

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

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7 **Abstract**

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

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

18 Ophiolites represent slices of oceanic-type lithosphere that have been tectonically
19 incorporated into continental margins in convergent plate settings (e.g. Dewey & Bird 1970;
20 Coleman 1971). Many ophiolites were formed in supra-subduction zone (SSZ) arc-forearc
21 settings shortly prior to orogenesis and then thrust (obducted) onto a colliding passive
22 margin (e.g. Davies & Jaques 1984; Searle & Cox 1999). The metamorphic soles present
23 beneath many ophiolites are thought to have resulted from high-temperature (\pm pressure)
24 metamorphism of a subducting oceanic slab beneath hot sub-ophiolitic mantle and later
25 accretion onto the base of the ophiolite during obduction (e.g. Dewey & Casey 2013 and
26 references therein). One of the most significant findings of the last 20 years of research into
27 modern ocean basins has been the identification of oceanic core complexes that expose
28 upper mantle and lower crustal rocks on the seafloor (e.g. Cannat 1993; Tucholke & Lin
29 1994). Although not recorded in many modern fore-arc settings, they have been identified
30 in some SSZ ophiolites (e.g. Suhr & Cawood 2001; Tremblay et al. 2009), thus representing
31 important opportunities for direct study of oceanic core complexes which are otherwise
32 difficult to access.

33 The Unst Ophiolite in the Shetland Islands (Fig 1) is the northernmost and best
34 exposed of a chain of SSZ ophiolites within the Laurentian Caledonides of Scotland and
35 Ireland (e.g. Williams & Smyth 1973; Flinn 1999 and references therein). Other examples
36 include the Bute, Ballantrae and Tyrone ophiolites and the Clew Bay Complex (Fig 1a). These
37 ophiolites were emplaced during the early to mid-Ordovician Grampian orogenic event

38 which records the initial stages in closure of the Iapetus Ocean (Chew et al. 2010). The
39 collision of a juvenile oceanic arc with the Laurentian passive margin resulted in NW-
40 directed ophiolite obduction (present reference frame) with regional deformation and
41 metamorphism of footwall metasedimentary successions (e.g. Dewey & Shackleton 1984;
42 Chew et al. 2010). The formation and subsequent obduction of the Unst Ophiolite is
43 currently bracketed between its magmatic age of 492 ± 3 Ma (U-Pb zircon, Spray & Dunning
44 1991) and K-Ar mineral ages obtained from the metamorphic sole which range from 479 ± 6
45 Ma to 465 ± 6 Ma (Spray 1988). In this paper we: 1) report the results of U-Pb dating of
46 metamorphic zircon from the sole of the ophiolite, thus placing a lower limit on the timing
47 of obduction, and 2) re-evaluate published geological data from the ophiolite to show that
48 these are consistent with an origin as a SSZ oceanic core complex with limited magma
49 supply.

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

63 The lower nappe consists of three steeply-inclined to sub-vertical layers with a total
64 maximum thickness of c. 7 km measured normal to lithological boundaries (Fig 1c; Flinn
65 1999, 2001). From northwest (structurally lowest) to southeast (highest) these comprise
66 metaharzburgite, metadunite (both extensively serpentinized) and metagabbro (Fig 1c),
67 which have been affected by greenschist facies metamorphism. A plagiogranite within a
68 metagabbro yielded a U-Pb zircon age of 492 ± 3 Ma (Spray & Dunning 1991). Uppermost
69 parts of the metagabbro in SE Unst are intruded by sub-parallel mafic dykes, separated by
70 narrow screens of host metagabbro (Prichard 1985; Flinn 1999). The dykes have been
71 interpreted as the base of a sheeted dyke complex (Prichard 1985; Spray 1988). The
72 boninitic chemistry of some of these dykes provides evidence that the ophiolite formed in a
73 SSZ setting (Prichard & Lord 1988; Spray 1988; Flinn 2001). Trace element discrimination
74 diagrams indicate a range of volcanic arc basalt/island arc-like compositions (Flinn 2001).
75 However, these dykes depart from classic sheeted dykes described from many other
76 ophiolites because they never form >50% of the rock by volume in any area and are nearly

77 *parallel* to the ophiolite layering than normal to it, so at present they occur as sub-vertical
78 sheets (Flinn 2001).

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

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

102 **Sample descriptions** To further constrain the timing of formation of the metamorphic sole,
103 we investigated two samples: 11-SH-10 (collected from beneath the lower nappe on Unst at
104 HP 5635 0067) and 11-SH-12 (collected from beneath the upper nappe on Fetlar at HU 6450
105 9206). Both samples are amphibolites, with 11-SH-10 being medium-grained and 11-SH-12
106 being medium to fine grained. Both amphibolites possess a well-developed foliation defined
107 by alternating mafic and felsic layers. Mafic layers are dominated by aligned grains of
108 hornblende, which in the case of sample 11-SH-10 wrap skeletal garnets up to 5 mm in
109 diameter. Titanite and apatite occur as visible accessory minerals. Felsic layers are
110 dominantly composed of sericitised plagioclase with minor quartz. There is evidence for
111 widespread replacement of hornblende by actinolite and biotite, and garnet (where
112 present) by chlorite. Scattered grains of secondary epidote are also common.

113 **U-Pb zircon data: results and interpretation.** Cathodoluminescence (CL) images of zircon
114 grains show sector and fir-tree zoning patterns and large parts of grains which are
115 homogeneous in CL, all of which is consistent with a single phase of metamorphic growth
116 (Corfu et al. 2003), which most likely occurred during upper amphibolite facies P-T

117 conditions as previously recognised from the early mineral assemblage (Spray 1988) . No
118 inherited cores and no igneous zircon were evident, compatible with a MORB-type protolith
119 (Spray 1988; Flinn 2001). Representative zircons from both samples were analysed by LA-
120 ICP-MS following the operating conditions of Crowley et al. (2014 see supplementary files
121 for data table) and were found to have low U concentrations (generally <3ppm for sample
122 11-SH-10 and <0.3ppm for sample 11-SH-12). 69 zircon analyses from 11-SH-10 are
123 concordant to near-concordant (Fig 2) and do not appear to have suffered any significant
124 Pb-loss. A weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age calculation from these data returns a date of 483.7
125 ± 4.4 Ma (MSWD=1.18), interpreted as the age of formation of the metamorphic zircon.
126 Despite a large number of analyses, only one age determination was possible from 11-SH-
127 12. This grain was found to have relatively elevated U concentration (c. 5ppm) compared to
128 other zircons from the same sample and yielded a $^{206}\text{Pb}/^{238}\text{U}$ age of c. 482 ± 18 Ma (i.e.
129 within error of the age determination for sample 11-SH-10).

130 **An oceanic core complex model for the Unst Ophiolite** Oceanic core complexes are known
131 to develop in conjunction with some intermediate to slow spreading centres and represent
132 exhumed lower crustal and upper mantle rocks in the footwalls of extensional detachments
133 (Whitney *et al.* 2013; Platt *et al. in press*). They typically exhibit a dome-shaped massif,
134 which may display considerable topographic relief relative to the surrounding sea-floor.
135 Oceanic core complexes are usually cored by gabbro, or gabbro and peridotite and do not
136 exhibit the classic Penrose-type ophiolite sequence (Miranda & Dilek 2010). They have been
137 described from segments of the Australian-Antarctic Discordance, Caribbean–North
138 American Ridge, Mid-Atlantic Ridge, Southwest Indian Ridge and from several back-arc
139 spreading centres (Whitney *et al.* 2013 and references therein). Several oceanic core
140 complexes occur in the inner-bend of ridge-transform intersections, so there is likely an
141 interplay of magmatic and tectonic processes, the mutual rates of which are critical to
142 initiating accelerated extension in magma-poor areas of divergent zones (Whitney *et al.*,
143 2013). Slow spreading rates of $<55 \text{ mm yr}^{-1}$ may result in episodic and spatially variable
144 magma supply, which in turn facilitates an increase in tectonic partitioning, accelerated
145 lithospheric thinning and denudation of oceanic crust (Miranda & Dilek, 2010). Furthermore,
146 the rate of magma supply (i.e. not too rapid and not too slow) appears to be a critical factor
147 in the development of oceanic core complexes (Tucholke *et al.* 2008). From a structural
148 perspective, the presence of pre-existing inward-dipping normal fault systems, coupled with
149 lithospheric thinning results in a passive rotation or flexure of faults (Morris *et al.* 2009),
150 resulting in the formation of low angle detachments. Tucholke *et al.* (2008) state that if
151 dykes and plutons collectively account for 50% or more of the total extension, then
152 detachment faults initiated at spreading centres may accumulate large displacements.

153 Any model for the formation of the Unst Ophiolite has to account for the steep
154 orientation of the internal layering relative to the assumed gently-dipping basal thrust (Fig
155 1c), the intrusion of the boninitic dykes sub-parallel to the steep layering within host upper
156 gabbro and their absence from underlying units. In contrast to Spray (1988) who envisaged

157 formation within an SSZ setting *sensu stricto*, Flinn (2001) suggested that it was derived
158 from a subducting oceanic slab. In this latter interpretation, the SSZ chemistry was derived
159 from flushing the subduction zone with fluids from the dehydrating descending slab. These
160 fluids promoted partial melting of the overlying wedge, forming basaltic melts that were
161 intruded into the upper gabbro as dykes from above and to the side. Flinn (2001) envisaged
162 that during arc-continent collision, a thrust that initiated in the overlying mantle wedge cut
163 down through the 'stratigraphy' of the subducting plate, then more parallel to it, eventually
164 acting as the obduction thrust for the ophiolite as it was thrust onto the Laurentian margin.
165 The Muness Phyllites were thought to have been deposited on the ophiolite *after* it was
166 obducted.

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

192

193 **Discussion and implications** The new interpretations and data presented here support a
194 new tectonic model for the Unst Ophiolite. The timing of initiation of oceanward-subduction
195 is poorly constrained but is likely to have occurred by c. 510 Ma (Chew et al. 2010), followed
196 by subduction roll-back and development of the precursor to the SSZ ophiolite at 492 ± 3
197 Ma (Spray & Dunning, 1991). A c. 18 m.y. duration between subduction initiation and

198 formation of the Unst Ophiolite and its boninites can be explained by ophiolite development
199 in an infant arc environment which subsequently became the fore-arc after initiation of
200 steady-state normal oceanic subduction. Spreading rates within the Unst Ophiolite
201 precursor are likely to have been low to moderate, as documented by the relatively low
202 volume of sheeted dykes. Crystallisation of the Unst Ophiolite protolith is immediately prior
203 to development of the metamorphic sole (484 ± 4 Ma; this paper) suggesting that they
204 formed as part of the same subduction system. Such a continuum of magmatism associated
205 with sea-floor spreading events followed by obduction initiation within a few m.y. is
206 documented by other SSZ ophiolites (e.g. Semail Ophiolite; Styles et al., 2006). Continued
207 subduction roll-back resulted in development of detachment faults, rotation of the oceanic
208 lithosphere, and exhumation of mantle and lower crust on the sea floor, accompanied by
209 deposition of the sedimentary protoliths of the Muness Phyllites. We agree with Spray
210 (1988) that arc-continent collision and ophiolite obduction onto the Laurentian passive
211 margin was probably initiated by c. 480 Ma, as constrained by our new zircon age for
212 subduction initiation in the Unst Ophiolite. There is a c. 20 m.y. gap between formation of
213 the metamorphic sole to the ophiolite and peak Barrovian metamorphism in the footwall
214 Dalradian Supergroup metasediments in Shetland at c. 462 Ma (Cutts et al. 2011). This
215 confirms the notion of Chew et al (2010) and Cutts et al (2011) that peak P-T conditions
216 recorded in Dalradian pelites are more likely to have been caused by Grampian orogenic
217 collisional thickening and convective heat transfer within this over-thickened crust, rather
218 than being directly related to burial following ophiolite obduction.

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301 **Figure captions**

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

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310 Fig 2. U-Pb plots for zircons from sample 11-SH-10 (a) Concordia plot. (b) weighted mean
311 $^{206}\text{Pb}/^{238}\text{U}$ age plot.





