

## Volcanic rocks in the Devonian Solund Basin, Western Norway: large landslides of Silurian (439 Ma) rhyolites

EBBE H. HARTZ<sup>1,2</sup>, MARK W. MARTIN<sup>3</sup>, ARILD ANDRESEN<sup>1</sup> & TORGEIR B. ANDERSEN<sup>1</sup>

<sup>1</sup>Department of Geology, University of Oslo, 0316 Oslo, Norway

<sup>2</sup>Present address: Geological Survey of Norway, Pb.5348 Majorstua, 0304 Oslo, Norway (e-mail: ebbe.hartz@ngu.no)

<sup>3</sup>Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, USA

**Abstract:** The Devonian Solund Basin in SW Norway is the only Old Red Sandstone basin in the Scandinavian Caledonides from which volcanic rocks have been described. These rhyolitic lavas yield a concordant ID TIMS U–Pb zircon age of  $439 \pm 1$  Ma (Early Silurian) interpreted as dating the crystallization of the rhyolite, and are thus too old to be syndepositional with the Devonian sedimentary rocks. This new age constraint, combined with field observations, suggests that the volcanic rocks form part of a 30 km<sup>2</sup> landslide of rhyolite, granite, gabbro and metasedimentary rocks, rather than extension related volcanism as previously suggested. The landslides show an inverted stratigraphy with respect to the nearby Solund–Stavfjord Ophiolite Complex and its overlying volcanic arc, which was formed immediately prior to obduction of Caledonian exotic terranes onto Baltica. The dated rocks thus appear to record the youngest pre-collisional volcanism in the Scandinavian Caledonides. Tectonic models involving Devonian Old Red volcanic rocks in the Scandinavian Caledonides must thus be revised.

**Keywords:** Devonian, Silurian, U–Pb, western Norway, landslides

The rhyolites in the Devonian Solund Basin (Kolderup 1926) are unique in the Old Red Sandstone basins of Scandinavia (Fig. 1a), and have been invoked in various tectonic models involving thrusting (Roberts 1983), rifting (Nielsen 1968; Furnes & Lippard 1983); strike slip faulting (Steel *et al.* 1985), orogenic collapse (Norton 1983, 1986; Andersen & Jamtveit 1990) and collision of a volcanic arc with Baltica (Sturt & Braathen 2001). All of these models have assumed a late Early Devonian age for the basal section of the basin based on plant fossils from the nearby island of Bulandet (e.g. Kolderup 1915) (Fig. 1b). The rhyolitic lavas at Hersvik occur at the base of large landslides (Bryhni 1976), and are thus here interpreted as syndepositional lahars (Furnes & Lippard 1983), involving the overlying slide units of metasediments and Devonian gabbro and granite (Norton 1983).

Precise dating of the rhyolite combined with field studies would thus address the on-going discussion regarding the relative age of the rhyolites versus the age of the sedimentary rocks, timing of landslide activity, and thereby clarify the tectonic significance of the volcanism.

### Geological setting and field relationships

Silurian to Devonian continental collision of Baltica and Laurentia resulted in stacking and deep subduction of the Baltic crust and cover and emplacement of composite terranes onto the margin of Baltica (Andersen *et al.* 1990). The final closure of Iapetus formed, deformed, and accreted late Ordovician to Silurian basins and volcanic arcs, including including the  $443 \pm 3$  Ma Solund–Stavfjord Ophiolite Complex, the  $440 \pm 5$  Ma subduction related granites intruding a mélange at Bremanger, the Sunnfjord Mélange, and the Herland Group (Dunning & Pedersen 1988; Furnes *et al.* 1990; Andersen *et al.* 1990; Hansen *et al.* 2001) (Fig. 1b, c).

The deformed Solund–Stavfjord Ophiolite Complex and its volcano-sedimentary cover are unconformably overlain by 5.2 km of Devonian boulder conglomerates and minor sandstones in the 800 km<sup>2</sup> Solund Basin (Kolderup 1926; Nielsen 1968) (Fig. 2a). The basin fill was derived from the depositional substrate composed of gabbro/diorite, meta-basalt, greywackes, quartzites and rhyolite derived from the west, and granitoids primarily derived from the east (Nielsen 1968). Sedimentary rocks in the Devonian basin are locally folded and cleaved (Nielsen 1968; Sturt & Braathen 2001) and metamorphosed to 230–320 °C (Svensen *et al.* 2001).

The Devonian rocks are, however, generally less deformed and rest with a profound unconformity on their substrate. Mafic and felsic volcanic rocks associated with fine-grained gabbros and diorites within the Devonian sedimentary rocks near Hersvik were first reported by Kolderup (1926), who inferred that these were emplaced as thrust-sheets of the substrate into the basin sediments (Fig. 2a, b). Nielsen (1968), however, found no evidence for major faults and suggested that the igneous rocks had been intruded into and extruded onto the sedimentary units contemporaneously with basin development. Subsequently, however, Bryhni (1976) argued that the large bodies of brecciated gabbro were deposited as debris flow deposits. Bryhni's (1976) main arguments were that the breccia is not entirely monomict but includes clasts of diorite, granite, rhyolite and metasediments, and that discontinuous beds of sandstone occurred at some localities. In an unpublished report, Norton (1983), mapped two exotic bodies in the Hersvik–Hagevatn area (Fig. 2b). The largest of these, here referred to as the Hagevatnet landslide, is an irregular body of metasediments, intruded successively by gabbro/diorite, granite and rhyolite (Fig. 3a). The body reaches a thickness of 80 m at Hagevatnet and thins towards the south (Norton 1983) (Fig. 2b). East of Hagevatnet the landslide rests on a matrix supported conglomerate, which consists of

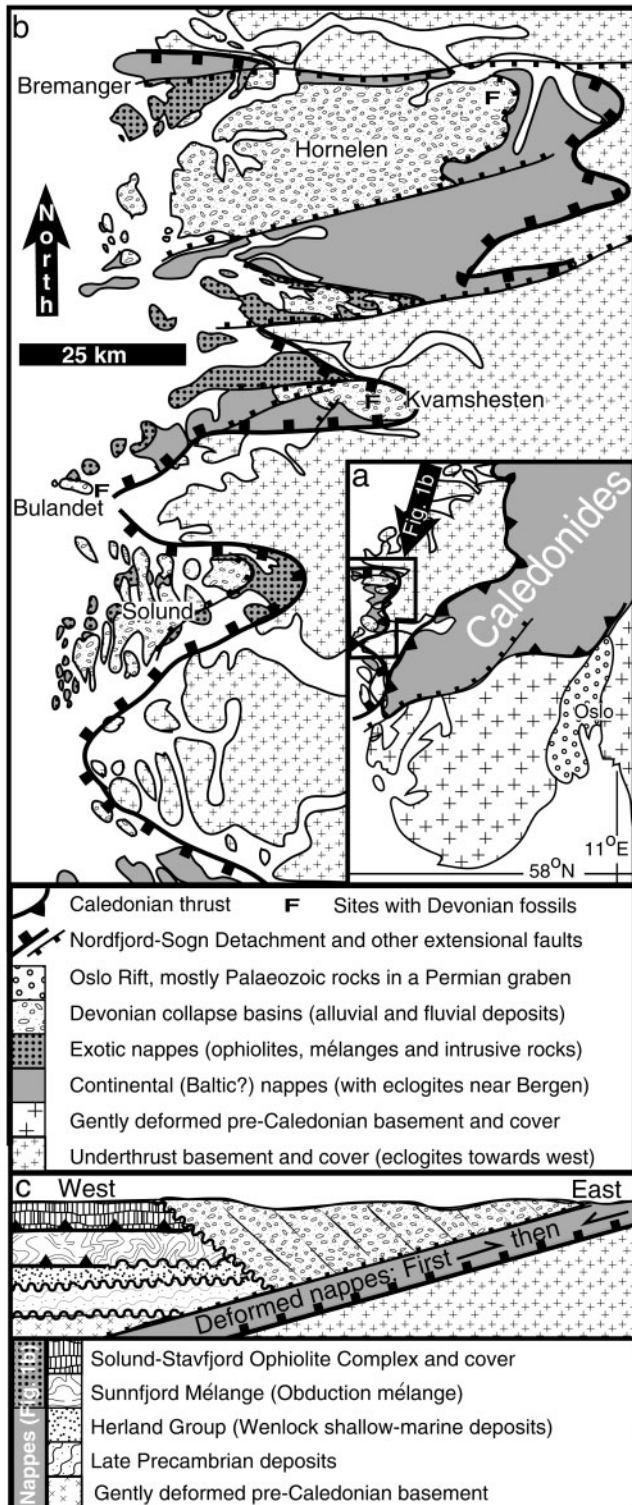


Fig. 1. (a) Index map showing the Scandinavian Caledonides in Southern Norway. (b) Simplified tectonic map of western Norway, showing the Nordfjord-Sogn detachment that separates late Caledonian high-grade gneisses from greenschist facies nappes with overlying Devonian continental sedimentary rocks. (c) Schematic cross-section illustrating the general tectonic setting of the Devonian basins.

unstratified siltstone with igneous clasts, and grades upwards into the clast-supported megabreccia of the slide. Siltstone dykes from the substrata penetrate locally up to 1 m into the base of the gabbro megabreccia. Hagevatnet landslide is one of at least four bodies occurring at different stratigraphic levels. These are all eroded towards the north and dip below conglomerates towards the south (Fig. 2a). Most bodies consist of monomict gabbro breccia resting directly on polymict conglomerate. In total the bodies are exposed over *c.* 30 km<sup>2</sup>. As these bodies are partially eroded and covered, their presently exposed volume of *ca.* 1.5 km<sup>3</sup> is probably a significant underestimate of the original volume. Large (700 m wide and 50 m thick) bodies of brecciated basic and felsic tuff, rhyolite, gabbro and metasediments also occur near Kråkevåg along the southern margin of the Solund Basin (Fig. 2b) (Norton 1983). Our reconnaissance studies corroborates Norton's (1983) observations that these rocks generally compare to the Hagevatnet landslide, but they occur at a much higher stratigraphic level, near the top of the Solund Basin stratigraphy.

Norton (1983) suggested that the Hagevatnet landslide and similar bodies consist of metasediments that were intruded by Devonian gabbros and granites prior to rapid exhumation, erosion and emplacement as landslides involved in lahars. The rhyolitic layer at the base of the bodies, were suggested to be evidence of the volcanogenic nature of the landslides. The thick and well-exposed rhyolitic lavas south of Hersvik are interpreted as the best evidence of such Devonian volcanism (Nielsen 1968; Bryhni 1976; Norton 1983; Furnes & Lippard 1983), and thus the target for this study. This body, here named the Kvernhusdalen landslide, occurs stratigraphically below the Hagevatnet landslide (Fig. 2b). The field relationships and geochemistry is described in detail by Furnes & Lippard (1983), who differentiated a lower body of brecciated flow-banded and 'flow-folded' trachytic lava, interlayered with microgabbro. The lower body is separated by a thin sedimentary breccia from the upper body consisting of brecciated, massive and flow banded to 'flow-folded' rhyolite and gabbro.

The traverse studied in this paper extends southwards from the small lake in Kvernhusdalen (Fig. 2b), and differs from the section along the fjord studied by Furnes & Lippard (1983). The landslide rests on a polymict conglomerate and the lower 2 m of the landslide is intensely brecciated (Fig. 3b). Along the measured profile the landslide is 60 m in thickness. It consists of a lower 35 m thick rhyolitic (Fig. 3c) overlain by 2 m of brecciated rhyolite and 23 m of monomict variably brecciated gabbro. The contact between the landslide and the overlying boulder conglomerates marks an irregular surface with up to 10 m of topography. Fractures in the landslide are filled with sand penetrating up to 2 m into the breccias (Fig. 3e). The boulder conglomerates both above and below the body is polymict and contains clasts of greenstone, metadiorite, rhyolite, gabbro, granite and metasediment (Fig. 3f).

Both gabbros and rhyolites are brecciated; however, the nature of brecciation is not always easily recognized in the field, apart from the obvious depositional sandfilled breccias at the base and top of the body. The internal brecciation could in theory result from post-volcanic deposition (incoherent debris flow) or as a syn-volcanic autobrecciation, as typical for viscous rhyolitic flows (lahars) as suggested by Furnes & Lippard (1983) and Norton (1983). The latter conclusion is supported by three layers 5–20 cm thick of rhyolite that can be traced undisturbed for tens of meters across brecciated

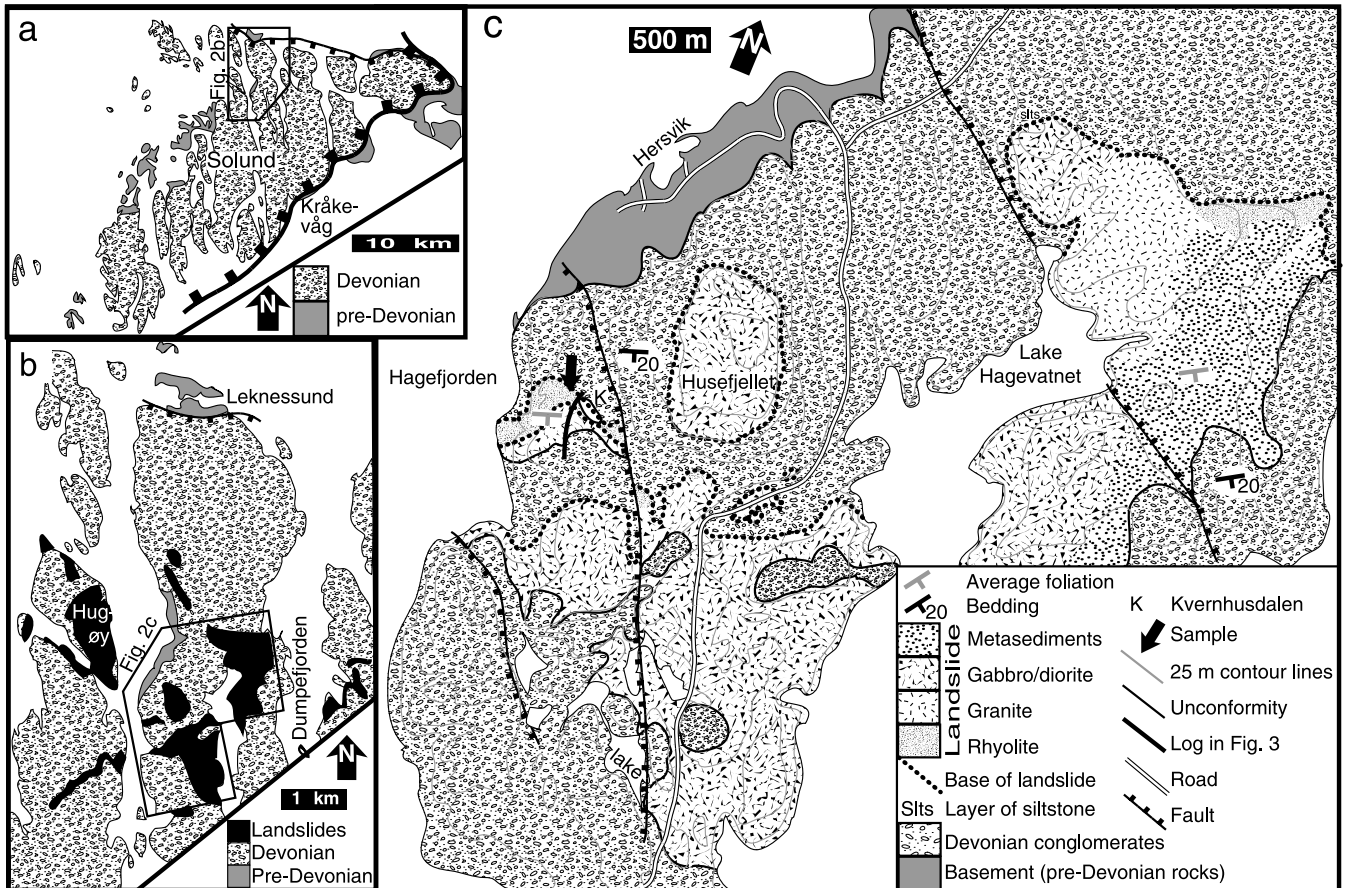


Fig. 2. (a) Map of the Solund Basin. (b) Map of the northernmost Solund Basin, showing the distribution of metasediments, gabbro, granite and rhyolite bodies. All of these are included in the Herveik landslides, based on the conclusions from this study. (c) Detailed map of the Herveik–Hagevatnet area showing the internal architecture of the Herveik landslides. The maps are modified from Nielsen (1968) and Norton (1983).

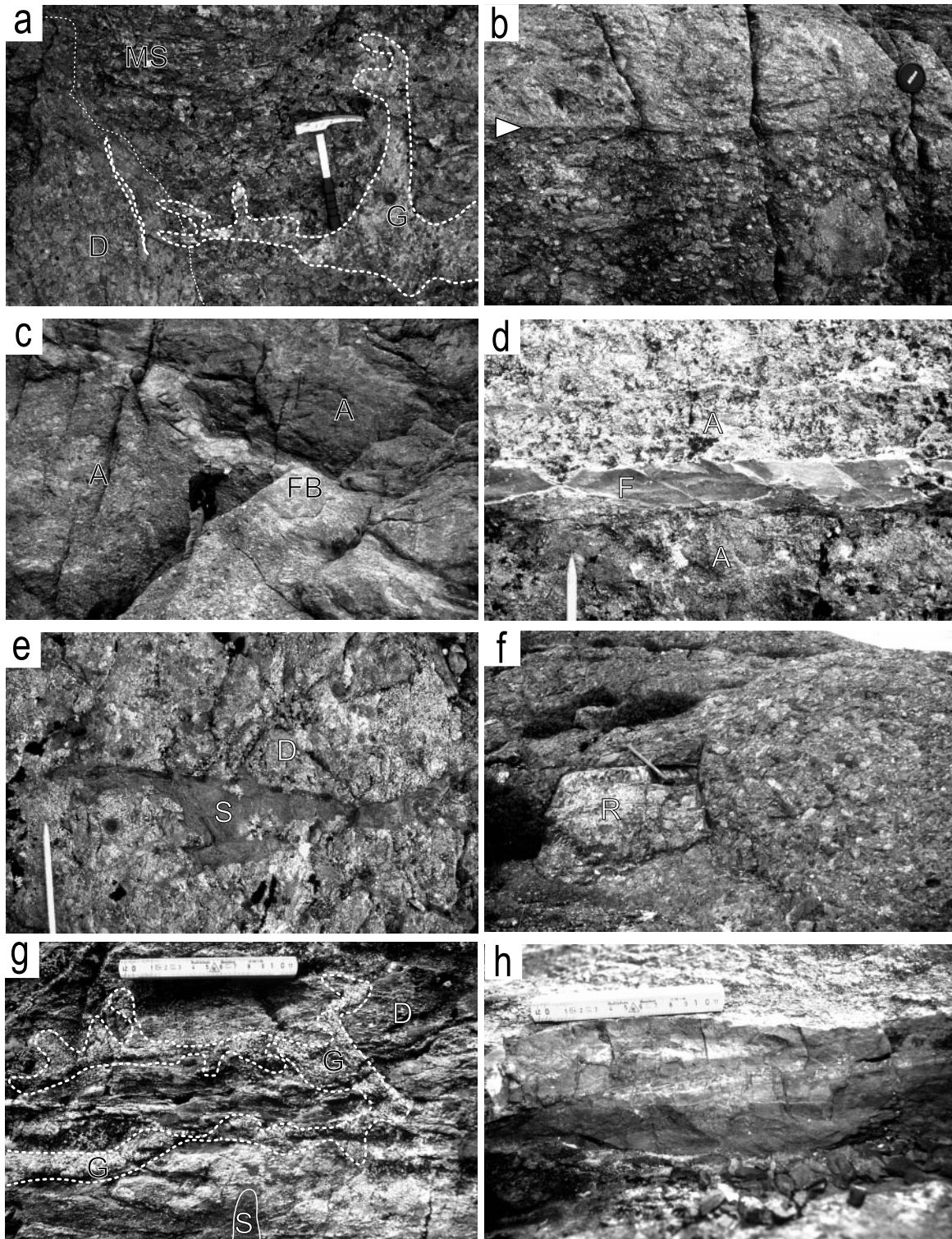
rhyolites (Fig. 3d), thereby proving that the brecciation is at least partly a synvolcanic feature in an otherwise coherent block. This observation is further corroborated by abundant felsic net veining in the brecciated gabbros/diorites (Fig. 3g) confirming that parts of the brecciation is synmagmatic. These abundant felsic net veins furthermore confirm a close primary relationship between gabbroic and granitic rocks (Furnes & Lippard 1983), and is consistent with the intrusive relationships observed within the Hagevatnet landslide (Figs 3a, 2b).

Collectively these field-relationships suggest an exotic (landslide) origin of the igneous rocks. We could not, however, unequivocally determine, even after close examination of critical localities in the field, if the rhyolitic lava and overlying rocks was emplaced as synvolcanic lahars (Nielsen 1968; Norton 1983; Furnes & Lippard 1983), or as post-volcanic landslides.

#### Sample description and U–Pb age data

The rhyolite sampled for this study was collected from the central part of the Herveik Lavas, 50 m SW of the small lake in Kvernhusdalen. In order to prevent sampling of exotic zircons unrelated to the rhyolite, the autobrecciated parts of the flow were avoided, and the second thin (5–20 cm) undisturbed layer of flow-banded rhyolite described above was sampled (Fig. 3h). Zircons were extracted using standard separation

techniques from 10 kg of sample. Five of these zircons were studied using backscattered electron imagery with a scanning electron microscope, and, showed regular oscillatory zoning interpreted as a primary igneous feature. The remaining zircons were air-abraded and analysed by isotope dilution thermal ionization mass-spectrometry (IDTIMS) at MIT; each analysis contained from 1 to 5 zircon crystals (Table 1). The U–Pb systematics suggests that the zircons display both inheritance and Pb loss. The eight analyses define three groups; z1–z3 are discordant and contain an inherited component, z4 and z5, composed of one or two crystals are concordant near 439 Ma; and z6–z8 are slightly discordant and are probably affected by minor Pb loss (Fig. 4). The crystallization age of the rhyolite is interpreted to be represented by the weighted mean of the  $^{206}\text{Pb}/^{238}\text{U}$  dates of z4 and z5 which gives an age of  $439.0 \pm 1.0$  Ma (MSWD = 0.3). Zircon fractions z4, z5, z7 and z8 fall on line forced through 0 Ma, giving an upper intercept of  $439 \pm 2$  Ma (MSWD 0.04) indicating that z7 and z8 are controlled by recent lead loss, whereas fraction z6 probably have both a minor inherited component, and recent lead loss. Fractions z2 and z3 plot on a discordia between *c.* 439 and *c.* 1052 Ma, and a single zircon (z1) have a upper intercept of *c.* 1638 Ma when regressed through 439 Ma. Although the data are limited and highly discordant it is worth noticing that both *c.* 1052 and 1638 Ma. are well known ages in western Norway (Skår *et al.* 1994; Bingen & van Breemen



**Fig. 3.** Photographs of rocks associated with the Hersvik landslides. (a) Plagiogranite (G) veins cutting diorite (D) that intrude metasediments (M). Outcrop at the base of Hagevatnet landslide, east of Hagevatnet. (b) Flow-banded rhyolite at the base of landslide (marked by triangle), resting on polymict breccia. (c) Two distinct layers of autobrecciated rhyolite (A) are separated by a c. 1 m thick layer of flowbanded rhyolite (FB). Central part of Kvernhusdalen landslide. (d) Two distinct layers of autobrecciated rhyolite (A) are separated by a thin layer of flow-banded rhyolite (F), that can be traced undisturbed for tens of metres, thereby showing that the brecciation is synmagmatic. Notice that the upper autobrecciated layer is lighter (more felsic) than the lower layer, which together with the flowbanded rhyolites show that the layering represents a volcanic stratigraphy. (e) Sand (S) fills fractures into brecciated diorite (D) at the top of landslide. (f) Polymict sedimentary breccia with clasts of rhyolite (R), diorite and metasediments. (g) Plagiogranitic (G) veins in diorite (D), near the top of Kvernhusdalen landslide. The undeformed granitic veins partly follow the fracture pattern, and in part cut the diorite, illustrating that the brecciation is igneous. Minor cross-cutting fractures are filled with sand (S). (h) Dated rhyolite. This flowbanded layer can be followed across autobrecciated rhyolite for 40 m.

Table 1. U-Pb isotopic data for zircons

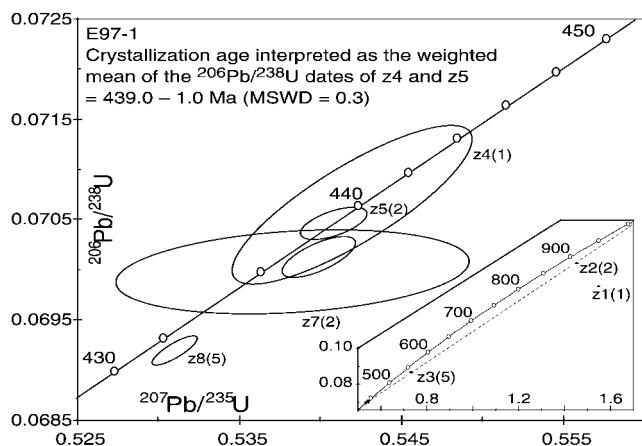
| Fractions(†)                   | Weight (µg) | Concentration |           |                      | Errors (% 2 σ)                       |                                      |                                      |       | Dates (Ma)                           |       |                                       |       | Corr. coef |                                     |                                     |                                      |
|--------------------------------|-------------|---------------|-----------|----------------------|--------------------------------------|--------------------------------------|--------------------------------------|-------|--------------------------------------|-------|---------------------------------------|-------|------------|-------------------------------------|-------------------------------------|--------------------------------------|
|                                |             | U (ppm)       | Pb* (ppm) | Pb <sup>s</sup> (pg) | <sup>206</sup> Pb/ <sup>204</sup> Pb | <sup>208</sup> Pb/ <sup>206</sup> Pb | <sup>206</sup> Pb/ <sup>238</sup> U† | % err | <sup>207</sup> Pb/ <sup>235</sup> U† | % err | <sup>207</sup> Pb/ <sup>206</sup> Pb† | % err |            | <sup>206</sup> Pb/ <sup>238</sup> U | <sup>207</sup> Pb/ <sup>235</sup> U | <sup>207</sup> Pb/ <sup>206</sup> Pb |
| <b>E97-1 (UTM 32VKN792886)</b> |             |               |           |                      |                                      |                                      |                                      |       |                                      |       |                                       |       |            |                                     |                                     |                                      |
| z1(1)                          | 2.6         | 1039.6        | 152.5     | 3.1                  | 7481.5                               | 0.191                                | 0.133685                             | (.11) | 1.5457                               | (.15) | 0.08386                               | (.10) | 808.9      | 948.8                               | 1289.2                              | 0.74                                 |
| z2(2)                          | 3.0         | 131.3         | 20.6      | 2.5                  | 1523.6                               | 0.176                                | 0.146460                             | (.17) | 1.4517                               | (.25) | 0.07189                               | (.17) | 881.1      | 910.6                               | 982.8                               | 0.72                                 |
| z3(5)                          | 2.9         | 350.5         | 34.6      | 4.9                  | 1165.0                               | 0.245                                | 0.086593                             | (.30) | 0.7324                               | (.39) | 0.06134                               | (.23) | 535.4      | 557.9                               | 651.1                               | 0.81                                 |
| z4(1)                          | 3.8         | 106.4         | 9.4       | 4.9                  | 394.6                                | 0.343                                | 0.070655                             | (.92) | 0.5420                               | (.11) | 0.05564                               | (.39) | 440.1      | 439.7                               | 437.9                               | 0.85                                 |
| z5(2)                          | 2.8         | 135.4         | 11.4      | 1.4                  | 1243.9                               | 0.332                                | 0.070463                             | (.18) | 0.5408                               | (.32) | 0.05567                               | (.25) | 438.9      | 439.0                               | 439.0                               | 0.62                                 |
| z6(4)                          | 2.9         | 150.6         | 12.2      | 2.4                  | 849.1                                | 0.282                                | 0.070126                             | (.24) | 0.5398                               | (.35) | 0.05583                               | (.24) | 436.9      | 438.3                               | 445.7                               | 0.74                                 |
| z7(2)                          | 3.2         | 87.2          | 7.1       | 2.8                  | 464.4                                | 0.294                                | 0.069981                             | (.50) | 0.5382                               | (.16) | 0.05578                               | (.14) | 436.0      | 437.3                               | 443.7                               | 0.47                                 |
| z8(5)                          | 2.5         | 447.8         | 36.1      | 2.6                  | 1941.4                               | 0.296                                | 0.069197                             | (.17) | 0.5310                               | (.20) | 0.05565                               | (.11) | 431.3      | 432.5                               | 438.6                               | 0.83                                 |

a \*Radiogenic Pb.

b † Number of zircon grains analysed. Sample weights are estimated by using a video monitor and are known to within 40%.

c ‡ Corrected for fractionation, spike, blank, and initial common Pb. Mass fractionation correction of 0.15‰/amu (atomic mass unit) was applied to single-collector Daly analyses.

d § Total common Pb in analyses. Total procedural blank for Pb ranged from 0.65 to 3.0 pg and < 0.5 pg for U. Blank isotopic composition: <sup>206</sup>Pb/<sup>204</sup>Pb = calculated by using the 19.10 ± 0.1, <sup>207</sup>Pb/<sup>204</sup>Pb = 15.71 ± 0.1, <sup>208</sup>Pb/<sup>204</sup>Pb = 38.65 ± 0.1. Age calculations based on the decay constants of (1)Steiger & Jäger (1971). (2) Common Pb correction model of Stacey & Kramers (1975) and the interpreted age of the sample.



**Fig. 4.** U–Pb concordia plot of analysed zircons from the Hersvik rhyolite, Solund Basin. Dashed line in inset concordia plot represents linear regression that includes z2, z3, z4, and z5 and has a lower intercept at  $440 \pm 2.5$  Ma and upper intercept of  $1056 \pm 6$  Ma (MSWD = 1.3).

1998), and thus corroborate that the lavas formed during obduction onto Baltica (Furnes *et al.* 1990)

### Discussion and tectonic implications of the U–Pb data

The crystallization age of the Hersvik rhyolite is interpreted to be  $439.0 \pm 1.0$  Ma which is Early Silurian (Llandovery). If the extrusion of the rhyolites are syndepositional with the alluvial conglomerates, then the lower part of the Solund Basin is Early Silurian in age, and not Devonian as suggested by fossils in the northern part of the basin (Fig. 1b). This interpretation is problematic for a number of reasons.

(1) There is no evidence of a structural or depositional discontinuity between the lava and the fossil-bearing strata (Nielsen 1968).

(2) The Solund–Stavfjord Ophiolite Complex has a protracted history postdating the dated  $443 \pm 3$  Ma pegmatitic diorite (Dunning & Pedersen 1988) as it was subjected to felsic intrusion, polyphase deformation, metamorphism and exhumation prior to deposition of the overlying Old Red deposits in the Solund Basin (Furnes 1974; Skjerlie & Furnes 1990). The age of the Hersvik rhyolite (439 Ma) is only slightly younger than a dated diorite in the Solund–Stavfjord Ophiolite Complex (*c.* 443 Ma), and it is not likely that intrusion, deformation, exhumation and erosion took place during 4 million years.

(3) There are no fossils in the substratum below the Solund Basin, but the Kvamshesten Devonian basin (Fig. 1b,c) rest on marine deposits of Wenlock age (429–424 Ma), generally suggest that the Old Red Sandstone basins are post-Wenlock in age.

Therefore, the 439 Ma age for the rhyolites in the Solund Basin requires an alternative explanation. Kolderup (1926) argued for basement–cover repetition by thrusting in the area. Although we acknowledge the local presence, of small-scale thrust faults within the basin (Sturt & Braathen 2001), there are no signs of a post-depositional thrust at the base of the landslides (Nielsen 1968; Bryhni 1976; Furnes & Lippard 1983). One could argue that the rhyolites are Middle Devonian in age, and that the zircons dated in this study are inherited from the volcanic rocks in the substrata. This possibility

cannot be ruled out, but it is regarded as highly unlikely as there is no evidence of Devonian zircons in the dated sample, and the imaged zircons were all undisturbed primary crystals.

Therefore the most reasonable interpretation, based on our own and previous field studies, and the new zircon U–Pb data presented here, is that the Hersvik volcanic rocks represent landslides from a Silurian volcanic sequence in the basin substrate. Some of the brecciation, at the base of the slide may be linked to post-volcanic transport of the slide. However, brecciation within the central sections of both the rhyolitic and gabbroic blocks is clearly primary igneous deformation as shown by Furnes & Lippard (1983), as igneous features (rhyolite flows and netveining) crosscut these structures. The rhyolitic lava were thus emplaced as a coherent landslide, and is the first of several landslides including Hagevatnet landslide (Fig. 2c). Collectively this indicate to us that all of the exotic bodies near Hersvik (Fig. 2a) are landslides, and here referred to as the Hersvik landslides.

As the volcanic rocks in the Solund Basin dated in this study were considered to be the only example of Devonian volcanism in Scandinavian Old Red Sandstone basins it is concluded that there is no evidence for extension related volcanism in the Scandinavian Caledonides. This implies that the new U–Pb data have no relevance for the regional extensional tectonic history as previously inferred (see above), but instead represent the youngest pre-Scandian volcanic rocks presently dated within the collapsed Scandinavian Caledonides.

The arc-volcanics in the cover sequence of the *c.* 443 Ma Solund–Stavfjord Ophiolite Complex are presently undated, but comprise both mafic and felsic volcanic rocks (Furnes *et al.* 1990). It is considered that these are the likely source of the landslide block. The composition of the Hersvik rhyolite is typical for both mature oceanic arcs and anorogenic rift basins (Furnes & Lippard 1983). The latter interpretation has previously been preferred on the basis of the inferred Devonian age and intermontane setting of the basins. However, based on the present interpretation that the Hersvik rhyolites formed in a mature arc setting prior to obduction of the Solund–Stavfjord Ophiolite Complex. This is supported by the late plagiogranitic veins in the gabbroic and dioritic sections of the landslides, suggesting that these rhyolitic/plagiogranitic rocks are intimately related to the mafic rocks, but slightly post-date them. The Solund–Stavfjord Ophiolite Complex is oceanic in nature, but the Pb–Pb dates of the inherited zircons from this study, suggest that the mature arc and rhyolites and other recently dated subduction related granites (e.g.  $440 \pm 5$  Ma, Hansen *et al.* 2001) formed on or near the Baltic continent from which these zircons were derived.

### Triggering and transport mechanisms of the Hersvik landslides

Interpretation of the Hersvik landslide as post-volcanic deposits opens new questions regarding triggering and transport mechanisms of the large rock masses. Terrestrial landslides occur in various topographic, climatic and tectonic settings, and their evolution varies accordingly. Landslides are the primary source of sediment discharge from humid mountains, where the landslides typically are triggered by major rainstorms or earthquakes (Hovius *et al.* 1997) or hill slope instability caused by ground water seepage (Hovius *et al.* 2000). Typically the ‘wet’ landslides occur in or above narrow

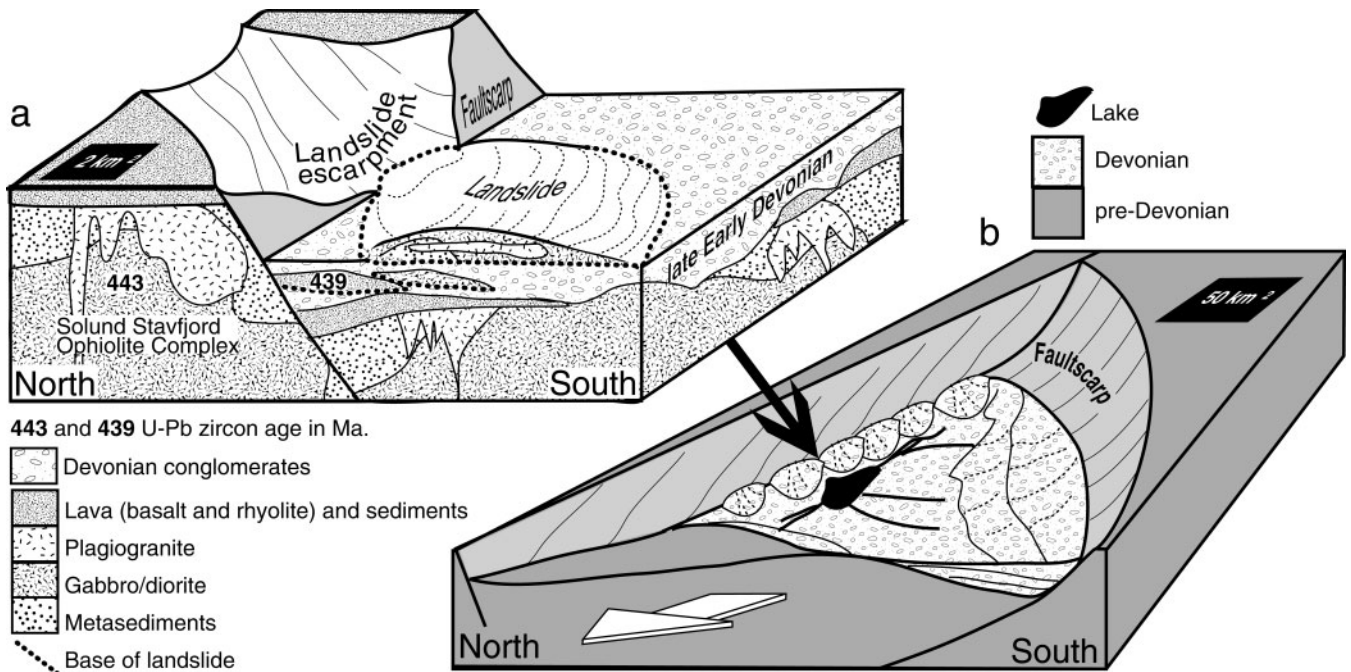


Fig. 5. Conceptual model for the formation of the Hersvik landslides (a) and the Solund Basin (b). The overall geometry of the Solund Basin (b) shows how the landslide formed near major basin bounding faults. Notice the depositional asymmetry of the basin (Nielsen 1968). The enlarged figure (a), illustrates a possible mode of deposition of the Hersvik landslides in late Early Devonian times. Note that the slides are separated by conglomerate, and generally show a reverse stratigraphy of the Solund–Stavfjord Ophiolite Complex in the footwall of the basin-bounding fault. The *c.* 443 Ma diorite in the Solund Stavfjord Ophiolite Complex (Dunning & Pedersen 1988) is cut by younger undated granite, and overlain by a volcano-sedimentary cover, which we suggest correlate with the *c.* 439 Ma rhyolites in the Hersvik landslides. The tectonostratigraphy of the ophiolites around and below the Solund Basin is adapted from Furnes *et al.* (1990). Deformation (brecciation, folding and faulting), of both landslides, Devonian deposits and the ophiolite is removed for simplicity. Particularly are the landslides not as internally coherent as shown in the simplified model.

gorges, and have limited preservation potential due to the major fluvial incision into the mass waste products (Hovius *et al.* 1997).

Landslides in arid regions are typically large flat megabreccias triggered by earthquakes as shown by recent events in southern California (Harp & Jibson 1996). These landslides are typical for the Basin and Range Region in southwestern USA, where large semi-coherent megabreccias slid from the elevated footwalls into the flat extensional basins (e.g. Forshee & Yin 1995).

The Hersvik landslide clearly shares most features with the latter type of slides. The Solund Basin formed in a climatic and tectonic setting resembling the high ridges (uplifted footwalls) and wide arid alluvial to fluvial basins of the Basin and Range Region (Steel *et al.* 1985; Osmundsen *et al.* 1998). Although the landslide is today incompletely exposed due to sedimentary cover and erosion, it is evident that it also internally comparable to Basin and Range landslides. It is compatible in size and has a similar minimum transport length, and comprises semi-coherent megabreccias with blocks that are several tens of metres across and show local preservation of internal pre-slide foliation parallel to bedding below the slide. These are all features found with slides near the Catalina Core Complex in southern Arizona (Dickinson 1991) or Whipple Mountain Landslide, California (Forshee & Yin 1995).

The triggering mechanism of non-historic landslides is obviously ambiguous, but the tectonic setting of the Hersvik landslides gives some clues. All of the Devonian basins in western Norway were bounded by major normal and oblique strike-slip and extensional faults with tens of kilometers of

displacement (Steel *et al.* 1985; Osmundsen *et al.* 2001) (Fig. 5b), suggesting that the region must have been highly seismically active. Ridge crests are highly prone to earthquakes, as the surface topography diffracts and thereby amplifies incident seismic waves, thereby releasing planar bedding-parallel landslides, that end up resting at hill-slope toes (Densmore & Hovius 2000). Both the tectonic setting, and internal structure thus compare to slides triggered by earthquakes.

The Hersvik landslide is proposed to have travelled on a thin layer of molten rhyolitic lava (Furnes & Lippard 1983; Norton 1983). Excluding this model in the light of the U–Pb dating presented here, other alternatives are discussed. Forshee & Yin (1995) suggest that landslides may either move by slow creep or catastrophically by fluidization of the breccia by entrapped air, interstitial dust, translation on ball-bearing like granular flow or water-saturated sediments. More detailed studies are needed to determine the transport mechanisms of the Hersvik landslide, however, again the slide compares to the Whipple Mountain landslide (Forshee & Yin 1995), in that pervasive fracturing suggest catastrophic emplacement, and that sub-parallelism between internal foliation, and substrate bedding suggest an overall translation of sheet-like bodies rather than mega-scale granular flow (not thereby excluding granular flow along the base or along internal horizons).

The lithologies within the slides compares to the nearby ophiolite complex at Leknessund (Furnes 1974), which constituted the eroded footwall of the closest basin bounding fault (Nielsen 1968) (Fig. 5a). The reversed internal stratigraphy of the slides (slides with volcanic rocks below slides with plutonic

rocks), also favour a transport model similar to a sliding deck of cards (Fig. 5a). The largest contrast between the Hersvik landslide and the Wipple Mountain landslide is that the latter traveled across thick water-laden sediments or lakes (Forshee & Yin 1995). Deposits in the Solund Basin consist almost entirely of alluvial fan conglomerates. The one exception is, however, the basal two meters of the main Hersvik landslide, consisting of unstratified siltstone and sandstone with rare outsized clasts to the east of Hagevatnet (Fig. 2c). This layer may represent an example of otherwise rarely preserved fine-grained water-laden sediments. The overall north to north-westwards sediment transport direction in the basin, and the half graben geometry (Nielsen 1968), suggest that the immediate hanging wall of the south dipping fault along the northern basin margin was a topographic low, where water, and thus water-laden sediment, would collect (Fig. 5b). We suggest that the Hersvik landslide may have travelled across rare fine-grained water-laden sediments, and that this could explain why the landslide made it so far from the margin.

## Conclusions

The rhyolitic lava at Hersvik, which has been regarded the only Scandinavian example of Devonian syn-depositional volcanism, is Silurian in age ( $439.0 \pm 1.0$  Ma) as shown by U–Pb ID TIMS analyses of zircons. The rhyolites and other large-scale blocks of gabbro and schist are interpreted as large landslides from the Caledonian nappes forming the depositional substrate and faulted margins of these basins. The U–Pb age reflects the youngest dated subduction related arc volcanism in Scandinavia, rather than the oldest, post-Scandian rift-volcanism. Preliminary studies suggest that the Hersvik landslides compare to landslides in arid regions that were triggered by earthquakes, and emplaced catastrophically partly along layers of water saturated sediments.

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