Timing of sedimentation, metamorphism, and plutonism in the Helgeland Nappe Complex, north-central Norwegian Caledonides

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ABSTRACT

The Helgeland Nappe Complex consists of a sequence of imbricated east-dipping nappes that record a history of Neoproterozoic–Ordovician, sedimentary, metamorphic, and magmatic events. A combination of U-Pb dating of zircon and titanite by laser-ablation–inductively coupled plasma–mass spectrometry plus chemostratigraphic data on marbles places tight constraints on the sedimentary, tectonic, and thermal events of the complex. Strontium and carbon isotope data have identified Neoproterozoic marbles in the Lower Nappe, the Horta nappe, and Scandian-aged infolds in the Vikna region. The environment of deposition of these rocks was a continental shelf, presumably of Laurentia. Detrital zircon ages from the Lower Nappe are nearly identical to those of Dalradian sedimentary rocks in Scotland. Cambrian rifting caused development of one or more ophiolite-floored basins, into which thick sequences of Early Ordovician clastic and carbonate sediments were deposited. On the basis of ages of the youngest zircons, deposition ended after ca. 481 Ma. These basin units are now seen as the Skei Group, Sauren-Torghatten Nappe, and Middle Nappe, as well as the stratigraphically highest part of the Horta nappe and possibly of the Upper Nappe. The provenance of these sediments was partly from the Lower Nappe, on the basis of detrital zircon age populations in metasandstones and cobbles from proximal conglomerates. However, the source of Cambrian–Ordovician zircons in all of the Early Ordovician basins is enigmatic. Crustal anatexis of the Lower and Upper Nappes occurred ca. 480 Ma, followed by imbrication of the entire nappe sequence. By ca. 478 Ma, the Horta nappe was overturned and was at the structural base of the nappe sequence, where it underwent migmatization and was the source of S-type magmas. Diverse magmatic activity followed ca. 465 Ma, 450–445 Ma, and 439–424 Ma. Several plutons in the youngest age range contain inherited 460–450 Ma zircons. These zircons are interpreted to reflect a deep crustal zone in which mafic magmas caused melting, mixing, and hybridization from 460 to 450 Ma. Magmatic reheating of this zone, possibly associated with crustal thickening, resulted in voluminous, predominantly tonalitic magmatism from 439 to 424 Ma.

Keywords: geochronology, U-Pb, zircon, Caledonian, Norway.

INTRODUCTION

The Norwegian Caledonides consist of four allochthons that were emplaced on the Baltic craton during Silurian–Devonian closure of the Iapetus Ocean (Fig. 1). The allochthons are bounded by broadly west-dipping shear zones into, in structurally ascending order, the Lower, Middle, Upper, and Uppermost Allochthons (Fig. 1B; Gee and Sturt, 1985; Roberts and Gee, 1985). The Lower and Middle Allochthons have clear affinities to the Baltic craton (Gee and Sturt, 1985; Roberts and Gee, 1985), whereas

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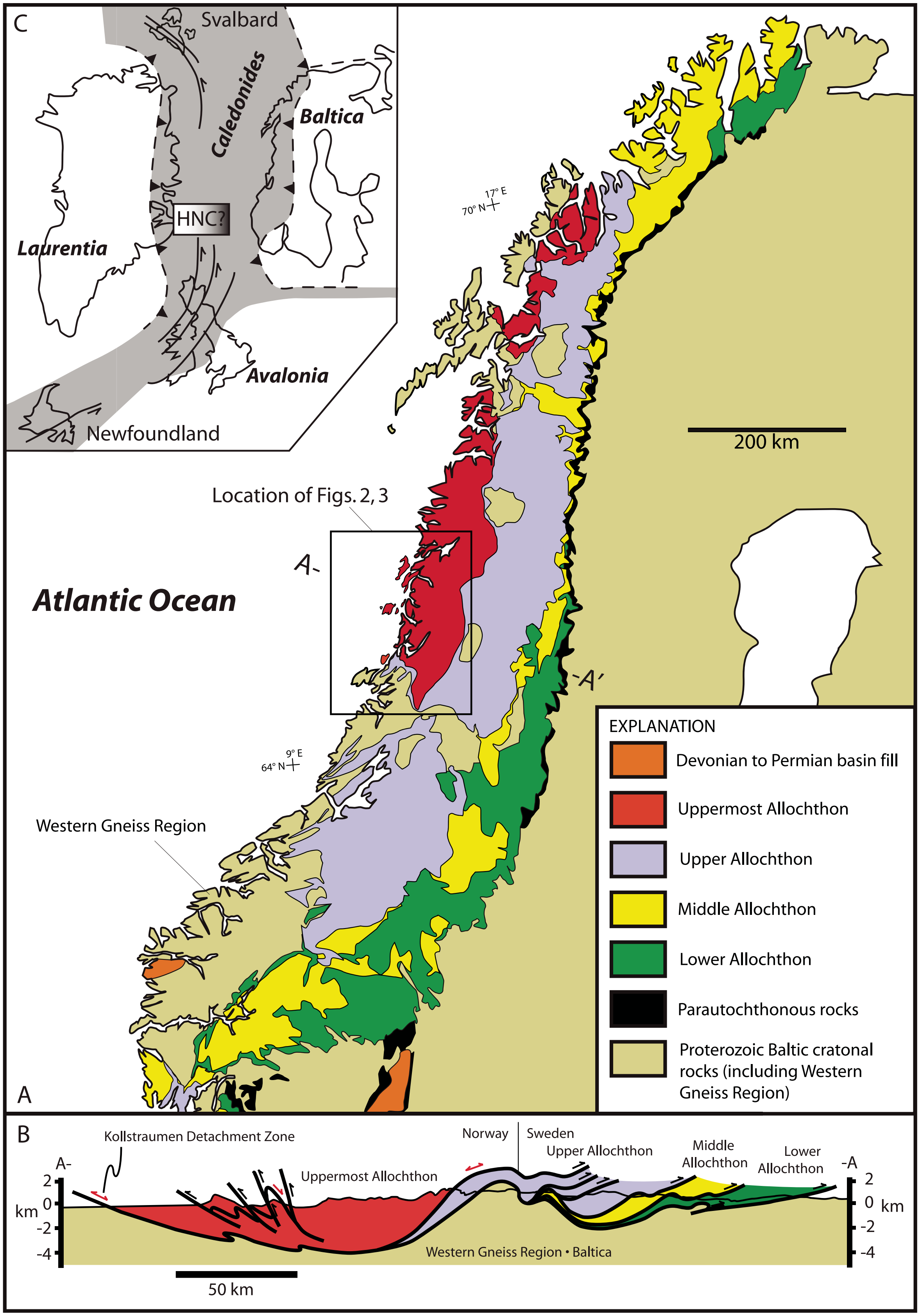

Figure 1. (A) Tectonostratigraphic map displaying the major units within the Scandinavian Caledonides (after Gee and Sturt, 1985). (B) Simplified geologic cross section across the Caledonides at the latitude of this study depicting the general structural style contained within the Uppermost Allochthon. (C) Speculative Ordovician–Silurian reconstruction of the Caledonide orogen and possible location of rocks within the Helgeland Nappe Complex (HNC). After Greiling and Garfunkel (2007).

nappes of the Upper Allochthon formed in the continent-ocean transition zone (Andréasson, 1994) and in the ancient Iapetus Ocean (e.g., Stephens et al., 1985; Grenne et al., 1999; Roberts et al., 2002a; Roberts, 2003).

The Uppermost Allochthon is generally interpreted to have formed in a continental margin setting (see review in Roberts et al., 2007). Our study concerns the Helgeland Nappe Complex, which is the structurally highest nappe complex of the Uppermost Allochthon (Figs. 1 and 2; Stephens et al., 1985; Roberts et al., 2007). The component nappes of the Helgeland Nappe Complex are of two types: one consists of calc-silicate, calcareous, and clastic metasedimentary rocks that overlie a basement of fragmentary ophiolitic rocks, and the other consists of interlayered pelitic, psammitic, carbonate, and calc-silicate units with no exposed crystalline basement. The former units are thought to represent deposition in continental-margin basins (e.g., Stephens et al., 1985; Heldal, 2001; Roberts et al., 2001), whereas the latter are typical of deposition in continental shelf environments. The near-continental depositional environments typical of the Helgeland Nappe Complex combined with structural data and sedimentary structures indicating a western sediment source have led to the suggestion that the Uppermost Allochthon originated along the margin of Laurentia (e.g., Roberts, 1980, 2003; Roberts et al., 2001, 2002a, 2007; Yoshinobu et al., 2002). Yoshinobu et al. (2002), in particular, documented pre-Scandian, west-vergent nappe-bounding structures in the Helgeland Nappe Complex that they ascribed to Taconian-age deformation in Laurentia.

The ability to correlate nappes in the Helgeland Nappe Complex with Laurentia, or any other Rodinian craton, is critically dependent on a knowledge and understanding of the ages of deposition and/or formation of the nappes, ages of metamorphism, and ages of magmatic activity. Such data have largely been lacking, except for the studies of Nordgulen et al. (1993) and Yoshinobu et al. (2002). Moreover, ages of metamorphism and magmatism should logically relate to Laurentian orogenic events (e.g., Yoshinobu et al., 2002). Because fossils were destroyed by high-grade metamorphism in much of the Helgeland Nappe Complex, it is necessary to use other methods to estimate ages of geologic events in the complex. In this paper we present U-Pb ages of zircon and titanite separated from clastic metasedimentary and plutonic rocks, and Sr and C isotope chemostratigraphic data for marbles. The data indicate that two of the nappes have evidence for deposition in Neoproterozoic time and at least four contain early Paleozoic metasedimentary rocks. Moreover, Precambrian zircons in early Paleozoic nappes can be explained by recycling of sediment from the adjacent older nappes. Conversely, the early Paleozoic nappes also contain Cambrian and Ordovician zircons that have no obvious sources in the Helgeland Nappe Complex. The combined age and provenance data provided by this study are consistent with a Laurentian depositional setting (e.g., Roberts et al., 2001; Yoshinobu et al., 2002) and they suggest a correlation of the oldest nappes with rocks of the Dalradian Supergroup of Scotland.

GEOLOGIC SETTING

In the study area, the Uppermost Allochthon consists of at least two nappe complexes (Fig. 2), the structurally lower Rödingsfjället Nappe Complex and the structurally higher Helgeland Nappe Complex. The latter consists of four named nappes, although the existence of additional nappes is likely and is discussed below. The most detailed studies, and those that developed nappe nomenclature, were in south and central Helgeland, particularly Thorsnes and Løseth (1991), Nordgulen et al. (1989), and Yoshinobu et al. (2002). In structurally descending order, the nappes are the Upper, Middle, Lower, and Sauren-Torghatten Nappes. The Sauren-Torghatten and Middle Nappes are sequences of metamorphosed clastic and carbonate rocks that unconformably overlie discontinuously exposed, tectonically fragmented basement of ophiolitic affinity. The basement rocks range from peridotite through gabbro and mafic metavolcanic rocks (Thorsnes and Løseth, 1991; Heldal, 2001). The overlying metasedimentary rocks include metawackes, metapelites, calc-silicate rocks, and marble, with interbedded conglomerates. Conglomeratic strata vary from monomict amphibolitic units to polymict beds (Thorsnes and Løseth, 1991; Nordgulen et al., 1992; Heldal, 2001). The metamorphic grade in these nappes is typically amphibolite facies.

In the Velfjord region, the Upper and Lower Nappes consist almost entirely of metasedimentary rocks, and no depositional basement has been recognized. Common rock types are migmatitic quartzofeldspathic and semi-pelitic gneiss, calc-silicate gneiss, and marble. The presence of migmatite is characteristic of the Upper and Lower Nappes and indicates regional metamorphism at upper amphibolite to lower granulite facies conditions (e.g., Barnes and Prestvik, 2000; Reid, 2004; Yoshinobu et al., 2005).

Two additional rock units may be separate nappes. The first is the Leka ophiolite (e.g., Prestvik, 1980; Furnes et al., 1988), which crops out in the southwestern Helgeland Nappe Complex (Fig. 2) and consists of a complete ophiolitic sequence, plagiogranites of which were dated as 497 ± 2 Ma (Dunning and Pedersen, 1988). The Leka ophiolite is one of a number of 500–470 Ma suprasubduction-zone ophiolitic units within the Norwegian Caledonides and elsewhere in the Appalachian–Caledonian orogen (e.g., Dunning and Pedersen, 1988; Pedersen and Furnes, 1991; Cawood and Suhr, 1992). The igneous rocks of the Leka ophiolite are unconformably overlain by the Skei Group, which consists of a lower unit of interbedded conglomerate, conglomeratic sandstones, and sandstones, and an upper unit of wackes, marble, and metapelitic rocks, all metamorphosed at lower greenschist facies (Sturt et al., 1985).

The second possible nappe unit crops out in the Horta archipelago and is host to the Hortavær igneous complex (Fig. 2; Gustavson and Prestvik, 1979; Barnes et al., 2003). The western part of the intrusive complex intrudes migmatitic quartzofeldspathic gneiss and smaller amounts of quartzite and marble, whereas the central and eastern parts intrude marble, calc-silicate gneiss, and calc-silicate stretched-pebble conglomerate. Leucosomes in the western migmatite are locally mingled with pillowed quartz dioritic dikes. In this report we informally refer to the entire sequence of metasedimentary rocks in the Horta archipelago as the Horta nappe. To the north, the island of Vega (Fig. 2) is underlain by an anatectic (S-type) pluton whose metasedimentary enclaves are identical to migmatites of the Horta archipelago. The Vega pluton intrudes marble, calc-silicate rocks, pelites, and quartzitic rocks that have yet to be assigned to a specific nappe. On the basis of regional geology, lithology, and metamorphic grade, we tentatively assign them to the Horta nappe (Fig. 2).

The Uppermost and Upper Allochthons are hosts to a number of plutonic complexes. The Bindal Batholith is the largest of these, and intrudes the Helgeland Nappe Complex (Nordgulen, 1992). This batholith consists of plutons that span the compositional range from gabbro to granite. Emplacement ages range from 482 to 424 Ma (Nordgulen et al., 1993; Yoshinobu et al., 2002; Nissen et al., 2006; this report). Isotopic studies showed that batholith magmas arose by partial melting of a range of crustal sources, with or without mantle-derived contributions (Nordgulen and Sundvoll, 1992; Birkeland et al., 1993; Barnes et al., 2004, 2005a). The early studies also hinted at a geographic distribution in which older, peraluminous, anatectic granites are most common in the western Helgeland Nappe Complex, whereas younger, tonalitic plutons are most common in its eastern parts (Nordgulen and Sundvoll, 1992; Nordgulen, 1992; Birkeland et al., 1993). The Nesåa Batholith intrudes the Gjersvik Nappe of the Upper Allochthon along the southeastern margin of

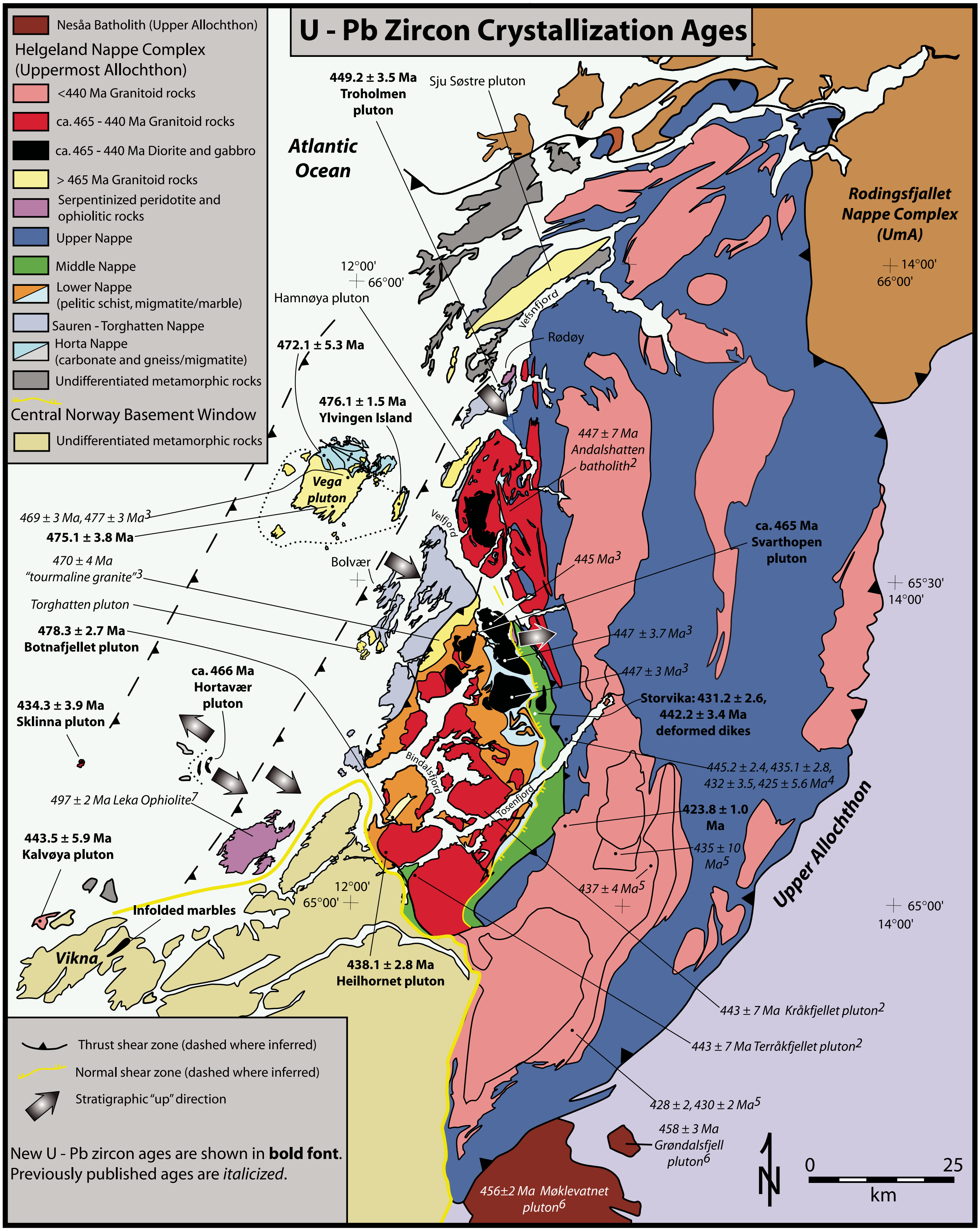

Figure 2. Regional geologic map showing the position of the Helgeland Nappe Complex in north-central Norway and the location and U-Pb ages of dated plutonic rocks. Ages reported in this paper are shown in bold print and previously published ages are in italics. Sources for published data: 1—T. Heldal, written communication, 2—Nordgulen et al. (1993), 3—Yoshinobu et al. (2002), 4—Yoshinobu et al. (2005), 5—Nissen et al. (2006), 6—Meyer et al. (2003), 7—Pedersen and Furnes (1991).

the Uppermost Allochthon. Meyer et al. (2003) dated the Grøndalsfjell Complex of the Nesåa Batholith at 458 ± 3 Ma and quartz monzodiorite of the Møklevatnet Complex was dated as 456 ± 2 Ma (Roberts and Tucker, 1991). In addition, Stephens et al. (1993) reported an age of 483 +5/–3 Ma for trondhjemite from the Gjersvik Nappe and 445 +24/–6 Ma for granite from the Storfjället Nappe, and Mørk et al. (1997) dated a gabbroic pluton in the upper Köli Nappes as ca. 437 Ma.

Although the overall structure and lithologic variation in the Helgeland Nappe Complex are well understood, a number of fundamental questions remain unanswered. Three topics of particular importance in reconstructing the geologic history of the complex are the depositional ages of metasedimentary rocks, the ages of regional metamorphism, and the crustal sources of the plutonic rocks. The importance of the first two questions is straightforward: meaningful paleotectonic reconstructions of the region are impossible without information concerning ages of sedimentation and metamorphism. The only depositional age data for the Helgeland Nappe Complex come from strontium and carbon isotopic data, which indicate Neoproterozoic and possible Ordovician ages for deposition of marbles in the Lower Nappe (Trønnes, 1994; Trønnes and Sundvoll, 1995; Sandøy, 2003). Knowledge of pluton age and source compositions additionally provides information about the crustal architecture during specific time frames in the development of the complex (e.g., Allen and Barnes, 2006).

METHODS

Strontium and Carbon Analysis of Marble

Samples of calcite marble were collected from the Horta archipelago, the Kalvøya and Lysøyvågen areas of Vikna, and at Solli, Leka (Table 1). Powdered samples were dissolved in HCl and diluted 250 times. The concentrations of Mg, Fe, Mn, Na, K, and Ba were then analyzed by inductively coupled plasma–atomic emission spectrometer (ICP-AES) at Texas Tech University using commercial standard solutions. For Sr isotopic analyses, acetic acid was added to ~0.1 g of sample to dissolve carbonate. An aliquot was spiked with a combined ^{84}Sr- and ^{87}Rb-enriched tracer. Sr and Rb were isolated by standard cation chromatography. Rb and Sr abundances were determined by isotope dilution and Sr isotope ratios measured on a VG Sector thermal ionization mass spectrometer at the University of Wyoming. Analyses (n = 18) of standard Sr NBS-987 during the period of analysis gave an average ^{87}Sr/^{86}Sr of 0.710247 ± 0.00002 (2 standard deviations). Initial ^{87}Sr/^{86}Sr values were calculated assuming an age of 500 Ma, although the correction for radiogenic growth is negligible for these samples, which have very low Rb contents. Stable isotope analysis of marble samples was carried out at Texas Tech University following the standard methods of McCrea (1950). The powdered samples were reacted with H_3PO_4 at 50 C for 1–2 days to release CO_2 gas, which was then analyzed for carbon and oxygen isotope ratios. Results (Table 2) are normalized to Vienna standard mean ocean water (V-SMOW) for oxygen and Vienna Peedee belemnite (V-PDB) for carbon.

Laser-Ablation (LA) ICP–Mass Spectrometer (MS) Dating of Zircon

Analytical Methods

U-Pb zircon LA-ICP-MS ages were obtained for 28 rock samples, information for which is listed in Table 1 and Appendix 1[1]. Zircon and titanite were separated from rock samples that varied in mass from 3 to 20 kg. Samples were crushed and then ground in a tungsten carbide swing mill (TTU) or a stainless steel disk mill (Wyoming). At Texas Tech, samples were then sieved and the heavy mineral fraction was concentrated with a gold pan, whereas at Wyoming, crushed samples were passed over a Wilfley table. The resultant heavy fractions were processed through Franz isodynamic magnetic separators and with heavy liquids. Zircon and titanite were then hand-picked. The mineral separates were placed into windows in the backing of double-stick tape on a glass base. After each sample was loaded, the window was covered and secured with tape before the next window was cut. Standards [zircon Temora 2 (416.8 ± 1 Ma; Black et al. 2004), zircon R33, and NIST 610 glass] were mounted in a similar fashion. A cylinder was placed on the tape after the backing was removed and epoxy resin was poured into the cylinder. The epoxy cast mount was polished to expose zircon interiors, which were photographed at 5× or 10× magnification on a polarizing microscope.

Zircons were analyzed in 4 sessions in a 24 month period and titanite in 2 sessions over the same period. The epoxy mounts were placed in a sample cell connecting a 193-nm-wavelength Lambda Physik LPX 120I ArF Excimer laser to an Agilent 7500S ICP-MS through a signal-smoothing manifold. The laser delivered 120 mJ constant energy at 5 Hz through two 45° mirrors, resulting in a power density at the sample surface of ~0.4 GW/cm^2. Ablated holes were 32 µm in diameter and ~20 µm deep after 40 s of drilling. Masses used in geochronology (^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th, ^{235}U, and ^{238}U) were each analyzed for 0.04 s, and other elements of geochemical interest (^{29}Si, ^{31}P, ^{91}Zr, ^{177}Hf, and seven rare earth elements, REE) were each collected for 0.01 s. Data were collected in peak jumping mode with a mass sweep time of 0.384 s. We employed ^{235}U only in the event that U concentration was >1000 ppm in order to avoid use of the analogue detector when the pulse (low count rate) detector is used for the standards, otherwise ^{235}U was calculated from ^{238}U/137.88. Data for ^{204}Pb were not collected because Hg backgrounds at that mass are 10 times that of Pb.

Backgrounds (laser off) were collected for ~20 s at the beginning of each analysis, after which the laser-on signal was collected for 40 s. The analysis cycle consisted of Temora (zircon standard), NIST 610, R33 (secondary zircon standard) and 10 unknowns. Unknowns were analyzed in a round-robin fashion so that the averages of the standards were applicable. Two standard materials were necessary because (1) precision for geochronology requires down-hole fractionation effects on isotope ratios be modeled from zircon and (2) homogeneous glass is required for calculating concentrations, especially Th/U (assumed to be 1.01870 in NIST 610). Zircon standards have uniform ages but are chemically heterogeneous. Fractionation factors for the isotope ratios of interest (known ratio/measured ratio), e.g., ^{206}Pb/^{238}U, derived from each mass sweep of the average Temora 2 standard were applied to the corresponding mass sweep in the unknown. Similarly, fractionation factors were derived from known/measured count ratios in NIST 610 for ^{232}Th/^{238}U and all elements, ratioed to ^{29}Si. Elemental concentrations were derived by multiplying element/^{29}Si count ratios according to the NIST 610 fractionation factor, and by 37.22 wt% SiO_2, the assumed concentration in stoichiometric zircon. This technique of applying fractionation factors from the average standard on a mass sweep-by-mass sweep basis corrects for down-hole mass fractionation.

Uncertainties quoted for an individual age analysis (one standard error, s.e.; Appendix 2; see footnote 1) include observed uncertainties plus a factor added in quadrature that is the standard deviation of the particular isotope ratio observed in average Temora 2. For example,

[1]If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00138.S1, http://dx.doi.org/10.1130/GES00138.S2, and http://dx.doi.org/10.1130/GES00138.S3 or the full-text article on www.gsajournals.org to view Appendices 1–3.

TABLE 1. LOCATIONS AND ROCK TYPES OF DATED SAMPLES

Sample	UTM* location (WGS84) Zone	Easting	Northing	Rock unit
Intrusive rocks				
VGWM-15A	32W	636125	7278490	Granodiorite, main stage, Vega pluton
F2	32W	619175	7281552	Granodiorite, main stage, Vega pluton
NWV-6	32W	630068	7289173	Tonalitic, migmatitic dike in the aureole of the Vega pluton
VCF05.06	33W	368909	7280756	Granite of the Ylvingen pluton
N259	33W	365595	7224945	Anatectic Botnafjellet granite
02.12H	32W	613873	7232416	Pegmatitic monzodiorite, sheet in the Hortavær intrusive complex
N03.05	33W	381435	7257083	Quartz monzonitic gneiss; hybrid zone in the Svarthopen pluton
KR-21	33W	386507	7300295	Granite; Troholmen pluton (near Rødøy)
N08.06	33W	368069	7223930	Granodiorite of the Heilhornet pluton
95.08V	32W	532419	7292993	Granite of the Sklinna pluton
04.07V	32W	586137	7206313	Granite from the Kalvøya area, outer Vikna
N03.06	33W	385393	7216115	Granitic gneiss of the Upper Nappe
NLV-19A	33W	393879	7244547	Boudinaged leucogabbro dike in the Middle Nappe (Storvika area)
NLV-17B	33W	393085	7244139	Boudinaged gabbroic dike in the Middle Nappe (Storvika area)
AND-61	33W	378618	7278195	Dioritic dike, Andalshatten pluton aureole, Sauren-Torghatten Nappe
Metamorphic rocks				
Horta nappe				
04.01H	32W	611717	7231596	Leucosome in semipelitic migmatite
05.02H	32W	609850	7236950	Leucosome in semipelitic migmatite
Enclave in Vega pluton				
VQ24	32W	633740	7279891	Kilometer-scale quartzite enclave in Vega pluton
Skei Group				
SCF-12.05	32W	629002	7220684	Quartz-rich metasandstone cobble in the Skei Group; Leka
Sauren-Torghatten Nappe				
B-45	33W	369378	7252342	Fine-grained metasandstone; Sømnes area
N01.04	33W	379015	7266874	Garnet schist in the Sauren-Torghatten Nappe
N251B	33W	365543	7256090	Hornblende biotite schist in the Sauren-Torghatten Nappe
AND-60	33W	378729	7277826	Metasandstone of the Sauren-Torghatten Nappe
BL11A	33W	364987	7266352	Metasandstone cobble in the Sauren-Torghatten Nappe; Bolvær area
KB-17	33W	365130	7265462	Metagabbro in the Sauren-Torghatten Nappe; Bolvær area
Lower Nappe				
N02.04	33W	382760	7257225	Leucosome-rich migmatitic gneiss of the Lower Nappe
B-46	33W	364854	7230503	Metasandstone interlayered in calcareous rocks; Fantvika, Holm Peninsula
Middle Nappe				
NLV-12	33W	390900	7243221	Schist of the Middle Nappe; Nordfjellmark area
NLV-23	33W	392055	7252510	Cobble in Middle Nappe conglomerate
Marbles				
Screens of the Horta nappe in the Hortavaer intrusive complex				
93.07H	32W	616125	7234125	White marble; Vågøya
93.47H	32W	615450	7234400	White marble; Burøya
93.74H	32W	613650	7233700	White marble; Ørnholmen
02.25H	32W	613025	7234550	White marble; Raudholmen
02.26H	32W	612751	7234287	White marble; Raudholmen
02.31HR	32W	614075	7234000	White marble; SE of Kvåholmen
02.47HR	32W	613825	7233650	White marble; eastern Ørnholmen
02.52HR	32W	614241	7233568	White marble; island between Bovarøya and Andersøya
02.56H	32W	615450	7232600	White marble; SE Kiklakken
05.45H	32W	619737	7233507	White marble; S Ørnholmen
Skei Group				
LEKA-1R	32W	625853	7220095	Pink marble; Leka
Caledonian rocks infolded into parautochthonous basement				
V05/02R	32W	585539	7207883	White marble; Vikna, NW of Kalvøya
V06/02R	32W	584400	7206500	White marble; Vikna, Møskja
VIK-8R	32W	599860	7203857	Graphitic marble; Vikna, Lysøya
VIK-9R	32W	599808	7203855	Graphitic marble; Vikna, Lysøya
VIK-10R	32W	599728	7203807	Graphitic marble; Vikna, Lysøya

* *Note*: UTM—Universal Transverse Mercator.

TABLE 2. TRACE ELEMENT AND ISOTOPIC COMPOSITIONS OF MARBLE

Sample	Ba (ppm)	Fe (ppm)	Mg (ppm)	Na (ppm)	Mn (ppm)	K (ppm)	Rb (ppm)	Sr (ppm)	Mn/Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Initial $^{87}Sr/^{86}Sr$*	$\delta^{13}C$**	$\delta^{18}O$**
White marble screens in the Hortavaer intrusive complex														
93.07H	dl	1039	284	249	23	119	0.03627	1770	0.01	0.0005929	0.70670	0.70670	4.3	24.3
93.47H	dl	257	96540	86	12	28	0.04731	512.0	0.02	0.0002670	0.70653	0.70653	3.8	23.2
93.74H	dl	338	35969	75	37	14	0.01567	205.4	0.18	0.0002210	0.70853	0.70853	-0.6	19.8
02.25H	139	6417	113500	130	286	2001	1.853	62.00	4.61	nd	nd	nd	-0.4	28.9
02.26H	325	1845	115600	322	94	2445	5.256	66.22	1.41	0.2298	0.71139	0.70976	0.7	31.0
02.31HR	7	1382	73980	173	238	15	0.02336	332.1	0.72	0.0002036	0.70881	0.70881	0.6	29.6
02.47HR	6	3550	53000	163	518	19	0.04380	130.8	3.96	0.0009690	0.70831	0.70831	4.9	29.5
02.52HR	2	2094	75237	204	752	47	0.1248	160.5	4.68	0.002250	0.70870	0.70869	3.9	29.6
02.56H	9	471	6799	232	83	518	0.9305	1086	0.08	0.002470	0.70658	0.70656	4.3	33.0
05.45H	nd	nd	nd	nd	nd	nd	0.02534	185.5		0.0003953	0.70845	0.70845	3.7	32.2
Skei group pink marble														
LEKA-1R	dl	3885	2458	81	337	74	0.1375	61.88	5.45	0.006429	0.70924	0.70920	2.7	25.5
Infolded white marble; outer Vikna														
V05/02R	dl	2383	1582	505	102	262	0.4330	1678	0.06		dnr		5.1	32.0
V05/06R	dl	508	6806	122	73	117	0.2859	415.4	0.18	0.001992	0.70905	0.70903	0.1	31.3
Infolded graphitic marble; Mellom Vikna														
VIK-8R	4	6806	2087	305	178	184	0.5953	1683	0.11	0.001020	0.70678	0.70677	5.4	26.5
VIK-9R	3	721	1545	118	85	46	0.0525	2772	0.03	0.00005480	0.70655	0.70655	6.6	30.1
VIK-10R	11	793	2787	56	123	9	0.0137	2003	0.06	0.00001970	~0.7066		4.2	30.6

Note: Ba, Fe, Mg, Na, Mn, and K were analyzed by inductively coupled plasma atomic emission spectrometry; Rb and Sr by isotope dilution. dl—below detection limits; dnr—did not analyze; nd—not determined. Sample labels in bold font are those whose isotopic compositions are consistent with Neoproterozoic deposition.

*Initial ratios calculated for depositional age of 500 Ma.

**Carbon isotope ratios in per mil relative to Vienna Peedee belemnite; oxygen in per mil relative to Vienna standard mean ocean water.

the standard deviation for $^{206}Pb/^{238}U$ observed in average Temora 2 in an analytical session is 1%–2%. This added term accommodates the imperfections of a natural standard as well as instrument drift, and dominates the 1 s.e. uncertainties quoted for individual analyses. The success of this data reduction method is demonstrated by analyses of in-house standard LP521. Daily averages are in agreement with one another, and with the laboratory's running average (42.2 ± 0.2 Ma; Campbell et al., 2006). It is noteworthy that the mean squares of weighted deviates (MSWDs) for these populations from a well-behaved natural zircon standard are only somewhat greater than unity (1.2–1.6). For further details of the technique, see Harris et al. (2004).

Individual analyses were inspected for obvious age changes during the ablation, and for inclusions, particularly with reference to P and La contents. Campbell et al. (2006) showed how inclusions express themselves, and how they can be avoided by visual selection of a portion of the ablation (minimum of 10 mass scans).

Common Pb corrections are calculated on the basis of both ^{207}Pb and ^{208}Pb values (Appendix 2); the latter is preferred and described here. Starting with the initial calculated $^{206}Pb/^{238}U$ age, the value of $^{232}Th/^{208}Pb$ at that age, and a model for common Pb composition of that age (Cumming and Richards, 1975), an amount of excess or deficit ^{208}Pb is calculated. This excess or deficit is used to calculate the amount of common ^{206}Pb, common ^{207}Pb, and the total common Pb content. The appropriate amounts of common ^{206}Pb and ^{207}Pb are then subtracted from (or added to) $^{206}Pb/^{238}U$, $^{207}Pb/^{235}U$, and $^{207}Pb/^{206}Pb$. The calculation is then repeated, using the newly calculated $^{206}Pb/^{238}U$ age, until the ratios stop changing at the 0.1 Ma equivalent level. The resulting ratios are called the ^{208}Pb-corrected ages. The $^{206}Pb/^{238}U$ ages are quoted for analyses of grains younger than 1000 Ma, and $^{207}Pb/^{206}Pb$ ages for those older.

Two filters were applied to evaluate the quality of the selected portion of an ablation. If observed uncertainty (s.e.) for the isotope ratio was more than twice that expected from counting statistics, the analysis was rejected because such variation could result from a mixed-age analysis, inclusions, and so forth. Then concordance was tested. For grains younger than 1000 Ma, a grain was deemed concordant if the $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ages, including uncertainty, agreed within 10%. Concordance in older grains was tested in a similar way using $^{206}Pb/^{238}U$ and $^{207}Pb/^{206}Pb$ ages.

Interpretation

Because zircon age spectra from migmatites and low-temperature granitic plutons generally represent a mixture of scarce primary and abundant inherited zircons, the compiled population for a sample was inspected for gross inheritance (ages many standard deviations older than the weighted mean). Once these were excluded, the remaining ablations were examined to determine if a single age population was represented by use of MSWD calculation and trends on cumulative probability plots (e.g., Fig. 3, insets H, I). If the MSWD was >2, and a distinct break in age distribution was obvious, slightly older zircons were interpreted to be inherited. In some instances, outliers younger than the main population were encountered and omitted from population age calculation. A weighted mean age for the rock was calculated from the remaining analyses, and the weighted means of the uncorrected $^{206}Pb/^{238}U$ ages are reported in Figures 3 and 4. Uncertainties reported for populations are ±2 s.e.

LA-ICP-MS Dating of Titanite

The titanite ages employed 9894 (Buick et al., 2003) as the standard for Th-Pb, U-Pb, and Pb/Pb ratios, and NIST 610 was the standard for all other element concentrations. This approach assumes that 9894 is concordant at 2015 Ma. This approach is used because on a mass sweep by mass sweep basis, the ratios of the $^{206}Pb/^{238}U$ values for standard titanite 9894 and NIST 610 change during the first 60% of the ablation, then approach a uniform value. An uncertainty of 0.3% for $^{206}Pb/^{238}U$ and 0.7% for $^{207}Pb/^{206}Pb$ was assigned to the standard analysis. The data (Appendix 3; see footnote 1) are used to construct discordia plots on the Tera-Wasserburg inverse concordia diagrams (Fig. 5) in which the upper intercept of the chord approximates the $^{207}Pb/^{206}Pb$ composition of common Pb and the lower intercept corresponds to a cooling age. The goodness of fit of the chord is an indication of whether the measured values are a mixture of a single common Pb component and radiogenic Pb that has grown since crystallization or metamorphism.

RESULTS AND INTERPRETATION OF AGES

Isotopic and Trace Element Data for Marbles

Marbles compose a large part of the Helgeland Nappe Complex being investigated. However, high-grade metamorphism and deformation have obscured or destroyed any fossil evidence of the ages of these units. Trønnes (1994) and Trønnes and Sundvoll (1995) showed that the Sr and C isotopic compositions of some marbles were characteristic of deposition during Neoproterozoic time. Melezhik and coworkers have documented the use of Sr and C isotope data to determine the age of deposition of metamorphosed calcite marbles (e.g., Melezhik et al., 2000, 2001, 2002, 2003; Sandøy, 2003; Slagstad et al., 2006) in several parts of the Norwegian Caledonides. Their work has shown the importance of geochemical screening in order to identify samples that have undergone minimal postdepositional alteration and/or diagenesis (see discussion in Melezhik et al., 2002). On the basis of this careful approach, Sandøy (2003) showed that marble in the Lower Nappe, his Saus marble, has a Neoproterozoic depositional age.

We analyzed marble samples from four locations during the course of this study. The largest number of samples was collected from screens in the Hortavær igneous complex. We also sampled Caledonian marbles from Mellom Vikna and from the Kalvøya area of outer Vikna. In the latter two instances, the marbles are part of infolded synforms into late Paleoproterozoic gneissic rocks of the northernmost Western Gneiss region (Bering, 1988). A sample of pink marble from the Skei Group on Leka was also analyzed.

Among the 16 samples analyzed, only 4 completely meet the criteria outlined by Melezhik et al. (2002) for chemostratigraphic use. These samples have low Mg and Fe contents, Sr contents >1000 ppm, low Rb contents, and Mn/Sr ratios < 0.1 (Table 2). A fifth sample (93.47H) is similar but has a Sr content of 512 ppm and Mg content of ~97,000 ppm, indicative of incipient dolomitization. The remaining samples show features indicative of alteration, including high Mg contents, low Sr contents, and high Mn/Sr ratios.

Of the five samples that meet, or nearly meet, the screening criteria, all have initial $^{87}Sr/^{86}Sr$ between 0.70653 and 0.70677 and $\delta^{13}C$ values between 3.8‰ and 6.6‰ (Table 2). Melezhik et al. (2002) compiled and summarized temporal variation of seawater $^{87}Sr/^{86}Sr$ during Neoproterozoic time, for which they presented two possible curves (their Fig. 8). According to their compilation, the $^{87}Sr/^{86}Sr$ values of all five marble samples are consistent with deposition from ca. 590 to 660 Ma using their (preferred) Curve 2. Use of their Curve 1 broadens the possible age range to ca. 700 Ma, with an additional age interval ca. 730–760 Ma. The carbon isotopic data are entirely consistent with the age ranges indicated by the Sr isotope data; however, they permit a wider range of ages, extending into Cambrian time. As noted by Melezhik et al. (2002) and as is apparent from the data in Table 2, alteration and radiogenic growth of ^{87}Sr has the effect of raising initial $^{87}Sr/^{86}Sr$ and lowering $\delta^{13}C$. This means that these marbles cannot be younger than ca. 590 Ma and are probably no older than 760 Ma. We conclude that Horta nappe marble screens in the Hortavær igneous complex, as well as infolded Caledonian marbles in the Vikna archipelago, are Neoproterozoic in age.

A sixth sample, Leka-1R (Table 2), was also considered as possibly retaining its original Sr and C isotopic characteristics. This sample has somewhat low Sr abundances and a high Mn/Sr ratio relative to the Melezhik et al. (2002) criteria. However, if the measured isotope ratios are representative of the time of deposition, the Leka marble is Ordovician in age.

U-Pb Ages of Zircon from Metamorphic and Related Rocks

Results of LA-ICP-MS dating of zircon are tabulated in Appendix 2 and age calculations for plutonic rocks are summarized in Table 3. Graphical representations of these data are in Figure 3 for metasedimentary rocks and in Figure 4 for plutonic rocks. In Figure 3, histograms are used to illustrate the entire data set, including concordant and discordant analyses. Probability plots in Figures 3 and 4 are used to illustrate the distribution of concordant data within specific age ranges.

Lower Nappe

Sample N02.04 is a migmatitic metasandstone collected along the west side of Velfjord (Fig. 3A). The sample lacks concordant zircons younger than ca. 900 Ma, shows prominent peaks ca. 1150, 1450, 1800, and 2800 Ma, and has a gap in zircon ages from ca. 2100 to 2600 Ma (Fig. 3A, inset A).

Sample B-46 is also a metasandstone; it was collected at Fantvika on the west side of the Holm Peninsula (Fig. 3A). On the basis of regional map patterns and rock associations it is thought to be part of the Lower Nappe. Detrital zircon from this sample shows age peaks similar to that of sample N02.04 and an age gap exists from ca. 1900 to 2600 Ma (Fig. 3A, inset B).

Sauren-Torghatten Nappe

Sample B45 is a metasandstone from the Sauren-Torghatten Nappe, collected near the village of Berg (Fig. 3A). The detrital zircon population (Fig. 3A, inset C) is generally similar to that of the two Lower Nappe samples, but the smaller number of zircons makes individual peaks less well defined.

Sample B11A is a metasedimentary cobble from a conglomerate in the Bolvær area (Fig. 3A). The conglomerate deposit is ~100 m stratigraphically above the depositional basement of Sauren-Torghatten metasedimentary rocks (Heldal, 2001). The rich detrital zircon

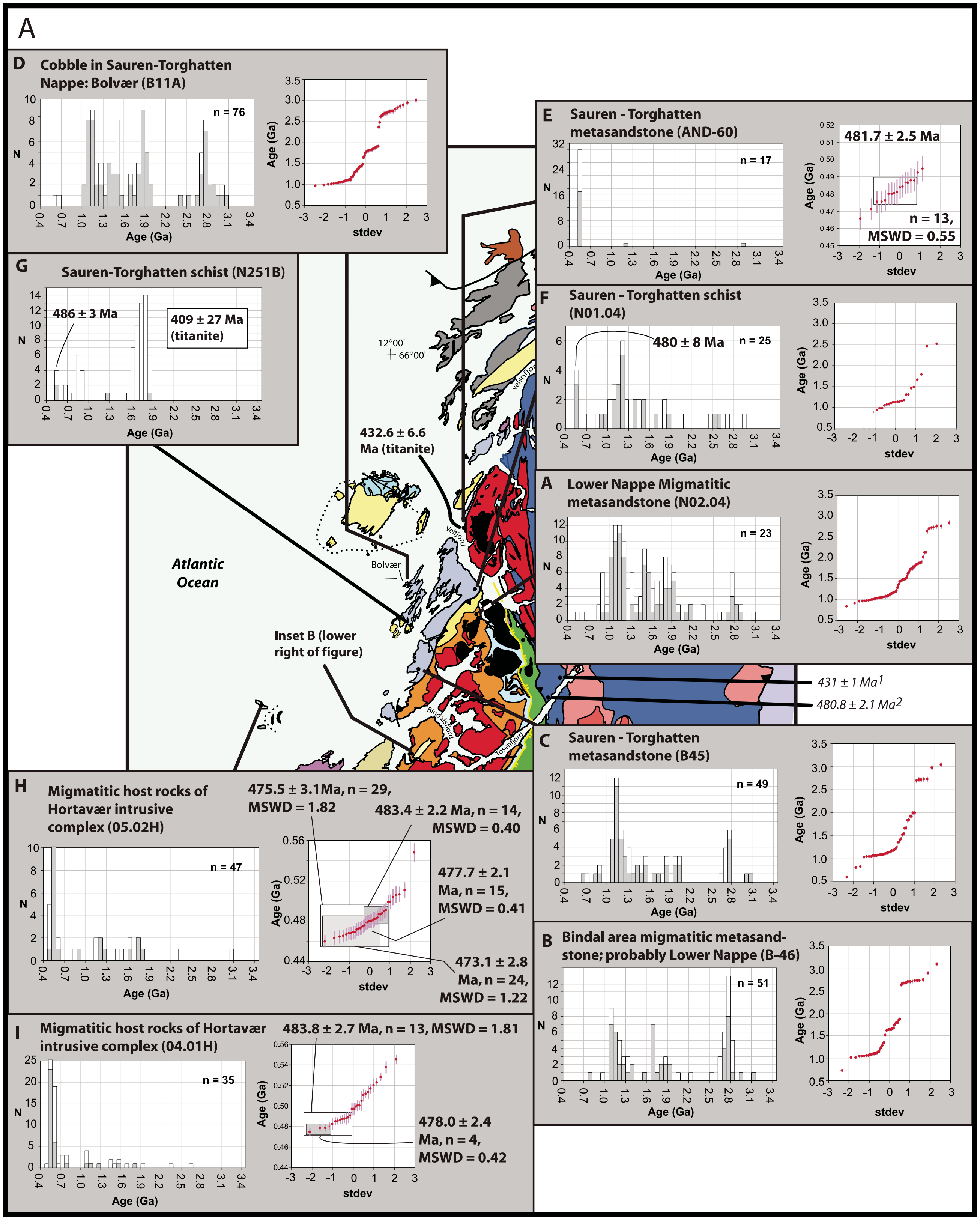

Figure 3 (*continued on next page*). **U-Pb results from detrital and magmatic zircon and metamorphic titanite for metasedimentary rocks and the S-type Vega pluton. Insets display histograms and probability plots of zircon age data. Probability plots for migmatites, the Sauren-Torghatten Nappe metasandstone, and the Vega pluton show age ranges (in boxes) used to interpret ages of migmatization, deposition, and magmatism, respectively. Ages reported in this paper are shown in bold print, others are in italics. Sources: 1—Bingen et al. (2002), 2—Yoshinobu et al. (2002). MSWD—mean square of weighted deviates. (A) Age data for the Lower Nappe, Sauren-Torghatten Nappe, and Horta nappe.**

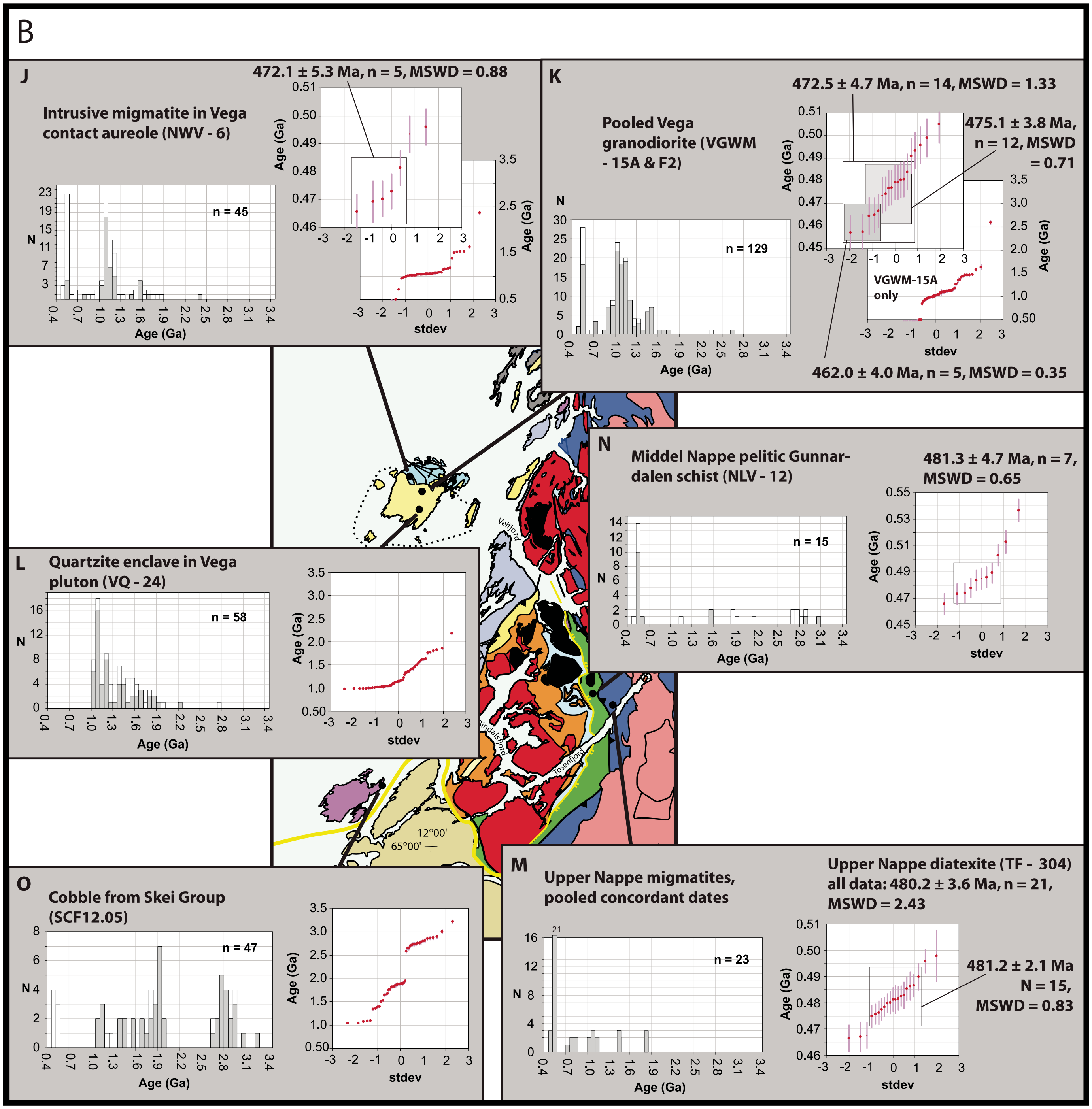

Figure 3 (*continued*). (B) Age data for the Middle and Upper Nappes, the Skei Group, and the Vega pluton.

population in this sample (Fig. 3A, inset D) is also strikingly similar to those of the Lower Nappe samples, although two zircons yielded dates within the 2000–2600 Ma range.

Sample AND-60 is a flysch-like metasedimentary rock collected in the Vevelstad area (Fig. 3A). Unlike the previous samples, the principal detrital zircon population is Ordovician, with two zircons of Late Cambrian age, one of Mesoproterozoic age, and one of Archean age (Fig. 3A, inset E). Except for the two youngest zircons, the early Paleozoic zircon population forms a relatively tight cluster in the probability plot, from 476 to 495 Ma.

Sample N01.04 is a garnet mica schist from a structurally high part of the nappe (Fig. 3A). The histogram shows a detrital zircon population unlike AND-60, but similar to those previously described (Fig. 3A, inset F). In addition, the sample contains three detrital zircons with ages of 471 ± 6, 482 ± 5, and 483 ± 5 Ma (mean = 480 ± 8 Ma). These data indicate that some of the detritus in this sample was derived from an Ordovician source.

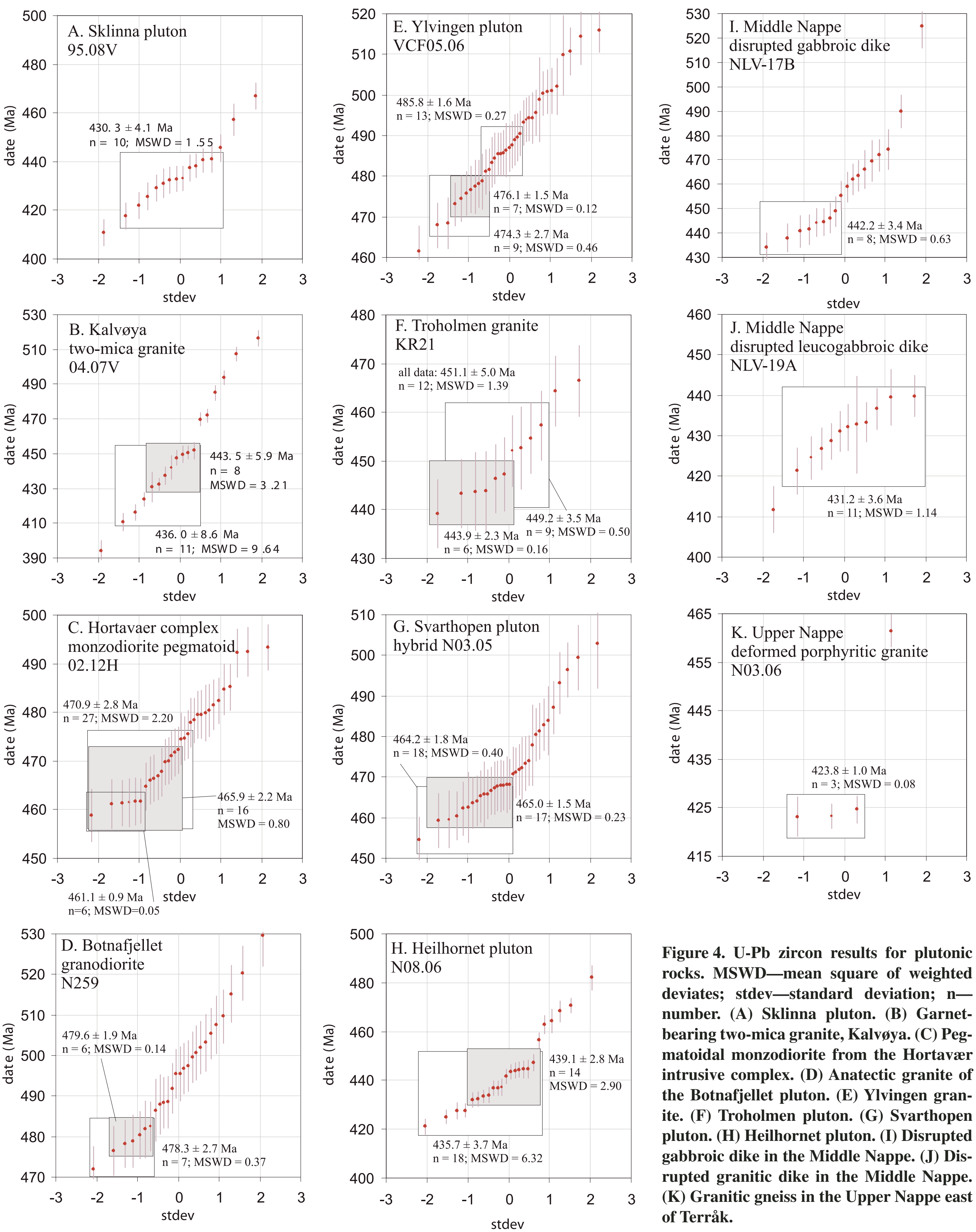

Figure 4. U-Pb zircon results for plutonic rocks. MSWD—mean square of weighted deviates; stdev—standard deviation; n—number. (A) Sklinna pluton. (B) Garnet-bearing two-mica granite, Kalvøya. (C) Pegmatoidal monzodiorite from the Hortavær intrusive complex. (D) Anatectic granite of the Botnafjellet pluton. (E) Ylvingen granite. (F) Troholmen pluton. (G) Svarthopen pluton. (H) Heilhornet pluton. (I) Disrupted gabbroic dike in the Middle Nappe. (J) Disrupted granitic dike in the Middle Nappe. (K) Granitic gneiss in the Upper Nappe east of Terråk.

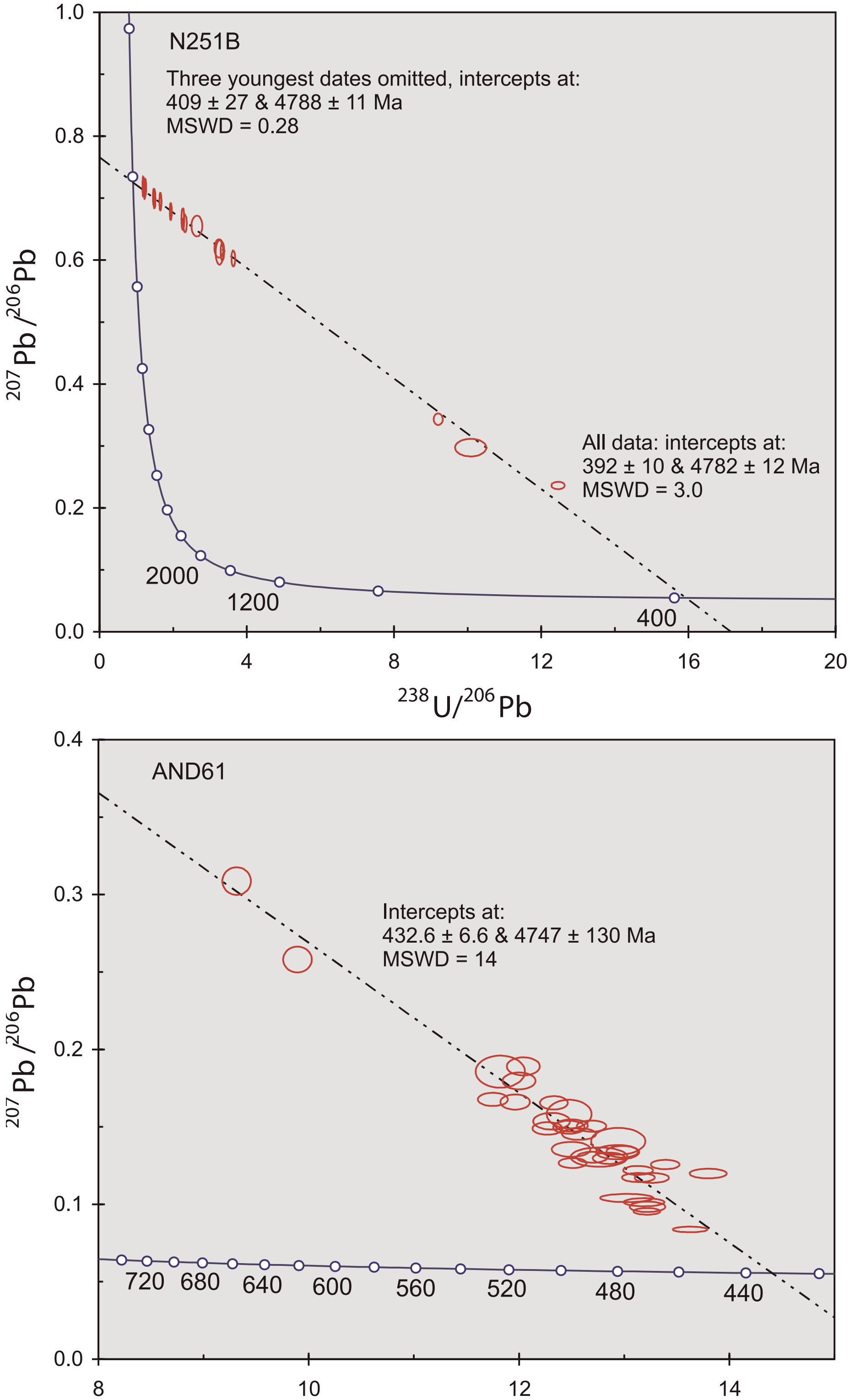

Figure 5. Concordia plots of U-Pb data for titanite. MSWD—mean square of weighted deviates.

Sample N251B is a garnet biotite hornblende schist from the Torget peninsula (Fig. 3A) and is within the structural aureole of the Torghatten pluton. Most zircons in this sample are discordant and lie on a crude chord (not shown). The two youngest concordant zircons yield an average age of 486 ± 3 Ma (Fig. 3A, inset G).

Horta Nappe

Two samples from the migmatitic host rocks west of the 466 Ma Hortavær igneous complex were dated (Fig. 3A). Both samples are leucosome portions of variably disrupted migmatitic gneiss. These samples are similar in having large Paleozoic zircon populations, smaller proportions of Proterozoic zircons, and no concordant Archean zircons (Fig. 3A, insets H, I). Therefore, the probability plots are scaled to show the ranges of Paleozoic zircons. In both samples, the youngest zircons are thin, clear, doubly terminated prisms. Sample 05.02H yielded concordant Cambrian and Ordovician zircon ages from 460 to 512 Ma and a single grain of 548 Ma. The probability plot (Fig. 3A, inset H) shows little change in slope between 460 and 495 Ma, and an average age over this range is 475.5 ± 3.1 Ma. Three subsets of this Paleozoic zircon population were made on the basis of slight changes of slope (Fig. 3A, inset H). These subsets may be interpreted to indicate older (483.4 ± 2.2 Ma; 477.7 ± 2.1 Ma) or younger (473.1 ± 2.8 Ma) ages of migmatization; however, neither subset is unique. Because the Hortavær igneous complex has an age of ca. 466 Ma (see below), we chose to restrict zircons in the age calculation to those older than 470 Ma. Moreover, because the sample contains Cambrian detrital zircons, it seems probable that, like samples from the Sauren-Torghatten Nappe, it contains Early Ordovician detrital zircons. We therefore chose to place an upper age limit of 485 Ma on zircons used to calculate the age of migmatization. This 470–485 Ma range results in an interpreted age of migmatization at 477.7 ± 2.1 Ma, with an MSWD of 0.41. In contrast, sample 04.01H contains only three zircons younger than 482 Ma (Fig. 3A, inset I). An average of these zircons yields 478.0 ± 2.4 Ma. If the range of zircons used in the age calculation is extended to 490 Ma, a date of 483.8 ± 2.7 Ma results (Fig. 3A, inset I). It is therefore possible to reach two distinct conclusions concerning the age of migmatization of these rocks, either ca. 477 Ma or ca. 483 Ma. Either age is consistent with the ca. 466 Ma age of the crosscutting Hortavær igneous complex. We note that the ca. 477 Ma age is probably not related to contact metamorphism because sample 05.02H was collected more than 3 km from the intrusive contact. Further support for the ca. 477 Ma age interpretation comes from the ages of the Vega pluton (see below).

No matter which age interpretation is accepted, if the Late Cambrian to Early Ordovician age zircons are detrital, then the migmatitic rocks of the Horta nappe west of the Hortavær igneous complex have a depositional age younger than Early Ordovician. This is in contrast to

TABLE 3. SUMMARY OF U-Pb (ZIRCON) AGES FOR PLUTONIC ROCKS

Sample	Age range (Ma)	Mean (Ma)	2 s.e.	MSWD	n
Sklinna biotite granite					
95.08V	415–445	**430.3**	**4.1**	**1.55**	**10**
	410–470	431.5	5.2	3.20	14
	Note: single grains at 987 ± 11, 1562 ± 29, 1610 ± 26, 1770 ± 18				
West Vikna (Kalvøya) deformed garnet two-mica granite					
04.07V	430–455	**443.5**	**5.9**	**3.21**	**8**
	410–455	436.0	8.6	9.64	11
Horta intrusive complex, pegmatitic monzodiorite					
02.12H	458–477	470.9	2.8	2.20	27
	458–474	**465.9**	**2.2**	**0.80**	**16**
	458–463	461.1	0.9	0.05	6
	Note: single grains at 504 ± 6, 506 ± 5, 524 ± 5				
Botnafjellet anatectic granite					
N259	475–485	479.6	1.9	0.14	6
	470–485	478.3	2.7	0.37	7
Ylvingen pluton, biotite granite associated with Vega pluton					
VCF05.06	480–493	485.8	1.6	0.27	13
	470–480	**476.1**	**1.5**	**0.12**	**7**
	465–480	474.3	2.7	0.46	9
Vega pluton					
VGWM-15A	455–486	472.5	4.7	1.33	14
	465–486	**475.1**	**3.8**	**0.71**	**12**
	455–470	462.0	4.0	0.35	5
	Note: broad peaks at 1027 ± 27 Ma and 1419 ± 62 Ma; one zircon at 2594 ± 39 Ma				
F2	none				
	Note: single zircon at 481 ± 7 Ma; two at 673.8 ± 5.5 Ma; peaks at 863 ± 16 Ma, 997 ± 11 Ma, 1100 ± 18 Ma, 1407 ± 30 Ma				
Troholmen granitic gneiss					
KR-21	440–462	**449.2**	**3.5**	**0.5**	**9**
	438–450	443.9	2.3	0.16	6
Svarthopen pluton					
N03.05	458–470	**465.0**	**1.5**	**0.23**	**17**
	454–468	464.2	1.8	0.40	18
Heilhornet pluton					
N08.06	430–450	**439.1**	**2.8**	**2.90**	**14**
	420–450	435.7	3.7	6.32	18
Disrupted gabbroic dike in the Middle Nappe					
NLV-17B	432–450	442.2	3.4	0.63	8
Disrupted leucogabbroic dike in the Middle Nappe					
NLV-19A	420–440	431.2	3.6	1.14	11
Deformed porphyritic granite (granitic gneiss) in Upper Nappe east of Terråk					
N03.06	420–425	423.8	1.0	0.08	3
	Note: one grain at 461 Ma				

Note: s.e.—standard error; MSWD—mean square of weighted deviates; n—number. Bold text indicates preferred age determinations.

the Neoproterozoic ages determined for marble screens within the Hortavær igneous complex. These results suggest that the Hortavær magmas intruded a contact between Neoproterozoic rocks on the east and early Paleozoic rocks on the west. This contact is now interpreted to be steeply west dipping. However, restoration of the original dip of layering in the Hortavær igneous complex (e.g., McCulloch et al., 2005; Barnes et al., 2006) shows that the Neoproterozoic rocks were structurally higher than the early Paleozoic rocks at the time of intrusion (see Fig. 6).

Vega Pluton and the Horta Nappe Migmatites

The Vega pluton (Fig. 2) is a classic S-type granite that underlies the southern part of the island of Vega and adjacent smaller islands (Gustavson, 1975, 1977; Nordgulen, 1992; Marko et al., 2005, 2006). The pluton is characterized by gneissic enclaves and enclaves and/or xenoliths of quartzite, marble, and calc-silicate rocks, as well as mafic magmatic enclaves. It can be thought of as an extreme diatexite (mobilized migmatite) with a paragneissic protolith. Therefore, the pluton should contain zircons inherited from its source rocks as well as primary magmatic zircon (e.g., see Williams, 2001; Miller et al., 2003).

Sample NWV-6 is a diatexitic migmatite dike that intrudes the host rocks of the Vega pluton, cuts two fold generations, and was emplaced during a third folding event (Anderson et al., 2005). The migmatite contains a few thin, clear, prismatic Ordovician-aged zircons (Fig. 3B, inset J) that we interpret to indicate an age of migmatization and intrusion ca. 472 Ma. The sample also contains two Cambrian-aged zircons, a large Grenvillian population, and sparse older Proterozoic grains.

Samples VGWM-15A and F2 are granodiorites from the main mass of the pluton. The histogram of pooled data for these two samples is shown in Figure 3B (inset K). Only two Paleozoic zircons were recovered from sample F2, so the probability plot shows data for sample VGWM-15A. This Paleozoic zircon population permits a number of interpretations of magmatic age. We prefer the 475.1 ± 3.8 Ma result, which is consistent with previous dating of the Vega pluton (Yoshinobu et al., 2002) and the age of the diatexitic dike described above. Sample VGWM-15A also contains inherited zircons of Cambrian age (491–506 Ma). The island of Ylvingen east of Vega is underlain by diatexitic rocks and biotite granite thought to be part of the Vega pluton (Fig. 2). As discussed below, the Ylvingen granite is interpreted to have a crystallization age of 474–476 Ma.

Sample VQ-24 (Fig. 3B, inset L) is from a kilometer-long disrupted quartzite xenolith in the Vega pluton (Vietti et al., 2005). The sample

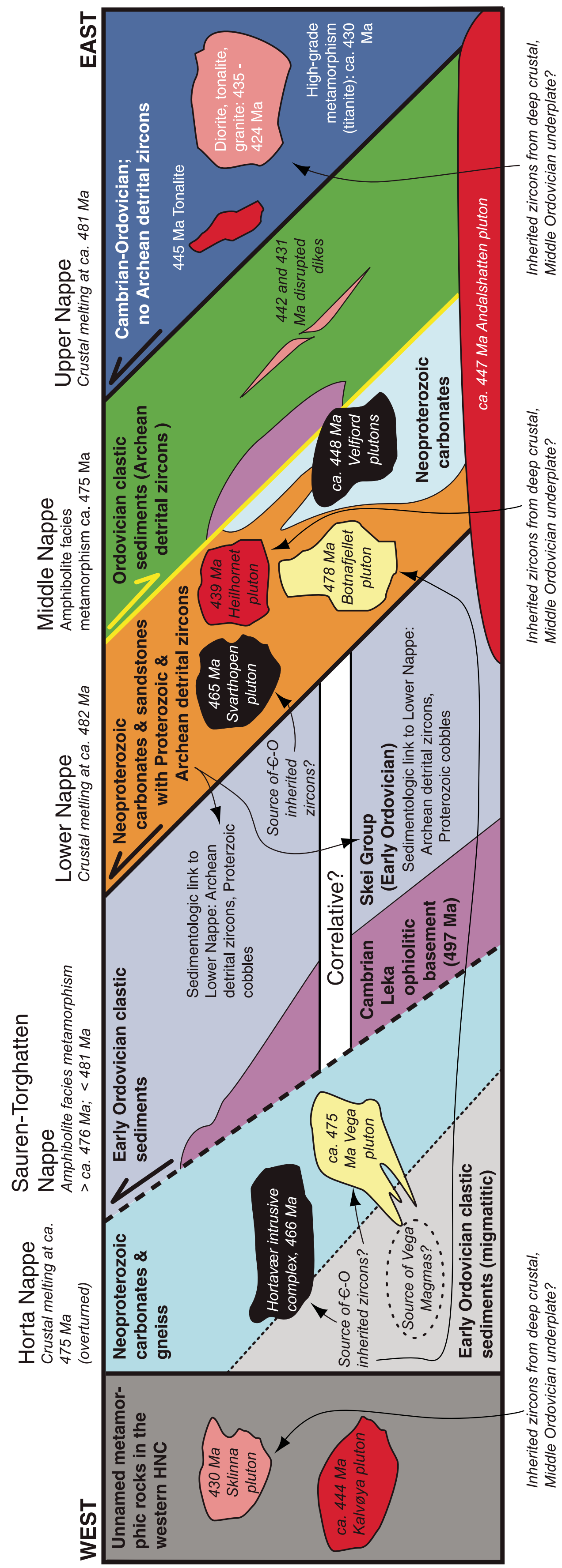

Figure 6. Schematic cross section of the Helgeland Nappe Complex nappes and plutons at the beginning of Scandian thrusting (ca. 430 Ma). Colors correspond to lithologies illustrated in Figures 1 and 2. The nature of the contact between the unnamed metamorphic rocks to the west and the Helgeland Nappe Complex (HNC) is unknown. The thrust that juxtaposes the Horta nappe against the Sauren-Torghatten Nappe is inferred. Possible sedimentologic and magmatic links between nappes and plutons (magma sources or contaminants) are shown with arrows.

lacks zircons younger than 1000 Ma, but is otherwise similar in zircon population to the Vega granodiorite (Fig. 3B, inset K).

Our mapping has shown that the rock types, diversity of enclaves, and isotopic composition of the Vega pluton are essentially identical to migmatites of the Horta nappe west of the Hortavær intrusive complex (Barnes et al., 2003; Marko et al., 2005). The lithologic types are characterized by a variety of gneissic rocks (particularly by a characteristic medium-grained, banded, pale gray siliceous gneiss), quartzite, marble, and calc-silicate rocks. Moreover, the Horta nappe migmatites were cut by synmigmatitic mafic dikes and the Vega pluton contains similar mafic magmatic enclaves. This striking similarity suggests that the migmatites of the Horta nappe are more or less intact equivalents of the Vega pluton, and that migmatization and mobilization of these rocks therefore occurred from ca. 475 to 478 Ma.

Middle Nappe

Detrital zircons were separated from pelitic Gunnardalen schist (Thorsnes and Løseth, 1991) from the central part of the Middle Nappe (sample NLV-12; Fig. 2). Few zircons were recovered, but Figure 3B (inset M) shows concordant ages that range from ca. 466 Ma to Archean. Among the Paleozoic detrital zircons, the youngest date is thought to be affected by Pb loss, because the age of metamorphism of this unit is ca. 475 Ma (see below). Nevertheless, the largest proportion of detrital zircons is of Ordovician age, with fewer Cambrian-age zircons and sparse Mesoproterozoic ones. The presence of Archean zircons suggests that some of the sediment contribution to the Middle Nappe was from rocks similar to those of the Lower Nappe.

Upper Nappe

Yoshinobu et al. (2005) presented U-Pb zircon ages for stromatic and diatexitic migmatites from the Tosen area of the Upper Nappe, the pooled dates for which are found in the histogram of Figure 3B (inset N). The migmatites contain a significant population of ca. 480 Ma zircons; this date was interpreted to be the age of migmatization. However, as shown in the probability plot for TF-304 (Fig. 3B, inset N), it is possible that zircons older than 490 Ma are detrital. The Precambrian detrital zircon population in these Upper Nappe migmatites is similar to those seen in the Horta nappe, but lacks the Archean component present in Lower Nappe metasedimentary rocks. These data indicate that deposition of Upper Nappe protoliths was no older than Neoproterozoic and may have extended into latest Cambrian time.

Skei Group

The depositional age of the Skei Group has been interpreted to be Ordovician (Sturt et al., 1985), which is consistent with the C and Sr isotope data for the Skei Group marble (see above). A metasandstone cobble from a Skei Group conglomerate yielded only Precambrian zircons (Fig. 3B, inset O), including a prominent group with Archean dates. The zircon population of this sample is very similar to that of the Lower Nappe metasandstones (Fig. 3A), which suggests that cobbles in at least some of the Skei Group conglomerates could have been derived from exposed Lower Nappe rocks.

U-Pb Ages of Zircon from Plutonic Rocks

The Sklinna pluton contains a cluster of zircons with dates in the 419–450 Ma range (Fig. 4A). If the youngest and oldest of these zircons are excluded from the population, a date of 430.3 ± 4.1 Ma results, with an MSWD of 1.55. Inclusion of the youngest and oldest data points does not affect the average (431.5 ± 5.2 Ma), but increases the MSWD to 3.2. The Sklinna sample also contains concordant inherited zircons of Ordovician, Mesoproterozoic, and Paleoproterozoic age.

Garnet two-mica granite is exposed in infolds of Caledonian rocks on Kalvøya, outer Vikna (Fig. 2). Zircons recovered from these rocks range from 392 to 514 Ma (Fig. 4B), with a distinct change in slope on the probability plot ca. 450 Ma. Figure 4B shows two possible age interpretations. We prefer the 443.5 ± 5.9 Ma value because of its lower MSWD.

A pegmatoidal monzodiorite from the Hortavær igneous complex (sample 02.12H; Fig. 2) contains a complex zircon population, from which a number of possible intrusive ages may be calculated. Figure 4C shows three such possibilities. In these calculations, zircons older than ca. 478 Ma are not used because we interpret the age of migmatization of the Horta nappe to be ca. 477 Ma (see above). It is noteworthy that the youngest calculated ages (465.9 ± 2.2 and 461.1 ± 0.9 Ma) are both within error of the 460.7 ± 5.1 Ma titanite age of another sample of the complex (see below).

Sample N259 was collected from an elongate, anatectic granodiorite exposed north of the Heilhornet pluton, herein referred to as the Botnafjellet pluton (Fig. 2). This pluton is locally crowded with metasedimentary enclaves, much like the Vega pluton. As with the Vega pluton, the Botnafjellet pluton contains a complex zircon population. Sample N259 yielded zircons with an age range from 471 to 530 Ma (Fig. 4D). We interpret the youngest of these zircons to record an igneous age of 478–480 Ma. (478.3 ± 2.7 or 479.6 ± 1.9 Ma; Fig. 4D). The remaining Ordovician- and Cambrian-aged zircons are interpreted to be inherited.

As indicated above, the Ylvingen pluton (Fig. 2) is thought to be the easternmost part of the large Vega pluton. Sample VCF05.06 of the Ylvingen pluton yielded a range of Paleozoic zircons from 461 to 515 Ma (Fig. 4E). In view of the discussion concerning the Vega plutonic system above, we interpret the Ylvingen pluton to have a crystallization age of 474–476 Ma. In this interpretation, zircons older than 480 Ma are considered inherited from an Ordovician-aged source and the ca. 461 Ma age is interpreted to result from Pb loss. Inheritance in the Ylvingen pluton also encompasses Mesoproterozoic and Paleoproterozoic zircons in a distribution similar to that seen in the migmatitic dike in the aureole of the Vega pluton (sample NWV-6; Fig. 3B, inset I).

A strongly foliated granitic pluton crops out on small islands (Troholmen) just west of Rødøy (Fig. 2). Sample KR21 from this pluton yielded zircons with ages from 439 to 467 Ma. If all analyses are included, an age of 451.1 ± 5.0 Ma results. Exclusion of the two oldest and the youngest zircons yields an age of 449.2 ± 3.5 Ma, and the youngest six crystals give an age of 443.9 ± 2.3 Ma. Because it is possible that deformation could be responsible for Pb loss in the youngest date, we tentatively accept 449.2 ± 3.5 Ma as the best age estimate for the Troholmen pluton. Either the 449 Ma or the 444 Ma age is consistent with the age of the Andalshatten batholith (447 ± 7 Ma; Nordgulen et al., 1993), which crops out 6 km to the south-southeast (Fig. 2) and locally shows similar deformation.

The Svarthopen pluton intrudes the Lower Nappe along the western side of Velfjord (Fig. 2). The pluton is characterized by zones of intimately mingled and mixed dioritic and granitic rocks. Hybrids range from heterogeneous enclave-rich zones to homogeneous intermediate rocks (Barnes et al., 2005b, 2006). Sample N03.05 is from a hybrid garnet granodiorite; it contains a suite of zircons with ages from 455 to 504 Ma (Fig. 4G). A distinct break in the probability plot occurs at 469 Ma, and we interpret the average of the younger crystals to represent the age of emplacement (ca. 465 Ma). The Svarthopen sample is unusual in that its inherited zircon population is entirely Ordovician and Cambrian; no Precambrian zircons were analyzed from this sample.

The Heilhornet pluton (Fig. 2) was previously dated as 444 ± 11 Ma (Nordgulen and Schouenborg, 1990). Because this pluton crosscuts important nappe structures (Nordgulen et al., 1989), we resampled it in an attempt to reduce the uncertainty of the age. Most of the zircons are stubby, doubly terminated prisms that range in age from 421 to 482 Ma (Fig. 4H). As with many plutons in the Bindal Batholith, a number of age interpretations are possible. We show two such possibilities and suggest that the 439.1 ± 2.8 Ma age is preferable in view of its lower MSWD and the likelihood that Pb loss may have affected the youngest dates.

Two samples of rocks intrusive into the Middle Nappe were collected from the Storvika area (Fig. 2). Both were deformed under amphibolite facies conditions and are disrupted into a sheared matrix. Sample NLV-17B is a mafic dike rock with a range of zircon ages from 434 to 526 Ma (Fig. 4I). We interpret the youngest zircons in this sample to represent an igneous age of 442.2 ± 3.4 Ma. This age is similar to the 445–448 Ma ages of the largest of the dioritic Velfjord plutons, which cut the Lower and Middle Nappes (Yoshinobu et al., 2002). Sample NLV-19A is a disrupted granitic dike. It yielded zircons in the age range 412–440 Ma (Fig. 4J), with one inherited grain at 528 Ma. If the youngest grain is excluded, the calculated age of the sample is 431.2 ± 3.6 Ma. Similar ages have been reported for granitic rocks that intrude the Upper Nappe east of the Storvika area (Nordgulen et al., 1993; Nissen et al., 2006; Yoshinobu et al., 2005).

Sample N03.06 is a deformed granite (granitic gneiss) collected 9 km east-southeast of Terråk (Fig. 2). On the basis of regional geologic patterns (Nordgulen et al., 1989), this unit is interpreted to be intrusive into Upper Nappe rocks. The sparse zircon population recovered from this sample contains a group of three crystals whose cores yield a mean age of 423.8 ± 1.0 Ma (Fig. 4K), a single grain at ca. 461 Ma, a group of zircons in the 706–842 Ma range, and a single grain at 1625 Ma. We interpret the ca. 424 Ma age as the age of intrusion, with the caveat that deformation may have resulted in resetting. However, an age of 424 Ma is consistent with ages of granitic intrusions elsewhere in the Upper Nappe (e.g., Nissen et al., 2006; Yoshinobu et al., 2005).

LA-ICP-MS Dating of Titanite

Sauren-Torghatten Nappe

Sample AND61 was collected from an undeformed dioritic dike in the western aureole of the Andalshatten batholith. Analyses of 14 titanite grains from this sample plot along a chord with a lower intercept of 432.6 ± 6.6 Ma and MSWD = 14 (Fig. 5). This result is tentatively interpreted to be the igneous age of the dike. Such an age is similar to ages of emplacement of mafic dikes in the Middle and Upper Nappes. However, the large MSWD may indicate resetting during this ca. 435–425 Ma event or could result from partial resetting during Scandian deformation.

The U-Pb data for titanite from sample N251B, the garnet biotite hornblende schist from the aureole of the Torghatten pluton, plot on a well-defined chord (Fig. 5). The complete data set yields a lower intercept at 392 ± 10 Ma with an MSWD of 3.0. If the three analyses with the highest $^{238}U/^{206}Pb$ are omitted, the resultant lower intercept is 409 ± 27 Ma with an MSWD of 0.28. These results are consistent either with resetting or with titanite growth during Scandian-age deformation. The data are not consistent with simple cooling from metamorphic conditions to titanite closure ca. 409 Ma, because titanites from other samples in the region preserve older ages; therefore, an interpretation of the age as representing resetting during late Scandian time is preferred.

Middle Nappe Metamorphism

A cobble collected from a polymict conglomerate of the Middle Nappe at Nyrud Pass (sample NLV-23) yielded metamorphic titanite with a lower intercept age of 475 ± 3 Ma from 21 spot analyses and an MSWD of 0.91. One of the analyzed grains had high common Pb compared to the others. If this analysis is omitted, the remaining 20 spots yield a lower intercept age of 472 ± 6 Ma with an MSWD of 0.91. In view of the probable 481 Ma minimum depositional age of Middle Nappe sedimentary rocks (see above), we interpret the ca. 475 Ma age to represent the age of regional metamorphism of the nappe.

Hortavær Igneous Complex

Barnes et al. (2003) reported a titanite age for a syenitic sample from the Hortavær complex. Titanite from sample 91.32H originally gave a lower intercept of 455.7 ± 8.4 Ma, with an MSWD of 1.05. This sample was reanalyzed and yielded a revised lower intercept date of 460.7 ± 5.1 Ma for 16 grains and an MSWD of 1.2. This date is within error of the 461–466 Ma U-Pb zircon age determined for the monzodiorite pegmatite from the complex (see above), so that the titanite date is interpreted to be an igneous cooling age that is somewhat younger than the U-Pb zircon age because of the lower closure temperature of titanite. In contrast, titanite from the sample dated by the U-Pb (zircon) method (sample 02.12H) gave a lower intercept of 430 ± 11 Ma for 19 spots, with an MSWD of 3.6. This result is perplexing in view of the igneous age of 461 Ma for sample 91.32H. We observe that titanite in sample 02.12H has an average Zr content of 94 ± 70 ppm, compared to more typical Zr contents of 1300–5200 ppm in igneous titanite from other Hortavær igneous complex samples. This distinction suggests that titanite in sample 02.12H may be metamorphic, thus recording a later thermal event.

DISCUSSION

This section begins with a discussion of the origins of metasedimentary and plutonic rocks of the Helgeland Nappe Complex. Possible sedimentologic and magma source links are schematically illustrated in Figure 6. We then consider some of the regional implications of the data, including the timing of nappe imbrication and possible tectonic connections with other Iapetan assemblages.

Metasedimentary Rocks

Lower Nappe

Sedimentary protoliths of the Lower Nappe are interpreted to be Neoproterozoic in age on the basis of chemostratigraphic evidence from marbles (Trønnes, 1994; Trønnes and Sundvoll, 1995; Sandøy, 2003). This is consistent with the absence of Paleozoic zircons in migmatitic metasandstones from Velfjord and Bindal (Fig. 3A). Sandøy (2003) found that some of the marbles in the Velfjord area were Ordovician in age; however, he did not assign these marbles to a particular nappe. We tentatively assign these Ordovician marbles to the Middle Nappe, but recognize that further work, particularly detrital zircon studies, is required to confirm this correlation. What is clear is that the provenance of the metasandstones in the Lower Nappe provided abundant Proterozoic and Archean detrital zircons. Migmatization of the Lower Nappe occurred before or during the ca. 480 Ma time frame, as is shown by small crosscutting anatectic granite plutons in the nappe (Yoshinobu et al., 2002).

Upper Nappe

The age of the Upper Nappe is less certain, primarily owing to sparse zircon data and a lack of dated marbles. However, the presence of sparse Cambrian-aged zircons in Upper Nappe migmatites (Fig. 3B, inset M) suggests Cambrian–Ordovician sedimentation. If so, the Lower and Upper Nappes are not correlative, despite their lithologic similarities. This interpretation is supported by the lack of Archean zircons and the paucity of Paleoproterozoic zircons in the Upper Nappe samples, compared to metasandstones of the Lower Nappe (Fig. 3). High-grade metamorphism and anatexis in the Upper Nappe occurred ca. 480 Ma. James et al. (1993) reported K/Ar (amphibole) ages of 470 ± 9 and 474 ± 5 Ma from a skarn zone along inner Tosenfjord. These ages probably represent cooling of Upper Nappe rocks through the amphibole closure temperature.

Other Neoproterozoic Nappe Units

Chemostratigraphic data (Table 2) suggest that marbles sampled from Caledonian infolds at Vikna and marbles of the Horta nappe are Neoproterozoic in age. In both cases, detrital zircon data would provide important additional information concerning the age and provenance of these deposits.

Other Early Paleozoic Units

Clastic metasedimentary rocks of the Sauren-Torghatten Nappe, the Middle Nappe, and the migmatitic paragneiss of the Horta nappe contain detrital zircon suggestive of deposition until Early Ordovician (Arenig) time (Fig. 3). Moreover, a number of metapelitic and metasandstone samples from these units and the Skei Group contain Archean zircons. These Archean zircons may be explained as having a source in the Lower Nappe or similar rocks (Fig. 6). A more complicated alternative may be proposed: that the Precambrian provenance of these Paleozoic sediments was identical to the provenance of the Lower Nappe sediments. We prefer the former interpretation because metamorphic cobbles in both the Sauren-Torghatten Nappe and Skei Group have zircon populations virtually identical to those of Lower Nappe metasandstones, as does the metasandstone unit near Berg (sample B45). These data suggest the presence of local sediment source(s) and short transport distances.

Deposition of Middle Nappe metasedimentary rocks is constrained to be older than the U-Pb age of metamorphic titanite ca. 475 Ma. Similarly, the age of deposition of the protoliths to Horta nappe migmatites must be older than the ca. 478 Ma age of migmatization. Apparently, all of the early Paleozoic units of the region underwent high-grade metamorphism within 5–10 m.y. of the end of deposition.

Sources of Plutonic Rocks

Few plutonic rocks in the Bindal Batholith can be ascribed to partial melting of mantle rocks without some assimilation and/or mixing of crustal material (Nordgulen and Sundvoll, 1992; Birkeland et al., 1993; Barnes et al., 2003, 2004). In some cases, this means that zircons inherited from the source region should provide information about crustal architecture. For example, the Vega-Ylvingen plutons carry a rich inherited zircon population that requires source rocks as young as Early Ordovician and/or Late Cambrian. These inherited zircon suites are entirely compatible with source rocks in the Horta nappe (see above; Fig. 6), in which case the Horta nappe migmatites represent a somewhat less mobilized correlative of the source of Vega-Ylvingen magmas.

The zircon assemblage of the pegmatoidal monzodiorite from the Hortavær igneous complex may be interpreted as the result of assimilation of migmatitic host rocks. This is a petrologically interesting result, because Barnes et al. (2005a) showed that magmatic evolution of Hortavær magmas required assimilation of both carbonate and silicate rocks.

Figure 4 also shows that the Svarthopen and Botnafjellet plutons contain Ordovician and Cambrian inherited zircons. The Svarthopen pluton clearly intrudes Neoproterozoic rocks of the Lower Nappe. Therefore, the presence of younger zircons must indicate the existence of underlying nappe(s) from which these younger zircons were inherited. The simplest explanation would be inheritance from, or assimilation of, rocks of the underlying Sauren-Torghatten Nappe or from the Skei Group (Fig. 6). If so, the ca. 465 Ma age of the Svarthopen pluton indicates nappe juxtaposition prior to that time. If, as regional geologic patterns indicate, the Botnafjellet pluton intrudes Lower Nappe rocks, nappe juxtaposition must be older than ca. 478 Ma.

The gabbroic dike rock from the Middle Nappe (NLV-17B) also contains Cambrian–Ordovician zircons. In this case, it is not possible to distinguish between inheritance from a deeper source versus in situ contamination.

Many of the plutons younger than ca. 440 Ma [e.g., Sklinna pluton, dioritic dikes in the Upper Nappe, deformed porphyritic pluton in the Upper Nappe (N03.06), and particularly the Heilhornet pluton] contain zircons from 450 to 460 Ma. Zircons of this age have not been found as detritus in any nappe of the Helgeland Nappe Complex. In some samples, these 450–460 Ma dates may represent ablation across two age zones, resulting in a spurious mixed age. However, in the Heilhornet sample (Fig. 4H), only one zircon yielded an age older than 468 Ma (496 ± 5.0 Ma; Appendix 2). Therefore, we interpret these 450–460 Ma zircons to be derived from deeper crustal levels; most probably from the source region of the magmas. Such a source could be a Middle Ordovician terrane, although we are unaware of any tectonic event that could explain the presence of such a terrane in the deep crust at this time. Alternatively, the source of the youngest Bindal magmas may have been a deep-crustal MASH zone, i.e., melting, assimilation, storage, and homogenization (Hildreth and Moorbath, 1988). Specifically, the MASH zone may have formed during the Middle Ordovician magmatic event responsible for the Hortavær and Svarthopen plutons in the Bindal Batholith and the Nesåa Batholith and related plutons in the Upper Allochthon. Because the MASH process recycles existing crust, the deep-seated products of the process may include quartz- and zircon-bearing rocks. If so, renewed mafic magmatism from 439 to 424 Ma could have partially melted the older MASH crust, with consequent mingling, hybridization, and inheritance of zircons that formed during the earlier MASH event. Such deep-seated rejuvenation of an older MASH zone is consistent with the tonalitic nature and low initial $^{87}Sr/^{86}Sr$ of many of the 439–424 Ma plutonic rocks (Nordgulen, 1992; Nordgulen and Sundvoll, 1992; Birkeland et al., 1993).

Toward a Regional Synthesis

Sediment Provenance

By Late Cambrian into Early Ordovician time, the Iapetus ocean basin was closing as Baltica approached Laurentia. In the outboard allochthons of the Norwegian Caledonides, as well as in the Newfoundland Appalachians, Early Ordovician tectonism involved development of oceanic arc sequences and related suprasubduction-zone ophiolites (e.g., Prestvik, 1980; Pedersen and Furnes, 1991; Cawood and Suhr, 1992; Roberts et al., 2002b). In the Upper and Uppermost Allochthons, ophiolite development predominantly ranged from 497 to 469 Ma (see compilation in Slagstad, 2003). In the Helgeland Nappe Complex, this activity is represented by the Leka ophiolite and ophiolitic fragments in the Sauren-Torghatten and Middle Nappes.

Following uplift and erosion of the ophiolitic sequences, high-energy conglomeratic sediments were unconformably deposited on them. The detrital zircon data from the Skei Group and the Sauren-Torghatten Nappe strongly suggest that one sediment source had a detrital zircon assemblage very similar to the Lower Nappe. Moreover, the abundance of conglomerates and coarse sandstones in these sediments implies significant topographic relief and/or short transport distance from the sediment source (e.g., Heldal, 2001). It thus seems likely that the Lower Nappe was the source of these sediments. This would provide a proximal source for coarse clastic sediments and cobbles with distinctive, Lower Nappe–like detrital zircons. It is also possible that the Neoproterozoic part of the Horta nappe is a rift fragment, conceivably related to the Lower Nappe. This topic is returned to in the following section.

Although a Lower Nappe–type provenance explains the Proterozoic zircon populations, the origins of the Cambrian–Ordovician detrital zircons (485–520 Ma) in the Sauren-Torghatten and

Middle Nappes are enigmatic. Volcanic activity may have accompanied ca. 480 Ma crustal melting in the Lower and Upper Nappes, but this activity cannot explain zircons with earliest Ordovician and Cambrian ages. The zircons may have been associated with magmatism related to ophiolites such as the Leka ophiolite, but this possibility is problematic because ophiolitic magmatism produces few zircons. On a regional basis, the 470–500 Ma age range is characterized by numerous suprasubduction-zone ophiolites and primitive arcs, many of which occur in the Upper Allochthon (summary in Slagstad, 2003). These may also be zircon sources; if so, then close proximity of the Uppermost Allochthon with the structurally highest nappes of Upper Allochthon is necessary by Arenig time.

Timing of Nappe Imbrication

Yoshinobu et al. (2002) showed that nappes of the Helgeland Nappe Complex were imbricated after migmatization of the Lower and Upper Nappes (ca. 480 Ma) and before emplacement of the ca. 447 Ma Andalshatten pluton. Our results provide significantly tighter constraints and indicate that imbrication had occurred by ca. 475 Ma. There are multiple reasons for this conclusion. (1) Regional metamorphism and anatexis of the Horta nappe ca. 478 Ma produced the parental magmas to the S-type Vega and Ylvingen plutons. The presence of peridotitic xenoliths in the Vega pluton (Marko et al., 2005) indicates that the Vega magmas engulfed peridotite from an overlying nappe (Leka ophiolite or Sauren-Torghatten Nappe), presumably by stoping. (2) We interpret the ca. 478 Ma Botnafjellet granite (sample N259) to be emplaced into the Neoproterozoic Lower Nappe. Thus, the Cambrian–Orodovician zircons in this granite must be inherited from a deeper, younger nappe. On the basis of its similarity with the Vega pluton, we interpret the source of the Botnafjellet magma to be the Horta nappe. (3) Cooling below amphibolite facies conditions occurred in the Middle Nappe ca. 475 ± 3 Ma, the age of metamorphic titanite in sample NLV-24B, and ca. 474 ± 5 Ma in the Upper Nappe, the age of skarn-related amphibole (James et al., 1993). Evidently, metamorphic activity in the Helgeland Nappe Complex began with high-grade metamorphism of the Lower and Upper Nappes and was followed by imbrication of all nappes ca. 478 Ma. Migmatization of the Horta nappe and consequent S-type granitic magmatism followed, along with amphibolite facies metamorphism of the structurally higher nappes. This sequence and timing of events is consistent with the regional metamorphic assemblages in the nappes: kyanite is present in the Sauren-Torghatten Nappe (Heldal, 2001) and the Vega magmas must have originated at pressures >~800 MPa (Marko et al., 2006). In contrast, sillimanite is the regional aluminosilicate phase in the Lower, Middle, and Upper Nappes and metamorphic cordierite is common only in the Upper Nappe (Yoshinobu et al., 2002; Reid, 2004). The cause of nappe amalgamation is currently speculative. Meyer et al. (2003) interpreted docking of the Uppermost and Upper Allochthons to be before the 456–458 Ma emplacement of the Nesåa Batholith. It is thus possible that docking of the Uppermost and Upper Allochthons was responsible for collapse and amalgamation of the Helgeland Nappe Complex.

Post-Amalgamation Magmatic Activity, 465 Ma

Following nappe amalgamation, volumetrically minor mafic magmatism occurred, as exemplified by the Hortavær igneous complex and the Svarthopen pluton ca. 465 Ma. Although plutonism of this age is uncommon in the Bindal Batholith, suprasubduction-zone magmatism in the 470–455 Ma age range is common elsewhere, including in the adjacent Upper Allochthon (e.g., Roberts and Tucker, 1991; Meyer et al., 2003). If, as suggested above, the Uppermost and Upper Allochthons were juxtaposed by 475 Ma, then the upper mantle and lower crust beneath the entire Helgeland Nappe Complex were probably affected by this magmatic activity.

450–444 Ma

The greatest volume of Bindal Batholith activity began ca. 450 Ma with emplacement of the Troholmen pluton and the large, diverse Andalshatten pluton (ca. 447 Ma; Nordgulen et al., 1993), the Velfjord plutons (448–445 Ma; Yoshinobu et al., 2002), and a number of smaller masses such as the Kalvøya pluton (ca. 444 Ma) and a tonalitic intrusion in the Upper Nappe (447.8 ± 1.7 Ma; Yoshinobu et al., 2005). Emplacement conditions at this time were ~600–700 MPa for plutons emplaced into the Lower Nappe (Barnes and Prestvik, 2000; Yoshinobu et al., 2002) and ~330 ± 45 MPa for the tonalite emplaced into the Upper Nappe (Reid, 2004). However, exhumation accompanied this magmatic pulse, so that rocks of the Lower Nappe had reached pressures of ~400 MPa ca. 445 Ma (Barnes and Prestvik, 2000; Yoshinobu et al., 2002).

439–424 Ma

The next known magmatic event was emplacement of the Heilhornet pluton ca. 439 Ma, and plutonism continued until ca. 424 Ma. This 439–424 Ma time interval was characterized by mafic, intermediate, and felsic magmas and it affected all of the Helgeland Nappe Complex, from the Sklinna pluton in the west to large tonalitic masses in the east (Nordgulen, 1992; Nissen et al., 2006). The presence of 450–460 Ma inherited zircons in some of these plutons suggests that the deep crust of the Bindal Batholith was significantly reworked (MASH zone; Fig. 6) despite the paucity of 450–460 Ma plutons in the exposed parts of the batholith.

Paleogeographic Considerations

The structurally highest nappe complexes in the Norwegian Caledonides have long been recognized as originating distal to the Baltic craton. Some, such as the Helgeland Nappe Complex, are thought to have origins in Laurentia on the basis of depositional and tectonic considerations (e.g., Roberts et al., 2001, 2002a; Yoshinobu et al., 2002; Roberts, 2003). Others, such as the Kalak Nappe Complex in northern Norway, have been interpreted as exotic to either Baltica or Laurentia (Corfu et al., 2007). Comparison of the lithologic features, thermal history, detrital zircon populations, and tectonic development of the Helgeland Nappe Complex with other circum-Iapetan terranes reveals many similarities but also significant differences.

Several recent studies have suggested that the Helgeland Nappe Complex originated near what is now East Greenland (e.g., Roberts et al., 2001, 2002a; Yoshinobu et al., 2002; Roberts, 2003). This location is partly based on reconstruction of plate geometries prior to opening of the North Atlantic. However, Ordovician magmatic activity that characterizes the Helgeland Nappe Complex is not evident in East Greenland (e.g., Kalsbeek et al., 2001). Moreover, Scandian-age thrusting in East Greenland was directed toward the northwest, but was in the opposite direction for the Helgeland Nappe Complex (Higgins and Leslie, 2000).

Detrital zircons from Lower Nappe metasandstone are comparable to those from the Dalradian Supergroup of Scotland, and specifically from the Argyll and Southern Highland Groups (Fig. 7; Cawood et al., 2003). This similarity suggests that the Dalradian units had a sediment provenance similar to that of the Lower Nappe. Although such zircons may be derived from either Laurentia or Baltica, Cawood et al. (2007b) interpreted the distinctive Dalradian zircon assemblage to result from deposition near the Laurentian (East Greenland) margin, north of the Grenville-Sveconorwegian deformation front. This location was apparently isolated from drainage systems that carried Pan-African–age zircons. The marked similarity of the Dalradian and Lower Nappe zircon suites appears to identify a distinctive set of sedimentary units deposited in a specific, geographically limited

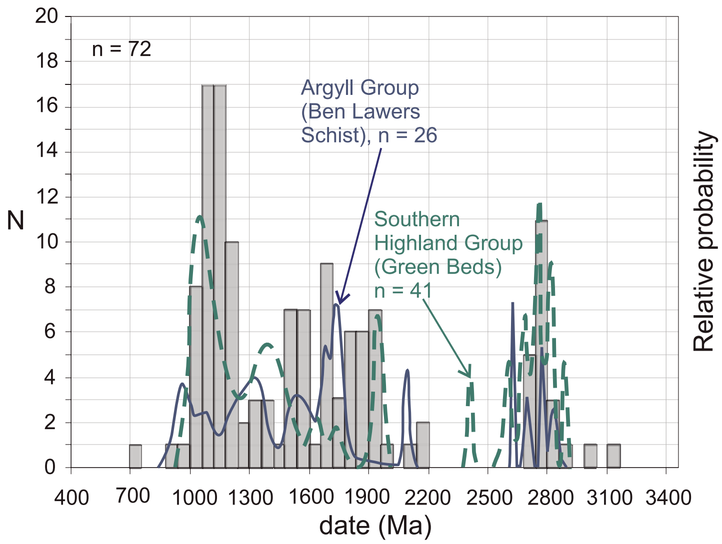

Figure 7. Comparison of the pooled concordant zircon ages from the Lower Nappe (histogram) with relative probability plots of zircons from the Argyll Group and Southern Highlands Group of the Dalradian Supergroup of Scotland (Cawood et al., 2003).

part of the Iapetus system: southeast Greenland, according to Cawood et al. (2007a).

Early Ordovician ophiolites that are present within the Caledonian–Appalachian mountain belt have been considered an inherent component of the Taconic or the time-equivalent Grampian orogenic events along the southeast margin of Laurentia (e.g., Pedersen and Furnes, 1991; Cawood and Suhr, 1992; van Staal et al., 1998). A number of models have been proposed for ophiolite development, the most recent of which involve rifting of the Laurentian margin to form peri-Laurentian microcontinent(s) separated from Laurentia by small ophiolite-floored ocean basins (e.g., Cawood et al., 1995; van Staal et al., 1998; Waldron and van Staal, 2001; Hatcher et al., 2004; Hibbard et al., 2007). In these models, the microcontinents serve as the basement for Cambrian–Ordovician arcs. Collision of the microcontinent and/or arc assemblages then resulted in Taconian deformation (e.g., Waldron and van Staal, 2001). High-pressure metamorphism and regional development of migmatites, reportedly related to collision of a volcanic arc with the Laurentian margin, occurred within a similar time frame in the Northern Highlands of Scotland (Kinny et al., 1999; Friend et al., 2000).

In the preceding section, we suggested that the Lower Nappe and the Neoproterozoic part of the Horta nappe may be rift fragments, presumably from the Laurentian margin. The ophiolitic basement rocks of intercalated nappes (Fig. 6) would then represent one or more ocean basins that separated these rift fragments. The similarities with Taconian and Grampian tectonic models prompt us to consider potential correlation between units of the Uppermost Allochthon with rock complexes elsewhere in the mountain belt that also underwent Early Ordovician tectono-magmatic evolution. In the peri-Laurentian arc complexes of Newfoundland, Early to Middle Ordovician magmatism was primarily tholeiitic to calc-alkaline (Zagorevski et al., 2006). In contrast, coeval magmatism in the Helgeland Nappe Complex was almost entirely granitic and was associated with regional migmatization and development of strongly peraluminous granites such as the Vega and Botnafjellet plutons. In northwestern Ireland, continental arc magmatism and associated metamorphism including anatexis took place ca. 475–463 Ma (Friedrich et al., 1999a, 1999b; Draut and Clift, 2001; Draut et al., 2004; Flowerdew et al., 2005), ages that slightly overlap the anatectic activity in the Helgeland Nappe Complex. In the Grampian terrane of Scotland, emplacement of gabbro (Dempster et al., 2002) and syntectonic peraluminous granitoids (Oliver et al., 2000) took place ca. 470 Ma.

In summary, the timing of ophiolite formation, metamorphism, and plutonism in the Helgeland Nappe Complex supports a general correlation between it and rock complexes in the northwestern British Isles and in Newfoundland. However, we caution that the dimensions and complexity of the Helgeland Nappe Complex in particular, and the Uppermost Allochthon in general, are similar to those of the much better studied regions in the Appalachians and British Caledonides. Thus, this contribution to the geochronology of the Helgeland Nappe Complex should be viewed as the basis for additional, detailed studies of the sedimentary, metamorphic, igneous, and tectonic development of this important part of the Scandinavian Caledonides.

CONCLUSIONS

Nappes of the Helgeland Nappe complex record a sequence of Neoproterozoic through Ordovician tectonic, sedimentary, magmatic, and metamorphic events. The two oldest nappes are the Lower Nappe and the Horta nappe. The Lower Nappe was deposited in a shelf environment during Neoproterozoic time, presumably along the margin of Laurentia. Its detrital zircon population is similar to that of Dalradian sedimentary rocks in Scotland, which is consistent with deposition near modern southeastern Greenland. The Horta nappe consists of Neoproterozoic basement similar to the Lower Nappe, which is overlain by a sequence of clastic-dominated early Paleozoic rocks. We have too few data to evaluate possible correlations between the Lower Nappe and Horta nappe; however, the distinctive zircon population of metasandstones from the Lower Nappe provides a valuable tool for future work. Rifting of these Neoproterozoic crustal rocks during Cambrian time resulted in ophiolite-floored basins and possible development of one or more microcontinental blocks. The intervening basins were filled with a combination of locally derived coarse clastic sediments, pelitic sediments, and carbonates. The basin deposits are represented by the upper part of the Horta nappe, the Skei Group on Leka, and the Sauren-Torghatten, Middle, and possibly the Upper Nappes. The sediments are younger than the 497 Ma age of depositional basement; deposition ended by ca. 481 Ma. High-grade metamorphism of the Lower and Upper Nappes occurred ca. 480 Ma and was followed by imbrication of the entire nappe sequence. By ca. 478 Ma, the Horta nappe was at the bottom of the nappe sequence, where it underwent migmatization and was the source of S-type magmas. This sequence of nappe juxtaposition and crustal melting occurred no more than a few million years after deposition of the Early Ordovician sediments, which implies deposition in an active continental margin setting. Younger magmatic activity consisted primarily of plutonism ca. 465 Ma, 450–444 Ma, and 439–424 Ma. Exhumation accompanied the 450–444 Ma plutonism. The youngest plutons contain zircons in the 460–450 Ma age range, which is interpreted to indicate an origin from a deep-seated MASH zone near the base of the Helgeland Nappe Complex crust.

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