U–Pb age and Lu–Hf signatures of detrital zircon from Palaeozoic sandstones in the Oslo Rift, Norway

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Short title: Detrital zircon from sandstones in the Oslo Rift

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Abstract

U–Pb and Lu–Hf isotope analyses of detrital zircon from the latest Ordovician (Hirnantian) Langøyene Formation, the late Silurian Ringerike Group and the late Carboniferous Asker Group in the Oslo Rift were obtained by LA–ICPMS. Overall the U–Pb dating give ages ranging from 2861 to 313 Ma. The U–Pb age and Lu–Hf isotope signatures correspond to virtually all known events of crustal evolution in Fennoscandia, as well as syn–orogenic intrusions from the Norwegian Caledonides. Such temporally and geographically diverse source areas likely reflect multiple episodes of sediment recycling in Fennoscandia, and highlights the intrinsic problem of using zircon as a tracer–mineral in 'source to sink' sedimentary provenance studies. In addition to its mostly Fennoscandia–derived detritus the Asker Group also have zircon grains ranging in age from late Devonian to late Carboniferous. Since no rocks of these ages are known in Fennoscandia these zircons are inferred to be derived from the Variscan Orogen of central Europe.

Keywords: Oslo Graben; Zircon; U/Pb; Lu/Hf; Asker Group; Ringerike Group; Langøyene Formation.

1. Introduction

Zircon is a very robust mineral — both physically and chemically, whose U–Pb and Lu–Hf systems can survive processes beyond high grade metamorphism and crustal anatexis (Williams, 2001; Hawkesworth & Kemp, 2006). Because of its robustness, it has become a popular indicator mineral in sedimentary provenance analysis, based on the assumption that the age and Hf isotopic composition can be related to distinct source rocks (e.g. Veevers *et al.* 2005, 2006; Augustsson *et al.* 2006; Yang *et al.* 2006; Veevers & Saeed, 2007). Ideally, such data could be used to map the routing of detritus deposited in a sedimentary basin 'from source to sink'. Detrital zircon can additionally give approximate ages of deposition, and shifts in U–Pb ages and Hf signatures through a stratigraphical section can potentially be used to highlight shifts in provenance. However, the refractory nature of zircon also causes an important problem for the interpretation of detrital zircon data: What can be identified by isotopic data is not necessarily the immediate source of detritus, but the rock in which the zircon originally crystallized — i.e. the protosource. Residence in any intermediate repository will normally not leave an imprint on the isotopic systems of detrital zircon, although diagenetic effects (Willner *et al.* 2003) and contact metamorphism (Andersen, 2013) have been indicated to cause lead loss in detrital zircon. The importance of recycling of older sedimentary rocks has been highlighted in several studies (e.g. Thomas *et al.* 2004; Dickinson, Lawton & Gehrels, 2009).

Sedimentary rocks of Cambrian to Permian age are preserved within the down–faulted blocks (half–grabens) of the Oslo Rift in southern Norway. Previous provenance studies on sandstones from the Oslo Rift have indicated protosources that include most of Fennoscandia, but also that specific temporally immediate, local and distal sources have contributed, some of which do not have identifiable counterparts within the Fennoscandian Shield or the Caledonian mountain chain (Dahlgren & Corfu, 2001; Andersen *et al.* 2011). This paper presents the results of a combined U–Pb and Lu–Hf isotope study of detrital zircon from Palaeozoic sandstone units in the Oslo Region. Problems that will be addressed include the importance of Fennoscandian protosources, recycling of sediments from older deposits within and outside of the rift, identifiable temporally immediate, distal sources and possible post–depositional effects on the U–Pb system of the analysed zircon grains by Permian magmatism in the Oslo Rift.

2. Geological setting

Geologically the Oslo Region comprises the onshore half–graben segments of the Oslo Rift (e.g. Larsen *et al.* 2008). It covers an area of c. 10 000 km^2, is about 40–70 km wide, and extends 115 km north and south of the city of Oslo (Bruton, Gabrielsen & Larsen, 2010). In this half–graben segment a pre–rift lower Palaeozoic (Cambrian–Silurian) succession is preserved, together with rift related Palaeozoic (late Carboniferous–Permian) sedimentary and magmatic rocks.

The increased influx of siliciclastic material to the Oslo Region recorded in the Darriwilian Elnes Formation (Fig. 2; e.g. Hansen, 2008; Candela & Hansen, 2010) and in the latest Ordovician period (Brenchley, Newall & Stanistreet, 1979; Brenchley & Newall, 1980**)** has been interpreted to reflect sedimentary response to the growing Caledonian mountain chain to the northwest of the Oslo Region (Bjørlykke, 1974; Bruton, Gabrielsen & Larsen, 2010). A land area to the northwest, which sheltered the foreland basin from the Iapetus Ocean (Hansen, 2008, 2009), is thought to have been an important source of siliciclastic material to the Oslo Region during the late Ordovician period (Størmer, 1967; Brenchley, Newall & Stanistreet, 1979; Bruton, Gabrielsen & Larsen, 2010). This view was however opposed by Braithwaite, Owen & Heath (1995) (see also Størmer, 1967), who instead favored a model in which the sediments of the Oslo Region were dominantly shed from sources located in the adjacent Precambrian basement to the east, as well as sources to the north and northeast**.**

The Hirnantian (latest Ordovician; Cocks, 1982; Owen, 1981, 1982) Langøyene Formation (Fig. 2) consists mainly of laminated sandstones, but also include interbedded shales and thin limestone beds (Brenchley & Newall, 1975; Owen *et al.* 1990), deposited during a major regressive phase of glacio–eustatic origin (Brenchley & Newall, 1980).

During the early Silurian period marine conditions prevailed with the deposition of the carbonates of the Steinsfjorden Formation (Fig. 2). A transition to non–marine and red–bed facies (Sundvollen Formation) occurred at or just below the Wenlock–Ludlow boundary (Bruton, Gabrielsen & Larsen, 2010). The 'Old Red Sandstone' sediments of the late Silurian to earliest(?) Devonian Ringerike Group (Fig. 2) are found discontinuously throughout the Oslo Region, and were deposited in the foreland basin to the rising Caledonian mountain range in the northwest (Worsley *et al.* 1983; Davies, Turner & Sansom, 2005b). Davies, Turner & Sansom (2005a) revised the lithostratigraphy of Turner (1974), adding the Store Arøya Formation to the existing — Sundvollen, Stubdal and Holmestrand — formations (Fig. 2).

The Sundvollen Formation — the oldest formation of the Ringerike Group — sits conformably on top of the Stubdal Formation (Fig. 2), both of these formations are restricted to the northern part of the Oslo Region (Davies, Turner & Sansom, 2005a). The base of the Store Arøya Formation is defined as the first terrigenous sediments that can be seen above the Steinsfjorden Formation south of Sylling (Fig. 1b; Davies, Turner & Sansom, 2005a), whereas the Holmestrand Formation (Fig.2) refers to the uppermost 100 m of the Ringerike Group. Where its base is exposed it sits conformably on top of the Store Arøya Formation (Fig2; Davies, Turner & Sansom, 2005a). While the age of the Ringerike Group is somewhat disputed, a late Wenlock to early Ludlow age for the Sundvollen Formation and a Ludlow to Pridoli age for the Holmestrand Formation have been suggested (Davies, Turner & Sansom, 2005a). The Sundvollen and Stubdal Formations are interpreted to have been sourced from the Jotun Nappes of the Norwegian Caledonides to the northwest (Bjørlykke, 1974; Turner & Whitaker, 1976; Davies, Turner & Sansom, 2005b). The Jotun Nappes are also considered to have been an important source area for the sediments of the Store Arøya and Holmestrand Formations, but Davies, Turner & Sansom (2005b) have argued that the late Neoproterozoic nappes of the Lower Allochton and the autochthonous basement have been important source rocks for these formations.

The Asker Group which was deposited during the proto–rift and initial–rift stages of the development of the Oslo Rift, consists of the Kolsås, Tanum and Skaugum formations (Fig. 2; Larsen *et al.* 2008). The proto–rift Tanum Formation consists, in the Asker area (Fig. 1b), of grey, carbonate–cemented sandstones of fluvio–marine origin (Olaussen, Larsen & Steel, 1994; Larsen *et al.* 2008). Fossils found in the formation indicate a late Bashkirian to late Moscovian (late Carboniferous) age for its deposition (Olaussen, 1981; Olaussen, Larsen & Steel, 1994; Larsen *et al.* 2008). Detrital zircon of Neoproterozoic, Cambro–Ordovician and early Carboniferous ages led Dahlgren & Corfu (2001) to suggest a southerly source (Variscan Orogen) for the Asker Group.

3. Analytical methods

10 sandstone samples were crushed, and their heavy mineral fractions were extracted by Wilfley table washing and heavy liquid (sodium polytungstate) separation. No magnetic separation was performed to avoid introducing an artificial bias (Sircombe & Stern, 2002; Andersen *et al.* 2011). Zircon grains were hand–picked, cast in epoxy resin, polished and imaged by cathodoluminescence (CL) using a JEOL JSM 6460LV scanning electron microscope at the Department of Geosciences, University of Oslo. U–Pb and Lu–Hf analyses were done by LA–ICPMS, using a Nu Plasma HR multi–collector mass spectrometer equipped with a NewWave LUV 213 Nd–YAG laser microprobe at the Department of Geosciences, University of Oslo. All plots were made using the R programming language and

statistical computing environment (R Development Core Team, 2012)**,** and ggplot2 (Wickham, 2009). Ages given are ^{206}Pb–^{238}U ages if younger than or equal to 600 Ma, otherwise the ^{207}Pb–^{206}Pb ages have been used. Only grains with less than ±10% central discordance have been included. Kernel density estimates (KDEs) were calculated using the algorithm of Botev, Grotowski, & Kroese (2010); gaussian KDEs with bandwidth=25 were also produced.

3.a. U–Pb isotope analysis

For U–Pb the analytical protocols of Andersen *et al.* (2009) and Rosa *et al.* (2009) were followed. Ablation conditions were: beam diameter 40 μm (aperture imaging mode), pulse frequency 10 Hz and beam fluence c. 0.06 J/cm^2, using static ablation. Data reduction was done using an interactive, in–house Microsoft Excel 2003 spreadsheet program. For an analysis with $^{207}Pb/^{235}U$=x, $^{206}Pb/^{238}U$=y and ^{207}Pb–^{206}Pb age=t, the central discordance (%) was calculated using the equation

$$disc = 100\left(\sqrt{\frac{x^2+y^2}{\left(e^{t\lambda_{235}}-1\right)^2+\left(e^{t\lambda_{238}}-1\right)^2}}-1\right),$$

where λ_{235} and λ_{238} are the decay constants of ^{235}U and ^{238}U (Steiger & Jäger, 1977), respectively.

GJ–1 (^{207}Pb–^{206}Pb age=609±1 Ma; Jackson *et al.* 2004), 91500 (^{207}Pb–^{206}Pb age=1065±1 Ma; Wiedenbeck *et al.* 1995) and A382 (concordia age=1876±2 Ma; Lauri *et al.* 2011) were used as standards. Repeated analyses of the in–house reference zircon C (weighted average ^{207}Pb–^{206}Pb age=556.4±1.5 Ma; J. Lamminen, pers. comm.) during the period the samples were analysed gave a weighted average ^{207}Pb–^{206}Pb age of 556.0±1.6 Ma (2SD, n=168).

3.b. Lu–Hf isotope analysis

For Lu–Hf the analytical protocols of Elburg *et al.* (2013) were followed. Ablation conditions were: beam diameter 50–60 µm (aperture imaging mode), pulse frequency 5 Hz and beam fluence c. 2 J/cm^2 , using static ablation. Data reduction was done using Nu Instruments online software.

During the period the samples were analysed, repeated analyses of the Mud Tank zircon yielded arithmetic mean ^{176}Hf/^{177}Hf=0.282511±47 (2SD; n=225) and Temora–2 ^{176}Hf/^{177}Hf=0.282680±48 (2SD; n=145), the latter (±2 ε_{Hf}) is accepted as a conservative estimate of the precision of the method. A decay constant value for ^{176}Lu of 1.867 × 10^{-11} (Söderlund *et al.* 2004) has been used in all calculations. For ε_{Hf} calculations we used the present-day chondritic ^{176}Hf/^{177}Hf=0.282785 and ^{176}Lu/^{177}Hf=0.0336 (Bouvier, Vervoort & Patchett, 2008). We have adopted the depleted mantle parameters of Griffin *et al.* (2000), this model, modified to the aforementioned decay constant and CHUR parameters, gives present-day ^{176}Hf/^{177}Hf=0.28325 (+16.4 ε_{Hf}; similar to average mid-ocean ridge basalt) from chondritic initial ^{176}Hf/^{177}Hf at 4.56 Ga and ^{176}Lu/^{177}Hf=0.0388.

4. Sample description, and U–Pb and Lu–Hf results

The sample localities are given in Figure 1b and Table 1**,** and their approximate stratigraphic position are given in Figure 2. U–Pb data are given in Figure 3 and online Supplementary Table S1 at http://journals.cambridge.org/geo, while Lu–Hf data are given in Figure 4 and online Supplementary Table S2 at http://journals.cambridge.org/geo.

4.a. AA11–41

This sample is a medium grained calcareous sandstone belonging to the Langøyene Formation (Figs. 1b, 2). Zircon from AA11–41 are dominated by the c. 1000–1100 Ma age group. This age group has an ε_{Hf} range of +4 to -2, in which only 1 of 25 grains are negative.

4.b. R2010–2

This sample is a fine grained sandstone from the Sundvollen Formation (Figs. 1b, 2). The major zircon age groups in this sample are c. 400–500 Ma, c. 900–1200 Ma, c. 1300–1500 Ma and c. 1600–1700 Ma. Grains in the c. 400–500 Ma group range in ε_{Hf}–values from +2 to -9. The c. 900–1200 Ma group has largely positive (21/27) ε_{Hf}–values, ranging from +8 to -4, while grains in the c. 1600–1700 Ma group range from +6 to -2.

4.c. R2010–6

This sample is a fine grained sandstone from the lower part of the Stubdal Formation (Figs. 1b, 2). The main age groups in this sample are found at c. 400–500 Ma, c. 1000–1200 Ma and c. 1600–1700 Ma. In the c. 400–500 Ma age group only one grain with a positive ε_{Hf}–value (+1) is found, the others range from -3 to -12. Two negative grains (-4 ε_{Hf}, -2 ε_{Hf}) are found in the c. 1000–1200 Ma, the rest (n=18) range in ε_{Hf}–values from +10 to 0. The c. 1600–1700 Ma group range in ε_{Hf}–values from +5 to -3.

4.d. R2010–8

This sample is a fine grained sandstone from the upper part of the Stubdal Formation (Figs. 1b, 2). The combined histogram and KDE plot (Fig. 3) is dominated by an age group at c. 1000–1100 Ma. This age group have an ε_{Hf} range of +7 to 0.

4.e. MK12–1

This sample is a calcareous medium grained sandstone from the Holmestrand Formation (Fig. 1b, 2). Peaks in the KDE plot (Fig. 3) define age groups at c. 400–500 Ma, c. 900–1200 Ma, c. 1400–1500 Ma and 1600–1700 Ma. The c. 400–500 Ma age group range in ε_{Hf}–values from -7 to +11, a small majority (6/10) of which are negative. In the c. 900–1200 Ma group the ε_{Hf}–values range from +6 to -6, in the c. 1400–1500 Ma group they range from +3 to -4, while in the c. 1600–1700 Ma group they range from +7 to -1. These three groups have largely positive ε_{Hf}–values (32/40, 9/12 and 9/12, respectively).

4.f. MK–2010–5

This sample is a fine grained calcareous sandstone from the lower part of the Tanum Formation (Figs. 1b, 2). The main zircon age group is found at c. 1000–1100 Ma. In this group the ε_{Hf}–values range from +4 to -1, with a majority being positive (17/20).

4.g. MK–2010–4

This sample is a fine grained calcareous sandstone from the upper part of the Tanum Formation (Figs. 1b, 2). The largest peak in the KDE plot (Fig. 3) is found at c. 1050 Ma, defining a c. 1000–1100 Ma age group. This age group have ε_{Hf}–values ranging from +6 to 0.

4.i. MK–2010–3

This sample is a fine grained sandstone overlain by conglomerate, most likely belonging to the Tanum Formation of the Asker Group (Figs. 1b, 2). The combined histogram and kernel density estimate plot (Fig. 3) define age groups at c. 300–400 Ma, c. 900–1100 Ma and c. 1450–1800 Ma. Zircon grains in the 300–400 Ma age group have ε_{Hf}–values from +10 to -5, the 900–1100 Ma group have ε_{Hf}–values from +7 to -6 and the 1450–1800 Ma age group have ε_{Hf}–values from +10 to -11.

4.h. MK–2010–2

This sample is a medium grained sandstone belonging to the Tanum Formation (Figs. 1b, 2). The zircon grains in the overall 2847–333 Ma age range define two major age groups at c. 900–1200 Ma and c. 1400–1800 Ma. Zircons in the c. 900–1200 Ma group range in ε_{Hf}-values from +5 to -4, of which a majority (29/46) are positive. In the c. 1400–1800 Ma group all zircons, except for an unusually negative (ε_{Hf}=-25) grain at 1599 Ma, range in ε_{Hf}-values from +10 to -7.

4.j. MK–2010–7

This sample, which likely belongs to the Tanum Formation, is a cross stratified sandstone comprising the upper part of the sandstone sequence of the Asker Group in the Ringerike area (Figs. 1b, 2). A zircon age peak at c. 350 Ma, defining a c. 300–400 Ma age group, dominates this sample. The zircons in this age group have ε_{Hf}–values ranging from +6 to -15, with a majority (18/27) being negative.

5. Discussion

5.a. Potential protosources

Zircon from the Langøyene Formation, the Ringerike Group and the Asker Group record ages ranging from Archaean to Carboniferous. From Figures 3 and 4 it is evident that these ages largely coincide with known events of crustal evolution in Fennoscandia, as well as magmatism in the Caledonides.

ε_{Hf}–values of the Archaean grains (Fig. 4) in this study falls within the range recorded for rocks in the Archaean Domain (Fig. 1a) of Fennoscandia (Patchett *et al.* 1981; Lauri *et al.* 2011, 2012; Heilimo *et al.* 2013), which today are exposed in northern Norway, northeastern Sweden, eastern Finland and northwestern Russia.

Ages of 1800–1700 Ma are recorded in several samples (Fig. 3). In Fennoscandia this period is characterized by the TIB–1 stage of the formation of the Transscandinavian Igneous Belt (TIB; Fig. 1a; Gorbatschev, 2004). A majority (19/26) of the grains (Fig. 4) found from this time period are consistent, within error (±2SE), with the evolutionary trend of the crustal source of TIB granite magmas (Andersen *et al.* 2009). With the exception of three grains with ε_{Hf}–values of +12, -6 and -11, the remaining grains plot close (±2 ε_{Hf}) to this trend. Zircons with negative ε_{Hf} are compatible with values reported from Palaeoproterozoic granitic intrusions within the Archaean Domain (Patchett *et al.* 1981), inherited zircon in rocks from

TIB (Andersen *et al.* 2009) and Palaeoproterozoic leucogranites in southern Finland (Kurhila, Andersen & Rämö, 2010).

Zircon grains falling in the 1700–1600 Ma age range are common to all samples (Fig. 3). The formation of TIB stages TIB–2 and TIB–3 (Larson & Berglund, 1992; Andersson *et al.* 2004; Gorbatschev, 2004), the formation of Rapakivi granites in southern Finland (e.g. Heinonen, Andersen & Rämö, 2010), and island arc magmatism related to the Gothian orogeny (Åhäll & Connelly, 2008) produced zircon–fertile rocks in Fennoscandia during this time period. With the exception of one unusually negative grain (-10 ε_{Hf}), the 1700–1600 Ma zircon span a range of ε_{Hf}–values from +12 to –4 (Fig. 4). These values are consistant with reported values from middle–Proterozoic calc–alkaline gneiss complexes from the Kongsberg–Marstrand, Bamble–Lillesand and Randsfjord–Lyngern blocks of the Southwestern Scandinavian Domain (SSD; Fig. 1a), and late (1670 Ma) TIB granites from the southwestermost part of the Transscandinavian Igneous Belt (Andersen, Griffin & Pearson, 2002); negative values are consistent, within error (±2 ε_{Hf}; 2SE), with the evolutionary trend of TIB (Andersen *et al.* 2009).

Voluminous magmatism in the Hardangervidda–Rogaland, Telemark, Bamble–Lillesand, Kongsberg–Marstrand and Randsfjord–Lyngern blocks of the SSD (Fig. 1a) affected the Fennoscandian Shield early in the Mesoproterozoic era (Andersen *et al.* 2004; Bingen *et al.* 2005b; Åhäll & Connelly, 2008; Bingen & Solli, 2009). Magmatism in the Telemark block, which included a sequence of bimodal volcanic rocks and correlative supracrustal rocks (Bingen & Solli, 2009), peaked at 1500 Ma (Bingen *et al.* 2005b). At c. 1440–1380 Ma the crust of southwestern Sweden was affected by the Hallandian thermo–magmatic event (Christoffel, Connelly & Åhäll, 1999; Söderlund *et al.* 2002). In this study zircon with ages 1500–1300 Ma are common to all samples, but this age group is most dominant in sample R2010–2 (Fig. 3). ε_{Hf}–values in this age group vary from -10 to +7 (Fig. 4), the majority

(98/117) of which plot on or above the CHUR curve. Detrital zircon with similar ε_{Hf}–values (-8 to +9) have been reported from the Mesoproterozoic Rjukan Rift Basin (Lamminen & Köykkä, 2010). Whereas the generally juvenile ε_{Hf} composition is consistent with values recorded in similarily aged rocks in the Bamble–Lillesand and Kongsberg–Marstrand blocks (Andersen *et al.* 2004), in c. 1521–1485 Ma deformed gneisses and granitoids in the Hardangervidda–Rogaland block (Roberts *et al.* 2012), and in the 1555 Ma Åsen metatonalite in the Telemark block (Pedersen *et al.* 2009).

The largest age group in this study, which contain 48% of all U–Pb data, is found at 1300–900 Ma (Fig. 3). Fennoscandia was in this time period affected by magmatism and metamorphism related to the Sveconorwegian orogeny (e.g. Bingen & van Breemen, 1998; Andersen *et al.* 2004; Bingen *et al.* 2008; Pedersen *et al.* 2009). With the exception of two zircon grains with ε_{Hf}–values of -25 and -32, the zircons from this age range vary in ε_{Hf} from -15 to +11 (Fig. 4), which is consistent with values reported for magmatic and inherited zircon from Sveconorwegian granitoids (Andersen, Griffin & Pearson, 2002; Andersen *et al.* 2004; Andersen, Griffin & Sylvester, 2007; Pedersen *et al.* 2009).

Fennoscandia is characteristically poor in middle–to–late Neoproterozoic magmatism, with little or no production of zircon fertile rocks between c. 900 Ma and the onset of magmatism related to the Caledonian orogeny (e.g. Bingen & Solli, 2009). Nevertheless, detrital zircon of this age has been reported from the Neoproterozoic Hedmark Basin (Bingen *et al.* 2005a) and the Asker Group (Dahlgren & Corfu, 2001). Middle–to–late Neoproterozoic grains found in this study make a contribution to the populations of samples MK–2010–2, MK–2010–3, MK–2010–7, R2010–2 and AA11–41. Bingen *et al.* (2005b) attributed their Neoproterozoic zircon to magmatism in the Egersund area, but as only zircon–free, low–volume mafic dykes are recorded in this area (e.g. Bingen & Solli, 2009) it is highly unlikely to be the source of any Neoproterozoic detrital zircon recorded in Fennoscandia. As discussed in Andersen *et al.*

(2011) the Neoproterozoic grains recorded by Bingen *et al.* (2005b) have U–Pb ages and ε_{Hf}–values fully compatible with Neoproterozoic loss of radiogenic lead in zircon from Sveconorwegian granitoids. While the Neoproterozoic zircon grains reported herein are broadly compatible with a similar lead loss scenario, it is deemed unlikely as these grains are all close to concordant. Dahlgren & Corfu (2001) suggested that the late Neoproterozoic zircons recorded in the Asker Group were derived from a southern source located somewhere in the newly uplifted northern or central Variscan Mountains in Central Europe. As late Neoproterozoic zircon grains are also recorded in sediments underlying the Asker Group (this study) they are more likely to be of Fennoscandian origin. One possible source area could be the Middle Allochton Seve nappe or the exotic (Corfu *et al.* 2007; Kirkland, Daly & Whitehouse, 2008) Kalak nappe of the Scandinavian Caledonides (e. g. Bingen & Solli, 2009 and references therein). Neoproterozoic zircon could also be derived from a northern source located in the Timanides (Fig. 5) where rocks of compatible ages are recorded (e.g. Gee *et al.* 2000; Larionov, Andreichev & Gee, 2004). Detrital zircon inferred to have been sourced from the Timanides in a forland basin setting have been found in the Cambrian Dividal Group (Andresen, 2013) in northern Norway. The southern limit of this foreland basin is, as of yet, undetermined — late Neoproterozoic zircon are however common in late Cambrian to early Ordovician sedimentary deposits in the St. Petersburg area (Miller *et al.* 2011) — but recycling of sediments in such basins could be a viable mechanism for introducing Neoproterozoic detrital zircon to southern Norway in the Palaeozoic era.

Grains of Caledonian age make a significant contribution to the Ringerike Group samples and a smaller contribution to the Asker Group samples, these grains are most likely derived from the c. 500–430 Ma synorogenic intrusions in the Caledonides (Gee *et al.* 2008 and references therein; Bingen & Solli, 2009 and references therein).

From the youngest Caledonian related magmatism (c. 425 Ma) to the initiation of magmatism in the Oslo Rift at 300±1 Ma (Corfu & Dahgren, 2008) no magmatic activity is known in Fennoscandia. Nevertheless do a fairly large fraction of the zircon from the Asker Group have ages in the range c. 380–313 Ma, peaking at c. 350 Ma (Fig. 6a). Similarily aged grains were reported from the Asker Group by Dahlgren & Corfu (2001), who argued that these grains were derived from the Carboniferous Variscan Orogen.

5.b. Recycling of detrital zircon and input from the Variscan Mountains?

Detrital zircon with U–Pb and Lu–Hf signatures covering sources as diverse both temporally and geographically as is recorded here are unlikely to have been derived directly from primary magmatic rocks. Recycling of older sediments, which could be the product of recycling of even older sediments, is therefore highly probable.

The Langøyene Formation and the Ringerike Group have age spectra broadly compatible with (meta)sediments in the Caledonides (Bingen & Solli, 2009 and references therein), and modern river sediments known to sample (meta)sediments and primary magmatic rocks in the Caledonides (Morton, Fanning & Milner, 2008). The proposed difference in source area of the Sundvollen and Stubdal Formations on one hand, and the Holmestrand Formation (Davies, Turner & Sansom, 2005b) on the other hand, is not reflected in the detrital zircon patterns.

The close similarity of U–Pb age and Lu–Hf signatures of the Ringerike Group and the Asker Group suggests that the Fennoscandia–derived detritus in the Asker Group could be sourced from a Silurian cover sequence similar to the Ringerike Group. Such sequences, which are now only preserved as small remnants outside the Oslo Rift, are likely to have covered much of central and western Fennoscandia in late- and post–Caledonian time. As the late Devonian to late Carboniferous aged zircons in the Asker Group have no known Fennoscandian source it seems likely that they were, as suggested by Dahlgren & Corfu (2001), derived from the

Variscan Mountains of western or central Europe, which is the area in closest proximity to the Oslo Rift with rocks of such ages. Several areas in the Variscan Mountains are possible source areas: the Armorican Massif which was affected by major intrusions (the Brittany granites) and associated metamorphism in the period 354–330 Ma (Martínez & Rolët, 1988), the Massif Central where c. 380–300 Ma rocks have been reported (e.g. Faure *et al.* 2010 and references therein) and the Bohemian Massif where c. 370–290 Ma rocks have been reported (e.g. Košler, Aftalion & Bowes, 1993; Siebel & Chen, 2009). The dearth of Hf–data from the Variscan Mountains make a definite source region for the late Devonian to late Carboniferous aged zircon grains difficult to ascertain, but reported ε_{Hf}-values from the central Variscan Mountains (the central Alps and the Bohemian Massif) are generally low ranging from 0 to -8 (Schaltegger & Corfu, 1992; Siebel & Chen, 2009). This trend of crustal contamination is also seen in our data (Fig. 6b) where 29 of the 39 zircons younger than 390 Ma have ε_{Hf}–values ranging from 0 to -15. The remaining zircon grains have a more juvenile composition (+1 to +10 ε_{Hf}), suggesting that at least two sources have contributed to the < 390 Ma grains.

If the youngest grains in the Asker Group (313 ± 4 Ma and 316 ± 2 Ma) were derived from a local source, they could mark the beginning of volcanism in the Oslo Rift, but since no rocks of such an age have, as of yet, been found in the Oslo Rift it is more likely that they come from the Variscan Mountains. Dahlgren & Corfu (2001) attributed the Neoproterozoic and Cambro–Ordovician grains in the Asker Group to the Variscan source, but since zircon of these ages are also found in the latest Ordovician Langøyene Formation and the late Silurian Ringerike Group they are equally likely to be Fennoscandia–derived. The Variscan source may have contributed zircon of other ages found in our samples, but if it did the detrital zircon must have had U–Pb ages and Lu–Hf signatures indistinguishable from typical Fennoscandian sources. The late Devonian to late Carboniferous zircon grains are compatible with post–depositional lead loss from Caledonian grains, assuming a felsic protolith with

$^{176}Lu/^{177}Hf$=0.010, possibly caused by contact–metamorphism related to the overlying latest Carboniferous to Permian lavas. But this is deemed unlikely as the CL–images of the grains show no evidence of metamictization or lead–loss.

These findings highlight the intrinsic problems of using zircon as a tracer-mineral from ‘source to sink’ in sedimentary provenance studies. Its refractory nature means that it will, in most case, keep a record of its protolith through the U–Pb and Lu–Hf isotopic systems. But this robustness is also the reason why zircon can survive multiple episodes of recycling — as is recorded in this study — causing us to only being able to trace zircon populations back to the rocks in which they crystallized and not the immediate precursor rocks which is the ultimate goal of ‘source to sink’ sedimentary provenance studies.

6. Conclusion

The latest Ordovician Langøyene Formation, the late Silurian Ringerike Group and the late Carboniferous Asker Group contain detrital zircon ranging in age from Mesoarchaean to Carboniferous. The U–Pb age and Lu–Hf signatures of the detrital zircon correspond to virtually every known event of crustal evolution in Fennoscandia as well as syn-orogenic intrusions in the Scandinavian Caledonides. Such diverse — both temporally and geographically — sources for the Palaeozoic sandstones in the Oslo Rift are likely caused by several episodes of recycling of sediments in Fennoscandia.

The Asker Group contains, in addition to its mostly Fennoscandia–derived detritus, late Devonian to late–Carboniferous aged zircon grains. As no rocks of such ages are known from Fennoscandia the most likely source area is located in the Variscan Orogen of central Europe. ε_{Hf}-values indicative of crustal to more juvenile sources suggests that at least two igneous sources have contributed to the < 390 Ma zircon group. If the Variscan source contributed detrital zircon of other ages to the studied samples they are indistinguishable from Fennoscandia–derived zircon.

Detrital zircon U-Pb age and Lu-Hf isotope data from Palaeozoic sandstones in the Oslo Rift highlight the difficulty of using zircon as a tracer-mineral in 'source to sink' studies. Its refractory nature causes it to survive multiple episodes of recycling, meaning that only the proto-sources of the various age groups are recorded and not the immediate (meta)sedimentary precursor source.

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Declaration of interest

None.

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List of figures

Figure 1.

(a) Geological map of the main crustal domains in Fennoscandia (modified after Gaál & Gorbatchev, 1987; Högdahl, Andersson & Eklund, 2004). The black box shows the approximate position of b. TIB: Transscandinavian Igneous Belt; SSD: Southwest Scandinavian Domain. (b) Simplified geological map of the central Oslo Graben (modified after Larsen *et al.* 2008) with sample locations indicated.

Figure 2.

(Colour online) Simplified lithostratigraphic column of the Oslo Graben at Ringerike (modified after Henningsmoen, 1978; B. T. Larsen & S. Olaussen, unpub. field guide, *The Oslo Region – A study in classical Palaeozoic Geology*, Norsk geologisk forening, 2005), with approximate sample positions indicated. Inset shows the idealised stratigraphy of the Ringerike Group and the Steinsfjorden Formation from Ringerike (Fig. 1) in the northern part of the central Oslo Graben to Skien in the southern Oslo Graben (after Davies, Turner & Sansom, 2005a). Cam: Cambrian; L: Lower; M: Middle; U: Upper; C: Carboniferous; P: Pennsylvanian; (m): metres above the Precambrian basement.

Figure 3.

(Colour online) Combined histogram and kernel density estimator plots of all samples stacked according to their (presumed) stratigraphic positions. (Red solid curves) are KDEs calculated using the algorithm of Botev, Grotowski, & Kroese (2010); (blue dashed curves) are gaussian KDEs with bandwidth=25. Depositional ages: AA11–41: Hirnantian (latest Ordovician); R2010–2: Wenlock (middle Silurian); R2010–6 and R2010–8: Ludlow (middle–late Silurian); MK12–1: Pridoli (latest Silurian); MK–2010–5, MK–2010–4, MK–2010–3, MK–2010–2 and

MK–2010–7: Pennsylvanian (late Carboniferous). Data from online Supplementary Table S1 at http://journals.cambridge.org/geo. Ages are given as ^{206}Pb–^{238}U ages if equal to or younger than 600 Ma, otherwise the ^{207}Pb–^{206}Pb ages have been used. Note that the individual panels use different y–axis scaling.

Figure 4.

Initial ε_{Hf} plotted against age (data from online Supplementary Table S2 at http://journals.cambridge.org/geo) with comparative fields for magmatic rocks in Fennoscandia (Patchett *et al.* 1981; Vervoort & Patchett, 1996; Andersen, Griffin & Pearson, 2002; Andersen & Griffin, 2004; Andersen *et al.* 2004; Andersen, Griffin & Sylvester, 2007; Andersen, Graham & Sylvester, 2009; Andersen *et al.* 2009; Pedersen *et al.* 2009; Heinonen, Andersen & Rämö, 2010; Kurhila, Andersen & Rämö, 2010; Lauri *et al.* 2011, 2012; Heilimo *et al.* 2013), and detrital zircon from the Orsa and Brumundal sandstones (Andersen *et al.* 2011). Ages are given as ^{206}Pb–^{238}U ages if equal to or younger than 600 Ma, otherwise the ^{207}Pb–^{206}Pb ages have been used.

Figure 5

Late Carboniferous palinspastic map of the central North Sea area (modified after Coward *et al.* 2003; Ziegler *et al.* 2004), with outlines of the main Variscan massifs and possible older source areas indicated.

Figure 6.

(a) (Colour online) Combined histogram and kernel density estimator plot of late Devonian to late Carboniferous aged zircon grains from the Asker Group. (Red solid curve) is a KDE calculated using the algorithm of Botev, Grotowski, & Kroese (2010); (blue dashed curve) is a gaussian KDE with bandwidth=5. Data from online Supplementary Table S1 at http://journals.cambridge.org/geo. Ages are given as ^{206}Pb–^{238}U ages. (b) Initial ε_{Hf} plotted against age for the late Devonian to late Carboniferous aged zircon grains from the Asker Group. Vertical error bars are 2SE; errorbars in the lower right corner shows the arithmetic mean of the age errors (2SD) and of the ε_{Hf} errors (2SE). Data from online Supplementary Table S2 at http://journals.cambridge.org/geo. Ages are given as ^{206}Pb–^{238}U ages.

Table 1. Sample positions.

Sample	Unit	UTM coordinates WGS84, zone 32V	
MK-2010-7	Asker Group/Tanum Formation (?)	573960	6658381
MK-2010-2	Tanum Formation	578525	6636168
MK-2010-3	Asker Group/Tanum Formation (?)	574152	6641204
MK-2010-4	Tanum Formation	584145	6644437
MK-2010-5	Tanum Formation	584145	6644437
MK12-1	Holmestrand Formation	592655	6595051
R2010-8	Stubdal Formation	573952	6658382
R2010-6	Stubdal Formation	576687	6665705
R2010-2	Sundvollen Formation	572464	6660314
AA11-41	Langøyene Formation	596536	6639347