic for assimilation with a minor influence of fractional crystallisation. The trachyandesites follow partly the tholeiitic trend and partly the alkaline trend. The rhyodacitic magma was probably formed when a continental tholeiitic magma that had already experienced assimilation and fractional crystallisation (AFC) processes at a deeper level, split off and experienced further crystal fractionation and intensive assimilation of continental crust in new high-level magma chambers. Assimilation of upper crustal material is supported by the presence of zircons from the Pearya basement rocks in the rhyodacite and its Mesoproterozoic Nd model ages (Table 3). The deeper magma reservoir is probably the source of the large magnetic anomaly over Yelverton Bay (Fig. 3A). Thus, the silicic igneous rocks of the Yelverton Bay area – despite their similar REE and multi-element patterns – have experienced a quite complex petrogenetic development involving mixing of OIB-type mantle components with two isotopically different evolved components of crustal and/or lithospheric mantle origin.

Discussion

Depth of anomalies and structure of the passive margin – implications from magnetic modelling

The variability of the amplitudes and frequency content of the magnetic anomalies shown in Fig. 3A suggest that different magnetic sources are present in the Yelverton Bay area. We have used a magnetic 2D modelling approach to estimate the possible source body distributions that might explain the observed magnetic data. The magnetic modelling was conducted on two approximately N–S-trending transects (Figs. 2, 15). Transect 1 (Fig. 15A) includes the southern part of the ‘ARTA’ refraction seismic line (lower panel, Fig. 16A; Funck et al., 2011) and extends from the shelf between Axel Heiberg and Ellesmere islands northwards onto the Alpha Ridge. The observed magnetic anomaly for transect 1 was interpolated from the gridded map in Fig. 15A (Teskey et al., 1995); therefore, the transect 1 model is restricted in the north to the extent of this aeromagnetic survey. Transect 2 was constructed along one of the 1991 GSC aeromagnetic survey flight lines (Fig. 15) approximately 150 km east of transect 1. The

Figure 14. (A) Sr–Nd isotope variation diagram for the Cretaceous igneous rocks of the Yelverton Bay area (data from Table 3). Two mixing trends are indicated (lilac dashed lines). The inset shows the Sr–Nd isotope diagram with the field of Pearya succession 1 orthogneiss at 100 Ma (unpublished BGR data) as an upper crustal end-member of mixing trend 2. (B) Variation of the Sr isotope ratio vs. SiO_2 as differentiation index. The fields of the Isachsen and Strand Fiord basalts (grey) and of the Hansen Point Volcanic Complex from Phillips Inlet South and Emma Fiord (HPVC/PI+EF, green) are shown for comparison (data from Estrada & Henjes-Kunst, 2004). Formation ages of 120 Ma for the Isachsen basalts, 100 Ma for the Strand Fiord basalts, and 80 Ma for the HPVC/PI+EF are assumed. Abbreviations: AFC – assimilation and fractional crystallisation, CHUR – chondritic uniform reservoir, WIC – Wootton Intrusive Complex.

line extends from the Sverdrup Basin to the Pearya Terrane and ends on the shelf adjacent to Ellesmere Island. All of transect 2 is likely continental crust; however, there is a large positive magnetic anomaly over the Yelverton Bay shelf area that looks very similar in its amplitude and irregular shape to the large anomalies of the Alpha Ridge (Fig 15B).

The transect 1 magnetic model (Fig. 16A) consists of two bodies: one representing highly magnetised Alpha

Ridge crust in the north with both induced and natural remnant magnetisation (NRM); and the other with only induced magnetisation representing upper continental crust. The high NRM values for the Alpha Ridge crust are the upper limit of magnetisation values of typical upper oceanic crust (e.g., pillow basalts; Gee & Kent, 2007) and the susceptibility values used in the models are in general agreement with our laboratory measurements on basaltic rocks (Table 2). The upper boundary of the highly magnetised (Alpha Ridge) body coincides with the top

Figure 15. Magnetic anomalies and the location of two transects that are modelled in Fig. 16. (A) Part of the Canadian aeromagnetic survey (Teskey et al., 1995) over Ellesmere and Axel Heiberg islands (see also Fig. 1). Transect 1 matches the southern part of the main refraction seismic line of Funck et al. (2011) (basically their ‘inner line’). Transect 2 coincides with the aeromagnetic line L91450 of Teskey et al. (1995). Bathymetric contours and elevations on land are from the DTU10 data set (Andersen & Knudsen, 2009; Andersen, 2010). (B) Index map showing the location of (A) on the magnetic anomaly map of the Arctic from Gaina et al. (2011).

of the volcanic upper crustal layer of Funck et al. (2011). Although the modelled thickness is ~6 km, it is strongly dependent on the assigned magnetisation values and it would also be possible to model the anomaly using lower magnetisation values for the lower parts of the crust for an even thicker source body. The model implies that the magnetic anomaly over the continental slope is caused by a thick (more than 6 km) magnetic layer with a strong bulk magnetisation that becomes thinner and deepens to the south in the area of the inferred transitional crust. The high-amplitude positive anomaly at km 180 and the wide minimum in the middle of the profile are partly an effect of the topography of the upper boundary of the magnetic layer but also an edge effect due to the thinning of the main magnetic body. Therefore, this model explains the magnetic anomalies basically as the edge effect of the thick igneous crust of the Alpha Ridge where it abuts against the Axel Heiberg/Ellesmere Island continental crust.

Transect 2 (Fig. 16B) crosses the long-wavelength anomaly at 82.5°N/85°W in its northern part and two

narrow but distinct chains of positive anomalies that are related to the Wootton Intrusive Complex (WIC; Fig. 3, anomaly chain C1) and the Hansen Point Volcanic

Figure 16. (A) Magnetic model calculation for the southern part of the seismic refraction line (transect 1) from Funck et al. (2011). Upper panel shows the measured data (draped flight altitude, approximately 300 m over ground) and the model anomaly for the magnetisation model in the middle panel. The lower panel shows the outline of the model bodies in relation to the velocity model of Funck et al. (2011, modified from their fig. 7). For the location of the line, see Figs. 2 & 15. (B) Magnetic model as in (A) for a synthetic line along aeromagnetic flight line L91450 from (Teskey et al., 1995). The lower panel shows the outline of the model bodies in relation to some geological features (see Fig. 2) and a hypothetical projection of the simplified seismic results from transect 1 between 0 and 100 km (adapted from Funck et al., 2011) assuming that transect 2 starts about 50 km landward of the shelf edge (see Fig. 15). Abbreviations: B. – Basin, EFFZ – Emma Fiord Fault Zone, HPVC/PI – Hansen Point Volcanic Complex/Phillips Inlet South, HVLC – high velocity lower crust, MPFZ – Mitchell Point Fault Zone, PBF – Petersen Bay Fault, WIC – Wootton Intrusive Complex.

Complex/Phillips Inlet South (HPVC/PI; see Fig. 2). The model utilises again a thick strongly magnetic layer in its northern part that simultaneously explains the strong positive anomaly at km 50 and the minimum at km 75 as a magnetic edge effect of the thick body that ends at km 70. An elongated body between km 80 and km 150 at about 7 km depth may represent a magma reservoir that fed the WIC and HPVC/PI high-level magma chambers. The simplified crustal section from Funck et al. (2011) in the lower panel of Fig. 16B between 0 and 100 km was projected from transect 1 to transect 2 using the observation that transect 2 starts about 50 km south of the slope. In accordance with the geology on land (Fig. 2) we subdivided transect 2 into three main sections, the Pearya Terrane, the Franklinian Basin and the Sverdrup Basin. The main source body of the model lies entirely within the Pearya Terrane continental crust at a depth between c. 6 and 15 km. It is unlikely that the body comes much closer to the surface because of the very smooth appearance of the anomaly at km 50. Therefore, and since the southern margin of the large anomaly covers Pearya rocks (Fig. 3), we propose that the igneous body represents a (? repeated) intrusion of mafic magma into the Pearya Terrane rocks. However, the offshore extent of the Pearya Terrane is unknown and thus we cannot completely exclude an extrusive origin for the northernmost part of the igneous body.

The joint interpretation of the magnetic models for transects 1 and 2 (Fig. 15) gives strong indications for large volumes of highly magnetised igneous material on both lines but in rather different settings. On transect 1 the main part of the magnetic source body lies seaward of the continental slope where the existence of oceanic or thickened igneous crust provides a natural explanation for large volumes of mafic igneous material. A similar and even stronger magnetic anomaly on transect 2 also requires large volumes of igneous material but within continental crust.

Such large intrusions should be accompanied by updoming of the continental crust. This was probably realised by the uplift of Pearya succession 1, the oldest known rocks of the Pearya Terrane, which form a NE–SW-trending belt in the Yelverton Bay area.

Comparison between the Cretaceous magmatism of northern Ellesmere Island and other parts of the HALIP

Canadian Arctic

In the Yelverton Bay area of northern Ellesmere Island, HALIP-related igneous activity is dated at c. 122 Ma to 74 Ma (Table 1). Tholeiitic magmatism lasted until c. 83 Ma and overlapped with the mildly alkaline magmatism that started at c. 93 Ma (WIC), or already at c. 96 Ma including the Hassel basalts of northeast Ellesmere Island. The main interval of magmatism took place in the

Yelverton Bay area at around 100 Ma with emplacement of rhyodacitic lava and volcanoclastic rocks preserved east and west of Yelverton Bay. They form part of a volcanogenic sequence that is magnetically expressed as a NNE–SSW-striking lineament crossing Yelverton Bay (anomaly chain C2; Fig. 3A). Geological units are displaced east and west of the magnetic lineament (Fig. 3). Pearya succession 2 is exposed north of succession 1 on Wootton Peninsula, but south of succession 1 east of Yelverton Bay. This is probably a consequence of a segmentation of the northern margin of Ellesmere Island, as described for the magma-rich rifted margin of central East Greenland (Karson & Brooks, 1999). During and after the emplacement of the rhyodacitic rocks, intrusion of mafic dykes occurred east of Yelverton Bay, but with a change towards a more transitional composition. West of Yelverton Bay, mildly alkaline intrusive magmatism (WIC) started at c. 93 Ma and was followed c. 10 m.y. later by extrusive activity, the youngest stage of the HPVC. The emplacements of the SW–NE-striking chain of WIC rocks (anomaly C1) and of the NNE–SSW-striking chain of HPVC-type rocks (anomaly C2) was controlled by segments of the Mitchell Point Fault Zone, a pre-existing structure of the Pearya Terrane (Fig. 3).

In the central Sverdrup Basin (Axel Heiberg Island, Ellef Ringnes Island), intrusive and extrusive basaltic magmatism (sills and dykes, Isachsen and Strand Fiord basalts, minor pyroclastic rocks) was active between c. 130 and 100 Ma. On Ellef Ringnes Island, a main intrusive interval at 126–121 Ma is related to the emplacement of saucer-shaped sills (Evenchick et al., 2015). This time interval overlaps with Isachsen basalt volcanism on Axel Heiberg Island. In the Yelverton Bay area, some mafic dykes cutting Pearya rocks can be related to this interval. The source for the large magnetic high (H1) over the coastal and shelf area of Yelverton Bay is modelled as a deep, thick, mafic intrusion (see section ‘Depth of anomalies...’ and Fig. 16). The age of the intrusion is unknown. Rhyodacitic lava was formed by AFC processes from a deeper, tholeiitic magma as indicated by geochemical and Sr–Nd isotope data. It contains zircons with Cretaceous ages up to c. 120 Ma, older than the age of extrusion at c. 104 Ma. Although such zircon ages are rare in the dated samples, they are a hint that the intrusion took place probably during the first main stage of HALIP activity in the Canadian Arctic.

Renewed volcanic activity on Ellef Ringnes Island between c. 105 and 101 Ma (Evenchick et al., 2015) overlaps with the Strand Fiord basalt volcanism on Axel Heiberg Island and western Ellesmere Island (Fig. 2) as well as with the rhyodacite volcanism and the ongoing intrusion of mafic dykes in the Yelverton Bay area. However, Late Cretaceous mildly alkaline magmatism that started in northern Ellesmere Island at c. 96 Ma (Estrada, 2015) is not recorded in the central Sverdrup Basin.

Alpha–Mendeleev Ridge

To date, only a few samples have been collected from the submarine Alpha–Mendeleev Ridge and age data are very rare. A basalt sample (PS051–041–1cc) taken by gravity coring from the central Alpha–Mendeleev Ridge during the ARCTIC–98 expedition was formerly dated at 82 Ma (Jokat, 2003), but yielded a new Ar–Ar plateau age of 89 ± 1 Ma (Jokat et al., 2013). A trachybasalt sample drilled from the northern Mendeleev Ridge (Trukshin Seamount) during the Russian expedition “*Arctic-2012*” was dated at 127 ± 3 Ma (SHRIMP U–Pb age of a single zircon, 2 spots; Morozov et al., 2013). Preliminary Ar–Ar ages grouping around 112 Ma, 100 Ma and 85–73 Ma are reported from basaltic samples from the southern Mendeleev Ridge and Chukchi Borderland, which were recovered during the US Coast Guard Icebreaker Healy expedition in 2008 and the Russian expedition “*Arctic-2012*” (Brumley, 2014; Mukasa et al., 2015). These ages are in the same range as the ages of HALIP-related igneous rocks from onshore areas of the Canadian margin. First geochemical and isotope data indicate a continental origin for the magmas from the southern Mendeleev Ridge and Chukchi Borderland (Morozov et al., 2013; Mukasa et al., 2015). However, in contrast to the Canadian margin, the older basalts (112 Ma) show alkaline, continental-rift related characteristics and the younger basalts (100–73 Ma) are comparable with continental flood basalts (Mukasa et al., 2015). This fact argues against a synchronous tectonic development of these marginal areas of the Amerasia Basin.

North Greenland

The magnetic anomaly pattern over North Greenland and the adjacent Lincoln Sea bears many similarities with the magnetic anomalies of the Yelverton Bay area (Damaske et al., 1997; Nelson et al., 1998; Damaske & Estrada, 2006; Jokat et al., 2015). A broad magnetic high over the southern Lincoln Sea (Damaske & Estrada, 2006, fig. 3) is comparable with the magnetic high H1 over Yelverton Bay. It is part of the magnetic high over the Morris Jesup Rise (Fig. 1) that is characterised as ‘thick mafic crust’ (Saltus et al., 2011). The Lincoln Sea magnetic high is bordered to the south along the coast by a chain of small-scale, positive, high-amplitude anomalies. On land, the anomaly chain includes the outcrops of the volcanic rocks of the Kap Washington Group on the north coast of Greenland (Damaske & Estrada, 2006).

Compared with the Canadian Arctic, the onset of Cretaceous igneous activity in North Greenland took place definitively later. The intrusion of N–S and E–W-trending alkaline mafic dykes with Ar–Ar ages of 85–80 Ma (Kontak et al., 2001) and U–Pb ages of 82–81 Ma (Thórarinnsson et al., 2015) was followed by the bimodal volcanism of the Kap Washington Group with U–Pb zircon and Ar–Ar amphibole ages in the range of 71–61 Ma (Tegner et al., 2011; Thórarinnsson et al., 2011). This igneous activity is temporally close to another event of

alkaline volcanism on northeast Ellesmere Island that is recorded by volcanic pebbles with Ar–Ar ages of 61–58 Ma, which form part of the clastic infill of Palaeogene pull-apart basins along the Canadian side of Nares Strait (Estrada et al., 2010).

Whereas the mildly alkaline magmatism of northern Ellesmere Island (c. 96–75 Ma) is genetically closely related with the formation of the passive margin of the Alpha Ridge segment of the Amerasia Basin as discussed above, the Late Cretaceous alkaline magmatism of North Greenland and the Nares Strait region is interpreted as initial continental rifting related to the formation of the Eurasia Basin and movements along a temporary plate boundary between North America and Greenland (e.g., Estrada et al., 2010; Tegner et al., 2011; Thórarinnsson et al., 2015). However, E–W extension in North Greenland might have caused reactivation of NE–SW-striking fault zones in northern Ellesmere Island, even accompanied by new igneous activity (e.g., Soper & Higgins, 1991; Piepjohn & von Gosen, 2001). Thus, it is possible that late tectonic processes related to the formation of the Amerasia Basin overlap with tectonic processes during the early evolution of the Eurasia Basin.

Conclusions

At the north coast of Ellesmere Island around Yelverton Bay, the metamorphic basement rocks of the Pearya Terrane were strongly affected by Cretaceous HALIP-related magmatism that is responsible for characteristic positive magnetic anomaly patterns. A large long-wavelength magnetic high (H1) covers the shelf and coastal areas of Yelverton Bay (Fig. 3A). Magnetic modelling revealed a thick intrusive body at a depth of c. 6–15 km as the source of the anomaly (Fig. 16B). The strong but spatially restricted magnetic anomaly in the Yelverton Bay area occurs just where the centre of Alpha Ridge in either the bathymetric map (Fig. 1) or the magnetic map (Fig. 15B) meets the Canadian continental margin.

A chain of small, short-wavelength, positive anomalies crosses Wootton Peninsula from southwest to northeast (C1), continues NNE-wards across Yelverton Bay forming a c. 50 km long lineament (C2) and turns again northeastwards offshore Ayles Fiord (C3) (Fig. 3A). On land, the anomaly chain covers outcrops of mafic to felsic extrusive and intrusive rocks of the HPVC and WIC. We infer that these igneous rocks and related volcanoclastic rocks are the source for the complete anomaly chain including the seaside areas. The magmatism and thus the anomaly chain follow in part pre-existing fault zones in the Pearya Terrane. Rhyodacite that forms large parts of the HPVC originates from a tholeiitic magma series and is probably genetically related to the deep intrusive body (anomaly H1). It was dated at c. 104 to 97 Ma (LA–ICP–

MS U–Pb zircon ages). Thus, the HPVC of the Yelverton Bay area was formed between c. 104 Ma and 74 Ma and includes the c. 93–92 Ma old WIC (Table 1). Based on still sparse age data, HALIP-related magmatism in the Yelverton Bay area altogether lasted from c. 122 to 74 Ma and changed from tholeiitic to dominantly mildly alkaline since c. 96 Ma.

The Yelverton Bay area (close to the Alpha Ridge of the Amerasia Basin) shows some characteristic features, which are known from magma-rich rifted (volcanic) margins worldwide (e.g., Menzies et al., 2002) and from central East Greenland (e.g., Karson & Brooks, 1999; Tegner et al., 2008):

- Continental flood basalt volcanism including related mafic dykes and a deep, thick, intrusive body below the shelf (anomaly H1; Figs. 3, 16);
- Presence of high-velocity lower crust (HVLC; see Fig. 16; Funck et al., 2011);
- Segmentation of the margin by a coast-oblique lineament (anomaly chain C2; see Fig. 3);
- Coexistence of tholeiitic and alkaline magma series.

Our interpretation of the Yelverton Bay area as a magma-rich rifted margin coincides with the tectonic model from Funck et al. (2011) in which the Alpha–Mendeleev Ridge was formed by seafloor-spreading parallel to the Canadian Polar margin (the Iceland model). A discussion about the involvement of a mantle plume that is postulated by Funck et al. (2011) and many other authors is out of scope of this paper.

The long duration of HALIP magmatism in the Canadian High Arctic (c. 130 to 75 Ma) indicates that the Amerasia Basin was formed over a longer time span and that its several segments probably opened at different times. The temporal relationship between the main intervals of the HALIP magmatism at c. 130–120 Ma and c. 100 Ma and the rift–drift transition in the segments of the Amerasia Basin is still unclear and the time for the onset of seafloor spreading is discussed controversially (summarised by Dove et al., 2010; Funck et al., 2011; Shephard et al., 2013). The geochronological database is still very limited and more samples and data from the basement of the Amerasia Basin and the Alpha–Mendeleev Ridge as well as from the possible conjugate Siberian margin are necessary to understand the geological history of the Amerasia Basin of the Arctic Ocean.

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