U-Pb zircon constraints on obduction initiation of the Unst Ophiolite: an oceanic core complex in the Scottish Caledonides?

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Abstract

The Unst Ophiolite is the best exposed of a chain of early Ordovician ophiolites in the Scottish Caledonides and is widely regarded as having formed in a supra-subduction zone setting within the Iapetus Ocean. Reinterpretation of sheeted dykes suggests that it formed as an oceanic core complex, presumably during subduction roll-back immediately prior to obduction onto the Laurentian margin. A new U-Pb zircon age of 484 ± 4 Ma for development of the metamorphic sole places a lower limit on the timing of obduction, which was subsequently followed by regional-scale crustal thickening and Barrovian metamorphism during the Grampian orogenic event.

Supplementary material: Analytical methods, CL images of representative zircon grains, and a LA-ICP-MS U-Th-Pb zircon data table are available at www.geolsoc.org.uk/SUPXXXXX.

Ophiolites represent slices of oceanic-type lithosphere that have been tectonically incorporated into continental margins in convergent plate settings (e.g. Dewey & Bird 1970; Coleman 1971). Many ophiolites were formed in supra-subduction zone (SSZ) arc-forearc settings shortly prior to orogenesis and then thrust (obducted) onto a colliding passive margin (e.g. Davies & Jaques 1984; Searle & Cox 1999). The metamorphic soles present beneath many ophiolites are thought to have resulted from high-temperature (± pressure) metamorphism of a subducting oceanic slab beneath hot sub-ophiolitic mantle and later accretion onto the base of the ophiolite during obduction (e.g. Dewey & Casey 2013 and references therein). One of the most significant findings of the last 20 years of research into modern ocean basins has been the identification of oceanic core complexes that expose upper mantle and lower crustal rocks on the seafloor (e.g. Cannat 1993; Tucholke & Lin 1994). Although not recorded in many modern fore-arc settings, they have been identified in some SSZ ophiolites (e.g. Suhr & Cawood 2001; Tremblay et al. 2009), thus representing important opportunities for direct study of oceanic core complexes which are otherwise difficult to access.

The Unst Ophiolite in the Shetland Islands (Fig 1) is the northernmost and best exposed of a chain of SSZ ophiolites within the Laurentian Caledonides of Scotland and Ireland (e.g. Williams & Smyth 1973; Flinn 1999 and references therein). Other examples include the Bute, Ballantrae and Tyrone ophiolites and the Clew Bay Complex (Fig 1a). These ophiolites were emplaced during the early to mid-Ordovician Grampian orogenic event which records the initial stages in closure of the Iapetus Ocean (Chew et al. 2010). The collision of a juvenile oceanic arc with the Laurentian passive margin resulted in NW-directed ophiolite obduction (present reference frame) with regional deformation and metamorphism of footwall metasedimentary successions (e.g. Dewey & Shackleton 1984; Chew et al. 2010). The formation and subsequent obduction of the Unst Ophiolite is currently bracketed between its magmatic age of 492 ± 3 Ma (U-Pb zircon, Spray & Dunning 1991) and K-Ar mineral ages obtained from the metamorphic sole which range from 479 ± 6 Ma to 465 ± 6 Ma (Spray 1988). In this paper we: 1) report the results of U-Pb dating of metamorphic zircon from the sole of the ophiolite, thus placing a lower limit on the timing of obduction , and 2) re-evaluate published geological data from the ophiolite to show that these are consistent with an origin as a SSZ oceanic core complex with limited magma supply.

Regional geological setting and internal architecture of the Unst ophiolite The Unst Ophiolite consists of two main nappes of ultramafic and mafic rocks and associated low-grade metasedimentary rocks and melanges (Fig 1c; Flinn 1999, 2000, 2001 and references therein). The overall structure is a down-folded klippe as suggested by the open synformal folding of the upper nappe on the island of Fetlar, 5 km south of Unst, and modelling of gravity and magnetic anomalies (Flinn 2000). The maximum depth to the tectonic base of the ophiolite is unlikely to be greater than 3 km (Flinn 2000). On Unst, the lower ophiolite nappe rests on metasedimentary rocks of the mid-Neoproterozoic to Cambrian Dalradian Supergroup (Fig 1b; Flinn 1999). A U-Pb monazite age of 462 ± 10 Ma obtained from Dalradian rocks in Unst is thought to date peak metamorphism during the Grampian orogenic event (Cutts et al. 2011). The gently-dipping to steep fault that underlies the lower nappe is not the original obduction thrust but the result of later reworking of the nappe pile (Cannat 1989; Flinn & Oglethorpe 2005; Cutts et al. 2011).

The lower nappe consists of three steeply-inclined to sub-vertical layers with a total maximum thickness of c. 7 km measured normal to lithological boundaries (Fig 1c; Flinn 1999, 2001). From northwest (structurally lowest) to southeast (highest) these comprise metaharzburgite, metadunite (both extensively serpentinized) and metagabbro (Fig 1c), which have been affected by greenschist facies metamorphism. A plagiogranite within a metagabbro yielded a U-Pb zircon age of 492 ± 3 Ma (Spray & Dunning 1991). Uppermost parts of the metagabbro in SE Unst are intruded by sub-parallel mafic dykes, separated by narrow screens of host metagabbro (Prichard 1985; Flinn 1999). The dykes have been interpreted as the base of a sheeted dyke complex (Prichard 1985; Spray 1988). The boninitic chemistry of some of these dykes provides evidence that the ophiolite formed in a SSZ setting (Prichard & Lord 1988; Spray 1988; Flinn 2001). Trace element discrimination diagrams indicate a range of volcanic arc basalt/island arc-like compositions (Flinn 2001). However, these dykes depart from classic sheeted dykes described from many other ophiolites because they never form >50% of the rock by volume in any area and are nearly *parallel* to the ophiolite layering than normal to it, so at present they occur as sub-vertical sheets (Flinn 2001).

The Unst Ophiolite does not contain any pillow lavas. Instead, the metagabbro layer of the lower nappe, together with its 'sheeted dykes', is overlain by dark, locally graphitic, quartz-sericite-chlorite meta-siltstones of the Muness Phyllites (Fig 1c). The presence of metagabbro and plagiogranite clasts, as well as elevated concentrations of Cr, suggests that the underlying ophiolitic rocks contributed detritus to the sedimentary protoliths of the Muness Phyllites. These phyllites are therefore thought to have been deposited unconformably on the lower nappe (Flinn 1985). Exposed contacts between the Muness Phyllites and the metagabbro occur as faults, but none of these are thought to be regionally significant structures. The structurally overlying upper nappe is dominated by metaharzburgite, and thought to result from the tectonic duplication of the lower nappe, most probably during the c. 435-430 Ma Scandian orogenic event (Flinn & Oglethorpe 2005).

Both nappes are underlain by discontinuous tectonic slices of amphibolite and hornblende schist which were interpreted by Williams & Smyth (1973) as a metamorphic sole to the ophiolite. Geochemical studies indicate a MORB-type basaltic origin for the igneous precursors (Spray 1988; Flinn 2001). On Unst an early amphibolite facies mineral assemblage of hornblende + plagioclase + titanite ± apatite shows evidence for greenschist facies retrogression to actinolite + albite + epidote (Spray 1988). On Fetlar, texturally early garnet-clinopyroxene assemblages preserve evidence for upper amphibolite facies P-T conditions of 700-780°C and 9-11 kbar (Spray 1988; Flinn et al. 1991). At several localities, the metabasic rocks contain a well-developed gneissic fabric defined by cm-scale trondjhemitic layers interpreted to be the result of high-temperature segregation. K-Ar hornblende ages obtained from various samples of the metamorphic sole (lower and upper nappe) range from 479 ± 6 Ma to 465 ± 6 Ma.

Sample descriptions To further constrain the timing of formation of the metamorphic sole, we investigated two samples: 11-SH-10 (collected from beneath the lower nappe on Unst at HP 5635 0067) and 11-SH-12 (collected from beneath the upper nappe on Fetlar at HU 6450 9206). Both samples are amphibolites, with 11-SH-10 being medium-grained and 11-SH-12 being medium to fine grained. Both amphibolites posess a well-developed foliation defined by alternating mafic and felsic layers. Mafic layers are dominated by aligned grains of hornblende, which in the case of sample 11-SH-10 wrap skeletal garnets up to 5 mm in diameter. Titanite and apatite occur as visible accessory minerals. Felsic layers are dominantly composed of sericitised plagioclase with minor quartz. There is evidence for widespread replacement of hornblende by actinolite and biotite, and garnet (where present) by chlorite. Scattered grains of secondary epidote are also common.

U-Pb zircon data: results and interpretation. Cathodoluminescence (CL) images of zircon grains show sector and fir-tree zoning patterns and large parts of grains which are homogeneous in CL, all of which is consistent with a single phase of metamorphic growth (Corfu et al. 2003), which most likely occurred during upper amphibolite facies P-T conditions as previously recognised from the early mineral assemblage (Spray 1988) . No inherited cores and no igneous zircon were evident, compatible with a MORB-type protolith (Spray 1988; Flinn 2001). Representative zircons from both samples were analysed by LA-ICP-MS following the operating conditions of Crowley et al. (2014 see supplementary files for data table) and were found to have low U concentrations (generally <3ppm for sample 11-SH-10 and <0.3ppm for sample 11-SH-12). 69 zircon analyses from 11-SH-10 are concordant to near-concordant (Fig 2) and do not appear to have suffered any significant Pb-loss. A weighted mean $^{206}Pb/^{238}U$ age calculation from these data returns a date of 483.7 ± 4.4 Ma (MSWD=1.18), interpreted as the age of formation of the metamorphic zircon. Despite a large number of analyses, only one age determination was possible from 11-SH-12. This grain was found to have relatively elevated U concentration (c. 5ppm) compared to other zircons from the same sample and yielded a $^{206}Pb/^{238}U$ age of c. 482 ± 18 Ma (i.e. within error of the age determination for sample 11-SH-10).

An oceanic core complex model for the Unst Ophiolite Oceanic core complexes are known to develop in conjunction with some intermediate to slow spreading centres and represent exhumed lower crustal and upper mantle rocks in the footwalls of extensional detachments (Whitney *et al.* 2013; Platt *et al*. *in press*). They typically exhibit a dome-shaped massif, which may display considerable topographic relief relative to the surrounding sea-floor. Oceanic core complexes are usually cored by gabbro, or gabbro and peridotite and do not exhibit the classic Penrose-type ophiolite sequence (Miranda & Dilek 2010). They have been described from segments of the Australian-Antarctic Discordance, Caribbean–North American Ridge, Mid-Atlantic Ridge, Southwest Indian Ridge and from several back-arc spreading centres (Whitney *et al*. 2013 and references therein). Several oceanic core complexes occur in the inner-bend of ridge-transform intersections, so there is likely an interplay of magmatic and tectonic processes, the mutual rates of which are critical to initiating accelerated extension in magma-poor areas of divergent zones (Whitney *et al*., 2013). Slow spreading rates of <55 mm yr^{-1} may result in episodic and spatially variable magma supply, which in turn facilitates an increase in tectonic partitioning, accelerated lithospheric thinning and denudation of oceanic crust (Miranda & Dilek, 2010). Furthermore, the rate of magma supply (i.e. not too rapid and not too slow) appears to be a critical factor in the development of oceanic core complexes (Tucholke *et al*. 2008). From a structural perspective, the presence of pre-existing inward-dipping normal fault systems, coupled with lithospheric thinning results in a passive rotation or flexure of faults (Morris *et al*. 2009), resulting in the formation of low angle detachments. Tucholke *et al.* (2008) state that if dykes and plutons collectively account for 50% or more of the total extension, then detachment faults initiated at spreading centres may accumulate large displacements.

Any model for the formation of the Unst Ophiolite has to account for the steep orientation of the internal layering relative to the assumed gently-dipping basal thrust (Fig 1c), the intrusion of the boninitic dykes sub-parallel to the steep layering within host upper gabbro and their absence from underlying units. In contrast to Spray (1988) who envisaged formation within an SSZ setting *sensu stricto*, Flinn (2001) suggested that it was derived from a subducting oceanic slab. In this latter interpretation, the SSZ chemistry was derived from flushing the subduction zone with fluids from the dehydrating descending slab. These fluids promoted partial melting of the overlying wedge, forming basaltic melts that were intruded into the upper gabbro as dykes from above and to the side. Flinn (2001) envisaged that during arc-continent collision, a thrust that initiated in the overlying mantle wedge cut down through the 'stratigraphy' of the subducting plate, then more parallel to it, eventually acting as the obduction thrust for the ophiolite as it was thrust onto the Laurentian margin. The Muness Phyllites were thought to have been deposited on the ophiolite *after* it was obducted.

In contrast, we suggest that all the main features of the Unst Ophiolite are more easily accounted for if it is viewed as having originated as an oceanic core complex that developed in association with a spreading ridge in the SSZ plate. It should be noted that Flinn (1999) in considering field relationships within the ophiolite suggested: "It is possible that the slice of crust and upper mantle forming the Lower Nappe slipped and rotated into a recumbent position on a listric fault at the constructive margin prior to the intrusion of the dykes" but did not develop this model any further. We justify our interpretation as follows; firstly, the steep orientation of the internal layering of the ophiolite (Fig 1c) can largely be accounted for by rotation of the oceanic lithosphere in the footwall of a major detachment that was initiated as a steep normal fault but was rotated to a gently-dipping orientation during progressive extension. Rotations of 50-80° have been demonstrated at slow-spreading sectors of mid-ocean ridges (e.g. Garcés & Gee 2007; Morris et al. 2009) and could also account for the otherwise enigmatic steep orientation of the earliest and dominant magnetization direction within the Unst Ophiolite (Taylor 1988). Secondly, the occurrence of dykes sub-parallel to the layering within the upper gabbros, and their absence from underlying units, can be accounted for by ongoing intrusion and dyke emplacement into this part of the ophiolite after rotation had been substantially completed. Garcés & Gee (2007) documented this process along the Fifteen-Twenty Fracture Zone in the Central Atlantic Ocean. Thirdly, no evidence precludes the possibility that the Muness Phyllites were deposited on the ophiolite *before* obduction. The generally low energy nature of the fine-grained sedimentary protoliths, albeit interrupted occasionally by high-energy debris flows, is entirely consistent with deposition in a deep-water oceanic setting on (and derived from) gabbros that were exposed on the sea-floor following detachment faulting and initiation of core complex formation. Such a predominantly low-energy sedimentary facies is unlikely to be deposited on top of an obducted ophiolite.

Discussion and implications The new interpretations and data presented here support a new tectonic model for the Unst Ophiolite. The timing of initiation of oceanward-subduction is poorly constrained but is likely to have occurred by c. 510 Ma (Chew et al. 2010), followed by subduction roll-back and development of the precursor to the SSZ ophiolite at 492 ± 3 Ma (Spray & Dunning, 1991). A c. 18 m.y. duration between subduction initiation and formation of the Unst Ophiolite and its boninites can be explained by ophiolite development in an infant arc environment which subsequently became the fore-arc after initiation of steady-state normal oceanic subduction. Spreading rates within the Unst Ophiolite precursor are likely to have been low to moderate, as documented by the relatively low volume of sheeted dykes. Crystallisation of the Unst Ophiolite protolith is immediately prior to development of the metamorphic sole (484 ± 4 Ma; this paper) suggesting that they formed as part of the same subduction system. Such a continuum of magmatism associated with sea-floor spreading events followed by obduction initiation within a few m.y. is documented by other SSZ ophiolites (e.g. Semail Ophiolite; Styles et al., 2006). Continued subduction roll-back resulted in development of detachment faults, rotation of the oceanic lithosphere, and exhumation of mantle and lower crust on the sea floor, accompanied by deposition of the sedimentary protoliths of the Muness Phyllites. We agree with Spray (1988) that arc-continent collision and ophiolite obduction onto the Laurentian passive margin was probably initiated by c. 480 Ma, as constrained by our new zircon age for subduction initiation in the Unst Ophiolite. There is a c. 20 m.y. gap between formation of the metamorphic sole to the ophiolite and peak Barrovian metamorphism in the footwall Dalradian Supergroup metasediments in Shetland at c. 462 Ma (Cutts et al. 2011). This confirms the notion of Chew et al (2010) and Cutts et al (2011) that peak P-T conditions recorded in Dalradian pelites are more likely to have been caused by Grampian orogenic collisional thickening and convective heat transfer within this over-thickened crust, rather than being directly related to burial following ophiolite obduction.

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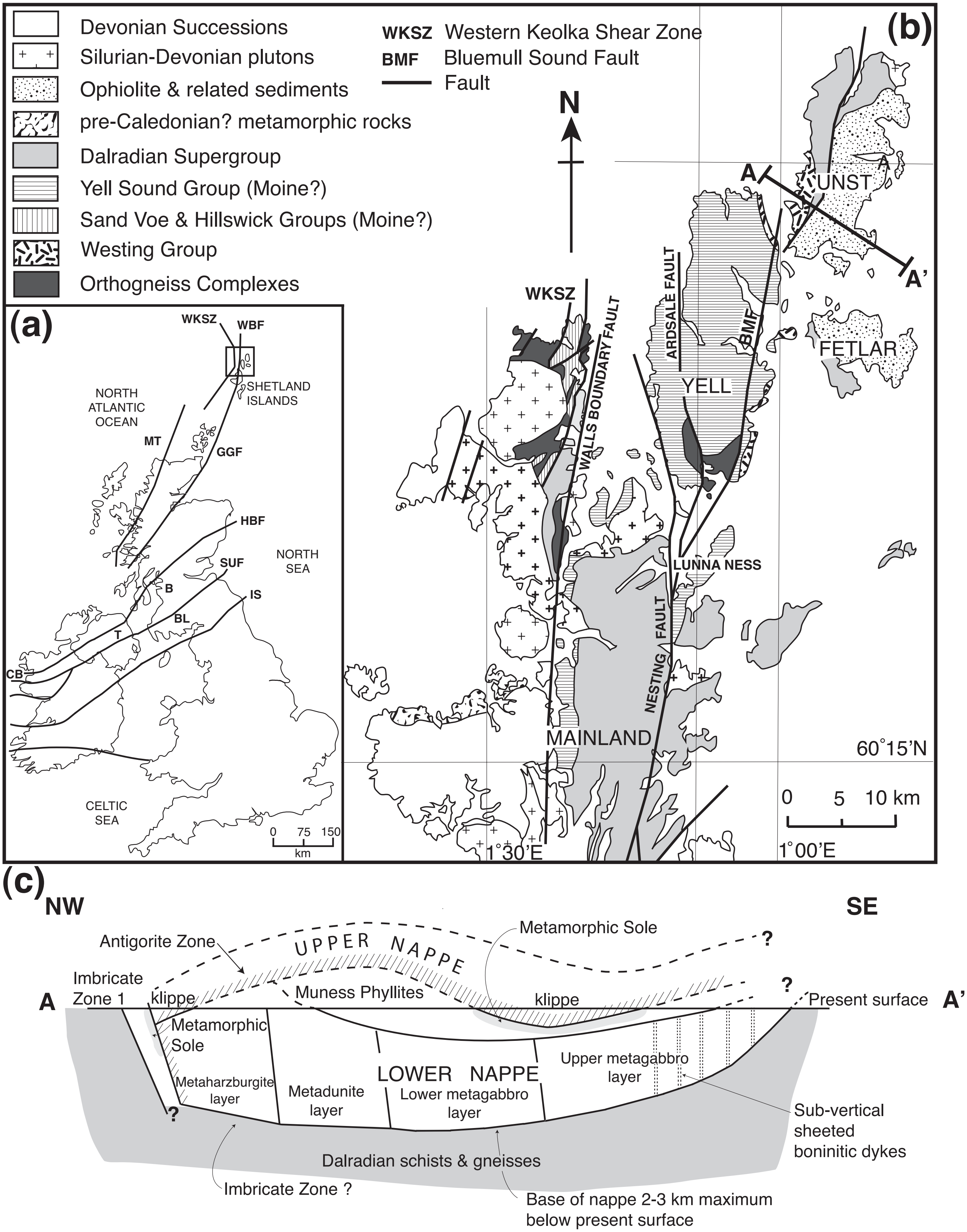
Figure captions

Fig. 1. (a) Map showing the location of the Shetland Islands in the British Isles: box indicates the location of (b). Abbreviations showing locations of other ophiolites: CB, Clew Bay; T, Tyrone; BL, Ballantrae; B, Bute. Major faults: IS, Iapetus Suture; SUF, Southern Uplands Fault; HBF, Highland Boundary Fault; GGF, Great Glen Fault; MT, Moine Thrust; WBF, Walls Boundary Fault; WKSZ, Wester Keolka Shear Zone. (b) Simplified geological map of the Shetland Islands ((a) and (b) modified from Cutts et al. 2011). (c) Diagrammatic cross-section of the ophiolite nappes on Unst and Fetlar (modified from Flinn 2001).

Fig 2. U-Pb plots for zircons from sample 11-SH-10 (a) Concordia plot. (b) weighted mean $^{206}Pb/^{238}U$ age plot.



Devonian Successions
Silurian-Devonian plutons
Ophiolite & related sediments
pre-Caledonian? metamorphic rocks
Dalradian Supergroup
Yell Sound Group (Moine?)
Sand Voe & Hillswick Groups (Moine?)
Westing Group
Orthogneiss Complexes
WKSZ Western Keolka Shear Zone
BMF Bluemull Sound Fault
Fault
(a)
WKSZ
WBF
SHETLAND ISLANDS
NORTH ATLANTIC OCEAN
MT
GGF
HBF
SUF
NORTH SEA
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(b)
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ARDSALE FAULT
BMF
WKSZ
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FETLAR
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LUNNA NESS
NESTING FAULT
MAINLAND
60°15'N
0 5 10 km
1°30'E
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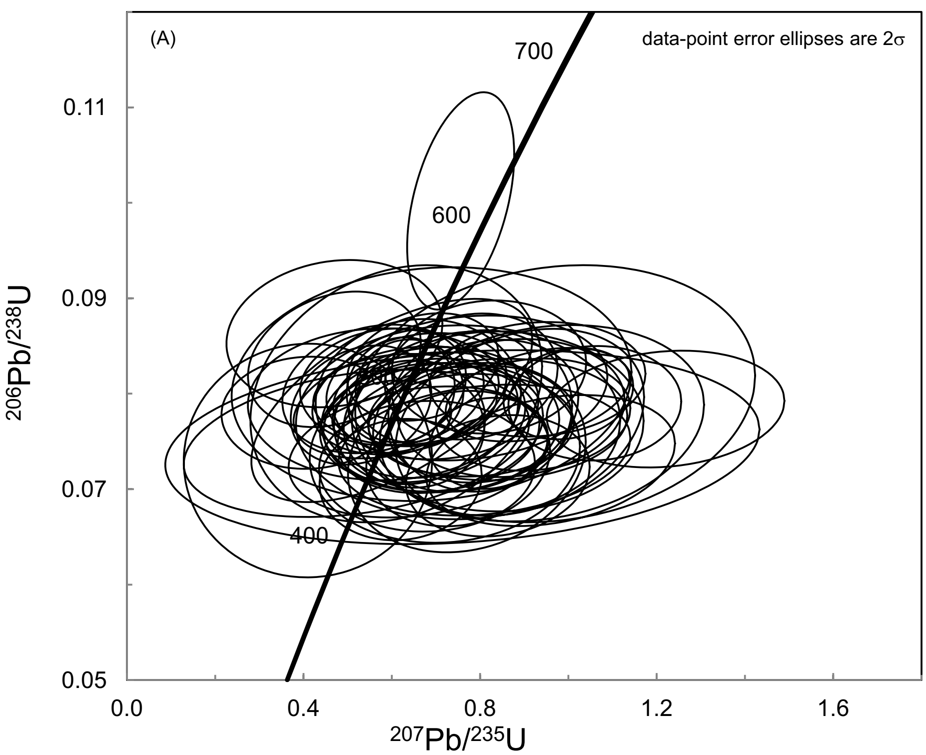
(c)
NW
SE
Metamorphic Sole
Antigorite Zone
UPPER NAPPE
?
Imbricate Zone 1
klippe
Muness Phyllites
klippe
Present surface
A
A'
Metamorphic Sole
Upper metagabbro layer
LOWER NAPPE
Metaharzburgite layer
Metadunite layer
Lower metagabbro layer
?
Sub-vertical sheeted boninitic dykes
Dalradian schists & gneisses
Imbricate Zone ?
Base of nappe 2-3 km maximum below present surface



(A)
data-point error ellipses are 2σ
700
600
400
0.11
0.09
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0.0
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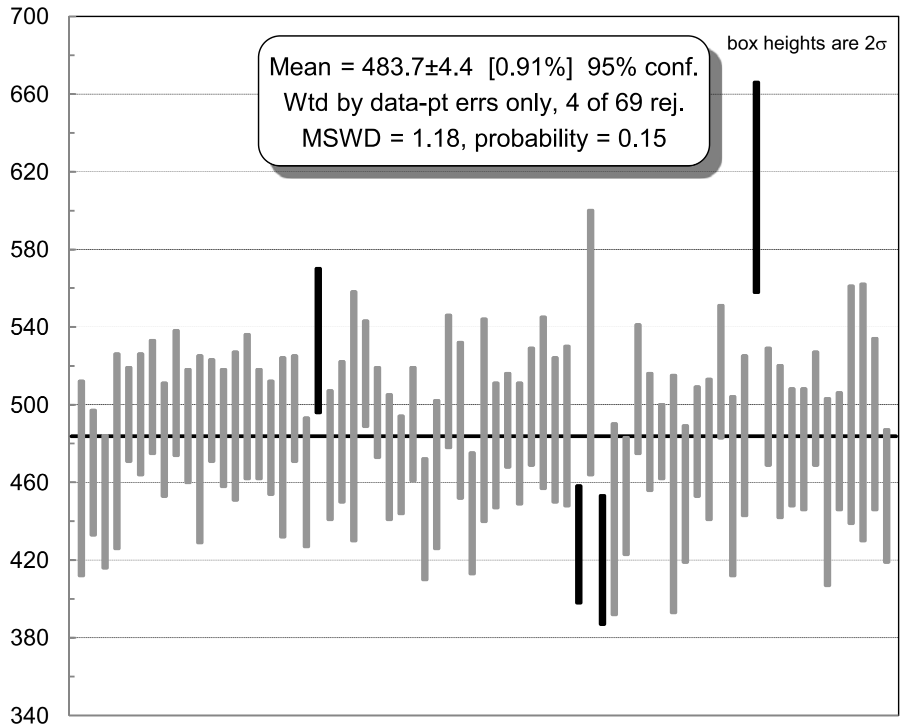
1.2
1.6
$^{206}Pb/^{238}U$
$^{207}Pb/^{235}U$


700
660
620
580
540
500
460
420
380
340
box heights are 2σ
Mean = 483.7±4.4 [0.91%] 95% conf.
Wtd by data-pt errs only, 4 of 69 rej.
MSWD = 1.18, probability = 0.15