Conditions of pluton emplacement and anatexis in the Caledonian Bindal Batholith, north-central Norway

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The Velfjord plutons are mafic-intermediate plutons in the Bindal Batholith that were emplaced into pelitic and calcareous rocks of the Helgeland Nappe Complex of the Uppermost Allochthon at ~448 Ma. Al-in-hornblende barometry suggests emplacement in the range of 6 to 8 kbar, which is consistent with the presence of magmatic epidote. Pelitic rocks in the aureoles underwent partial melting and leucosomes locally accumulated adjacent to the plutons to form "contact granites". The results of garnet-aluminosilicate-silica-plagioclase (GASP) thermobarometry combined with a recently published petrogenetic grid for dehydration partial melting of pelites suggest a clockwise P-T path for the wall rocks. The GASP results, plus the presence of early kyanite, suggest initial melting in the 7 to 8 kbar range. Temperatures close to the plutons reached at least 750°C to 800°C. Nearly isothermal decompression to ~4 to 5 kbar resulted in the hottest wall rocks reaching cordierite stability while they were still partly molten. Cooler wall rocks could not reach conditions of cordierite stability because of the positive slope of the cordierite-in-equilibrium. The state of HNC crust prior to pluton emplacement cannot yet be determined. However, rapid exhumation of the plutons and their wall rocks suggests that the youngest plutons of the Bindal Batholith were probably emplaced at lower pressures.

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Introduction

The Bindal Batholith is the largest Caledonian-age batholith in Norway (Fig. 1). It consists of dozens of plutons emplaced into the Helgeland Nappe Complex (HNC) from about 481 Ma to 430 Ma (Nordgulen et al. 1993; Pedersen et al. 1999). The Batholith contains a remarkably wide variety of rock types and compositions, from gabbro to granite (Nordgulen 1993). Elemental data (Nordgulen 1993; Barnes et al. 1992) and isotopic studies (Nordgulen & Sundvoll 1992; Birkeland et al. 1993) indicate that many of the plutons are either the products of crustal melting or that they contain a significant crustal component. Therefore, an understanding of the origin of the Batholith, and the evolution of the HNC, requires knowledge of the style and conditions of crustal melting during Caledonian time. In Velfjord, heat from mafic to intermediate plutons caused partial melting of their metapelitic wall rocks at about 448 Ma. Detailed study of these rocks can provide knowledge of melting reactions, the effect of deformation on melt extraction, and heat transfer in the crust. To that end, this paper presents the results of thermobarometric studies of pluton emplacement. The results suggest that the Velfjord plutons were emplaced during a clockwise P-T path and that exhumation followed while some of the wall rocks were still partly molten.

Geologic setting

The Bindal Batholith (Fig. 1) is the northernmost of two Ordovician-Silurian batholiths in north-central Norway; the other is the Smøla-Hitra Batholith west of Trondheim (Gautneb & Roberts 1989). The host rocks of the Batholith are part of the Helgeland Nappe Complex (HNC, Fig. 2) (Barnes et al. 1992; Gustavson 1978; Kollung 1967; Myrland 1972; Thorssnes & Løseth 1991), which is the structurally highest unit of the Uppermost Allochthon. The nappe is composite and consists of Late Proterozoic (Trønnes 1994), medium-to high-grade pelitic, semipelitic, calc-silicate, and calcareous rocks, and Early Ordovician ophiolite complexes along with their unconformable cover sequences of medium-to low-grade schists, metasedstones, metaconglomerates, and calc-silicate rocks (Thorssnes & Løseth 1991). Amalgamation of these units in the Velfjord area took place prior to emplacement of the Andalshatten pluton at about 448 Ma (Nordgulen et al. 1992; Ø. Nordgulen, pers. comm.) and therefore prior to intrusion of the coeval Velfjord plutons (see below), Eastward translation of the HNC, including the Bindal Batholith, across low-grade rocks of the Upper Allochthon took place in the Late Silurian to Devonian Scandan phase of the Caledonian orogeny (e.g. Stephens et al. 1985). Plutonic rocks in the Bindal Batholith range from gabbro to granite, with intermediate to felsic calc-alkaline and alkali-calcic compositions most common. Although most plutons can be classified as I-type (Nordgulen 1993), reconnaissance Pb, Sr, and Nd isotopic studies indicate the presence of metasedimentary and meta-igneous crustal sources and a mantle component. The large Velfjord plutons are, from north to south, Hilstadfjellet, Akset-Drevli, and Sausfjellet (Fig. 2, Barnes et al. 1992). The Sausfjellet pluton is predominantly dioritic whereas the Akset-Drevli pluton is Fe-rich.
Fig. 1. Regional geologic setting of the Bindal Batholith. HNC, Helgeland Nappe Complex; RNC, Rödingsfjället Nappe Complex; KNC, Kidnappe Complex.
Fig. 2. Simplified geologic map of the Velfjord plutons, showing locations of analyzed samples.
diorite and gabbro (Fig. 2). A 1-km-long body of olivine pyroxenite that crops out south of the Akset-Drevli pluton is considered to be part of that plutonic sequence. The Hillstadfjellet pluton was emplaced in two stages: an early gabbroic Stage 1 and a later, more voluminous, monzonitic Stage 2 (Fig. 2).

The small monzonitic Aunet pluton underlies an area north of the Akset-Drevli pluton (Fig. 2) and crosscuts a sequence of gabbroic and dioritic dikes from that pluton and from Stage 1 of the Hillstadfjellet pluton. The Aunet monzonite is compositionally similar to Stage 2 of the Hillstadfjellet pluton (Barnes et al. 1992). Other small mafic bodies crop out southeast of Sausfjellet and west of Sørfjord (Myrland 1972; Thorsnes & Løsseth 1991).

U-Pb ages on zircon fractions from the Velfjord plutons yield ages of 447 Ma and 448 Ma (Pedersen et al. 1999). These ages are comparable with other U-Pb ages from the batholith, which range from 447 Ma to 430 Ma (Nordgulen & Schouenborg 1990; Nordgulen et al. 1993). In addition to this Scandian activity, new U-Pb ages of granites from the Velfjord area are as old as 481 Ma (Pedersen et al. 1999), which indicates that granitic magmatism in the Bindal Batholith lasted longer and was more complex than originally thought. Furthermore, two fractions of elongate, prismatic zircons from a porphyritic granite adjacent to the southeastern side of the Hillstadfjellet pluton near Lisjøen (Fig. 2) yielded a concordant zircon age of 475 ± 3 Ma (Pedersen et al. 1999). Whether this age indicates the age of emplacement of the granite or whether the zircons are inherited is a topic that will be partially addressed in this paper and is the subject of on-going ion probe analysis.

At the level of exposure, the host rocks of the Velfjord plutons consist entirely of the high-grade Late Proterozoic unit (Fig. 2). Pelitic lithologies in this unit are commonly migmatitic, but some quartz-rich lithologies lack evidence of partial melting. The age of the (apparent) regional migmatization event has been thought to predate emplacement of the Velfjord plutons, but its precise age awaits detailed geochronological data.

Contacts between the Velfjord plutons and calcareous wall rocks are sharp and locally brecciated, but where pelitic gneisses would normally be in contact with a pluton, the contact zone consists of kilometer-length sections of a distinctive assemblage of porphyritic granitic rocks, diatexite (disrupted migmatite that lacks internal structure), and stromatic migmatite (Fig. 2). Contacts between these distinctive assemblages and the Velfjord plutons are generally sharp. Dikes of the porphyritic granite cut the Akset-Drevli pluton and Stage 1 of the Hillstadfjellet pluton and decimeter-scale dikes of diatexitic intrude the margins of the Hillstadfjellet and Sausfjellet plutons. The porphyritic granitic rocks range from granite to tonalite, but they are herein referred to as 'contact granites'.

The degree of migmatization decreases with distance from plutonic contacts. Adjacent to the plutons, either contact granite or highly disrupted diatexite is typical. These rocks grade into zones in which pelitic compositions are diatexitic and more siliceous rocks are stromatic migmatite. Further from contacts, pelitic migmatites are stromatic and siliceous rocks are unmigmatized or are vein migmatite. These distinct types of migmatite occupy layers within the wall rock sequence. These field relationships suggest that the contact granites resulted from in situ anatexis of pelitic wall rocks. The heat necessary to form the contact granites was apparently supplied by the mafic Velfjord magmas.

As the plutonic contacts are approached and the proportion of leucosome in the wall rocks increases, migmatitic veins and dikes more commonly crosscut compositional layers. The leucosomes range from granitic to tonalitic in composition and most are equigranular. However, a few leucosomes contain K-feldspar megacrysts (up to 2 cm long), with textures similar to those of the porphyritic contact granites. Where crosscutting relationships were observed, the K-feldspar-phyllic leucosomes are foliated and pinch out along strike of the foliation. These pinch-outs are tapered, not boundinaged, which suggests entrapment of the granitic magma during deformation. They are crosscut by later leucogranitic dikes.

Petrography

Plutonic rocks

Petrographic descriptions of the Velfjord plutons were presented in Barnes et al. (1992) and the mineral assemblages of analyzed samples are presented in Table 1. For the purposes of this account, amphibole petrography is particularly important. Amphibole varies in habit; it occurs as a rimming phase on pyroxene, as distinct prismatic to intergranular crystals, and as dispersed scrappy grains in granites. It is locally rimmed by biotite. In many samples, amphibole is color zoned, with olive green cores and blue-green rims. The width of these rims is variable, from a few microns to several tenths of a millimeter. In many evolved rocks, amphibole is entirely blue-green.

In addition to amphibole, many samples contain accessory epidote (Table 1), which occurs in two types. The first is as euhedral to subhedral weakly pleochroic (yellow) grains with allanite cores. These crystals are typically intergrown with biotite ± amphibole ± ilmenite. They commonly have euhedral faces against biotite and anhedral surfaces against plagioclase and quartz. The second type of epidote forms clear, equant to elongate, prismatic inclusions in plagioclase. Such inclusions are most common in plagioclase cores. In a few samples, the latter type of epidote occurs in plagioclase cores, whereas the former type is present as inclusions in plagioclase rims.

Migmatitic rocks

Stromatic migmatites are medium grained gneisses banded on a cm scale. They typically contain sillimanite, muscovite, biotite, quartz, and plagioclase ± garnet ± staurolite ± K-feldspar. Robust sillimanite prisms are sparse and
some are partly replaced by muscovite. Acicular to granular sillimanite is more common; it occurs along crystal boundaries and as inclusions in muscovite. Staurolite and garnet are euhedral against micas and ragged where in contact with quartz and plagioclase. Muscovite typically replaces sillimanite and is intergrown with biotite. In the latter case, it commonly crossescuts foliation formed by the biotite.

Diatexitic migmatites are medium grained and contain the assemblage plagioclase + quartz + biotite + sillimanite ± K-feldspar (microcline) ± kyanite ± cordierite ± muscovite. Foliation is generally weak and is defined by discontinuous banding, with or without orientation of biotite and sillimanite. A ‘matrix’ of quartz, granular to interstitial plagioclase, sparse poikilitic K-feldspar, biotite, and sillimanite surrounds porphyroblasts of garnet and cm-scale sillimanite. Quartz is generally interstitial to subpoikilitic.

Sillimanite is present in four habits: as sparse, stubby, cm-scale grains, as mm-scale elongate prisms, as acicular inclusions in other silicates (particularly biotite), and less commonly as fibrolitic mats. The coarse sillimanite crystals typically show deformation-induced subgrain development. Prismatic sillimanite is generally subparallel to bedding or foliation and is locally bent or broken. Acicular sillimanite is generally unoriented, undeformed, and intergrown with biotite. Where present, kyanite occurs as ragged crystals, typically surrounded by biotite and quartz ± sillimanite. Some kyanite occurs as inclusions in cordierite.
Where cordierite is present, it varies from 4-mm long blocky crystals to \( \sim 0.5 \) mm diameter intergranular grains mostly enclosed in quartz (± sillimanite ± plagioclase). In all cases, cordierite is present in the leucocratic parts of the rock rather than in biotite-and garnet-rich zones. The cordierite is locally replaced by acicular sillimanite + pale biotite.

Garnet is poikiloblastic; its cores contain inclusions of quartz ± biotite ± sillimanite. Garnet rims are typically free of inclusions except for local concentrations of acicular sillimanite. Many crystals appear to be resorbed, whereas others have idiomorphic outlines. Individual crystals commonly show both of these features.

Reddish brown biotite rims and replaces cordierite and is intergrown with acicular and prismatic sillimanite. Much of the intergrown biotite and sillimanite are interstitial. Where muscovite is present, it replaces feldspars or forms discrete grains that crosscut the gneissic banding.

Anhedral ilmenite, tourmaline, zircon, monazite, apatite, chalcopyrite, and staurolite are accessories. The staurolite locally occurs as ragged mm-scale crystals. It is more common as fine euhedral prisms associated with intergranular biotite.

No petrographic break between regional and contact migmatites was observed. With approach to the plutons, the abundance and size of staurolite decreases, size and abundance of sillimanite increases, and the abundance of prograde muscovite decreases (is absent near contacts). Kyanite and cordierite were only observed in samples collected within 200 m of a plutonic contact. The lack of such a petrographic break does not necessarily imply synchronicity of regional and contact migmatization, but indicates that retrogression of a regional assemblage did not occur prior to emplacement of the Velfjord plutons. Ion probe studies of the timing of migmatitic events are in progress.

Leucosomes are typically medium-grained, leucocratic, garnet-bearing biotite granodiorite and granite. Garnet, muscovite, and sillimanite are variably present. A few leucosomes contain megacrysts of K-feldspar ± plagioclase that reach 2 to 3 cm in length. Original hypidiomorphic granular texture shows variable degrees of synmagmatic to subsolidus deformation. Quartz is interstitial to sub-poikilitic, with prominent subgrain development. Biotite and muscovite form ragged books. Primary (?) muscovite crystals commonly crosscut the weak foliation and are locally poikilitic. Secondary muscovite replaces K-feldspar and sillimanite. Garnet is poikilitic (or poikiloblastic?) and encloses quartz, or is present as crystal clusters intergrown with quartz. In addition to sillimanite, ilmenite and blocky zircon are accessory minerals.

Contact granites

Most contact granites are K-feldspar-phyric two-mica granite to granodiorite. Some also contain plagioclase phenocrysts. Phenocrysts reach at least 3 cm in length. Groundmass texture is hypidiomorphic granular with variable hypersolidus deformation. The contact granites are typically heterogeneous in terms of the proportion of K-feldspar phenocrysts. Many also contain accessory garnet and sillimanite, some of which occurs in clusters with biotite ± muscovite. The garnet is anhedral and sillimanite ranges from acicular inclusions to sparse euhedral crystals as much as 0.5 mm across. Ilmenite, zircon, and monazite are accessory minerals; sericite and chlorite are sparse secondary minerals.

Sample N17.91, a garnet-bearing biotite orthopyroxene tonalite, is an unusual contact "granite". Plagioclase phenocrysts up to 2 cm long are set in a groundmass of plagioclase, quartz, and sparse interstitial K-feldspar. Orthopyroxene (opx) crystals are generally ragged and locally have biotite rims. Garnet is anhedral (resorbed?) with straight contacts adjacent to quartz and ragged contacts adjacent to biotite. A symplectic garnet-quartz rim surrounded by biotite separates garnets from adjacent plagioclase phenocrysts. Zircon, ilmenite, and green spinel are accessory minerals.

Analytical methods

Major and trace element contents were determined by x-ray fluorescence analysis at the Norwegian University of Science and Technology (Barnes et al. 1992). Initial wavelength-dispersive electron microprobe analysis of amphibole and plagioclase in the plutonic rocks was done at the Mineralogical Museum, University of Oslo and at Southern Methodist University (Dallas). All other microprobe analyses were done at the Department of Geology and Geophysics, University of Wyoming. Analytical conditions were at a nominal accelerating voltage of 20 kV, beam current of 20 nA (0 for feldspar), with a focused beam. Natural and synthetic standards were used; the data were reduced using the ZAF correction.

Analyses of amphibole, garnet, biotite, and plagioclase used in thermobarometric calculations were, whenever possible, made on adjacent rims. In general, mantle and core compositions were also determined. In migmatitic samples, biotite crystals adjacent to garnet, as well as ones at some distance from garnet, were analyzed.

Mineral compositions

Representative compositions of minerals analyzed for thermobarometric calculations are presented in Tables 2 to 6. Plagioclase anorthite contents are summarized in Table 1 and compositions used in the calculations are given along with the appropriate amphibole or garnet rim composition in Tables 3 and 5. For Al-in-hornblende barometry, plagioclase rim compositions were used. Plagioclase in the migmatitic rocks ranges widely in composition, but zoning in individual samples is commonly weak. Where present, zoning in plagioclase crystals adjacent to garnet is reversed, with slightly more calcic
Table 2. Representative average pyroxene compositions.

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1. Plutons identified as: H1, Hillstadfjellet stage 1; H2, Hillstadfjellet stage 2; A, Aunet; Akset-Drevli, Akset-Drevli; S, Sausfjellet.
2. Sample N17.91 is a garnet biotite opx tonalite 'contact granite', see text.
3. n.a. indicates element not analyzed.

Table 3. Representative amphibole analyses.

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1. Plutons are H1, Hillstadfjellet stage 1; H2, Hillstadfjellet stage 2; A, Aunet; Akset-Drevli, Akset-Drevli; S, Sausfjellet.
2. Sample N17.91 is a garnet biotite opx tonalite 'contact granite', see text.
3. Composition of coexisting plagioclase, given as Ca/(Ca + Na).
Table 4. Representative epidote analyses.

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Cations per 25 oxygens:

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<tr>
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</table>

1 Pluton abbreviations as in Table 2. p indicates a primary epidote. -s indicates a secondary epidote. -pi indicates a primary epidote inclusion in plagioclase rim, -si indicates a secondary epidote in plagioclase core.

Minerals from the plutonic rocks

Pyroxene and olivine. – Average pyroxene compositions are given in Table 2 and plotted in Fig. 3. Clinopyroxene (cpx) is augitic and orthopyroxene (opx) is intermediate in more Mg-rich olivine could have equilibrated with Fe-rich composition (Fig. 3). Cpx and opx from a gabbroic dike in residual liquid during slow cooling, whereas opx did not.

Table 5. Representative garnet rim compositions.

<table>
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<tr>
<th></th>
<th></th>
<th></th>
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Cations per 12 oxygens:

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1 Samples N17.91 and N19.91 are contact 'granites'. Other samples are diatexite (diatex.) and layer-bound migmatite (migm.).
2 Composition of coexisting plagioclase, given as Ca1(Ca+Na).
3 n.a. indicates not analyzed.
Table 6. Representative biotite analyses.

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Rock type: tonalite - granite, diatex. - dike.

Cations per 22 oxygens

| Si     | 5.363 | 5.280 | 5.305 | 5.304 | 5.351 | 5.304 | 5.397 | 5.230 | 5.317 |
| Ti     | 0.493 | 0.375 | 0.138 | 0.284 | 0.253 | 0.326 | 0.182 | 0.194 | 0.204 |
| Fe     | 2.795 | 2.585 | 1.939 | 2.185 | 1.985 | 2.209 | 2.008 | 2.737 | 2.766 |
| Mg     | 2.274 | 1.847 | 2.879 | 2.447 | 2.693 | 2.293 | 2.781 | 2.094 | 1.929 |
| Mn     | 0.005 | 0.003 | 0.008 | 0.011 | 0.004 | 0.018 | 0.011 | 0.000 | 0.004 |
| Ca     | 0.001 | 0.000 | 0.002 | 0.000 | 0.003 | 0.000 | 0.000 | 0.000 | 0.000 |
| Na     | 0.009 | 0.030 | 0.099 | 0.050 | 0.086 | 0.064 | 0.047 | 0.053 | 0.064 |
| K      | 1.910 | 1.941 | 1.706 | 1.824 | 1.756 | 1.872 | 1.718 | 1.766 | 1.832 |
| Cl     | 0.036 | 0.010 | 0.012 | 0.013 | 0.002 | 0.004 | 0.023 | 0.013 | 0.068 |
| Total  | 15.790 | 15.722 | 15.677 | 15.617 | 15.591 | 15.591 | 15.767 | 15.695 | 15.782 |
| Fe(Fe + Mg) | 0.551 | 0.583 | 0.402 | 0.472 | 0.424 | 0.491 | 0.419 | 0.567 | 0.589 |

Amphibole. - In the classification of (Leake et al. 1997), amphiboles are predominantly ferropargasite and ferro-edenite (Fig. 4). Amphiboles from relatively Mg-rich samples also encompass edenitic and paragasitic compositions. Amphibole from the Hillstadfjellet, Akset-Drevli, and Aunet plutons are Na-rich relative to amphibole from the Sausfjellet pluton, with >0.4 cations per formula unit (pfu) in the former plutons and <0.44 cations pfu in the latter (Table 3). Among samples that show core-to-rim zoning, Al is generally enriched in the rims relative to the cores.

Epidote. - Epidote compositions (Table 4) show little variation within single crystals except adjacent to allanite.

Fig. 3. Pyroxene compositions plotted in the pyroxene quadrilateral. Compositions of coexisting olivine are shown along the base of the upper figure.

Fig. 4. Classification of amphiboles from the Velfjord plutons and related granitic dikes (Leake et al. 1997).
Minerals from migmatites and contact granites

Temperature calculated on the basis of two-pyroxene-component from core to rim. Zoning as is apparent involves a 2-3% increase in grossular replaces plagioclase has Ps values from 0.0 to 0.2 (see inclusions in the core of a plagioclase phenocryst). From the Hillstadfjellet pluton (N40.91) clear epidote crystals ranges from 0.15 to 0.26 (ave. 0.21). In a sample from the Hillstadfjellet pluton (N40.91) clear epidote typically has Ps contents in the 0.23 to 0.29 range (Zen & Hammarstrom 1984), whereas epidote that replaces plagioclase has Ps values from 0.0 to 0.24 (see Brandon et al. 1996).

Minerals from migmatites and contact granites

Garnet. — Garnets from migmatitic wall rocks and contact granites are predominantly almandine (Table 5). The almandine component ranges from 69% to 84% (molecular), grossular component from 1% to 9%, pyrope component from 7% to 21%, and spessartine component from 1% to 15%. Garnets from the contact granites and diatexites show little zoning. For example, the value of Fe/(Fe+Mg) in garnet, which varies from 0.68 to 0.92 among all analyzed samples, typically has a standard deviation of less than 0.02 in an individual sample. Such zoning as is apparent involves a 2–3% increase in grossular component in garnet rims (Table 5). In such crystals these rims must be discontinuous because other analyses of rims from the same crystals are identical to the cores.

In contrast, garnets from a stromatic migmatite collected 1.5 km from the Sausfjellet pluton (sample N51.91B) show the opposite sense of zoning, with a decrease in grossular component from core to rim.

Biotite. — Representative biotite analyses are given in Table 6. Individual biotite crystals lack compositional zoning. However, biotites adjacent to garnet typically have lower values of Fe/(Fe+Mg) than those at some distance from garnet grains. The differences in Fe/(Fe+Mg) range from 0.04 to 0.10. The biotites are relatively Al-rich (Al >3.4 pfu), as is expected of biotite from peraluminous compositions. An exception to this is biotite from the opx-bearing contact granite (N17.91) in which Al contents are <3.0 pfu.

Temperature and pressure estimates from plutonic rocks

Pyroxene – olivine equilibria

Temperature estimates were made on the basis of two-pyroxene and olivine + pyroxene equilibria using the QUILF4 program (Frost & Lindsley 1992; Lindsley & Frost 1992). Ferric iron contents were estimated by recalculation to four cations (Cebriá Gómez 1990). Temperature calculated on the basis of two-pyroxene – olivine equilibria for samples from the Akset-Drevli pluton were 1060°C. Estimates made on the basis of two-pyroxene equilibria range from 900°C to 1015°C for Sausfjellet, 790°C to 1070°C for Akset-Drevli, and 810°C to 970°C for Hillstadfjellet.

Apatite saturation thermometry

The apatite saturation thermometer (Harrison & Watson 1984; Watson & Harrison 1984) permits calculation of the temperature at which apatite would be stable in a melt of the composition of a particular rock. This method assumes that apatite stability was not controlled by kinetic factors (Bacon 1989) and was not a cumulate phase. It also assumes that a given rock composition is representative of a melt composition.

The thermometric data suggest apatite saturation in the Hillstadfjellet pluton at about 975°C, in the Akset-Drevli pluton at about 1020°C, and in the Sausfjellet pluton at about 950°C. Although these temperatures are consistent with those determined with pyroxene-olivine equilibria, they do not necessarily indicate the temperature of emplacement.

Amphibole – plagioclase equilibria

The Al content of amphibole in equilibrium with plagioclase, quartz, biotite, Fe-Ti oxide, titanite, and K-feldspar has been shown to be a function of the pressure of crystallization (Hammarstrom & Zen 1986; Hollister et al. 1987; Schmidt 1992). Furthermore, Holland & Blundy (1994) showed that the compositions of coexisting amphibole and plagioclase are controlled by temperature-dependent equilibria. Anderson & Smith (1995) combined the two schemes to develop a temperature-corrected Al-inhornblende barometer.

Several samples from the Velfjord plutons contain the appropriate equilibrium assemblage and others lack only K-feldspar; however, Anderson & Smith (1995) indicated that the presence or absence of K-feldspar and titanite had no apparent effect on the results of the barometer. Application of the Anderson & Smith (1995) barometer results in P estimates that range from 4.1 to 7.7 kbar (Table 7). Amphiboles that yield pressure estimates ≥5.7 kbar are characteristically blue-green. Among the three samples that yield pressures <5.7 kbar, the amphiboles are olive green with blue-green rims a few microns wide. For two of these samples (91.13 and 91.14), it was not possible to determine rim compositions. Textural study of the other sample (gabbro NK74.99) suggests that amphibole was not in equilibrium with the complete buffer assemblage at the solidus.

Temperature estimates based on the pressure-corrected Holland & Blundy (1994) algorithm (Anderson & Smith 1995) range from 673°C to 757°C (Table 7). These values are consistent with final amphibole-plagioclase equilibria at near-solidus conditions.
Table 7. Results of Al-in-hornblende barometry.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Setting</th>
<th>P (kbar) 1</th>
<th>sd</th>
<th>P (kbar) 2</th>
<th>sd</th>
<th>T (°C) 3</th>
<th>sd</th>
<th>N 4</th>
<th>Notes 5</th>
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<tbody>
<tr>
<td>Sausfjellet</td>
<td>Granitic dike</td>
<td>7.6</td>
<td>0.4</td>
<td>7.0</td>
<td>0.3</td>
<td>710</td>
<td>20</td>
<td>8</td>
<td>Primary Ep</td>
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<td>91.22A</td>
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<td>7.7</td>
<td>0.7</td>
<td>686</td>
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<td>6.6</td>
<td>0.5</td>
<td>725</td>
<td>10</td>
<td>4</td>
<td>Primary Ep</td>
</tr>
<tr>
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<td>0.6</td>
<td>5.9</td>
<td>0.6</td>
<td>696</td>
<td>13</td>
<td>11</td>
<td>Ilm primary; Ttn secondary</td>
</tr>
<tr>
<td>NK74.99</td>
<td>Stage 1, Hill.</td>
<td>7.5</td>
<td>1.0</td>
<td>6.8</td>
<td>0.9</td>
<td>720</td>
<td>13</td>
<td>11</td>
<td>Lacks Kfs</td>
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<td>Hillstadfjellet, 2</td>
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<td>0.6</td>
<td>7.2</td>
<td>0.5</td>
<td>734</td>
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<td>8</td>
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</tr>
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<td>0.3</td>
<td>701</td>
<td>21</td>
<td>10</td>
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<tr>
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<td>0.2</td>
<td>4.5</td>
<td>1.5</td>
<td>673</td>
<td>8</td>
<td>4</td>
<td>Secondary Ttn</td>
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<tr>
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<td>Quartz monzonite</td>
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</tr>
<tr>
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<td>9</td>
<td>Hbl after Cxp &amp; Opx</td>
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<td>Hillstadfjellet, 2</td>
<td>Rocks in contact zones</td>
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<td>N40.91</td>
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<td>693</td>
<td>16</td>
<td>18</td>
<td>Primary Ep</td>
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<td>Hill. contact hybrid</td>
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<td>1.6</td>
<td>5.7</td>
<td>1.4</td>
<td>731</td>
<td>6</td>
<td>5</td>
<td>Primary Ep</td>
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<tr>
<td>N79.99A</td>
<td>Hill. contact hybrid</td>
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<td>2.4</td>
<td>7.0</td>
<td>2.0</td>
<td>757</td>
<td>14</td>
<td>11</td>
<td>Primary Ep; lacks Ttn, Kfs</td>
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<td>Aunet pluton and  dikes of Aunet-type monzonite</td>
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<td></td>
<td></td>
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<tr>
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<td>Monzonite</td>
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<td>0.2</td>
<td>6.8</td>
<td>0.2</td>
<td>678</td>
<td>3</td>
<td>2</td>
<td>Hbl after Cxp &amp; Opx</td>
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<tr>
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<td>Monzonite</td>
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<td>0.2</td>
<td>4.1</td>
<td>0.3</td>
<td>712</td>
<td>10</td>
<td>4</td>
<td>Hbl after Cxp &amp; Opx</td>
</tr>
<tr>
<td>N65.99</td>
<td>Monzonite dike</td>
<td>5.5</td>
<td>0.6</td>
<td>4.8</td>
<td>0.6</td>
<td>726</td>
<td>19</td>
<td>8</td>
<td>Hbl after Cxp &amp; Opx</td>
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<tr>
<td>Samples lacking equilibrium assemblage or clearly hybrid</td>
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<tr>
<td>75.01</td>
<td>Hillstadfjellet gabbro</td>
<td>5.8</td>
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<td>0.7</td>
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<td>Lacks Qtz, Ttn, Kfs</td>
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<td>Hillstadfjellet monzodiorite</td>
<td>4.2</td>
<td>0.1</td>
<td>3.9</td>
<td>0.3</td>
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<td>3</td>
<td>Lacks Qtz</td>
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<td>0.8</td>
<td>10.0</td>
<td>0.9</td>
<td>730</td>
<td>27</td>
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<td>Lacks Qtz, Ttn, Kfs</td>
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<td>91.24</td>
<td>Sausfjellet diorite</td>
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<td>4.3</td>
<td>0.5</td>
<td>812</td>
<td>10</td>
<td>3</td>
<td>Lacks Qtz, Ttn, Kfs</td>
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<tr>
<td>Mingled rocks in Hillstadfjellet contact zone</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>91.10A</td>
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<td>0.8</td>
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<td>5</td>
<td>Lacks Ttn, Kfs</td>
</tr>
<tr>
<td>91.10B</td>
<td>Quartz diorite</td>
<td>9.3</td>
<td>0.1</td>
<td>9.5</td>
<td>0.2</td>
<td>656</td>
<td>19</td>
<td>3</td>
<td>Lacks Ttn, Kfs</td>
</tr>
</tbody>
</table>

1 Pressure estimates based on rim compositions and Schmidt (1992) calibration; all Fe as Fe2+.
3 Temperature estimate based on Holland & Blundy (1994).
4 Number of samples used in calculation.
5 Mineral abbreviations are Kfs, K-feldspar; Hbl, calcic amphibole; Cpx, augite; Opx, orthopyroxene; Bi, biotite; Ttn, titanite; Qtz, quartz; Ep, epidote.

Epidote stability

Experimental studies of epidote stability (Naney 1983; Schmidt 1992, 1993) show that magmatic epidote in equilibrium with amphibole is stable at pressures greater than about 6 kbar. The widespread presence of such primary epidote in quartz-bearing rocks of the Hillstadfjellet and Sausfjellet plutons is consistent with the Al-in-hornblende pressure estimates >5.7 kbar cited above.

Temperature and pressure estimates from migmatitic rocks and contact granites

Zircon saturation temperatures

Zircon saturation thermometry (Watson 1979; Watson & Harrison 1983) can be applied to the leucosomes in the wall rock migmatites and the contact granites. This method assumes that the bulk rock compositions represent melt compositions and that no zircon in the rock is residual. Neither assumption is fully justified. For example, residual zircon can be entrained in both leucosomes and contact granites. In addition, if melting and melt extraction is rapid, zircon may not reach equilibrium with the partial melt (Watson 1996). Finally, leucosomes may be partial cumulates. Entrainment of residual zircon results in higher T estimates than expected, whereas lack of equilibrium and accumulation of other crystals result in lower T estimates. Given these caveats, the contact granites yield zircon saturation temperatures from 780°C to 900°C and granitic leucosomes yield temperatures from 710°C to 780°C.

GASP barometry

Migmatitic metapelites adjacent to the Velfjord plutons and some contact granites contain the assemblage garnet + sillimanite ± kyanite + biotite + plagioclase + quartz. This ‘GASP’ assemblage can be described by two equilibria: the Mg-Fe exchange between biotite and garnet and the discontinuous reaction:

\[ 3\text{CaAl}_2\text{Si}_2\text{O}_8 \leftrightarrow \text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 2\text{Al}_2\text{Si}_3\text{O}_5 + \text{Si}_2 \text{O}_3 \]

anorthite grossular Als quartz.

Simultaneous solution of these reactions results in an estimate of the P and T of equilibrium. The reactions were solved using the program TWQ (Berman 1991) and the

The results of GASP thermobarometry on contact granites and migmatites vary widely (Fig. 5). Because garnets in many migmatitic samples have rims with variable composition, it is possible to determine more than one pressure for a single sample. The highest pressure estimates (~7 to 9 kbar) result from the use of the most Ca-rich garnet rim compositions and plot on or near the kyanite = sillimanite boundary (Fig. 5), which is consistent with the presence of early kyanite in some samples. It is noteworthy that the sample that yields P ≥ 9 kbar (sample N103.99) is a diatexitic dike in the Sausfjellet pluton. Thus, this sample may have equilibrated at higher P and risen to its depth of emplacement.

In cases where plagioclase adjacent to garnet is zoned, reverse zoning is most common. The lower An core compositions result in this highest pressure estimates, which is consistent with the proposed P-T path (see below).

Many samples give GASP results in the 4 to 5 kbar range and a few give results that plot in the andalusite stability field (Fig. 5). Because the GASP thermobarometer is commonly affected by re-equilibration of biotite during retrogression, and because andalusite is absent in these samples, the GASP results are thought to reflect low-T equilibration of biotite. This is supported by the fact that the garnet – plagioclase equilibrium boundary for these samples projects toward high-P values at higher temperatures, it is the garnet – biotite equilibrium that has been reset. For all other samples, the equilibration temperature determined by GASP calculations ranges from 620°C to 730°C.

Discussion

A proposed P-T-t path

On the basis of Al-in-hornblende barometry and the presence of magmatic epidote, the pressure of emplacement of the Velfjord plutons is thought to be ≥7 kbar (Fig. 6A). If this is the case, the migmatitic wall rocks should also record pressures near 7 kbar.

The migmatitic wall rocks adjacent to the plutons contain a metamorphic assemblage that is unlikely to represent a single equilibrium P and T. This is particularly apparent in samples that contain kyanite, sillimanite, garnet, and cordierite. In the melting interval, mutual stability of kyanite and cordierite is unlikely (e.g. Spear et al. 1999; Fig. 6A). The assemblage suggests either a polymetamorphic history or metamorphism along a P-T loop. The latter interpretation is preferred because the two discrepant minerals, kyanite and cordierite, are absent in pelitic rocks collected outside the contact aureole of the plutons. This in turn suggests that these minerals grew in response to contact metamorphism (and anatexis) associated with Velfjord plutonism. Thus the metamorphic history recorded in the contact migmatites is thought to reflect variations in P and T during the late Ordovician intrusive event.

A recent contribution by Spear et al. (1999) outlined the equilibria that are likely to control melting of pelitic rocks in which a free H₂O phase is absent during anatexis (dehydration melting). Their petrogenetic grid (Fig. 6A) provides a useful context for discussion of the various estimates of P and T. Features of the contact migmatites that must be explained include:

1. Localized occurrence of kyanite and cordierite adjacent to the plutons and the fact that these minerals probably were out of equilibrium with each other.
2. Local growth of K-feldspar megacrysts in early stages of migmatization.
4. Discontinuous zoning of garnet rims and preservation of rim sections, some of which apparently grew at P ≥ 7 kbar and others that grew at ~4 to 5 kbar.
5. Evidence for resorption of garnet.
7. Local retrogression of sillimanite and feldspar to muscovite, especially in leucosomes and migmatites collected at some distance from the pluton.
8. Late-stage growth of fine-grained prisms of staurolite.

A clockwise P-T metamorphic path can explain these observations. Specific evidence that points to a clockwise,
rather than anti-clockwise, P-T path can be summarized as follows. Kyanite formed early in the metamorphic history, as shown by development of sillimanite beards on ragged kyanite and by kyanite inclusions in cordierite. Cordierite was a late-stage mineral, as shown by its idiomorphic to intergranular habit and its concentration in leucosomes. In samples in which residual plagioclase is zoned, the most sodic zones are in the cores or mantles of the crystals, not the rims. Reverse zoning is consistent with a clockwise path, whereas normal zoning suggests an anti-clockwise path.

Possible clockwise P-T paths are illustrated in Fig. 6B and are described below. So little is known about the regional pre-intrusion metamorphic gradient that the P-T conditions of the wall rocks prior to magmatism cannot yet be properly determined. Further work in this regard is in progress. However, because P of emplacement can be independently estimated by Al-in-hornblende barometry, it is possible to determine the initial P-T conditions of contact melting with reasonable accuracy.

As the Velfjord magmas were emplaced, the wall rocks immediately adjacent to the plutons were heated to temperatures $>700^\circ$C. Kyanite grew in the hottest rocks, probably by the reaction (Spear, 1993):

$$\text{staurolite} + \text{chlorite} = \text{biotite} + \text{kyanite} + \text{H}_2\text{O}$$

at temperatures above about 625 °C.

If interstitial vapor was present, melting was initiated along reaction I (see Fig. 6B); otherwise melting occurred by reactions II and III. Each of the latter two reactions produces K-feldspar. Formation of K-feldspar-phyric leucosomes can be explained by partial melting according to these reactions, and the K-feldspar-phyric contact granites probably originated in the same way. At the highest temperatures of contact melting, reaction IV produced opx and sillimanite. It seems likely that the biotite opx tonalite 'contact granite' sample N17.91 resulted from melting by this reaction. The higher temperature implied by the assemblage in sample N17.91 suggests that the magma formed in a deeper, hotter part of the contact zone and rose to the level of emplacement.

Following initial melting at high-P, conditions must have changed such that cordierite became stable. Most probably, this involved decompression (due to exhumation?). As reaction V was crossed, biotite and sillimanite reacted to form garnet, cordierite and melt (Fig. 6B). In the highest grade diatexites, garnet has a value of $\text{Fe}/(\text{Fe} + \text{Mg}) = 0.7$. This composition of garnet in equilibrium with cordierite, biotite, sillimanite, K-feldspar, and melt suggests conditions of cordierite growth of slightly greater than 750°C at about 5.5 kbar (Fig. 6B). Samples collected at greater distances from the plutons would have followed a tighter P-T loop and would not have entered the cordierite stability field.

The retrograde path requires precipitation of biotite and sillimanite, but not cordierite. Crystallization (isobaric cooling?) on equilibria III and V would result in resorption of garnet, cordierite, and K-feldspar and precipitation of biotite and sillimanite as interstitial, intergrown patches. Crystallization at about 4 to 5 kbar is consistent with the numerous P-T estimates from these rocks on the basis of GASP thermobarometry (Fig. 5). The widespread appearance of euhedral, interstitial staurolite in these migmatites suggests that staurolite was stable at the solidus. Spear et al. (1999) proposed a staurolite dehydration reaction:

$$\text{staurolite} + \text{muscovite} = \text{garnet} + \text{biotite} + \text{Al}_2\text{Si}_3\text{O}_8 + \text{liquid},$$

and proposed a range of pressure for this reaction from about 5 to 7 kbar. The validity of this reaction awaits further experimental and field study.

Samples from which significant amounts of melt had been extracted underwent retrogression without evolution of significant H$_2$O (e.g. Spear et al. 1999). Such rocks would be effectively isolated from further down-temperature retrogression because of the lack of H$_2$O. In contrast, samples that contained a significant volume of anatectic melt probably evolved significant H$_2$O during crystallization, which was then available for reaction with sillimanite and K-feldspar to form muscovite. This is indeed the case adjacent to the Velfjord plutons. Sillimanite- and garnet-rich (restitic) samples typically contain little or no retrograde muscovite. However, leucosomes and leucosome-rich diatexites commonly display significant retrogression in which muscovite replaces feldspars, sillimanite, and cordierite. This abundant late-stage muscovite probably results from evolution of H$_2$O from the leucosome melt at the solidus.

**Thermal energy for wall rock anatexis**

Migmatization due to heat from the Velfjord plutons affects rocks at least 1000 m from the pluton contacts. This implies emplacement of the plutonic magmas at high T, or initial high wall rock temperature, or both. Available data for olivine-pyroxene equilibria suggest pluton emplacement temperatures $\leq$1070°C. Such temperatures are too low to represent crystal-free liquidus of H$_2$O-poor magmas, which are in the range of 1250°C for the Akset-Drevli pluton (K. Chapman and C. Barnes, research in progress). Thus, the temperature estimates may reflect cation exchange among pyroxene±olivine during cooling. Alternatively, the plutons could have been emplaced as crystal-rich mushes. However, the presence of fine-grained mafic dikes in the wall rocks suggests near-liquidus emplacement of at least some of the magmas, thereby supporting higher emplacement temperatures.

A complete understanding of the heat transfer to the wall rocks and its effects on migmatization depends on more detailed two-pyroxene thermometry. These data will permit modeling of the thermal evolution of the magmas and their wall rocks.
Fig. 6. Results of P-T estimates compared to a petrogenetic grid for dehydration partial melting of metapelites (after Spear et al. 1999). (A) Shaded rectangles indicate the location of P-T estimates from GASP thermobarometry that are consistent with the mineral assemblage. The large rectangle with patterned fill is the range of P and T estimates for emplacement of the Velfjord plutons using the temperature-corrected Al-in-hornblende barometer (Anderson & Smith 1995) and the hornblende-plagioclase thermometer of Holland & Blundy (1994). The black bar in this box represents the average of the results. The bar on the right side of the diagram gives the range of pressures determined using the Schmidt (1992) calibration of the Al-in-hornblende barometer, the black bar is the average. Bars at the
Implications for regional geology

This work shows that a number of thermobarometric methods can be used to determine the conditions of emplacement of the Bindal Batholith. In addition, it shows that at least some of the older plutons in the Batholith underwent nearly isothermal decompression from about 7 kbar to 4 to 5 kbar. The duration of such a decompression event is uncertain, but was probably rapid in view of the fact that the migmatitic wall rocks remained above their solidi.

Radiometric ages of Bindal plutons range from ~450 Ma to ~430 Ma (Nordgulen et al. 1993) with some plutons as old as 481 Ma (Pedersen et al. 1999). If exhumation following ~448 Ma emplacement of the Velfjord plutons were rapid, then younger plutons in the Batholith would have been emplaced at lower pressure. The success of Al-in-hornblende barometry in this work suggests that judicious application of the barometer on other samples of the Bindal Batholith can provide important constraints on the P-T history of the entire HNC during late Ordovician and early Silurian time.

At present it is not certain what tectonic event was responsible for burial of the HNC prior to emplacement of the Velfjord magmas. The host rocks to the plutons are in thrust contact with low-to medium-grade rocks to the east (Fig. 2) such that juxtaposition of these units could be responsible for the high emplacement pressure. However, previous workers indicated that a break in metamorphic conditions exists between the structurally higher rocks and the host rocks to the Velfjord plutons. Further work must be focused on the nature of this contact and the age of movement along it. At least two possible explanations are possible. First, a metamorphic break may not exist between the two thrust sheets. It is possible that the apparent higher grade of the host rocks of the Velfjord plutons is a ‘regional contact metamorphic’ effect. Second, if a metamorphic break does exist, the last displacement on the thrust could have been in a low-angle normal sense, possibly reactivation of an original thrust fault. This would explain the normal sense of displacement (low-grade over high-grade) and could provide a mechanism for exhumation of the underlying high-T unit.

Summary

The Velfjord plutons were emplaced into carbonate and pelitic host rocks at about 448 Ma. The pressure of emplacement was approximately 7 kbar. Heat transferred from the plutons to the host rocks resulted in dehydration partial melting of the pelites, which produced melts capable of crystallizing K-feldspar megacrysts. Some of these anatectic magmas escaped the zone of melting, thereby depleting the source rocks in fertile components and H₂O. A few of the anatectic magmas accumulated and solidified along the contacts of the plutons to form ‘contact granites’.

Nearly isothermal decompression followed, such that high-T wall rock migmatites entered the stability field of cordierite. In contrast, lower-T wall rocks at greater distances from the plutons did not reach cordierite stability. Cooling and solidification occurred at 4 to 5 kbar and resulted in interstitial intergrowths of biotite and acicular sillimanite.

The results of this study imply rapid exhumation of the HNC following emplacement of the Velfjord plutons. Thermobarometric studies on younger plutons in the Bindal Batholith should provide a means to test and better quantify this P-T-t path.

References


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