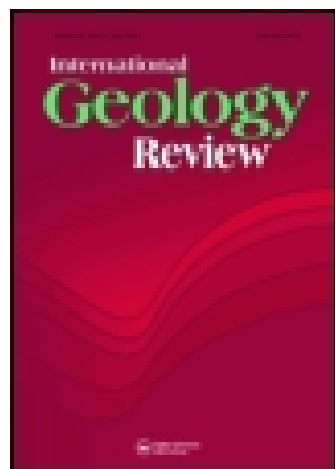




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Characteristics of, and a Tectonic Model for, Ultrahigh-Pressure Metamorphism in the Overriding Plate of the Caledonian Orogen

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Abstract

Ultrahigh-pressure (UHP) metamorphism in the Laurentian margin of North-East Greenland formed in the overriding plate of the Caledonian collision with Baltica. The Greenland UHP terrane exhibits a number of characteristics that are incompatible with accepted models for UHP metamorphism in the downgoing slab during continental subduction. UHP metamorphism occurred very late in the collision, and was a long-lived event. Peak temperatures in excess of 950°C are well above the norm for subduction-zone UHP terranes. Exhumation rates, based on new U-Pb SHRIMP results for two boudin neck pegmatites and one hornblende-bearing leucosome, are slower (~3–5 mm/year) than for many other UHP terranes. These atypical features can be explained by intracratonic subduction of Laurentia in the waning phase of the Caledonian collision, analogous to deformation that may be occurring in the Tibetan Plateau today.

Introduction

PHANEROZOIC ULTRAHIGH-PRESSURE metamorphism is now recognized from over 20 localities around the world, based on the presence of the index minerals coesite and/or diamond (Chopin, 2003; Liou et al., 2004). The vast majority of UHP terranes are formed in continental crust, demonstrating that continental crust can indeed be subducted to mantle depths. A new paradigm, modeled after the Himalayan collision between Asia and India, is emerging to explain the common features of these continental UHP terranes (Liou et al., 2004; Ernst, 2005; 2006). The idea is that UHP metamorphism occurs at the leading edge of the continent in the downgoing plate of a continent-continent collision; consequently, UHP metamorphism occurs early in the collision. Significant amounts of metasedimentary rock, in addition to quartzofeldspathic basement gneiss, commonly reach depths in excess of 100 km. Peak temperatures (T) are typically < 800°C—i.e., low for such elevated pressures (Ernst et al., in press), but easily explained by the depressed isotherms found within subduction zones (Peacock, 1990). The entire round trip from the Earth's surface down the subduction zone and back is of short duration, on the order of 10–20 m.y. (Ernst, 2006). The subduc-

tion channel provides a viable pathway for exhumation; therefore, exhumation rates are approximately equal to subduction rates, inasmuch as material returns up the same channel it went down.

The Caledonian collision between Laurentia and Baltica presents a challenge to this paradigm because UHP metamorphism is found in the continental margins of both colliding plates (Fig. 1). The well-known Western Gneiss Region (WGR) of southwestern Norway contains UHP rocks whose protoliths are correlated with the continental crust of Baltica's Fennoscandian shield, whereas UHP rocks in the North-East Greenland eclogite province (NEGEP) are part of the Laurentian margin. Laurentia is assumed to be the overriding plate in this collision, analogous to Asia in the Himalayan collision. In this paper, we show that several features of the Greenland UHP rocks do not match those predicted by the continental subduction paradigm. Specifically, T exceeds 900°C, peak P is achieved very late in the collision, and exhumation is apparently slower than the norm. We present new U-Pb-sensitive high-resolution ion microprobe (SHRIMP) zircon ages from two pegmatites in boudin necks between eclogite blocks and an amphibolite-facies leucosome in order to calculate better exhumation rates. We suggest that the anomalous features seen in North-East Greenland may be characteristic of

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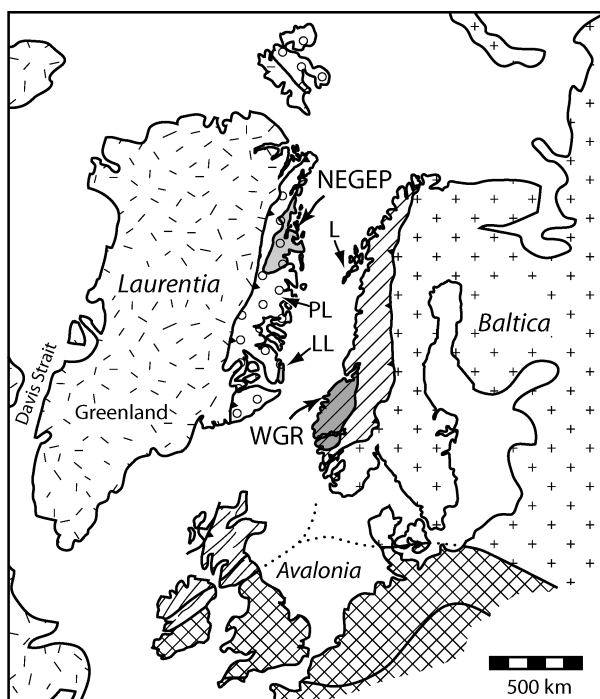


FIG. 1. Pre-rift reconstruction of Baltica and Laurentia based on Soper and Woodcock (2003). Abbreviations: NEGEP = North-East Greenland eclogite province; L = Lofoten Islands; PL = Payer Land; LL = Liverpool Land; WGR = Western Gneiss Region.

UHP metamorphism occurring in the overriding plate of a continent-continent collision. Finally, we present a tectonic model that can explain high-pressure (HP) and UHP metamorphism on both margins of the Caledonian orogen, as well as in the overriding plate in general.

HP and UHP Metamorphism in the Continental Margins of Baltica and Laurentia

Large HP eclogite terranes and associated smaller UHP terranes are found on the continental margins of both Baltica and Laurentia (Fig. 1). The WGR in southwestern Norway covers an area of approximately 50,000 km² and is composed of quartzofeldspathic gneisses with abundant meter- to kilometer-scale lenses of eclogite and garnet peridotite (e.g., Krogh and Carswell, 1995, and references therein). The protolith to the gneisses is a sequence of Paleoproterozoic (1685–1650 Ma) granitoid intrusive rocks with younger components

of rapikivi granites and gabbros that are interpreted as a continuation of the Baltic shield (Corfu and Andersen, 2002; Austrheim et al., 2003; Tucker et al., 2004). HP metamorphism in the WGR occurred between ~420 and 400 Ma (Griffin and Brueckner, 1980, 1985; Mørk and Mearns, 1986; Krogh et al., 2003). Eclogites are also found in the islands of Lofoten (Markl and Bucher, 1997) within Paleoproterozoic continental crust that may or may not be part of the Baltica crystalline basement (Corfu, 2004). The eclogites are common along shear zones and are likely to be Caledonian (Boundy et al., 2003; Steltenpohl et al., 2006), but an accurate age is not known. Eclogite terranes also formed outboard of Baltica in multiple, earlier subduction events involving arcs, ophiolites, and microcontinental fragments prior to the continent-continent collision proper (Brueckner, 2006). These HP/UHP terranes, which include parts of the Bergen Arcs, Seve Nappe Complex (see Brueckner and van Roermund, 2004 for a review) and the Troms Nappe (Ravna and

Roux, 2006), reside in the stack of thrust sheets preserved in the Scandinavian Caledonides.

Starting with the discovery of coesite in eclogite (Smith, 1984) and continuing with the recognition of rare diamonds (Dobrzynetska et al., 1995; Godard et al., 2003), the WGR UHP terrane is now recognized in three areas along more than 100 km of the outer coast of southwestern Norway (Carswell and Cuthbert, 2003, and references therein). If the three main structural culminations are connected at depth, the UHP terrane may be a 5 km thick sheet covering an area > 8,000 km² (Root et al., 2005). The UHP terrane is a similar mixture of quartzofeldspathic gneisses enclosing mafic and ultramafic eclogite pods as seen in the WGR proper. It is also assumed to be part of the continental margin of Baltica (Andersen et al., 1991; Tucker et al., 2004). A number of attempts have been made to determine the timing of UHP metamorphism in the WGR (Terry et al., 2000a; Root et al., 2004), producing a range of ages from 415 to 400 Ma. The most robust result is a 402 ± 2 Ma U-Pb TIMS age from a population of metamorphic zircon in which three coesite inclusions were found, although coesite apparently was not present in the analyzed grains (Carswell et al., 2003; Tucker et al., 2004). UHP metamorphism occurred near the end of HP metamorphism in the WGR.

Eclogite-facies metamorphism has affected the crystalline basement of Laurentia along the length of the orogen. The largest area of HP metamorphism is the NEGEP (Gilotti, 1993; Elvevold and Gilotti, 2000), which extends approximately from 76° to 81°N and covers roughly the same area as the WGR (Figs. 1 and 2A). Abundant layers and lenses of eclogite *sensu stricto*, garnet-clinopyroxenite, garnet-websterite, websterite, and coronitic gabbro enclosed within Precambrian gneisses attest to widespread eclogite-facies metamorphism (Gilotti, 1994). The earlier geochronology of Brueckner et al. (1998) on the age of metamorphism has been refined, and is superseded by new results that place HP metamorphism between 415 and 390 Ma (Gilotti et al., 2004). This range of ages is based on corresponding Sm-Nd mineral isochrons and U-Pb SHRIMP analysis of zircon in four eclogite samples located across the NEGEP. The U-Pb ages are tied to the HP metamorphism through trace element chemistry of the dated zircons, where the lack of a Eu anomaly is attributed to zircon growth in plagioclase-free (i.e., eclogite-facies) assemblages (Rubatto, 2002). HP metamorphism is also recog-

nized in Laurentian basement in the southern part of the orogen. The Payer Land gneisses at 74°30' N are HP granulites of mafic, felsic, and pelitic lithologies derived from continental crust of both Archean and Paleoproterozoic protoliths (Gilotti and Elvevold, 2002; Elvevold et al., 2003). The HP metamorphism is ~405 Ma, based on U-Pb SHRIMP ages from zircons formed in HP leucosomes of kyanite + garnet + K-feldspar + plagioclase + quartz from anatectic metapelites and zircon rims from interbedded metapsammities with the same HP assemblage (McClelland and Gilotti, 2003). Less is known about the Liverpool Land localities (71°N), but it appears that there are abundant eclogites formed at ~400 Ma (Hartz et al., 2005) that reside in crystalline rocks of the same deep structural level as Payer Land (Henriksen, 2003).

UHP metamorphism in Greenland is documented on an island in Jøkelbugten (78°00' N, 18°04' W), informally known as "Rabbit Ears Island" (Fig. 2B). The true extent of the UHP rocks is unknown, but the slice of UHP basement must be at least 5 km thick based on the structural thickness exposed on the island. The UHP gneisses have the same compositions and protoliths as the adjacent gneisses in the NEGEP. Polycrystalline quartz pseudomorphs after coesite are common inclusions within garnets and pyroxenes from eclogites (Gilotti and Ravna, 2002). Coesite inclusions in zircon have been identified with laser Raman spectroscopy in four samples, including a host gneiss (Fig. 2B; McClelland et al., 2006). Peak metamorphic conditions calculated from the assemblage garnet + omphacite + kyanite + coesite/quartz + phengite are 3.6 GPa at 970°C (Gilotti and Ravna, 2002). The UHP metamorphism was dated at 365–350 Ma by identifying the coesite-bearing zircon domains in cathodoluminescence (CL) images and then analyzing those parts of the zircon with the U-Pb SHRIMP method (McClelland et al., 2006). The age of UHP metamorphism in the NEGEP is much younger than the HP metamorphism in Greenland and UHP metamorphism in the WGR.

The genealogy of the Laurentian and Baltica continental margins hinge on linking the now allochthonous basement terranes in the Caledonian orogen to their respective cratons. The evidence connecting the WGR to Baltica and the HP terranes in North-East Greenland, Payer Land, and Liverpool Land to Laurentia is mostly based on ages of the protoliths and similarities in Precambrian geologic histories. In Laurentia, the contacts between

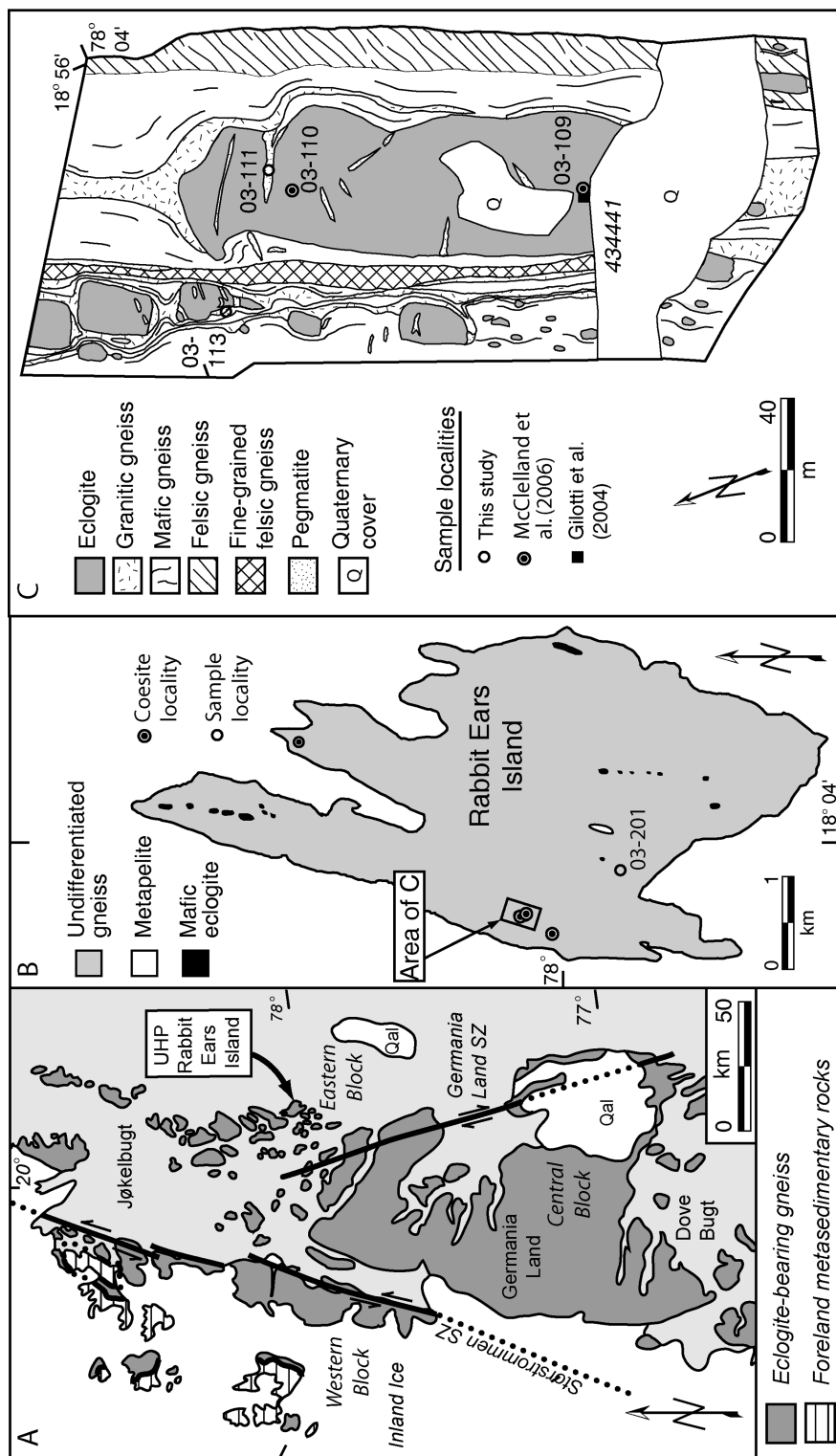


FIG. 2. A. Geological map of the North-East Greenland eclogite province. B. Geological sketch map of Rabbit Ears Island. C. Detailed geologic map showing location of pegmatite samples 03-111 and 03-113 in the necks of UHP eclogite boudins discussed in the text, modified from Figure 3 in McClelland et al. (2006).

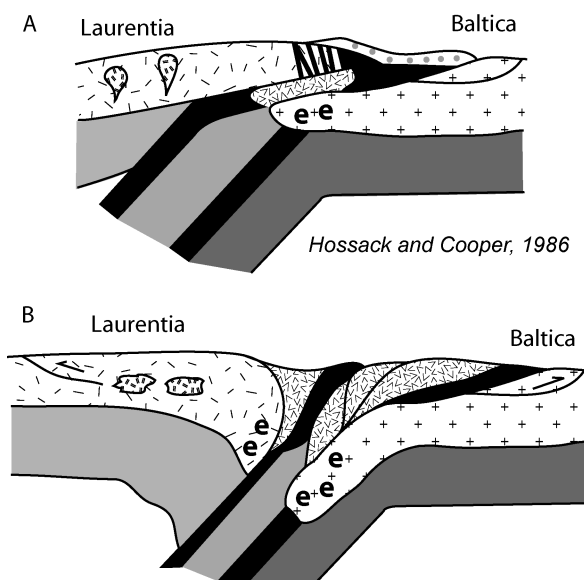


FIG. 3. Old (A) and revised (B) tectonic model for Caledonian collision based on Hossack and Cooper (1986).

Archean and Proterozoic basement are essentially E-W across the Canadian shield (Hoffman, 1988) and can be followed across to West Greenland. Thrane (2002) has established an E-W boundary at $\sim 73^\circ\text{N}$ between Paleoproterozoic gneisses to the north and Archean gneisses to the south in the East Greenland Caledonides that can be connected to the northern Archean–Paleoproterozoic boundary in West Greenland. The quartzofeldspathic gneisses in North-East Greenland are derived from juvenile Paleoproterozoic calc-alkaline arcs intruded at 2.0–1.8 Ga that contain a component of 1.75 Ga anorogenic granites (Kalsbeek et al., 1993; Kalsbeek, 1995). These Paleoproterozoic gneisses are found in the foreland of the thrust belt and match the lithologies and ages of the gneisses in the hinterland thrust sheets, including the NEGP in the uppermost thrust sheet. On the other margin, the WGR is correlated with the Baltic shield because Precambrian gneisses occur in thrust culminations across the Scandinavian Caledonides (Gee, 1975), and the WGR is the westernmost of these basement windows. Protolith ages of approximately 1.65 Ga in the WGR match the age, if not the geochemistry, of the Transcandinavian igneous province of the Fennoscandian shield (Tucker et al., 2004). In addition, a Sveconorwegian overprint (~ 990 Ma) is recognized in some of the gneisses of the WGR (e.g., Corfu and Andersen, 2002; Røhr et al., 2004) that

can be correlated with events of similar age on the Fennoscandian shield in southwest Sweden. One problem with all these correlations is that the basement gneisses of Baltica and Laurentia may have previously collided to form the Neoproterozoic supercontinent of Rodinia (e.g., Hoffman, 1991; Dalziel, 1997; Cawood and Pisarevsky, 2006). Rifting of Rodinia could leave pieces of one craton on the other, as well as rifted microcontinental fragments in Iapetus, compounding the difficulties of determining the provenance of allochthonous basement in later orogens (Thomas, 2006). In the following discussion, we assume the NEGP is of Laurentian origin and the WGR is derived from Baltica.

Laurentia as the Overriding Plate

The Caledonian orogen is traditionally viewed as an asymmetric, continent–continent collision between Baltica and Laurentia, with Laurentia as the overriding plate (Gee, 1975; Hodges et al., 1982; Hossack and Cooper, 1986). The iconic view of the architecture of this collision (Fig. 3A) was based on three main lines of reasoning: (1) the recognition of east-directed, far traveled thrust sheets on the margin of Baltica, including an imbricated section of Baltic margin sedimentary rocks and a plethora of exotic terranes; (2) the

presumption that the granites in Greenland (Laurentia) were dominated by I-type plutons derived from partial melting of the downgoing, subducted oceanic slab attached to Baltica; and (3) the notion that the eclogites in the WGR were formed by doubling the thickness of the crust by emplacing Laurentia, which was thought to be eclogite free, over top. The general P-T pattern of eclogite equilibration, with T and P increasing westward across the WGR, supported this view (Krogh, 1977). The discovery of widespread Caledonian HP and UHP metamorphism in the Laurentian continental margin of Greenland requires a rethink of the collision geometry for the Caledonides.

A revised schematic model (Fig. 3B) for Caledonian collision incorporates recent work on the geology of East Greenland. West-directed displacement of basement and cover thrust sheets derived from the Laurentian margin is now well documented (Higgins and Leslie, 2000; Higgins et al., 2004), resulting in a doubly vergent collision geometry. Local suites of 466–435 Ma I-type granites are known south of 71°N, in Liverpool Land and eastern Milne Land (F. Kalsbeek, pers. commun., 2006); however, the vast majority of the Caledonian granitoids in East Greenland are crustally derived, S-type leucogranites (Kalsbeek et al., 2001a, 2001b) that range in age from 435 to 415 Ma (Gilotti and McClelland, 2005 and references therein). The leucogranites, previously cited as evidence for the polarity of subduction (Hodges et al., 1982; Hossack and Cooper, 1986), can form in either the overriding or subducting continental crust. The small volume of I-type granites in Greenland may point to limited magmatism resulting from subduction of oceanic lithosphere. Given that I-type granites are not known in the continental margin of Baltica proper, this is one line of evidence for subduction polarity. Continental collision is not expected to produce a significant magmatic arc (Ernst, 2005), so the scarcity of evidence for a large arc in either continent is not surprising.

The presence of UHP eclogites in the continental crust of both Baltica and Laurentia presents a bigger problem, because in a collision as large as the Caledonides, it is expected (by analogy with the Alps and Himalayas) that the polarity of continental subduction will be constant along the length of the orogen. Ryan (2001) raised the possibility of a flip in subduction across a hypothetical transform fault in the middle of the Caledonian orogen, but discarded the notion as unlikely. The revised model (Fig. 3B)

retains Laurentia as the overriding plate based on the paucity of mantle rocks in the NEGP (Gilotti, 1994) compared to the abundance of mantle-type garnet peridotites in the WGR of Norway (e.g., Krogh and Carswell, 1995). Garnet peridotites are incorporated into continental eclogite terranes primarily by interaction with the overlying mantle wedge (e.g., Brueckner and Medaris, 2000) during continental subduction. The precise mechanism is unknown, but may involve sinking of the more dense mantle material into the crust followed by exhumation and tectonic emplacement (Brueckner, 1998). Figure 3B shows a tail of Laurentian continental crust dragged into the subduction zone; other possibilities for forming UHP rocks in the upper plate of the Caledonian collision are described later in the paper.

Atypical Characteristics of the North-East Greenland UHP Terrane

The UHP rocks on Rabbit Ears Island in North-East Greenland display a number of anomalous characteristics when compared to other UHP terranes that fit the continental subduction paradigm (Liou et al., 2004; Ernst, 2005, 2006). As discussed above, UHP metamorphism in Greenland took place in the overriding plate of the Caledonian orogen, whereas UHP metamorphism formed in the WGR as part of the downgoing slab. The atypical characteristics of UHP metamorphism in Greenland include the late timing and high T of metamorphism, followed by slow exhumation. Each point is briefly discussed below, and an explanation is presented in the form of a tectonic model.

UHP metamorphism late in collision

UHP metamorphism is predicted to form early in continent-continent collision because the leading edge of the subducted continent should be the first crust to reach mantle depths. In the Himalayan example, collision began at ~55 Ma (Rowley, 1996), while UHP metamorphism, best dated by U-Pb SHRIMP analysis of coesite-bearing zircon domains, took place at 46 Ma (Kaneko et al., 2003). Continent-continent collision between Laurentia and Baltica, known as the Scandian event, is generally thought to span the Late Silurian to Early Devonian (Gee, 1975). The UHP eclogites in the WGR formed at the end of the Scandian, at around 400 Ma (Carswell et al., 2003; Tucker et al., 2004), which is somewhat late if the WGR was the leading edge of

Baltica continental crust. In sharp contrast to this, UHP metamorphism in North-East Greenland at 365–350 Ma is much later in the Caledonian collision (Gilotti et al., 2004; McClelland et al., 2006). In addition, zircons from the Greenland UHP rocks record continuous growth and recrystallization over a 15 million year time span that points to a long residence time at mantle depths (McClelland et al., 2006). Both these features differ significantly from the early and rapid subduction seen in other UHP terranes (e.g., Rubatto and Hermann, 2001; Kaneko et al., 2003).

High temperature

The continental subduction paradigm predicts that UHP rocks will have relatively low temperatures because isotherms are depressed in subduction zones. All but a few UHP terranes form at peak temperatures < 800°C, and thus fit this scenario (Ernst, 2006; Ernst et al., in press). For example, the Himalayan Kaghan Valley coesite eclogite experienced peak T ~690–750°C (O'Brien et al., 2001), whereas peak UHP temperatures in the Norwegian WGR are < 820°C (Terry et al., 2000b). In contrast, T calculated from kyanite eclogites in the North-East Greenland UHP rocks exceeds 950°C (Gilotti and Ravn, 2002). P-T estimates show that other high-temperature outliers are the diamond-bearing Erzgebirge unit in the Bohemian Massif (Massonne and O'Brien, 2003) and the Kokchetav Massif in Kazakhstan (Shatsky et al., 1995). The Greenland, Erzgebirge, and Kokchetav localities all show evidence of partial melting at UHP conditions (Massone, 2003; Korsakov et al., 2004; Korsakov and Hermann, 2006; Lang and Gilotti, 2007), which is another indication of quite high temperatures. These temperatures could have been achieved in a warm to hot slab subduction zone, but such environments are thought to be restricted to young lithosphere (Peacock, 1990).

Slow exhumation

Exhumation rates for UHP terranes that have undergone continental subduction tend to be in the same range as subduction rates, or faster (Liou et al., 2004; Ernst, 2006). This is because exhumation is most likely to occur in a subduction-channel environment and be buoyancy driven. Average rates calculated for the vertical ascent of rocks from UHP depths to the middle crust (10–15 km) compiled by Ernst (2006) range from 6–45 mm/year, with the Himalayan eclogites at the fast end. These rates do

not take into account that eclogite exhumation is commonly a differential process, with fast exhumation to the lower crust and decreasing rates to the middle crust (Maruyama et al., 1996; Rubatto and Hermann, 2001; Walsh and Hacker, 2004).

Exhumation of the North-East Greenland UHP rocks appears to have been slower than observed in the normal subduction-channel setting. A rough estimate of the exhumation rate for the Rabbit Ears Island UHP eclogites can be made using the Sm-Nd method, which gave a mineral isochron age of 342 ± 6 Ma from kyanite-eclogite sample 434441 (Gilotti et al., 2004). Sample 03-110 from the same pod (Fig. 2C) contained zircons with coesite-bearing CL domains that gave a U-Pb SHRIMP age of 350 ± 4 Ma (McClelland et al., 2006). A Sm-Nd closure temperature between 600 and 700°C (Mezger et al., 1992) puts the Greenland UHP rocks in the lower crust by 342 Ma (Fig. 4). Peak P at 3.6 GPa (Gilotti and Ravn, 2002) corresponds to a depth of approximately 120 km, whereas estimated pressures for the Sm-Nd age from a probable P-T path are 0.8–1.0 GPa, corresponding to a depth of approximately 30–35 km. The average vertical rate of ascent is then ~11 mm/year, or ~4 mm/year if we use the oldest obtained UHP age of 365 Ma. These rates are on the slow side of the exhumation rates compiled by Ernst (2006).

New Ages for Exhumation from Hornblende-Bearing and Boudin Neck Pegmatites

In order to obtain a better estimate of exhumation rates for the UHP rocks in North-East Greenland, we dated two pegmatites associated with eclogite boudins and a foliated hornblende-bearing leucocratic pod within the quartzofeldspathic host gneisses (Fig. 2). Boudinage is thought to occur during amphibolite-facies deformation of eclogitic bodies because their margins commonly show strong deformation fabrics composed of retrograde minerals, such as plagioclase and hornblende, compared to the better preserved centers of the eclogites (e.g., Sartini-Rideout et al., 2006). The detailed map of Figure 2C shows a number of boudinaged eclogite bodies with pegmatites in their necks. Sample 03-113 is a garnet-bearing pegmatite from a well-developed boudin neck (Fig. 5A). The largest block in Figure 2C has a pinch-and swell geometry with pegmatites filling fractures that eventually localize the boudin necks. Pegmatite sample 03-111 is from the largest

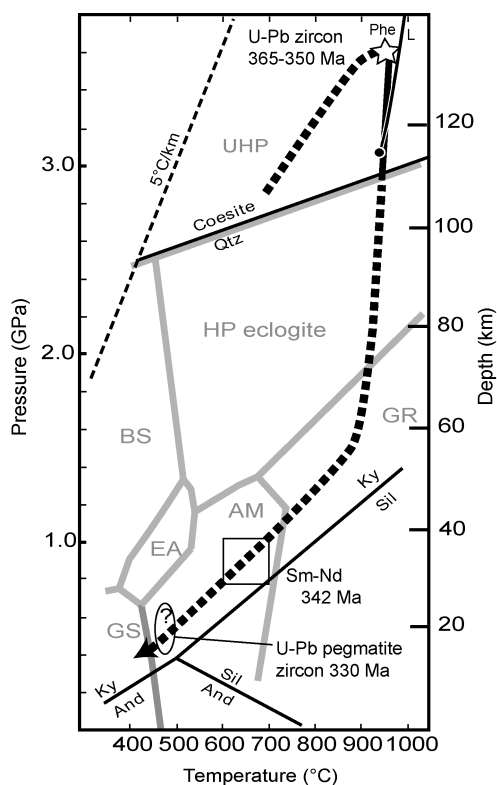


FIG. 4. Part of the pressure-temperature-time (P-T-t) path for the North-East Greenland UHP terrane. Peak UHP conditions at the star are from kyanite eclogites (Gilotti and Ravna, 2002). The dashed line is estimated; the solid path (Lang and Gilotti, 2007) is based on crossing the phengite (phe) melting reaction in nearby metapelites. The phengite melting reaction, phengite + coesite + clinopyroxene = garnet + kyanite + K-feldspar + melt, is from Hermann and Green (2001). The star marks the age of UHP metamorphism (McClelland et al., 2006); the Sm-Nd age (Gilotti et al., 2004) is represented by the box and the pegmatite age is given in this paper. Facies grid from Oh and Liou (1998).

fracture in this large eclogite block (Fig. 5B). Sample 03-201 was collected from a coarse-grained hornblende-garnet-biotite-quartz-feldspar leucosome that cuts the penetrative foliation defined by alignment of hornblende, quartz, and feldspar in the adjacent gneiss but is itself moderately foliated (Fig. 5C).

Analytical methods

Zircon grains were separated by standard gravimetric and magnetic techniques; hand-picked under alcohol on the basis of clarity, lack of inclusions, and integrity; mounted in 2.54 cm epoxy



FIG. 5. Outcrop photographs of dated pegmatite samples, circled. A. UHP eclogite boudin with isoclinally folded layers of kyanite eclogite, quartz eclogite, and bimineralic eclogite within quartzofeldspathic gneisses. Sample 03-113 from boudin neck pegmatite. Hammer (45 cm long) for scale. B. Pegmatite sample 03-111 from partially developed boudin neck in large eclogite block. C. Coarse-grained, granitic leucosome pod (sample 03-201) in quartzofeldspathic gneiss is weakly foliated. Large, black, euhedral crystals are hornblende.

rounds; and polished to expose grain interiors. Zircon U-Pb analyses were completed using the SHRIMP-RG (reverse geometry) at the United States Geological Survey–Stanford University Ion Probe Laboratory, Stanford, CA. Reflected/transmitted light and CL images were used to characterize the zircons and select spot locations (Fig. 6). The analytical routine followed Williams (1998) and Barth et al. (2001). Calibration concentrations of U and Th were based on analyses of zircon standard CZ3 (U = 550 ppm). Isotopic compositions were calibrated by replicate analyses of zircon standard R33 (419 Ma; Black et al., 2004). Calibration errors for the $^{206}\text{Pb}/^{238}\text{U}$ ratios of R33 for the three different analytical sessions were 0.45%, 0.51%, and 0.54% (2σ). Data reduction and plotting utilized the programs of Ludwig (2001, 2003). The U-Pb data for each sample are plotted on Tera-Wasserburg diagrams (Fig. 7) and presented in Table 1. Uncertainties in the isotopic ratios are reported at the 1σ level. Ages were assigned based on the weighted mean of observed $^{206}\text{Pb}/^{238}\text{U}$ ages corrected using the ^{207}Pb correction method. Common Pb compositions were estimated from Stacey and Kramers (1975). Weighted mean age uncertainties are reported at the 95% confidence level.

U-Pb results

Zircons from the boudin neck pegmatite within the eclogite pod (sample 03-111) are elongate euhedral to subhedral grains with low-U (57–84 ppm) and higher-Th/U (0.08–0.17) cores overgrown by oscillatory zoned, higher U (106–640 ppm) and lower Th/U (0.003–0.17) rims (Fig. 6A). Core analyses give an age of 347 ± 5 Ma (MSWD = 1.1; $n = 8$), whereas with the exception of one analysis that is interpreted to be too young due to Pb-loss, rim analyses give an age of 329 ± 3 Ma (MSWD = 1.4; $n = 15$).

Zircons from the granitic pegmatite in the boudin neck (sample 03-113) are elongate euhedral to subhedral grains with cores mantled by high-U (1012–3450 ppm) zircon and overgrown by thin (<30 μm), lower U (111–675 ppm) rims (Fig. 6B). Discordant analyses from the cores have ^{204}Pb -corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ages as old as ~1685 Ma. The cores are interpreted as inherited xenocrystic components in the sample. The high-U mantles and lower-U rims yield ages ranging from 336 to 365 Ma and 301 to 337 Ma, respectively. The five rim analyses that most clearly lie entirely within the low-U rim domain in CL images define a weighted mean

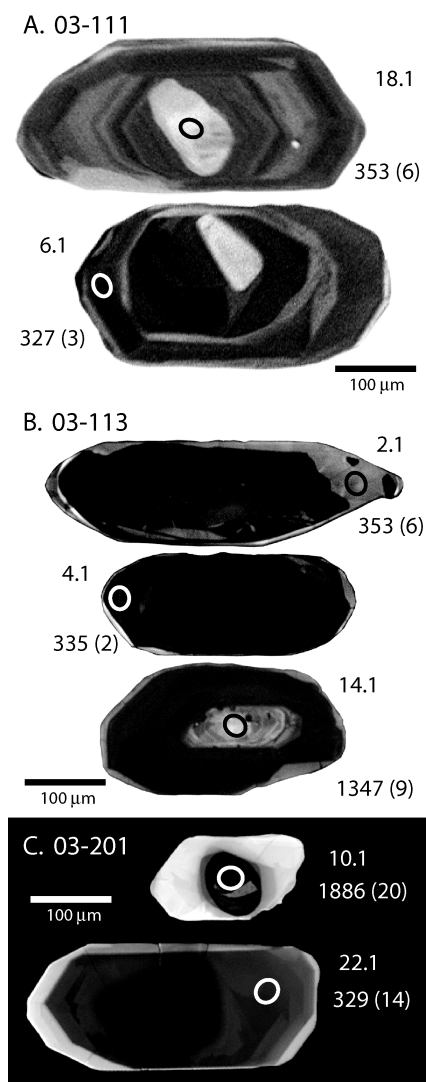


FIG. 6. Representative cathodoluminescence images of complex zircons from pegmatites (A) 03-111, (B) 03-113, and (C) 03-201. Ellipses indicate individual analysis spots for the sensitive high-resolution ion microprobe–reverse geometry (SHRIMP-RG). Each spot is labeled by grain number.analysis number (e.g., 4.1) and the corresponding U-Pb age ($\pm 1\sigma$ Ma).

$^{206}\text{Pb}/^{238}\text{U}$ age of ~330 Ma. Analyses from the high-U domain cluster around 347 Ma, with eight analyses yielding a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 347 ± 2 Ma (MSWD = 1.8). The scatter in ages observed in this sample is inferred to record a combination of mixture with older core material, long-lived rim growth, and Pb loss.

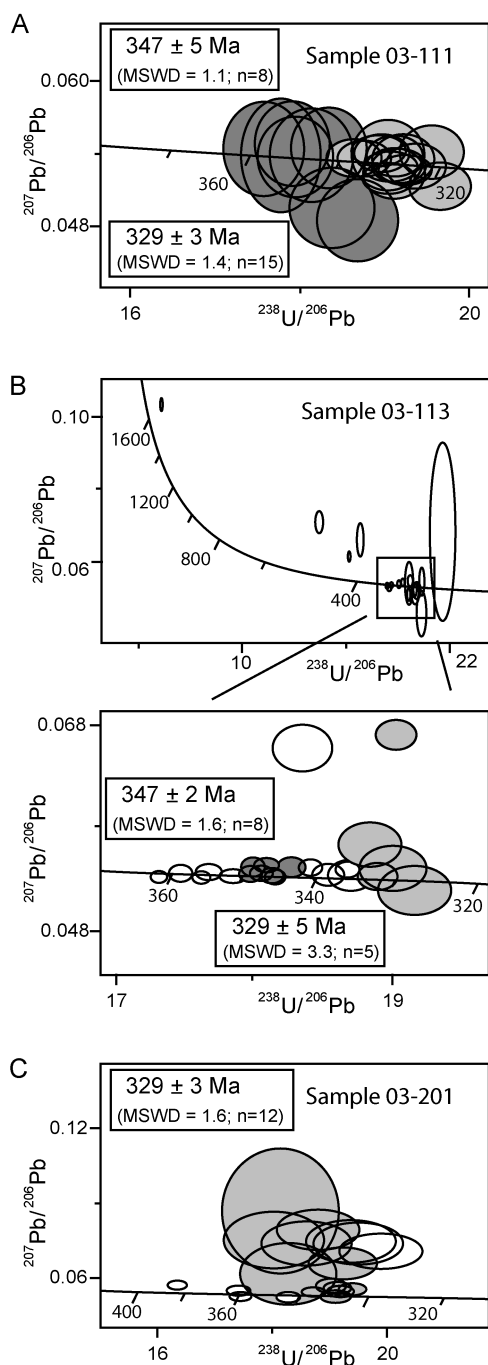


FIG. 7. Tera-Wasserburg plots of SHRIMP-RG U-Pb data of zircon for pegmatite samples (A) 03-111, (B) 03-113, and (C) 03-201. The data are presented as 1σ error ellipses uncorrected for common Pb. MSWD = mean square of weighted deviates. Ellipses of similar greytone shading correspond to analyses used in age calculation (Table 1).

The leucosome (sample 03-201) yielded euhedral zircons characterized by inherited xenocrystic cores overgrown by moderate-U (193–357 ppm) oscillatory zoned zircon, lower-U (13–42 ppm) inner rims and extremely low U (0.2–2 ppm) outer rims (Fig. 6C). The observed $^{206}\text{Pb}/^{238}\text{U}$ ages range from 381 to 309 Ma, excluding analyses of two xenocrystic cores and the extremely low U rims. Zircons in this age range have low Th concentrations (0.003–25 ppm) and very low Th/U (0.002–0.07) as well. The four oldest and three youngest ages of this group are interpreted to have incorporated inherited components and experienced Pb loss, respectively. The remaining 12 analyses give a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 329 ± 3 Ma (MSWD = 1.6), which is taken as the crystallization age for the leucosome. The extremely low U rims likely record a subsequent phase of growth or recrystallization.

Interpretation

A common assumption in high-grade terranes is that boudin neck pegmatites are intruded in the amphibolite facies, and represent exhumation to mid-crustal levels (e.g., Krogh et al., 2003). However, two periods of zircon growth—at ~347 and 329 Ma—are recorded in the pegmatite samples from Rabbit Ears Island, and no direct observations tie the pegmatite ages to points on the P-T path. A U/Pb age obtained from the coesite-bearing domain of zircon extracted from the kyanite eclogite at this locality (sample 03-110, Fig. 2C) is 350 ± 4 Ma (McClelland et al., 2006). The coesite domains are overgrown by rims that yield an age of 342 ± 3 Ma, which matches the Sm-Nd age of 342 ± 2 Ma. The two periods of pegmatite zircon crystallization bracket the observed Sm-Nd age and rim growth in the eclogite pod. We suggest that the ~347 Ma zircon growth in the pegmatite records the initial formation of decompression melts via phengite melting while still at the UHP conditions (Fig. 4). Lang and Gilotti (2007) have observed phengite melting reactions in polyphase “melt” inclusions preserved in garnets from nearby metapelites (Fig. 2B). Phengite was also an important UHP phase in the quartzofeldspathic host gneisses, and could have provided the source for the boudin neck pegmatites. The hornblende-bearing leucosome appears to record melting at a late stage relative to deformation in the adjacent host gneiss. The foliation and lineation in the gneiss are defined by hornblende, suggesting deformation under amphibolite-facies conditions. Growth of zircon rims in the boudin neck pegmatites

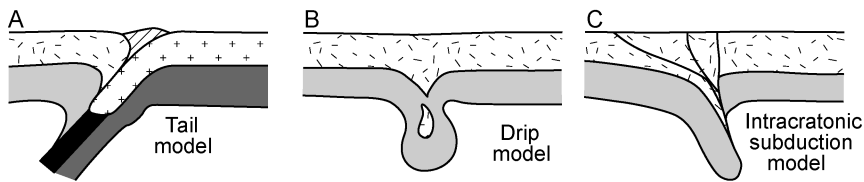


FIG. 8. Diagrammatic sketch of possible mechanisms for forming UHP metamorphism in the overriding plate of a continent-continent collision. A. A tail in the main subduction zone. B. A mantle drip. C. Intracratonic subduction.

and crystallization of the leucosome at 329 Ma are interpreted to have occurred at amphibolite-facies conditions in the middle crust.

Using these new ages to calculate exhumation rates reinforces the original inference that exhumation was relatively slow in North-East Greenland. The average rate of vertical ascent from 120 km at 365–350 Ma to the middle crust (20 km) at 329 Ma was 3–5 mm/year. In reality, the exhumation of the UHP rocks must have been at least a two-stage process, with exhumation first to the lower crust (35–40 km) at 4–11 mm/year, followed by a slower phase to approximately 20 km at 329 Ma at rates of ~1 mm/year.

Mechanisms for Forming UHP Metamorphism in the Overriding Plate

UHP metamorphism in the overriding plate of a collisional orogen cannot be explained by the continental subduction paradigm (Ernst, 2006). Nor can it be explained by pure crustal thickening mechanisms, such as imbrication or pure shear. The thickest crust found on Earth is approximately 90 km under the highest part of the Tibetan Plateau in western Qiangtang (Wittlinger et al., 2004), in the overriding plate of the Himalayan collision. Finite element models (Ryan, 2001), with all the parameters maximized to produce thick crust with eclogites at the base, yielded crustal thicknesses on the order of 90 km, which means pressures below the coesite stability field. We suggest three possible mechanisms for producing UHP metamorphism in crustal rocks in the overriding plate (Fig. 8): (1) a tail of material from the overriding plate pulled down the subduction channel; (2) a mantle drip; and (3) intracratonic subduction.

The fate of the leading edge of the overriding continent in collisions is not well known, but one can envision a geometry where the continental margin becomes highly ductile and gets pulled down

the subduction zone as a long tail (Fig. 8A). A geometric requirement of this model is that the UHP terrane be physically located at the leading edge of the overriding plate. A second test is that UHP metamorphism on both margins should be roughly the same age. Roberts (2003) draws such a model for the Caledonian collision at 420–400 Ma; however, neither of these criteria are met in the Greenland Caledonides.

Mantle drips (Fig. 8B), formed by Rayleigh–Taylor instabilities in the subcrustal lithosphere (e.g., Houseman and Molnar, 1997), could potentially pull weak, hot crust into the mantle where it would experience UHP metamorphism. Mantle drips have been numerically modeled for a number of tectonic settings, including intracratonic environments (Neil and Houseman, 1999; Pysklywec and Beaumont, 2004), which may share some similarities with the North-East Greenland UHP terrane. A modern mantle drip beneath the southern Sierra Nevada batholith of California has recently been imaged with seismic tomography (Jones et al., 1994; Zandt, 2003; Boyd et al., 2004; Zandt et al., 2004). Mantle delamination and downwelling have been proposed as the mechanism to form the hot UHP rocks in the Erzgebirge massif late in the Variscan collision (Massonne, 2005). One problem with this model is how to release the crust from a mantle drip once the dense quartzofeldspathic material (Massonne et al., 2007) starts to sink into the surrounding mantle. Massonne (2005) solved the problem by entraining the UHP crust in granitic melts that bubble back up into the crust. Gravitational instability models are favored by weak lower crust and mantle—i.e. a *crème brûlée* rheology—which may not apply to the crust and lithosphere of an overriding plate (Burov and Watts, 2006).

Our preferred mechanism for creating UHP metamorphism in the overriding plate of a continent-continent collision is intracratonic subduction (Fig. 8C). Early recognition that steps in the Moho

TABLE 1. U-Pb SHRIMP Geochronologic Data and Apparent Ages¹

Spot ²		U, ³ ppm	Th, ppm	Th/U	206Pb/ ³ Pb, ppm		238U/ ²⁰⁶ Pb ⁴	207Pb/ ²⁰⁶ Pb ⁴	206Pb/ ²³⁸ U ⁵ , Ma	207Pb/ ²⁰⁶ Pb ⁵ , Ma
					f ²⁰⁶ Pb ³ _c					
Sample 03-111 (UTM27x WF 0548080; 8659958)										
1.1	r	220	1	0.003	9.7	0.2	19.56 (1.2)	0.0541 (2.7)	**321 (4)	
2.1	r	475	11	0.02	21.3	0.0	19.16 (1.0)	0.0529 (1.8)	**328 (3)	
3.1	r	488	7	0.01	22.3	0.1	18.76 (1.1)	0.0536 (1.8)	**335 (3)	
4.1	r	621	15	0.02	28.0	–	19.04 (1.0)	0.0519 (1.6)	**330 (3)	
5.1	r	337	4	0.01	15.1	0.1	19.12 (1.1)	0.0538 (2.1)	**328 (4)	
6.1	r	640	8	0.01	28.6	0.0	19.24 (1.0)	0.0526 (1.5)	**327 (3)	
7.1	r	351	7	0.02	14.5	3.6	20.75 (1.2)	0.0809 (1.9)	293 (3)	
8.1	r	455	4	0.01	20.6	0.1	18.97 (1.1)	0.0540 (1.9)	**331 (3)	
9.1	r	288	2	0.01	12.8	–	19.37 (1.1)	0.0529 (2.3)	**325 (4)	
10.1	r	219	1	0.003	9.9	–	19.11 (1.2)	0.0523 (2.7)	329 (4)	
11.1	r	623	18	0.03	28.3	0.1	18.93 (1.0)	0.0542 (1.5)	**331 (3)	
12.1	r	410	7	0.02	18.9	0.1	18.64 (1.1)	0.0536 (1.9)	**337 (4)	
13.1	r	224	1	0.01	10.0	0.1	19.22 (1.2)	0.0535 (2.6)	**327 (4)	
14.1	r	238	1	0.003	10.4	–	19.66 (1.2)	0.0512 (2.6)	**320 (4)	
15.1	r	625	11	0.02	28.1	–	19.07 (1.0)	0.0528 (1.6)	**330 (3)	
16.1	c	59	6	0.11	2.9	0.1	17.57 (1.8)	0.0544 (4.8)	*357 (6)	
17.1	c	79	6	0.08	3.8	0.03	17.94 (1.6)	0.0537 (4.2)	*350 (6)	
18.1	c	64	9	0.14	3.1	0.03	17.78 (1.7)	0.0538 (5.8)	*353 (6)	
19.1	r	106	17	0.17	4.8	0.1	19.04 (1.4)	0.0539 (3.6)	**330 (5)	
20.1	c	84	14	0.17	3.9	0.1	18.35 (1.6)	0.0545 (4.0)	*342 (5)	
21.1	c	57	5	0.09	2.7	0.1	18.14 (1.8)	0.0538 (4.8)	*346 (6)	
22.1	c	73	6	0.09	3.5	0.2	17.91 (1.6)	0.0548 (4.2)	*350 (6)	
23.1	c	75	11	0.16	3.5	–	18.38 (1.8)	0.0494 (4.5)	*343 (6)	
24.1	c	71	6	0.09	3.3	–	18.68 (1.7)	0.0484 (4.6)	*338 (6)	
Sample 03-113, boudin neck pegmatite (UTM 27x WF 0548046; 8659980)										
1.1	c	148	40	0.28	10	2.3	13.28 (1.0)	0.0748 (2.5)	458 (5)	
2.1	lr	206	3	0.01	9	–	19.16 (0.9)	0.0522 (3.0)	**328 (3)	
3.1	dr	1805	77	0.04	86	0.1	17.98 (0.3)	0.0544 (1.1)	*349 (1)	
4.1	lr	529	23	0.05	24	0.1	18.69 (0.6)	0.0536 (1.9)	336 (2)	

5.1	lr	636	28	0.05	29	—	18.88	(0.5)	0.0535	(1.6)	**332	(2)
6.1	c	452	76	0.17	25	2.8	15.54	(0.7)	0.0769	(1.6)	391	(3)
7.1	lr	294	5	0.02	14	1.6	18.34	(0.8)	0.0659	(2.3)	337	(3)
8.1	c	2410	89	0.04	137	1.0	15.12	(0.3)	0.0629	(1.1)	409	(1)
9.1	dr	1670	130	0.08	82	—	17.45	(0.3)	0.0538	(1.0)	359	(1)
10.1	lr	1012	28	0.03	47	—	18.53	(0.4)	0.0536	(1.3)	339	(1)
11.1	dr	3450	83	0.02	171	—	17.29	(0.2)	0.0534	(0.7)	363	(1)
12.1	dr	3292	141	0.04	161	—	17.60	(0.2)	0.0534	(0.7)	356	(1)
13.1	lr	250	12	0.05	11	0.4	18.83	(0.8)	0.0566	(2.5)	**332	(3)
14.1	c	217	110	0.52	44	2.1	4.21	(0.6)	0.1040	(1.0)	1349	(9)
15.1	dr	1199	9	0.01	56	0.1	18.26	(0.4)	0.0544	(1.2)	*343	(1)
16.1	lr	675	54	0.08	30	1.8	19.02	(0.5)	0.0673	(1.5)	**325	(2)
17.1	lr	111	6	0.05	5	7.1	19.47	(1.3)	0.1090	(3.1)	301	(4)
18.1	lr	241	5	0.02	11	0.2	19.00	(0.9)	0.0543	(2.7)	**330	(3)
19.1	dr	2335	88	0.04	110	—	18.17	(0.3)	0.0534	(0.8)	*345	(1)
20.1	dr	2027	35	0.02	96	—	18.15	(0.3)	0.0535	(0.9)	*346	(1)
21.1	dr	1802	31	0.02	84	0.1	18.42	(0.3)	0.0543	(1.0)	340	(1)
22.1	dr	1785	91	0.05	87	—	17.53	(0.4)	0.0539	(1.0)	357	(1)
23.1	c	5166	667	0.13	1160	1.3	3.83	(0.2)	0.1036	(0.3)	1480	(2)
24.1	dr	2175	85	0.04	103	0.1	18.10	(0.3)	0.0546	(0.9)	*346	(1)
25.1	dr	1769	84	0.05	84	0.04	18.07	(0.3)	0.0538	(1.0)	*347	(1)
26.1	dr	1682	34	0.02	77	0.1	18.69	(0.3)	0.0542	(1.0)	336	(1)
27.1	dr	1819	87	0.05	87	0.03	17.99	(0.3)	0.0537	(1.0)	*349	(1)
28.1	c	2289	107	0.05	111	—	17.77	(0.5)	0.0535	(0.8)	353	(2)
Sample 03-201, leucosome (UTM 27x WF 0548525; 8658834)												
1.1	dc	206	8	0.04	9	0.5	19.03	(0.9)	0.0573	(3.2)	*329	(3)
2.1	lr	28	0.11	0.004	1.2	2.3	19.90	(2.4)	0.0709	(6.6)	309	(8)
3.1	dc	347	25	0.07	16	0.2	18.71	(0.7)	0.0548	(2.1)	*335	(2)
4.1	dc	298	20	0.07	15	0.2	17.40	(0.8)	0.0554	(2.2)	359	(3)
5.1	dc	231	1.5	0.007	10	0.4	19.14	(0.8)	0.0564	(2.6)	*327	(3)
6.1	lr	24	0.02	0.001	1.0	2.7	19.38	(2.7)	0.0747	(7.4)	316	(9)
7.1	lr	24	0.04	0.002	1.1	3.3	18.80	(2.5)	0.0793	(6.8)	*323	(8)
8.1	dc	314	17	0.05	15	—	17.43	(0.7)	0.0530	(2.2)	360	(3)

Table continues

[illegible]

²Abbreviations: c = core; dc = relatively high U core; r = rim; dr = relatively high U rim; lr = relatively low U rim; wr = very low U rim.

⁴Reported ratios are not corrected for common Pb. Errors are reported in parentheses as percent at the 1 σ level.

⁵Ages calculated from ratios corrected for common Pb using $^{207}\text{Pb}/^{238}\text{U}$ ages and ^{204}Pb for the $^{207}\text{Pb}/^{236}\text{Pb}$ ages. Uncertainties in millions of years reported as 1σ . Annotation of ages denotes ages used in calculation of weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages.

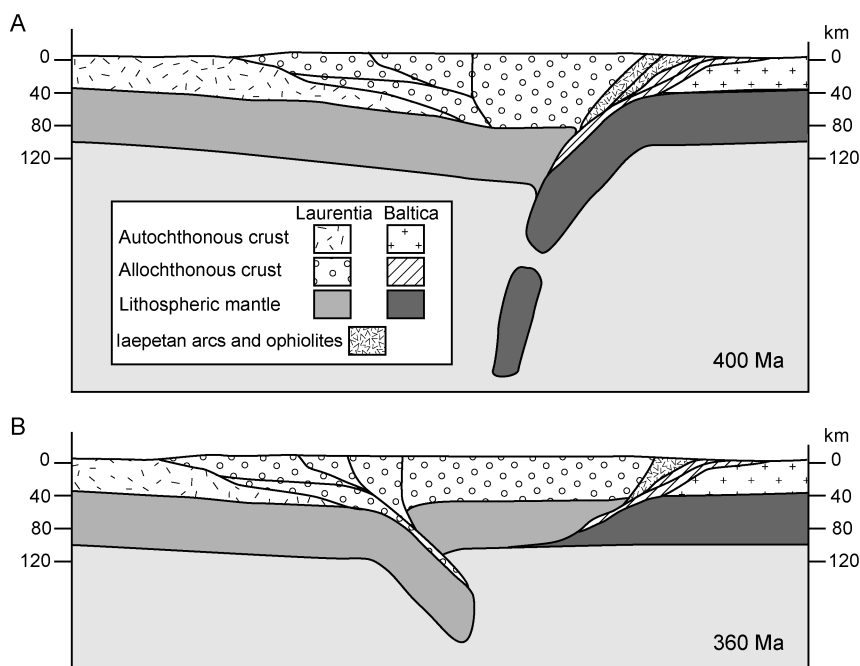


FIG. 9. Profiles of Laurentia–Baltica collision at the plate scale. A. At 400 Ma. B. At 360 Ma.

beneath Tibet roughly correlate with strike-slip faults at the surface (Hirn et al., 1984) led to the idea of oblique subduction zones within Asian continental crust (Tapponnier et al., 2001). Evidence for intracratonic subduction in Tibet is strongest in the Tien Shan region, and is based on teleseismic imaging (Poupinet et al., 2002; Shi et al., 2004; Wittlinger et al., 2004) and magnetotelluric data (Bielinski et al., 2003). Steep velocity anomalies beneath the Altyn Tagh fault are some of the best-imaged Moho steps, and southward-dipping mantle slabs are imaged down to 300 km (Wittlinger et al., 2004). It has been proposed that a number of the large strike-slip faults in Tibet, such as the Altyn Tagh, are steep at the surface but dip southward at depth and cut the entire lithosphere (Tapponnier et al., 2001; Wittlinger et al., 2004). Crust in the footwall of such structures could experience metamorphism at UHP conditions.

Tectonic Models for UHP Metamorphism in the Caledonides

Figure 9 presents two schematic models for the Caledonian collision between Laurentia and Baltica, one at 400 Ma and a second at 360 Ma, that

focus on explaining UHP metamorphism in the continental margins. Important, earlier HP and UHP events involving arcs, ophiolites, and microcontinents took place in subduction-zone environments outboard of Baltica (Brueckner and van Roermund, 2004). The onset of continent-continent collision however is generally taken to be the start of the Scandian event in the Silurian (Gee, 1975). In reality, the initial time of continental collision is difficult to assess. Leucogranites formed from partial melting of Mesoproterozoic metasedimentary rocks on the Laurentian margin by 435 Ma (Kalsbeek et al., 2001a, 2001b) are one indication that the collision was under way by then. In addition, the carbonate platform of North Greenland was drowned by latest Llandovery or Early Wenlock turbidite sedimentation (Hurst et al., 1983). On the Baltica margin, HP metamorphism of the WGR apparently began as early as 420 Ma (Griffin and Brueckner, 1980, 1985) and can be explained by either crustal thickening or continental subduction.

Figure 9A shows the culmination of the Scandian event at 400 Ma. This is the time favored for UHP metamorphism in the WGR (Carswell et al., 2003), and thus we show a portion of allochthonous Baltica subducted to mantle depths. This is rather late for

UHP metamorphism to form at the leading edge of the downgoing plate, suggesting that Baltica may not strictly conform to the simple continental subduction paradigm (Brueckner and van Roermund, 2004; Hacker and Gans, 2005). HP eclogites of this age (410–390 Ma) are widespread in the Greenland Caledonides. We envision them forming in overthickened crust of Laurentia by imbrication and pure shear. Both HP and UHP eclogites in the WGR are exhumed to the middle crust by 395 Ma (Tucker et al., 2004; Hacker and Gans, 2005), perhaps by a slab-breakoff event. Continental lithosphere across the WGR is assumed to return to near normal thickness following initial exhumation of the Baltica margin.

Figure 9B depicts the Caledonian orogen at 360 Ma, the well-documented time of UHP metamorphism in North-East Greenland (Gilotti et al., 2004; McClelland et al., 2006), approximately 60–70 million years after the initiation of collision and 40 million years after the culmination of UHP metamorphism in the downgoing plate. Thus, Caledonian plate convergence apparently continued or was renewed long after the official end of the Scandian event. The time frame fits with the now 55 million year old Himalayan collision; the geometry of intracratonic subduction is inferred to be similar to that suggested for present-day Tibet (Tapponnier et al., 2001). In Figure 9B, we show a sliver of Laurentia in the footwall of an intracratonic subduction zone undergoing UHP metamorphism. We prefer this model over the others discussed above (Fig. 8) for a number of reasons. The North-East Greenland UHP terrane lies at least 500 km west of the exotic nappes in Scandinavia (Fig. 1), some with Laurentian faunas (Tucker et al., 2004), and thus developed well within the overriding plate and well away from the leading edge of Laurentia—in marked contrast to the tail model. Differing ages of UHP metamorphism on both continental margins are inconsistent with the tail model as well. Residence time at the base of the overthickened crust can explain the elevated T of UHP metamorphism in Greenland. Finally, the two large strike-slip faults observed in the NEGEP (Fig. 2) are obvious candidates for zones of oblique convergence similar to those in Tibet, and others could be hidden offshore in the rifted continental margin. These structures are known to have been active during and after the time of UHP metamorphism (Hallett, 2005; Sartini-Rideout et al., 2006) and thus may be linked to the formation and exhumation of the UHP terrane.

Rates for UHP rocks exhumed out of an overriding plate environment may be relatively slow due to the lack of a fast conveyor belt such as a subduction channel. Nevertheless, we speculate that exhumation from an intracratonic subduction zone could be accomplished by return flow of buoyant crust, perhaps triggered by a change in plate motion. The strike-slip faults themselves may permit a component of vertical extrusion (Thompson et al., 1997), whereas the large horizontal component of displacement may be responsible for the slower exhumation rates calculated for the Greenland UHP rocks.

Conclusions

The UHP metamorphic terrane of North-East Greenland formed in the overriding plate of the Caledonian orogen and has several atypical characteristics that distinguish it from most other UHP terranes, which clearly formed in the downgoing continental plates of collisional orogens. The three important differences are higher peak T associated with peak P, metamorphism later in the collisional history, and slower exhumation rates than observed in the lower plate setting. These variations are inferred to be unique to the overriding plate in a collisional orogen, and thus may be useful in establishing the setting of other anomalous UHP terranes. By analogy with the modern Himalayan orogen, an intracratonic subduction setting is inferred for the North-East Greenland UHP metamorphic province.

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