The age and regional correlation of the Svecofennian Geitfjell granite, Vestranden, Norway

LEIF JOHANSSON, HANS SCHÖBERG & ZOLTAN SOLYOM


U-Pb zircon dating of granitoids from the northermost part of the Western Gneiss Region (Vestranden) has yielded the oldest ages so far recorded in central and southern Norway. The Geitfjell granite, a coarse-grained granite near Grong, central Norway yielded a U-Pb zircon age of 1828 ± 8 Ma, whereas a Rb-Sr determination yielded an age of 1770 ± 62 Ma. Its initial 87Sr/86Sr is 0.7094 and εNd (1830 Ma) value is −0.04. Petrographically similar granites occur also in large areas in the northern half of the Vestranden region. The petrography, isotope chemistry, age and chemical composition of the Geitfjell granite closely resembles that of granites of the Revsund type, which intrude marine metasedimentary rocks in the Svecofennian Bothnian Basin in north-central Sweden. A correlation of these granites is thus suggested. The occurrence of this type of granite in the northermost part of the Western Gneiss Region suggests that this region represents the westerly continuation of the Svecofennian Bothnian Basin rather than a part of the Southwest Scandinavian orogenic province or the Transscandinavian Granite Belt. Precambrian metasedimentary rocks of Bothnian Basin affinity may thus occur in Vestranden.


The Baltic Shield is composed of successively younger crustal segments from the Archean region in the northeast to the Southwest Scandinavian orogenic region in the southwest. In western and northern Scandinavia, all crustal segments, such as the Archean Domain (ca. 2900–2600 Ma), the Svecokarelian Domain (ca. 2000–1750 Ma) with Svecofennian subprovinces, the Transscandinavian Granite Belt (ca. 1800–1650 Ma) and the Southwest Scandinavian Region (ca. 1770–1550 Ma) are partly overlain by Caledonian nappes (Fig. 1). The western parts of these regions have also been more or less reworked during Caledonian tectonothermal event(s). Studies of the lithotectonic evolution of the western continuations of these regions are important for understanding the successive growth of the Baltic Shield and for the plate tectonic reconstruction of the Precambrian and early Paleozoic continents (i.e. the relationships between the Baltic and Laurentian shields). These studies must be based on work in the basement windows in the Caledonides, in the Western Gneiss Region of Norway, and on studies of far-travelled Caledonian nappes that contain tectonic slices of pre cambrian basement rocks.

The purpose of the present paper is to discuss the age of the Geitfjell granite in the northermost part of the Western Gneiss Region, the correlation of this granite with the similar granites in the Bothnian Basin in the Caledonian foreland and, finally, to briefly discuss the setting of the northern part of the Western Gneiss Region in the framework of orogenic segments in the Baltic Shield.

Fig. 1. Major tectonostratigraphic units of the western Baltic Shield (modified after Gål & Gorbatschev 1987). 1. North Svecofennian subprovince. 2. Central Svecofennian subprovince, including Bothnian Basin. 3. South Svecofennian subprovince. 4. Transscandinavian Granite Belt. 5. Southwest Scandinavian orogenic province. 6. Caledonides. WGR = Western Gneiss Region, VR = Vestranden, GOC = Grong–Olden Culmination, BW = Börgefjell Window, TW = Tömerdals Window, MZ = Mylonite zone, PZ = Protogine zone.
The Precambrian basement of the Scandinavian Caledonides is exposed in many windows along the mountain chain and in the Western Gneiss Region of Norway (Fig. 1). In the central Caledonides, the basement lithologies may be studied in the northern part of the Western Gneiss Region (Vestranden), the Grong–Olden Culmination, The Borgefjell Window and the Tömmerås Window (Figs. 1 and 2). In the Grong–Olden Culmination and in Vestranden, its western continuation, the Precambrian basement crops out in near continuous exposure over a distance of nearly 200 km across the strike of the Caledonian orogen. With regard to lithologies, deformation, metamorphism and age relationships this basement area can be subdivided into an eastern part (Grong–Olden Culmination) and a western part (Vestranden) (Johansson 1986). The eastern part is much less affected by the Caledonian orogeny than the western part.

**Grong–Olden Culmination**

Radiometric dating suggests that the granites and rhyolites of the Grong–Olden Culmination were formed around 1650 Ma ago (Stuckless et al. 1982; Wilson 1982; Stuckless & Troéng 1984; Wilson et al. 1985). Fossen & Nissen (1991), however, presented a very well defined Rb–Sr whole-rock age of 1356 ± 29 Ma for the Blåfjellhatten granite, a granite that is petrographically and chemically identical but isotopically distinctly different to the nearby 1650 Ma old Olden granite. This suggests that there are different generations of granites present within the Grong–Olden Culmination.

Reworking in the Grong–Olden Culmination is limited to the Caledonian Late Silurian to Early Devonian deformation and metamorphism, in contrast to the Western Gneiss Region, which is a polymetamorphic region with events in Gothian (1750–1550 Ma), Sveconorwegian–Grenvillian (1250–900 Ma) and Caledonian (650–370 Ma) times (see Kullerud et al. 1986; Gaál & Gorbatschew 1987, and references therein).

Based on the age, geochemistry and lithological similarities, the granites and volcanic rocks of the Grong–Olden Culmination and the Tömmerås Window are generally considered to belong to the Transscandinavian Granite Belt of I-type granites and associated acidic volcanic rocks (Fig. 1), which extends from southern Sweden to northern Norway (Gorbatschew 1985; Wilson et al. 1985; Gaál & Gorbatschew 1987; Patchett et al.). Its rocks were formed over a long period of time, between ca. 1800 Ma (Åberg & Persson 1984; Patchett et al. 1987; Mansfeld 1991) and 1650 Ma (Wilson et al. 1985; Patchett et al. 1987). Larson et al. (1991) subdivided the granites of the Transscandinavian Granite Belt into two age groups, a younger group comprising granites ca. 1680–1650 Ma old and an older group of granites ca. 1810–1770 Ma. Granites of the younger group appear to be more common along the western margin of the belt.

**Vestranden**

Vestranden is, in this paper, defined as the broadly triangular shaped area of the Western Gneiss Region north of the Trondheim fjord, including also the coastal areas just south of the Trondheim fjord (Fig. 1).

Vestranden is dominated by granitic to granodioritic gneisses. Tonalites, monzonites and mafic rocks occur locally. Precambrian mafic dykes are common. These orthogneisses are metamorphosed and folded together with supracrustal rock sequences. This folding, deformation and metamorphism is Late Silurian to Mid-Devonian in age.

The age and origin of the supracrustal rocks is unclear at present. Dating of the Vestranden orthogneisses north of the Trondheim fjord has, until recently, been restricted to a few Rb–Sr determinations (Priem et al. 1968; Råheim et al. 1979). More precise and reliable ages based on U–Pb dating of zircons were presented by Johansson (1986) and Schouenborg et al. (1991). The results of these suggest that the oldest crust in the northern part of Vestranden is at least ca. 1820 Ma old, whereas granitoid rocks in the southwestern Vestranden are significantly younger, and were formed mainly around 1650 Ma (Tucker et al. 1987; Tucker & Krogh 1988; Johansson & Möller, in prep.).

Gorbatschew (1985) and Gaál & Gorbatschew (1987) correlated the Precambrian crust of the entire Western Gneiss Region, including Vestramden, with that of the Southwest Scandinavian orogenic province (Fig. 1). The correlation was based on geochronological and lithological similarities between the southern and central parts of the Western Gneiss Region, southern Norway and southwestern Sweden.

**Geitfjell granite**

The Geitfjell granite occurs in the easternmost part of Vestranden (Figs. 2 and 3). It forms a slice, less than 500 m thick, of coarsely porphyritic, very inhomogeneously deformed, grey rock consisting of microcline, quartz, oligoclase (An$_{20-25}$) and biotite. Accessory minerals are muscovite, hornblende, zircon, apatite, chlorite, garnet, allanite, opaque minerals, and rarely diopside. Muscovite, chlorite, diopside, garnet and at least some of the hornblende and magnetite are secondary metamorphic minerals. Flouroite occurs rarely as thin coatings on fracture surfaces. Myrmekite is common and in some samples string-perthitic feldspar is present. The K-feldspar megacrysts reach up to 6 cm in length. Brown biotite is the major Fe–Mg phase in the granite. Hornblende, biotite and elongated quartz domains define the gneissic foliation. In many places, intense deformation transformed the granite into augen gneiss. Sheared and almost isoclinally folded aplite layers, thin mafic dykes and tonalitic xenoliths are common within the augen gneiss. There are also large areas where the Geitfjell granite is well preserved (Fig. 4).
Stratigraphically above the granite slice there is a sequence of predominantly red and grey fine-grained gneisses interbanded by numerous amphibolite layers. Mica schists and quartzites are found occasionally. Sheets of coarse- to medium-grained augen gneiss, metres or tens of metres thick, occur in the upper part of the fine-grained rock series. Similar, but somewhat more reddish granitic gneisses occur beneath the Geitfjell granite slice. No tectonic or metamorphic break is found between the Geitfjell granite and the surrounding gneisses. Close to the top of Geitfjell, the granite is partly migmatitic. The migmatite neosomes are medium- to
coarse-grained with a patchy appearance. They consist almost exclusively of feldspars and quartz, with only minor amounts of biotite, which define a weak Caledonian foliation.

Sampling

Samples for chemical analysis and for isotopic dating were collected from fresh exposures along a road to the top of Geitfjell. In order to satisfy representative sampling of the coarse-grained Geitfjell granite (Fig. 4), five samples of 6–7 kg each were collected for chemical analysis. For U–Pb zircon dating, approximately 70 kg of the coarse-grained granite were sampled. The samples for zircon dating and geochemical analysis were taken within the same area and are petrographically identical. The whole-rock samples for Rb–Sr dating were collected farther up along the same road. The maximum distance between individual Rb–Sr sample points is approximately 20 m. Short petrographic descriptions of the Rb–Sr whole-rock samples are found in the Appendix. One of the samples collected for geochemical analysis was also used for the Sm–Nd whole-rock analysis.

Geochemistry

It must be emphasized that the geochemistry presented here is representative only for the coarse-grained grey granite type (Fig. 4) at Geitfjell. An extensive geochemical investigation of the Geitfjell granite is beyond the scope of this paper. Details of the analytical procedures are described in Solyom et al. (1984).

The small standard deviation seen in major and trace elements (Table 1) shows that the samples were large enough to be representative of the coarse-grained granite. The molar oxide ratio of \( \text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}) \) (Table 1) is slightly larger than 1.0, which indicates a weak peraluminous character for the Geitfjell granite. This is also reflected by the presence of normative corundum (Table 1). For comparison, analyses of the nearby Olden granite and the Revsund granite were carried out. Granite of the Olden type together with rhyolitic porphyries occupy the major part of the Grong–Olden Culmination east of Geitfjell (Figs. 1 and 2). The geochemistry and isotope geology of the Olden granite were discussed by Troeng (1982) and Stuckless et al. (1982) respectively. The Revsund granite occurs as large massifs east of the Caledonian Front and will be considered further below.

Results

U–Pb dating

Most zircons have subhedral outlines with well-developed prismatic shapes and less well-developed pyramidal terminations (Fig. 5a). The length/width ratio is approx-

| Table 1. Chemical composition (wt%) of the Geitfjell, Revsund and Olden granites. PERAL = the molar ratio of \( \text{Al}_2\text{O}_3 \) to \( (\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{CaO}) \); D.I. = Sum of normative quartz, albite and orthoclase (differentiation index of Thornton & Tuttle 1960); x = average value; SD = standard deviation. |
|--------------------------|--------------------------|--------------------------|--------------------------|--------------------------|
|                          | Geitfjell (5 samples)    | Revsund I (26 samples)   | Revsund II (99 samples)  | Olden (35 samples)       |
|                          | x                        | SD                       | x                        | SD                       |
| SiO₂                    | 71.48 ± 0.22             | 70.75 ± 1.02             | 69.89 ± 3.36             | 75.2 ± 2.20             |
| TiO₂                    | 0.33 ± 0.01              | 0.46 ± 0.22              | 0.52 ± 0.25              | 0.16 ± 0.09             |
| Al₂O₃                   | 14.10 ± 0.06             | 14.08 ± 0.72             | 14.08 ± 0.88             | 13.0 ± 0.09             |
| Fe₂O₃                   | 0.31 ± 0.03              | 0.41 ± 0.26              | 0.43 ± 0.31              | 0.49 ± 0.30             |
| FeO                     | 2.27 ± 0.08              | 2.80 ± 0.68              | 3.24 ± 1.26              | 1.10 ± 0.40             |
| MnO                     | 0.03 ± —                 | 0.07 ± 0.11              | 0.06 ± 0.05              | 0.04 ± 0.02             |
| MgO                     | 0.44 ± 0.02              | 0.59 ± 0.23              | 0.73 ± 0.45              | 0.18 ± 0.18             |
| CaO                     | 1.45 ± 0.03              | 1.64 ± 0.46              | 1.82 ± 0.69              | 0.60 ± 0.40             |
| Na₂O                    | 2.77 ± 0.02              | 2.76 ± 0.42              | 2.78 ± 0.54              | 3.70 ± 0.40             |
| K₂O                     | 5.72 ± 0.10              | 4.88 ± 0.45              | 4.61 ± 0.67              | 5.10 ± 0.40             |
| Rb (ppm)                | 248 ± 5                  | —                        | —                        | —                        |
| Sr (ppm)                | 133 ± 2                  | —                        | —                        | —                        |
| Zr (ppm)                | 194 ± 6                  | —                        | —                        | —                        |
| CIPW NORM               |                          |                          |                          |                          |
| Qt                      | 28.2 ± —                 | 29.7 ± 28.8              | 29.7 ± 31.9              |
| Qtₚ                     | 33.7 ± —                 | 28.8 ± 27.2              | 27.2 ± 30.1              |
| Al₂O₃                   | 23.5 ± —                 | 23.4 ± 23.5              | 23.5 ± 31.3              |
| An                      | 7.2 ± —                  | 8.1 ± 9.0                | 9.0 ± 3.0                |
| Or                      | 0.7 ± 1.3                | 1.2 ± 1.3                | 1.2 ± 0.3                |
| Hy                      | 3.4 ± 4.2                | 4.9 ± 4.9                | 4.9 ± 1.5                |
| En                      | 1.1 ± 1.5                | 1.8 ± 0.5                | 1.8 ± 0.5                |
| Ilₚ                     | 0.6 ± 1.0                | 1.0 ± 0.3                | 1.0 ± 0.3                |
| Mtₚ                     | 0.5 ± 0.6                | 0.6 ± 0.6                | 0.6 ± 0.6                |
| PERAL                   | 1.05 ± 1.10              | 1.09 ± 1.02              | 1.09 ± 1.02              |
| D.I.                    | 85.4 ± 81.9              | 79.5 ± 93.3              |


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Fig. 5. a. Zircon from the Geitfjell granite. Length of scale is 100 µm. b. Backscatter electron image of a sectioned zircon. Note the concordance between crystal growth planes and the compositional zoning. Length of scale is 100 µm.

approximately 2.5. The colour is light brown to light grey. Backscatter electron images of polished sections of zircons reveal a compositional zoning as dark and light areas in Fig. 5b. This variation is caused largely by differences in the average atomic weight of the elements present, with light areas being relatively enriched in heavier elements, probably hafnium. A majority of the zircons imaged are compositionally zoned and in most cases zoning is concordant to crystal growth planes, probably the result of successive growth of zircon. The isotopic data do not suggest the presence of any inherited components. No rounded cores indicating the presence of older detritial zircons were found. In some zircons there is a thin outer zone of zircon that is discordant to the crystallographic planes of the internal part of the grain. The discordant outer shell has an idiomorphic outline and possibly represents a recrystallized part or a metamorphic growth zone of the zircon.

The analytical data are shown in Table 2. The results from regression of different size-fraction combinations are summarized in Table 3. The oldest age is obtained by regression of all abraded zircon fractions. The age is 1828 ± 88 Ma (MSWD = 2.1) with a lower intercept of

Table 3. Summary of different alternatives.

<table>
<thead>
<tr>
<th>Alternative</th>
<th>( n )</th>
<th>Upper intercept age (Ma)</th>
<th>Lower intercept age (Ma)</th>
<th>MSWD</th>
</tr>
</thead>
<tbody>
<tr>
<td>All abraded zircon fractions</td>
<td>4</td>
<td>1828 ± 88</td>
<td>499 ± 188</td>
<td>2.1</td>
</tr>
<tr>
<td>All zircon fractions except 74–106 µm</td>
<td>8</td>
<td>1823 ± 22</td>
<td>487 ± 66</td>
<td>1.2</td>
</tr>
<tr>
<td>All zircon fractions</td>
<td>9</td>
<td>1795 ± 44</td>
<td>421 ± 114</td>
<td>4.1</td>
</tr>
<tr>
<td>All non-abraded zircon fractions except 74–106 µm</td>
<td>4</td>
<td>1771 ± 75</td>
<td>362 ± 170</td>
<td>0.1</td>
</tr>
<tr>
<td>All non-abraded zircon fractions</td>
<td>5</td>
<td>1758 ± 153</td>
<td>235 ± 771</td>
<td>4.1</td>
</tr>
</tbody>
</table>

Fig. 6. U–Pb concordia plot of zircons from the Geitfjell granite. Open circles = non-abraded zircon fractions; filled circles = abraded fractions. The zircon fractions are numbered in accordance with Table 2. MSWD value is 2.1.

Table 2. U–Pb analytical data.

<table>
<thead>
<tr>
<th>Fraction Sample no. (µm)</th>
<th>Concentration ( \text{U} ) (ppm)</th>
<th>( \text{Pb}_{\text{tot}} ) (ppm)</th>
<th>( 206^{\text{Pb}}/204^{\text{Pb}} )</th>
<th>( 207^{\text{Pb}}/204^{\text{Pb}} )</th>
<th>( 208^{\text{Pb}}/204^{\text{Pb}} )</th>
<th>( 206^{\text{Pb}}/238^{\text{U}} )</th>
<th>( 207^{\text{Pb}}/235^{\text{U}} )</th>
<th>( 208^{\text{Pb}}/235^{\text{U}} )</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 &lt; 45</td>
<td>1181</td>
<td>247</td>
<td>41800</td>
<td>0.1025</td>
<td>0.0342</td>
<td>0.21410 ± 58</td>
<td>3.0260 ± 124</td>
<td>1251</td>
<td>1414</td>
</tr>
<tr>
<td>2 45–74</td>
<td>1163</td>
<td>247</td>
<td>62200</td>
<td>0.1028</td>
<td>0.0347</td>
<td>0.21729 ± 50</td>
<td>3.0786 ± 83</td>
<td>1268</td>
<td>1427</td>
</tr>
<tr>
<td>3 74–106</td>
<td>1102</td>
<td>233</td>
<td>7675</td>
<td>0.1030</td>
<td>0.0364</td>
<td>0.21608 ± 37</td>
<td>3.0682 ± 68</td>
<td>1271</td>
<td>1431</td>
</tr>
<tr>
<td>4 Abraded</td>
<td>1178</td>
<td>259</td>
<td>33300</td>
<td>0.1035</td>
<td>0.0364</td>
<td>0.22434 ± 47</td>
<td>3.2012 ± 77</td>
<td>1261</td>
<td>1425</td>
</tr>
<tr>
<td>5 &gt;106</td>
<td>1133</td>
<td>247</td>
<td>100000</td>
<td>0.1033</td>
<td>0.0346</td>
<td>0.22320 ± 51</td>
<td>3.1782 ± 153</td>
<td>1299</td>
<td>1452</td>
</tr>
<tr>
<td>6 Abraded</td>
<td>1115</td>
<td>245</td>
<td>64100</td>
<td>0.1036</td>
<td>0.0359</td>
<td>0.22466 ± 50</td>
<td>3.2009 ± 78</td>
<td>1303</td>
<td>1457</td>
</tr>
<tr>
<td>7 &gt;150</td>
<td>1164</td>
<td>251</td>
<td>39000</td>
<td>0.1029</td>
<td>0.0352</td>
<td>0.22014 ± 77</td>
<td>3.1240 ± 35</td>
<td>1283</td>
<td>1439</td>
</tr>
<tr>
<td>8 Abraded</td>
<td>1369</td>
<td>288</td>
<td>37200</td>
<td>0.1025</td>
<td>0.0313</td>
<td>0.21586 ± 108</td>
<td>3.0511 ± 153</td>
<td>1260</td>
<td>1420</td>
</tr>
</tbody>
</table>

*) Corrected for blank and mass spectrometer fractionation. **) Corrected for common lead, blank and mass spectrometer fractionation. Errors are ±2σ in the last digits. Mass spectrometer fractionation: \( \text{U} \), 0.10% per AMU; \( \text{Pb} \), 0.12% per AMU. Common lead: \( 206^{\text{Pb}}/204^{\text{Pb}} = 15.8 \), \( 207^{\text{Pb}}/204^{\text{Pb}} = 15.3 \), \( 208^{\text{Pb}}/204^{\text{Pb}} = 35.4 \) (Stacey & Kramers (1975) growth curve).
499 \pm 155 \text{ Ma} \text{ (Fig. 6), and is considered as the best age estimate of crystallization of the Geitfjell granite. When all nine zircon fractions are regressed, the discordia obtained has an upper intercept age of} 1795 \pm 42 \text{ Ma and a lower intercept age of} 421 \pm 19 \text{ Ma. The MSWD value for this line is 4.1. This slightly high MSWD is caused by the poor fit of the non-abraded 74–106 \mu m fraction. When the zircons of this fraction were abraded and analysed, the data obtained fitted the discordia very well. If the non-abraded 74–106 \mu m fraction is excluded, the age obtained is 1771 \pm 31 \text{ Ma, with a lower intercept age of} 362 \pm 170 \text{ Ma. The U and Pb contents in the different zircon fractions are quite uniform except for the non-abraded > 150 \mu m fraction, which has a considerably higher U and Pb content. All abraded zircon fractions, except the > 150 \mu m fraction are less discordant than non-abraded zircons of the same size fraction.}

There are very small differences in the U and Pb contents of the non-abraded zircons and the internal parts of the zircons (Table 2). However, the abrasion of the zircons results in a slightly higher upper intercept age, suggesting that there is a small age difference between the internal and outer parts of the zircons. The regression of the four abraded zircon fractions, yielding an age 1828 \pm 88 \text{ Ma, is therefore the best estimate of the age of the Geitfjell granite. The abrasion also reduces possible influence on the age calculation by recent lead loss in the outer parts of the zircons.}

**Rb–Sr dating and Sm–Nd analysis**

Among the 10 analysed whole-rock samples, three were collected from migmatitic neosomes, the other samples represent different types of granite or rocks clearly related to the formation of the Geitfjell granite.

The Rb–Sr data are listed in Table 4 and plotted on an isochron diagram (Fig. 7), where a clear difference between the migmatitic and the non-migmatitic samples is seen. The seven non-migmatitic rocks yield an age of 1770 \pm 62 \text{ Ma with an initial} ^{87}\text{Sr}/^{86}\text{Sr} \text{ ratio of} 0.7094 \pm 0.0024, \text{ MSWD = 4.0. The data from the three migmatitic samples fail to yield a well-defined regression line but scatter around a line corresponding to an age of} 1670 \text{ Ma (Fig. 7).}

The Geitfjell granite has not been dated using the Sm–Nd method, but a Sm–Nd whole-rock analysis was made in order to obtain some genetic information on the source material of the granite. The Sm–Nd analytical results are shown in Table 5.

**Discussion**

**Geitfjell granite**

The 1828 \pm 88 \text{ Ma upper intercept age is interpreted as the intrusion age of the Geitfjell granite. Zircon dating of rocks from northern Vestranden has given the lower intercept ages 434 Ma, 393 Ma and 366 Ma (Schouenborg et al. 1991) and in the Roan area 352 Ma, 372 Ma and 104 Ma (Johansson & Möller, in prep.). All these ages except the last one are within the time interval characterized by high grade metamorphic conditions or subsequent cooling (Johansson et al. 1987; Möller 1988; Dallmeyer et al. 1992). The 499 \pm 155 \text{ Ma lower intercept age (Table 3) most likely reflects the Caledonian disturbance of the U–Pb system.}

The age of the Geitfjell granite gneiss is so far the oldest documented age recorded in the Western Gneiss
Region. For the reconstruction of the westerly continuations of different orogenic regions in the Baltic Shield an important question is whether rocks of similar age also occur farther south in the Western Gneiss Region or whether they are restricted to the northernmost part of the region. O'Nions (1973) and Brueckner (1972) reported Rb-Sr ages of 1900 Ma and 1840 Ma respectively for gneisses in the central part of the Western Gneiss Region. These age determinations are, however, of doubtful significance as they have large uncertainties and are poorly documented. Kullerud et al. (1986) considers ages around 1750 Ma to be the maximum for gneisses from the southern and central part of the Western Gneiss Region.

The 1828±65 Ma age of the Geitfjell granite shows that it is not related to the nearby granites of the Grong–Olden Culmination or the Tömerrås Window. Furthermore, the Olden granite is chemically different from the coarse-grained Geitfjell granite.

In contrast, the Geitfjell granite is very similar to the Revsund granite, which intrudes metasedimentary and volcanic rocks of the Bothnian Basin in north-central Sweden (Fig. 2). The Revsund granite massifs are very large and consist of many smaller plutons, but they are petrographically and geochemically rather homogeneous over large areas. The type Revsund granite is a grey, coarse-grained porphyritic rock with K-felspar megacrysts up to 6–7 cm in size. Eqigranular and reddish varieties occur locally. The granite is located enriched in U, W, Sn and Mo (Gavelin 1955; Wilson & Åkerblom 1982; Wilson et al. 1985). Claesson & Lundqvist (1987, 1990) subdivided the granites within the Bothnian Basin into early, late, and post-orogenic granites. They consider the Revsund granite to be post-orogenic in relation to the Svecofennian orogeny. Geochemistry and oxygen isotope data indicate a major petelic component in the source of the granite (Wilson et al. 1985). Claesson & Lundqvist (1990) stated that, based on geochemical and isotopic data, it is not very likely that the Revsund granite formed by melting of the surrounding greywackes – an opinion proposed by Gavelin (1955) and Svensson (1970).

The geochemistry of different Revsund plutons has been studied by Svensson (1970, 1979), Einarsson (1978) and Persson (1978). The Revsund I group (Table 1) includes 26 selected analyses of grey coarse-grained porphyritic granite, with SiO₂ contents close to the average of the Geitfjell granite. The average composition of the Geitfjell granite is within the range of the Revsund average, plus or minus the standard deviation. The only exception is K₂O, which is somewhat higher in the Geitfjell samples. The Revsund II group comprises 99 analyses of grey Revsund gneisses from the entire outcrop area in the Bothnian Basin.

The ages of some Revsund granite massifs have been determined by the U–Pb zircon method. The granites formed in the interval 1770–1800 Ma ago (Wilson et al. 1985; Patchett et al. 1987; Claesson & Lundqvist 1990). These datings are in most cases based on non-abraded zircons and may therefore also include younger components of the zircon rims. The regression of non-abraded zircons from Geitjellet yielded and age of 1758±12 Ma (Table 3). If the age of the Geitfjell granite only is considered, it can be classified as a Late Svecofennian granite corresponding in time, for instance, to the 1825±5 Ma old Hännö granite (Claesson & Lundqvist 1990) of the Bothnian basin. The Hännö granite is, however, finer grained and equigranular, the εNd value is generally lower and the zircons are different in morphology compared to zircons of the Revsund granite.

The Δ²⁶⁷⁷⁰⁶⁸Sr initial ratio of the Revsund granite (0.7074±0.0006) (Welin et al. 1971) is close to the initial Δ²⁶⁷⁷⁰⁶⁸Sr ratio of the Geitfjell granite (0.7094±0.0024). The εNd values and depleted mantle model ages of Svecofennian granites intruding the Bothnian Basin and of the Geitfjell granite are summarized in Table 6. The εNd value of the Geitfjell granite is −0.04, which is close to the highest εNd values obtained for the Revsund granites, by Wilson et al. (1985), and somewhat higher than values obtained by Patchett et al. (1987) and Claesson & Lundqvist (1987). A minor change (a few per cent) in the proportions of a mantle component and an old crustal component (particularly if the crust is Archean) in the source of the granite may lead to a significant change in the εNd value of the granite. Results from several investigations suggest that northern Scandinavia during Mid-Proterozoic time was underlain by a deformed mantle (Wilson et al. 1985; Huhma 1986; Claesson 1987; Patchett et al. 1987). The negative εNd value suggests that the Geitfjell magma was derived from a source that had a long residence time within the crust. This is also supported by the comparatively high initial Sr ratio.

Occurrence of coarse-grained grey granite of the Geitfjell type also has been reported farther west and north-west in the Vestranden area. These granite bodies are probably much larger than the Geitfjell granite (A. Solli, pers. comm., Namsos map sheet; L. Johansson, unpublished results).

Farther north, in the Kolvereid area (Fig. 2), zircons from tonalitic gneiss have yielded an age of 1819±6 Ma, with a lower intercept age of 393±35 Ma (Schouenborg

### Table 6. εNd values and depleted mantle model ages of the Geitfjell granite and granites in the Bothnian Basin.

<table>
<thead>
<tr>
<th>Sample</th>
<th>εNd</th>
<th>TDM (Ma)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Geitfjell</td>
<td>−0.04</td>
<td>2170</td>
<td>This paper</td>
</tr>
<tr>
<td>Storuman 1336</td>
<td>−0.40</td>
<td>2090</td>
<td>Wilson et al. 1985</td>
</tr>
<tr>
<td>Storuman 1347</td>
<td>+0.29</td>
<td>2030</td>
<td>Wilson et al. 1985</td>
</tr>
<tr>
<td>Storuman 1368</td>
<td>+0.27</td>
<td>2080</td>
<td>Wilson et al. 1985</td>
</tr>
<tr>
<td>Joran 1395</td>
<td>−0.50</td>
<td>2090</td>
<td>Wilson et al. 1985</td>
</tr>
<tr>
<td>Grundfors 1138</td>
<td>−0.20</td>
<td>2150</td>
<td>Wilson et al. 1985</td>
</tr>
<tr>
<td>Sörbygden SW 49</td>
<td>−2.02</td>
<td>2300</td>
<td>Patchett et al. 1987</td>
</tr>
<tr>
<td>Gällö SW 50</td>
<td>−2.42</td>
<td>2280</td>
<td>Patchett et al. 1987</td>
</tr>
<tr>
<td>Pilgrimstad SW 51</td>
<td>−1.43</td>
<td>2180</td>
<td>Patchett et al. 1987</td>
</tr>
<tr>
<td>Hammadal SW 54</td>
<td>−2.78</td>
<td>2310</td>
<td>Patchett et al. 1987</td>
</tr>
<tr>
<td>Strømstad SW 55</td>
<td>−1.10</td>
<td>2230</td>
<td>Patchett et al. 1987</td>
</tr>
</tbody>
</table>

* At the age of intrusion (see original papers). **TDM = depleted mantle model age (DePaolo 1981).
et al. 1991), whereas three Rb–Sr datings of basement gneisses (Priem et al. 1968) gave ages in the interval 1763–1864 Ma (recalculated for a Rb decay constant of 1.42 10⁻¹¹ year⁻¹).

The correlation between Geitfjell and Revsund granites is also supported by the similarity between the granites in the Borgefjäll Window and the Revsund granites (Kaatsky 1948; Zachrisson 1969; Gustavsson 1973; Greiling 1974, 1982). The Borgefjäll Window (Fig. 2) is situated approximately halfway between the Revsund province in Västerbotten and Vestranden. The window is dominated by coarse porphyritic granite of the Revsund type (Greiling 1982). There are also Cambrian rocks of supracrustal origin, which may correspond to the metasedimentary and metavolcanic rocks of the Bothnian Basin farther east. Granodiorites and syenites occur in subordinate amounts.

The Rb–Sr isotope system in the Geitfjell migmatite samples is clearly disturbed. The scatter around the 1670-Ma reference line (Fig. 7) may be explained if the entire Vestranden–Grong–Olden region is considered. Numerous granites intruded the area in the time interval 1650 ± 30 Ma. The Olden granite (ca. 1650 Ma; Stuckless et al. 1982; Stuckless & Troeng 1984), the Ingdal granite (1664 Ma; Tucker & Krogh 1988), the Roan granitoids (ca. 1650 Ma; Johansson & Möller; in prep.) and granite dykes in the Osen region (Fig. 2) (ca. 1630; Schouenborg et al. 1991) were all intruded around 1650 Ma in the Grong–Olden and Vestranden areas. Also, within the allochthonous nappes in the central Caledonides, there occur granites broadly coeval with these ca. 1650-Ma-old granites (ca. 1685 Ma: Tännäs Augen gneiss, Claesson 1980). The formation of migmatites at Geitfjell may be caused by increased temperatures in the crust during the intrusion of the granites.

In conclusion, the Geitfjell granite is a slightly older counterpart to the late to post-Svecofennian Revsund granites intruding the Bothnian Basin in north-central Sweden.

**Vestranden**

The recognition of rocks of Svecofennian affinity in Vestranden supports the common opinion that Vestranden crystalline rocks belong to the Baltic Shield and not to an ‘exotic’ crustal block accreted to the shield during the Caledonian continent-continent collision as indicated in tectonic models by Mykkeltveit et al. (1980) and Gilotti & Hull (1993).

The Vestranden area has generally been considered as a part of the Southwest Scandinavian orogenic province (Fig. 1) (Gaål & Gorbatschev 1987) or as a continuation of the Transscandinavian Granite Belt (Tucker et al. 1991). The Southwest Scandinavian orogenic province is characterized by major crustal formation in the interval 1750–1550 Ma, and subsequent repeated deformation, metamorphism and igneous activity in the time intervals 1500–1400 Ma, 1250–900 Ma (Sveconorwegian–Grenvillian orogeny) and 600–370 Ma (Caledonian orogeny) (Gaål & Gorbatschev 1987). There is no evidence that the northern Vestranden region has been influenced by any orogenic event between 1600 Ma and 700 Ma. A large part of the Vestranden region north of the Trondheim fjord is made up of granitoids >1800 Ma old, similar to those of the central Svecofennian subprovince, intruded by younger granites ca. 1650 Ma old. Available age data on the younger granites suggest a very short and intense period of granite formation at around 1650 Ma.

Some of the ca. 1650-Ma-old intrusions in Vestranden include rocks of monzonitic composition (Roan area), which also are typical of granitoid suites of the Trans- scandinavian Granite Belt. These granites in Vestranden could either belong to the younger group, defined by Larson et al. (1991), of granites of the Transscandinavian Granite Belt, or be easterly located ca. 1650-Ma-old granites of the Southwest Scandinavian orogenic province. The choice between these two alternatives is not easy because the relationships between these two regions, especially with respect to the age of crust formation, are a matter of debate and far from clear.

Rocks that have been derived from areas west of Vestranden and which now occur in the far-travelled Sever Nappes of the Caledonian Upper Allochthon originated from a source area dominated by rocks with ages in the range 1730–1400 Ma, with a minor component at least as young as 1000 Ma (Williams & Claesson 1987). These data indicate the presence of a terrane typical of the Southwest Scandinavian orogenic province along the Atlantic margin of the Baltic Shield. This also fits well with the evidence of Sveconorwegian metamorphism and sedimentation on the Lofoten islands (Griffin et al. 1978) and magmatism and metamorphism in the Bergen Arc (Cohen et al. 1988).

In terms of the traditional subdivision of the Precambrian of the Baltic Shield the Vestranden area north of the Trondheim fjord is best described as a composite region built up of Svecofennian granitoids >1800 Ma old and intruded by granitoids ca. 1650 Ma old of probable Transscandinavian Granite Belt affinity, which finally were subject to partial reworking during the Caledonian orogenic event. This scenario is different from earlier models in that it considers parts of the northernmost Vestranden area as a westerly continuation of the Svecofennian orogenic domain rather than as the northernmost part of the Southwest Scandinavian orogenic province (Gorbatschev 1985, Gaål & Gorbatschev 1987) or a continuation of the post-Svecofennian Trans- scandinavian Granite Belt (Tucker et al. 1991).

There is a possibility that not only the granites of the Bothnian Basin but also greywackes, pelites, pillow lavas, metavolcanites and other Svecofennian supracrustal rocks may occur in Vestranden. Rocks of Bothnian affinity may also have been tectonically incorporated into the Caledonian nappes that were derived from the outer
parts of the Baltoscandian margin. These rocks would then be ‘hidden’ in the numerous ‘cover’ sequences that occur in many places in the Vestranden area, their true nature being obscured by Caledonian metamorphism and deformation. Since the Bothnian Basin may represent an outer arc accretionary prism environment (Wilson et al. 1985), the rocks may, even if they are completely petrogenetically unrelated, share many petrological features with the rocks of the much younger Caledonian outboard terranes. It would then be very difficult to distinguish between rocks from the two different sources, particularly in their present deformed and metamorphosed state.

In conclusion, there are petrographical, geochemical, isotopic and geochronological similarities between the Geitfjell and Revsund granites. Based on these facts, on the occurrence of other Svecofennian granitoids in Vestranden and the absence of ages and orogenetic events typical for the Southwest Scandinavian orogenic province, it is suggested that the northern and eastern part of Vestranden, north of the Trondheims fjord, is a western continuation of the Svecofennian subprovince of central Sweden, intruded by younger granites probably belonging to the youngest phase of granite formation within the Transscandinavian Granite Belt.

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Appendix

Analytical procedures for radiometric dating

U–Pb and Rb–Sr isotopic analyses were performed in 1984 at the Museum of Natural History in Stockholm and the sample for Sm–Nd was analysed in 1988 at Mineralogisk–Geologisk Museum in Oslo.

U–Pb. Zircons were separated from the granite sample using standard mineral separation techniques. The least magnetic fraction was further split into size fractions from which impurities, inclusion-rich or irregularly shaped crystals were removed by hand-picking. U and Pb were extracted from the zircons and separated for mass spectrometric analysis following the method of Krogh (1973), with some modifications which included electrolytic deposition of Pb (described in Christiansson 1982). Uranium was measured on a MAT 261 mass spectrometer, which for measurements of NBS 987 gave a mean value of 0.71025 ± 3 during the time of data collection. The data were corrected for fractionation by normalizing the 86Sr/88Sr ratio to 0.1194. From these measurements 87Rb/86Sr ratios were calculated with a precision of 0.6%. The Rb–Sr whole-rock ages and initial 87Sr/86Sr ratios were calculated following the method of Provost (1990) using a decay constant of 1.42 × 10^{-11} year^{-1} for 87Rb (Steiger & Jäger 1977). Errors in Rb–Sr ages are given as 2σ expanded errors Clifford (1973).

Sm–Nd. A Sm–Nd whole-rock analysis was made on one of the samples collected for geochemical analysis. The analytical procedure essentially followed that described by Meams (1986). Isotopic ratios for Nd were normalized to 146Nd/144Nd = 0.7219. The decay constant used for 147Sm = 6.54 × 10^{-12} year^{-1}. The eNd values are calculated relative to CHUR, with present-day 147Sm/144Nd = 0.1967 and 143Nd/144Nd = 0.512638.

Brief descriptions of the Rb–Sr whole-rock samples

All Rb–Sr samples consist of quartz, K-feldspar, plagioclase and biotite. Accessory minerals are muscovite, apatite, zircon, ± allanite, ± garnet, ± chloride. The samples for U–Pb dating and geochemical analysis were taken at UTM coordinate UM 685 478 and for Rb–Sr dating at UM 662 460. These coordinates refer to map sheet 1823 IV Grong 1:50,000.

Sample 1. Pegmatite, forming pods and patches in the granite. The pegmatite pod is not intruding the granite but represents a late magmatic phase in the crystallization of the granite. It is slightly foliated.

Sample 2. Reddish, well-foliated augen gneiss. The augens consist of numerous small microcline crystals. Many augens are strongly elongated and flattened. The average diameter of the least deformed augens are approximately 20 mm. Plagioclase, quartz and biotite occur as inclusions in the augens. The matrix is dominated by quartz, K-feldspar, plagioclase and biotite.

Sample 3. (Neosome) White, medium-grained and even-grained granite from a migmatite neosome. The granite is only slightly foliated. Biotite is very subordinate. Small-garnets (<0.5 mm) are common.

Sample 4. (Neosome) Reddish, medium-grained granite with well-developed granitic texture and almost no foliation visible. The sample comes from a migmatite neosome.

Sample 5. Hybrid rock occurring along the margin of a xenolith, partly assimilated in the granite. The contact zone is 10–15 cm wide. The rock is light grey, fine-grained and rich in biotite and plagioclase.

Sample 6. (Neosome) White to light grey granite from a migmatite neosome. The rock has a diffuse banding with centimetre-wide bands somewhat richer in biotite. Small garnets (<0.5 mm) are common.

Samples 7–10. Coarse augen gneiss with K-feldspar megacrysts up to 6 cm across. These four samples are almost identical to the rock sampled for the zircon dating.