Paradigm lost: Buoyancy thwarted by the strength of the Western Gneiss Region (ultra)high-pressure terrane, Norway

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ABSTRACT

Subduction and exhumation of ultrahigh-pressure (UHP) metamorphic terranes are typically envisaged as short-lived processes associated with the transition from oceanic subduction to continent-continent collision. Norway’s Western Gneiss Region, by comparison, is a giant, late-orogenic UHP terrane that underwent protracted residence at UHP conditions during the Scandian phase of the Caledonian orogeny followed by relatively slow exhumation. Here, we use two-dimensional numerical thermal-mechanical models to explore the tectonics of orogens of this type and the associated controls on the size of their UHP terranes and the duration of UHP metamorphism.

The models have four tectonic phases designed to capture the main stages of the Caledonian evolution: oceanic subduction and micro-continent accretion; continental margin subduction; plate quiescence; and postorogenic extension during plate divergence. Contrasting styles of exhumation are explored by varying the strength of the margin crust and investigating melt-induced weakening. The tectonic and metamorphic evolution of the Western Gneiss Region is consistent with a model in which continental margin crust was subducted beneath a thick orogenic wedge where it underwent metamorphism at (U)HP conditions for at least 15 million years (Myr) as subduction ended. The buoyant Baltic crust must have been especially strong in order to have stayed coupled to the underlying lithosphere during this phase, perhaps reflecting its refractory composition and/or a lack of fluids. Subsequent exhumation of the Western Gneiss Region can be explained by orogen-scale extension resulting from minor (~100 km) plate divergence, with removal of the orogenic wedge by combined top-to-the-hinterland transport, normal faulting, and erosion. We conclude that large, long-duration UHP terranes are fundamentally different from transient smaller ones. The latter are often explained by the paradigm of buoyant exhumation. This paradigm is incomplete, but both types can be explained by control of the system by the exhumation number (ratio of buoyancy force to basal traction). By implication, the existence of this type of large UHP terrane is a consequence of the high strength of the subducted crust.

INTRODUCTION

The Western Gneiss Region of southwestern Norway records widespread ultrahigh- and high-pressure metamorphism (UHP and HP) generally interpreted to have formed by continental subduction during the Caledonian orogeny (Lappin and Smith, 1978; Smith, 1984; Wain, 1997; Cuthbert et al., 2000; Root et al., 2005). Although it is one of Earth’s best-studied UHP terranes, the evolution of this system remains controversial (Andersen et al., 1991; Terry and Robinson, 2003; Hacker et al., 2010; Labrousse et al., 2011; Brueckner and Cuthbert, 2013; Robinson et al., 2014), in large part because the exhumation mechanism is not fully understood.

Geodynamic modeling has contributed to the current understanding of the formation and exhumation of UHP terranes by quantitatively linking crustal deformation and metamorphism to plate dynamics (e.g., Warren et al., 2008a; Yamato et al., 2008; Beaumont et al., 2009; Li and Gerya, 2009; Duretz et al., 2012; Sizova et al., 2012; Bottrill et al., 2014). The prevailing quantitative paradigm is that (U)HP terranes form by continental subduction and are exhumed by buoyancy acting on the continental crust at mantle depths (e.g., Chemenda et al., 1995; Burov et al., 2001; Raimbourg et al., 2007; Gerya et al., 2008; Yamato et al., 2008; Hacker and Gerya, 2013). What distinguishes the geodynamic models from each other is the plate-tectonic setting, the size of the exhumed terrane, the timing and duration of (U)HP metamorphism, and the style of deformation during exhumation. For example, UHP crust may detach from the underlying lithospheric mantle and rise buoyantly up the subduction channel/conduit either as a rigid block (Chemenda et al., 1995) or as a deforming “plume” (Gerya et al., 2008; Warren et al., 2008a; Beaumont et al., 2009; Butler et al., 2011; Sizova et al., 2012), with the latter producing coeval “within-orogen” extension and orogenscale shortening similar to that documented in the Western Alps and Himalayas (Beaumont et al., 2009; Butler et al., 2013a).

Alternatively, exhumation may occur by “plate ejection” (reverse motion of the subducted lithosphere; Andersen et al., 1991) in response to breakoff of the subducted oceanic slab (Duretz et al., 2012; Bottrill et al., 2014), or by diapirism of UHP crust during orogen-scale extension driven by plate divergence (Ellis et al., 2011). Each of these mechanisms has the potential to operate in nature. The question is whether any of these mechanisms explains the formation and exhumation of the Western Gneiss Region.

Most previous modeling studies have focused on the formation and exhumation of UHP terranes during the early stage of collision while plate-scale convergence is active (e.g., Warren et al., 2008a; Yamato et al., 2008; Beaumont et al., 2009; Li and Gerya, 2009; Butler et al., 2013a). The problem is that this setting does not apply to the Western Gneiss Region. Instead, the Western Gneiss Region was buried during the late stages of the Caledonian orogeny beneath a large orogen where protracted convergence had already accreted terranes and produced significant crustal thickening and associated heating (Roberts, 2003; Hacker and Gans, 2005). Moreover, the Western Gneiss Region was exhumed during waning Caledonian orogenesis, possibly after convergence ceased (Fossen, 2000), rather than during unabated subduction. How these
factors affected formation and exhumation of the Western Gneiss Region has been the subject of considerable debate. Numerous studies have speculated on the geodynamic mechanisms responsible for the formation and exhumation of the Western Gneiss Region (e.g., Andersen et al., 1991; Krabbendam and Dewey, 1998; Hacker, 2007; Tucker et al., 2004; Robinson et al., 2014). Most concur that UHP metamorphism resulted from subduction of Baltican crust to mantle depths, although certain anomalous high-pressure estimates have been attributed to localized tectonic overpressures (Vrijmoed et al., 2009). In contrast, exhumation of the Western Gneiss Region has been attributed to a variety of mechanisms ranging from postconvergent extension resulting either from plate divergence (Fossen, 1992; Rey et al., 1997; Krabbendam and Dewey, 1998; Bottrill et al., 2014) or slab breakoff followed by eduction (Andersen et al., 1991; Duretz et al., 2012), to syn-/post-convergent detachment and buoyancy-driven ascent of the subducted crust (Tucker et al., 2004; Hacker, 2007; Robinson et al., 2014), possibly enhanced by partial melting of the crust at UHP conditions (Labrousse et al., 2011; Gordon et al., 2013).

Here, we present two-dimensional (2-D) numerical geodynamic models designed to explore the mechanics of late- to postorogenic exhumation of UHP terranes in large and hot orogens. We first summarize the key characteristics of Western Gneiss Region to be reproduced by any successful geodynamic model. Next, we describe results from numerical model experiments designed to capture the broad aspects of Caledonian tectonics, namely: oceanic subduction and growth of a thick orogenic wedge by accretion of multiple terranes; subduction of a continental margin; and exhumation of the margin from UHP conditions during late- to postorogenic extension. Last, we compare the model results with Western Gneiss Region observations, discuss the implications for Scandian tectonics, and test the Western Gneiss Region exhumation processes against the buoyancy paradigm that is commonly used for smaller synconvergent UHP terranes.

THE WESTERN GNEISS REGION: A LARGE AND HOT LATE-OROCENIC UHP TERRANE

The Western Gneiss Region (Figs. 1 and 2) exposes the deep levels of the Greenland-Scandian Caledonides, a bivergent orogen formed during Ordovician–Devonian subduction of the Iapetus Ocean and subsequent Baltica–Laurentia transpressional-transtensional collision (Roberts and Gee, 1985; Roberts, 2003). We describe its evolution in terms of four phases (named Phases

![Figure 1](https://pubs.geoscienceworld.org/gsa/lithosphere/article-pdf/7/4/379/3039876/379.pdf)

**Figure 1.** (A) Simplified geological map of the Western Gneiss Region (WGR) showing Baltica basement overlain by Caledonian allochthons. See B for location. Red regions are recognized ultrahigh-pressure (UHP) domains. White and gray circles indicate locations of coesite/pseudomorph- and microdiamond-bearing rocks, respectively. Dashed lines represent contours of peak metamorphic temperature (red) and pressure (blue), from Hacker et al. (2010). (B) Simplified map of southwestern Norway showing Caledonian structures discussed in text. JD—Jotun detachment; LGF—Lærdal Gjende fault; HSZ—Hardangerfjord shear zone; MTFZ—Møre Tøndelag fault zone; NSDZ—Nordfjord-Sogn detachment zone. Dashed lines x–x’ and y–y’ indicate locations of cross sections shown in Figure 2. The Caledonian allochthons and Baltican basement are separated by an implied thrust sense tectonic contact (the Scandian basal detachment). Top-W basal detachments indicate where this contact was reactivated as a top-W extensional shear zone/fault.
Figure 2. Present-day cross sections and schematic tectonic reconstruction of the Western Gneiss Region. See text for explanation of numbers 1–6. (A) Schematic crustal-scale reconstruction of the Norwegian Caledonides at ca. 395 Ma, following subduction and partial exhumation of the metamorphosed Baltic basement (2, 3) from beneath the Caledonian allochthons (1). Figure is modified from present-day cross section of Milnes et al. (1997), x–x’ of Figure 1B. Bold black line shows position of the basal thrust (4) (black kinematic indicators), locally termed the Jotun detachment, which was reactivated as a top-(N)W ductile shear zone (white kinematic indicators) during exhumation of the ultrahigh-pressure (UHP) Baltic crust. The “Former UHP?” region (here and in B) is speculative, based on extrapolation from UHP domains further north (cross-section C). Eclogitization of the Western Gneiss Region lower crust is assumed based on its apparent continuity with overlying (U)HP Baltic basement. (B) Present-day cross section (modified from Milnes et al., 1997, their Fig. 1) showing inferred removal of allochthons owing to NW-directed extension on the detachment between the allochthons and Baltic crust (4), and later normal faulting within the Nordfjord-Sogn detachment zone (NSDZ) and on the Lærdal Gjende fault (LGF) (6) and the Hardangerfjord shear zone (HSZ). The Hardangerfjord shear zone is a top-W ductile shear zone that was overprinted by the Lærdal Gjende fault. The dashed line (HSZ) in B and C shows the correct dip of the Hardangerfjord shear zone/Lærdal Gjende fault at depth based on recent seismic data from Fossen et al. (2014). Progressive E-W cooling during exhumation (5) is indicated by ages younging from ca. 400 to 380 Ma toward the northwest. (C) Schematic present-day cross section (y–y’, Fig. 1B) through the Nordøyane UHP domain, modified from Milnes et al. (1997), illustrating the present cross-sectional width of the UHP domain–Western Gneiss Region (Hacker et al., 2010) and oblique sinistral shearing along the Møre Tøndelag fault zone (MTFZ).
1–4), which we later associate with tectonic regimes (Fig. 3). In Scandinavia (Figs. 1 and 2), the orogen consists of a stack of allochthons derived from Baltica, Iapetus ophiolites and volcanic arcs, and the Laurentian margin (Fig. 1A; cf. Roberts and Gee, 1985; Gee et al., 2013; Corfu et al., 2014), assembled by westward accretion (Phase 1) and later underthrusting of the Baltic margin during the Silurian–Devonian or “Scandian” phase of the orogeny (Phase 2; Hacker and Gans, 2005). In western Norway and eastern Greenland, these nappes are locally overlain by Devonian–Carboniferous conglomerate/sandstone basins formed during postorogenic extension (Phases 3–4; e.g., Osmundsen and Andersen, 2001).

Throughout the nappe stack, Ordovician–Silurian high-grade rocks, locally including UHP assemblages (Jannik et al., 2012, 2013; Root and Corfu, 2012), migmatites, and arc volcanics and plutons, point to a protracted pre-Scandian orogenic history (Roberts, 2003; Hacker and Gans, 2005; Gee et al., 2013). We refer readers to Roberts (2003), Hacker and Gans (2005), and Brueckner and Cuthbert (2013, and references therein) for details of these pre-Scandian events. For the present work, the crucial point is that by the onset of Phase 2, i.e., the Scandian (Middle Silurian, ca. 425 Ma, Fig. 3; Andersen et al., 2004; Kylander-Clark et al., 2004; Tucker et al., 2004) and their Fig. 1). Cooling of Western Gneiss Region between these ages preserved in eclogite and peridotite bodies, UHP parageneses also appear locally in the Proterozoic orogenes (referred to as the Western Gneiss Complex) that dominate the Western Gneiss Region (Wain, 1997; Wain et al., 2001; Walsh and Hacker, 2004; Root et al., 2005), implying that the entire region experienced burial to mantle depths. The highest-pressure (coesite ± diamond-bearing) eclogites are exposed within the Nordøyane UHP domain, reach- ing ~3.2 GPa and 820 °C and decreasing south-eastward to ~2.0 GPa and 650 °C (Fig. 1; Wain et al., 2000; Cuthbert et al., 2000; Terry et al., 2000b; Krogh-Ravna and Terry, 2004; Walsh and Hacker, 2004; Young et al., 2007; Butler et al., 2013b). Whether the transitions between UHP and non-UHP domains represent true metamorphic gradients (i.e., a structurally continuous record of coeval metamorphism at different P-T conditions) or diachronous metamorphism and/or later tectonic juxtaposition remains debated (cf. Cuthbert et al., 2000). However, the apparent absence of structural breaks within much of the Western Gneiss Region (Young et al., 2007; Hacker et al., 2010) has been interpreted by many to indicate that the Western Gneiss Region was buried and exhumed as a “coherent” body.

Decompression of the Western Gneiss Region (Phase 4) by ca. 395–390 Ma (Fig. 3; Tucker et al., 2004; Kylander-Clark et al., 2009; Krogh et al., 2011) is recorded by amphibolite- to granulite-facies metamorphism at P-T conditions from ~0.5 GPa and 600 °C to ~1.5 GPa and 800 °C (Labrousse et al., 2004; Root et al., 2005; Terry and Robinson, 2003; Walsh and Hacker, 2004; Butler et al., 2013b). The timing of this metamorphism appears to vary locally, with associated U-Pb zircon and titanite ages ranging from ca. 400 to 380 Ma (Tucker et al., 2004; Kylander-Clark et al., 2009; Krogh et al., 2011; Spencer et al., 2013).

In the west, decompression was accompanied by partial melting, possibly beginning at UHP conditions (Labrousse et al., 2002; Tucker et al., 2004; Krogh et al., 2011; Labrousse et al., 2011; Gordon et al., 2013; Ganzhorn et al., 2014), although some of these melts may prove to be Precambrian in age (Hacker et al., 2010, their Fig. 1). Cooling of Western Gneiss Region orthogencses through ~400 °C took place by ca. 400 Ma in the east, and by ca. 385–360 Ma in the west, with the younger ages preserved in the UHP domains (Root et al., 2005; Walsh et al., 2007, 2013; Steenkamp, 2012).

Deformation during and/or after decompression (Phase 4) was dominated by exten-
sion attributed to regional transtension (Krabbendam and Dewey, 1998; Barth et al., 2010). From north to south, amphibolite-facies NE-SW sinistral shearing within the Møre Trøndelag fault zone (Figs. 1B and 2C) gave way to E-W (constrictional) stretching and top-W shearing (Andersen, 1998; Milnes et al., 1997; Krabbendam and Dewey, 1998; Terry and Robinson, 2003). Across the Western Gneiss Region, the intensity of Scandian deformation decreases eastward toward the foreland (Hacker et al., 2010). Near Nordfjord, Baltican basement gneisses containing UHP eclogites were juxtaposed at amphibolite facies (~0.6–1 GPa) with overlying lower-pressure rocks across the top-W Nordfjord-Sogn detachment zone (Figs. 1B, 2A, and 2B) (Andersen, 1998; Johnston et al., 2007). Further east, the Jotun detachment was reactivated from ca. 402 to 394 Ma as a top-W extensional shear zone (Figs. 1B, 2A, and 2B; Fossen and Dunlap, 1998; Fossen, 2010). Crustal extension was accompanied by deposition of a series of deep Devonian–Carboniferous basins within the hanging wall of the Nordfjord-Sogn detachment zone (Fig. 2B; Osmundsen and Andersen, 2001). As extension continued, the Western Gneiss Region and overlying allochthons were cut by W-dipping crustal-scale normal faults (Fig. 2B; Fossen, 1992; Milnes et al., 1997; Andersen, 1998).

Exhumation of the Western Gneiss Region appears to have largely postdated orogen-core shortening (Fossen, 1992; Fossen and Dunlap, 1998). Some have proposed that exhumation of the Western Gneiss Region was accommodated by coeval shallow-level extension and shortening at depth (Andersen, 1993; Hacker et al., 2010), and in the foreland of the orogeny, there is clear evidence that Baltican basement was incorporated into shallow-level thrust sheets (Gee et al., 2010). However, whether a much larger-scale E-directed thrust exists beneath the Western Gneiss Region, which could have accommodated its exhumation, remains to be shown (Fig. 2B). At higher structural levels, thrusting along the Caledonian sole thrust (Jotun detachment) appears to have ceased by ca. 408 Ma (the interpretation of Fossen and Dunlap, 1998), while Baltica basement was still at UHP conditions (Carswell et al., 2003). Finally, in the foreland basin to the east, the youngest sediments affected by thrusting are Late Silurian (423–418 Ma; Bockelie and Nystuen, 1985). The sparse evidence of late shortening, except locally (Robinson et al., 2014; see following), coupled with the regional pattern of amphibolite-facies extension, has been explained by regional transtension owing to oblique plate divergence, with exhumation of the Western Gneiss Region attributed to extreme vertical thinning of the crust (Krabbendam and Dewey, 1998). This implies a change from shortening to extension over an up to 10 million years (Myr) interval, which we provisionally term Phase 3. However, given the uncertainties in the geochronological data, it is possible that this interval was significantly shorter.

These observations underscore several important characteristics (some sequential, others coeval) of Caledonian and Western Gneiss Region tectonic history that any successful geodynamic model should reproduce (Fig. 2):

1. terrane accretion and growth of a large orogenic wedge, corresponding to the pre-Scandian accretion and high-grade metamorphism of Caledonian allochthons (Fig. 2A [1]; e.g., Hacker and Gans, 2005);
2. subduction of the Baltican continental margin beneath this wedge, corresponding to ca. 415–400 Ma (and perhaps as early as ca. 420 Ma; Kylander-Clark et al., 2007) UHP metamorphism of the Western Gneiss Region at ~3.5 GPa and 800 °C, with P and T both decreasing eastward (Fig. 2A [2]; Wain et al., 2000; Cuthbert et al., 2000; Terry et al., 2008b; Krogh-Ravna and Terry, 2004; Walsh and Hacker, 2004; Young et al., 2007; Butler et al., 2013b);
3. slow exhumation to midorogenic crustal depths (Fig. 2A [3]) corresponding to decompression and ca. 400–380 Ma (Spencer et al., 2013) amphibolite-facies metamorphism of the Western Gneiss Region at ~1.5 GPa and 800 °C (Labrousse et al., 2004; Kylander-Clark et al., 2008; Tucker et al., 2004; Butler et al., 2013b) and associated partial melting (Labrousse et al., 2002; Schärer and Labrousse, 2003), with P-T conditions also decreasing eastward (Cuthbert et al., 2000; Walsh and Hacker, 2004; Spencer et al., 2013);
4. extensional deformation during exhumation and amphibolite-facies metamorphism, including E-W stretching and normal-sense shearing focused at the top of the exhuming UHP terrane, corresponding to the Nordfjord-Sogn detachment zone (Fig. 2A [4]; Andersen, 1998; Johnston et al., 2007; Hacker et al., 2010) and reactivation of the Jotun detachment (Fossen and Dunlap, 1998) with limited evidence for coeval shortening (see following);
5. progressive westward cooling of the Western Gneiss Region from amphibolite facies documented by titanite (Tucker et al., 2004; Kylander-Clark et al., 2008; Spencer et al., 2013) and mica ages (Fig. 2B [5]; Walsh et al., 2007); and
6. postorogenic upper-crustal extension corresponding to the formation of Devonian basins in the hanging wall of the Nordfjord-Sogn detachment zone, including later overprinting of amphibolite-facies structures by brittle normal faults (Fig. 2B [6]; Andersen, 1998; Cuthbert, 1991; Johnston et al., 2007).

In addition, some aspects of Western Gneiss Region evolution, in particular, the style of exhumation-related deformation at the regional scale and the role of partial melting during exhumation, remain controversial. These are considered in the discussion where model results are compared with observations.

METHODS

We used the nested version of the finite element software SOPALE to solve 2-D thermal-mechanical creeping flows at the mantle scale (cf. Butler et al., 2014; see also Appendix A1, Eq. A1–A3). The model domain (Fig. 4) is 4000 km x 1200 km (10 km x 2 km resolution) and contains a 140-km-deep nested subdomain in the vicinity of the subduction zone (from 400 to 2000 km) with a resolution of 2 x 2 km.

The models are computed in four phases considered relevant to the tectonic phases noted earlier herein (Fig. 3B). In the models, the phases are specified in terms of boundary conditions. The subducting plate (Fig. 4) consists of oceanic lithosphere, two microcontinents (outer and inner), and a procontinent margin, whereas the upper plate is a laterally homogeneous retrocontinent. The outer microcontinent and inner microcontinent are separated by outer and inner rift basins, and the assemblage represents a composite hyperextended margin as envisaged by Andersen et al. (2012). For simplicity, we assume subduction and accretion of terranes at a single subduction zone. The continental margins have relatively weaker upper and stronger lower crusts with maximum total thickness of 36 km. The microcontinents have a three-layer crust: upper crust, a decoupling layer comprising weakened upper crust, and lower crust (Fig. 4; Table 1). Their maximum thickness is 30 km. Weak cover sediment layers with thicknesses of 4 and 6 km, respectively, cover the continental margins, and the microcontinents and rift basins.

Model materials deform by either frictional-plastic (brittle) or power-law viscous (ductile) flow (Appendix A3, Eq. A4 and A5), which includes respective strain-softening/strain-weakening mechanisms. Frictional-plastic deformation depends on the dynamic pressure and effective angle of internal friction (φd, see also Appendix A1, Eq. A1–A3).

Viscous flow laws (Eq. A5) are derived from a small set of reliable laboratory flow laws scaled by the factor f to represent materials that are somewhat more or less viscous than the ref-
The effective viscosities (Eq. A5) are also linearly scaled between 1 and $W_q$ over the strain range $\varepsilon = 5-10$ (Appendix A3; see also Butler et al., 2014) to represent strain weakening. For the sediments, we use scaled wet quartzite ($W_q$; Gleason and Tullis, 1995), $W_q \times f$, where $f = 1$ and $W_q = 10$. Upper-crustal units are also based on $W_q$ values for retrocontinental upper crust ($f = 5$, $W_q = 10$), microcontinental upper crust ($f = 1$, $W_q = 10$), and the decoupling layer at the base of the upper crust ($f = 0.1$, $W_q = 1$; Fig. 4, inset), which are chosen to represent already strain-weakened detached pieces of the rifted continental margin or previously deformed terranes. The only property that is varied in the first set of models is the strength of the procontinental upper crust ($W_q \times f$, $f = 1$, 10, 25, 50). By increasing the viscous strength among models, we test the assumption that the upper crust represents incrementally stronger than the reference $W_q \times 1$. The lower basement rocks that dominate the Western Gneiss Region was subject to both plate convergence and divergence, with a possible intervening quiescent phase. The thermal boundary conditions are: basal heat flux 21.0 mW$m^{-2}$, a 0 °C surface, and insulating sides. Radioactive heat production, $A_{r'}$ is 1.3 and 0.4 µWm$^{-3}$ for the upper and lower crust, respectively, resulting in an initial steady-state Moho temperature of -560 °C, and surface heat flow of 57 mW m$^{-2}$ on the standard thickness continents. The value of upper-crustal
### TABLE 1. MODEL THERMAL AND MECHANICAL PARAMETERS

| Parameter                          | Units       | Cover including ORB/IRB | PC margin upper crust | RC margin upper crust | IMC/OMC upper crust | PC lower crust | RC lower crust | IMC/OMC lower crust | Oceanic crust | Continental lithospheric mantle | Oceanic lithospheric mantle | Sublithospheric upper mantle | Lower mantle |
|-----------------------------------|-------------|--------------------------|-----------------------|-----------------------|---------------------|------------------|---------------|---------------------|-----------------|-----------------------------|-----------------------------|-----------------------------|--------------|
| **Mechanical parameters**         |             |                          |                       |                       |                     |                  |               |                     |                 |                             |                             |                             |              |
| Thickness (max.)                  | km          | 4–6                      | 20                    | 20                    | 18 and includes 6 km cover | 12               | 12            | 12                  | 8               | 120                          | 90                          |                             | to 660 km depth | 660 to 1200 km depth |
| Reference density                 | kg m⁻³      | 2800                     | 2800                  | 2800                  | 2800                | 2900            | 2900          | 2900                | 2900            | 3350                        | 3370                        | 3370                       | 3630         |
| Reference density HP (0 °C)       | kg m⁻³      | 2900                     | 2900                  | 2900                  | 2900                | 3200            | 3200          | 3200                | 3400            | No change                   | No change                   | No change                   | No change    |
| Reference density UHP (0 °C)      | kg m⁻³      | 3000                     | 3000                  | 3000                  | 3000                | No change        | No change     | No change            | No change        | Changes to lower mantle at 660 km depth | No change                   |                             | No change    |
| Effective angle of internal friction (φ_eff) | deg         | 6–4                      | 15–4                  | 15–4                  | 15–4                | 15–4            | 15–4          | 15–4                | 15–4            | 15–4                        | 15–4                        | 15–4                      |              |
| Cohesion                          | MPa         | 2                        | 2                     | 2                     | 2                   | 2               | 2             | 2                   | 2               | 0                           | 0                           | 0                          |              |
| Flow law                          |             | WQ                       | WQ                    | WQ                    | WQ                  | DMD             | DMD           | DMD                 | DMD             | WOL                         | WOL                         | WOL                       | Constant viscosity 3 × 10² Pa s |
| Viscosity scaling factor (f)      |             | 1                        | 1                     | 10, 25, 50             | 5                  | 1               | 0.5           | 0.5                 | 0.5             | 0.1                         | 7.5                         | 2                         | 2             |
| Strain weakening factor (W_t)     |             | 10                       | 10                    | 10                    | 10                 | 10              | 10           | 10                  | 10              | 1                           | 1                           | 1                         |              |
| n                                 |             | 4                        | 4                     | 4                    | 4                  | 4               | 4            | 4                   | 4               | 3                           | 3                           | 3                         |              |
| A²                               | Pa⁻¹ s⁻¹    | 8.57 × 10⁻¹⁰             | 8.57 × 10⁻¹⁰          | 8.57 × 10⁻¹⁰          | 8.57 × 10⁻¹⁰       | 5.78 × 10⁻⁷⁻⁷   | 5.78 × 10⁻⁷⁻⁷ | 5.78 × 10⁻⁷⁻⁷ | 5.78 × 10⁻⁷⁻⁷ | 1.76 × 10⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³⁻³-
heat production is consistent with measurements from granitoid gneisses from western Norway (Slagstad, 2008).

Crustal materials undergo reversible pressure- and temperature-dependent density changes at the eclogite and coesite-eclogite phase transitions (Table 1; Appendix A5). The densities of the upper continental crust at low, high, and ultrahigh pressures are 2800, 2900, and 3000 kg m\(^{-3}\), consistent with the range of calculated densities for orthogneisses from the Western Gneiss Region (Walsh and Hacker, 2004). Similarly, lithospheric/upper-mantle materials undergo a reversible phase change at the upper-lower mantle boundary (Table 1; Appendix A5).

The dynamically consistent model topography (Appendix A7) is a product of the model calculation without any imposed constraints, except for water loading where the surface is below sea level, and local sediment loading. Slope-dependent surface erosion acts on the local subaerial topography. It varies spatially according to local slope of the model surface, and it is defined by a maximum erosion rate that operates on a slope of 45° and is scaled down linearly to the local slope of the surface. This results in nominal erosion rates of ~0.5–1 mm yr\(^{-1}\) during Phases 1 and 2. During Phases 3 and 4, slope-dependent erosion rates increase somewhat, but the maximum is limited to 2 mm yr\(^{-1}\). Other model properties are given in Table 1 and described in more detail in Appendix A.

**MODEL RESULTS**

The depth of crustal subduction and style of exhumation in the models depend critically on the initial strength of the procontinental margin crust and its subsequent evolution. To illustrate this point, we present a set of models with contrasting initial procontinental margin upper-crustal strengths of WQ \(\times f\), where \(f = 1, 10, 25\), and 50. Each of the models undergoes the same four-phase evolution described in the methods section, corresponding to the interpreted four phases of the Caledonian/Scandian orogeny.

**Model f50, Phase 1: Oceanic Subduction and Microcontinent Accretion**

Model f50, with the strongest procontinental margin crust (WQ × f, \(f = 50\)), is our preferred model. It begins with 15 Myr of oceanic subduction at \(V_s = -5\) cm yr\(^{-1}\), during which deformation is confined to the retrocontinental accretionary wedge and the subduction channel. \(V_s\) is reduced to \(-2.5\) cm yr\(^{-1}\) when the outer microcontinent (OMC) contacts the accretionary wedge (Fig. 3B). We define this as time of the onset of “collision” and measure model time and convergence with respect to this event (millions of years postcollision, for which we define the term Myr-pc). Following collision (Fig. 5A), the leading edge of the outer microcontinent is subducted beneath the retrocontinent. The outer microcontinent upper crust progressively decouples along the preweakened detachment layer and frontally accretes to the wedge, forming a stack of thrust sheets (Fig. 5B). Subsequent underthrusting of the outer rift basin (ORB) followed by the inner microcontinent (IMC) leads to shortening of the wedge and the addition of a second thrust stack. Phase 1 ends 32.5 Myr-pc (Fig. 3B), after the inner rift basin (IRB) underthrusts the orogenic wedge. By this time, the model has accreted a thick (~24–70 km), ~250-km-wide orogenic wedge composed of stacked thrust sheets and sediments derived from the outer microcontinent, inner microcontinent, and rift basins (Fig. 5C). At depth, the wedge consists of strongly deformed HP/UHP microcontinent crust that has been exhumed to lower-crustal depths, whereas at higher levels, the thrust sheets remain intact.

**Model f50, Phase 2: Subduction of the Procontinental Margin**

At 32.5 Myr-pc, the procontinental margin starts to underthrust the orogenic wedge at \(V_s = -2.5\) cm yr\(^{-1}\) (Fig. 3B). The wedge shortens, driving some HP/UHP microcontinental crust up and prowad over the subducting margin (Fig. 6A). With the exception of its sedimentary cover, which detaches and accretes to the front and base of the wedge, the margin crust remains coupled to the subducting lithosphere (Fig. 6A). Deformation is concentrated within a top-to-the-foreland shear zone separating the procontinental margin from the overlying orogenic wedge (Fig. 7A), with minor internal deformation of the wedge and the margin crust. By the end of Phase 2 (45 Myr-pc), the leading 150-km-wide portion of margin crust has reached HP/UHP conditions. Model tracking particles record relatively cool prograde P-T paths during Phase 2, with maximum P-T conditions at the leading edge of the margin of ~4 GPa and 600 °C, and with \(P\) and \(T\) decreasing systematically upward along the inclined margin (Fig. 6A). The model geometry at this time consists of the orogenic wedge, with stacked thrust sheets, including fragments of partly exhumed UHP crust derived from the microcontinents, underlain by the relatively undeformed procontinental margin within which P-T conditions increase gradually toward its leading edge.

**Model f50, Phase 3: Quiescent Phase**

Phase 3 starts 45 Myr-pc, when convergence ceases (\(V_s = 0\) cm yr\(^{-1}\); Fig. 3B). In response, the system thermally relaxes from a subduction-style state reflecting advective underthrusting of the cool procontinent, toward an equilibrium in which the underthrust margin, the subduction conduit, and the accretionary wedge heat up by diffusion and radioactive heating of accreted continental crustal units (Fig. 6). Surface erosion removes crust at an average rate of 1–2 mm yr\(^{-1}\) (1–2 km Myr\(^{-1}\)), but this does not prevent temperature increasing at depth. P-T paths during Phase 3 exhibit near-isobaric heating, with maximum temperatures at the leading edge of the margin reaching ~750–800 °C (Fig. 6B). At depth, the full effective slab pull now acts on the stationary prolithosphere. The subducted slab also heats up, and eventually slab pull overcomes its strength, determined by the stress threshold of 80 MPa corresponding to the onset of Peierls creep; at 54 Myr-pc, the slab detaches (Fig. 8).

From the tectonic perspective, Phase 3 is characterized by slab detachment and a minor amount of exhumation of (U)HP crust (Fig. 6B). This exhumation takes the form of rotation and uplift of the procontinental lithosphere, coupled with limited gravitational spreading of the orogen (Figs. 6 and 7). Despite heating to ~750–800 °C during thermal relaxation, the margin crust does not detach. Rather, the buoyant, upward rotation of the subduction channel results in limited decompression as this region starts to exhume. Strain during this phase is concentrated in the necking region of the slab, and in the top-to-the-foreland shear zone that separates the wedge and margin during gravitational spreading. At the end of Phase 3, 55 Myr-pc, the orogenic wedge retains its full crustal thickness and the HP/UHP margin crust has not yet detached from the underlying lithospheric mantle (Fig. 6B).

**Model f50, Phase 4: Extension**

Phase 4 starts 55 Myr-pc with the onset of lithospheric divergence at \(V_s = -1\) cm yr\(^{-1}\) (Fig. 3B). Continued extension leads to retrograde transport of the orogenic wedge, resulting in normal-sense reactivation of the former thrust-sense shear zone between the wedge and underlying margin (Figs. 6C and 7B). Surface erosion continues at a similar rate as in Phase 3 but gradually declines as the model orogenic wedge becomes more plateau-like during extension. At depth, the underthrust prolithosphere rebounds isostatically, ascending beneath the orogenic wedge together with its attached margin crust. By 60 Myr-pc, the upper surface of the HP/UHP margin has exhumed to lower-middle crustal depths, ~36 km (Fig. 6C). Slow exhumation and continued heating of the continental margin at depth by thermal relax-
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**EXHUMATION**

**Research**

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Figure 5. (A–C) Evolution of model f50, Phase 1: accretion of allochthons. (A) Following subduction of the oceanic lithosphere (Fig. 4), the outer microcontinent (OMC) collides with the retrocontinent at 0 Myr-pc and begins to decouple from its lower crust; (B) at 10 Myr-pc, the outer microcontinent has been accreted by decoupling from its lower crust to form thrust sheets, which are stacked and accreted to the retrocontinent; (C) at end of Phase 1, 33 Myr-pc, the outer microcontinent and inner microcontinent (IMC) have been accreted as stacked thrust sheets and the outer and inner rift basins (ORB and IRB) initially underthrust the orogen but are accreted at depth. Convergence rate is –2.5 cm yr⁻¹ following contact of the outer microcontinent with the retrocontinent, defined as “collision.” During the 33 Myr part of Phase 1 following collision, there is 1650 km of convergence, ending with the arrival of the procontinent margin at the toe of the accreted allochthons.
Figure 6. Evolution of model f50, Phases 2–4: margin underthrusting, quiescence, and extension, shown in four frames with current positions of pressure ($P$) and temperature ($T$) tracking particles (a–c, square boxes at left) and corresponding $P$-$T$ paths (right, ending with current values, circles a–c).

(A) Model at end of Phase 2 procontinent margin underthrusting, showing the procontinent margin separated from the overlying orogenic wedge by a thrust-sense shear zone (large black arrow). Current metamorphic conditions are ~3.5 GPa and 600 °C within the ultrahigh-pressure (UHP) part of the margin (red). (B) End of Phase 3, following 10 Myr of quiescence ($V_p = V_r = 0$ cm yr$^{-1}$). Gravitational spreading (velocity arrows) during this phase leads to minor thrusting in the foreland. The procontinent margin crust remains buried, with its (U)HP part undergoing near-isobaric heating to ~700–850 °C (right). (C) Mid-Phase 4 extension, showing the position of the exhuming margin after 50 km of plate divergence ($V_r = -1$, $V_p = 0$ cm yr$^{-1}$), and consequent isothermal decompression of the UHP part of the margin to ~1 GPa and 700–850 °C (right). Exhumation during this phase is primarily accommodated along the normal-sense shear zone developed between the procontinent margin and the orogenic wedge (see Fig. 7C). (D) End of Phase 4 extension showing final geometry of orogen after 100 km of extension and early cooling of the margin. Later cooling of the margin through 450–350 °C (approximate mica closure temperature) is achieved during postextension quiescence [Fig. 3] not shown here.
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The system then begins to cool in response to erosion and tectonic thinning of the orogenic wedge. As the orogen cools, the style of deformation shifts from viscous normal-sense shearing to more localized normal faulting (Fig. 7C) and associated subsidence/basin formation within the rear of the orogenic prowedge (hinterland), located above the highest-P portion of the exhumed margin crust.

The final geometry of model f50, after 100 km of extension, consists of residual microcontinent-derived thrust sheets resting above the now flat-lying procontinent margin (Fig. 6D). At the hinterland side of the margin, the normal-sense shear zone responsible for exhumation from mantle depths is transected by late-stage normal faults with an adjacent 2-km-deep syntectonic basin (Figs. 6D and 7D). The margin remains relatively undeformed, retaining its HP/UHP metamorphic field gradient. The final configuration (Figs. 6D and 7D) represents an orogen ~10 Myr into slow extension.

A combination of continued extension and protracted erosion will remove most of the remaining orogenic wedge and expose the surface of the underthrust continental margin. At this stage, the crust will have returned to a normal continental thickness with little or no high topography and a correspondingly low erosion rate. For the purposes of comparing the cooling history of the model with that of the Western Gneiss Region, we ran the model for an additional phase of postextension quiescence during which $V_p = 0 \text{ cm yr}^{-1}$ (Fig. 3B).

MODEL SENSITIVITY TO UPPER-CRUSTAL STRENGTH, MELT WEAKENING, AND OTHER INPUT VARIABLES

Variation in Initial Strength of the Procontinent Upper Crust

Model experiments with weaker procontinent margin crust ($f = 1, 10, 25$) provide sensitivity tests in regard to the subduction and decoupling of the model continental margin during Phases 2 and 3. All models ($f = 1, 10, 25$, and 50) result in the underthrusting/subduction of the margin to similar UHP conditions during Phase 2. Figure 9A shows a model result representative of all models at the end of Phase 2. All four models are essentially the same at this stage. The results show that even upper crust with the reference wet quartz flow law can be subducted at 2.5 cm yr$^{-1}$.

However, the behavior of the models during Phase 3 quiescence is quite different (Fig. 9). In model f1 (WQ × $f = 1$), detachment and upward buoyancy-driven flow of the pro-
continent margin follow immediately after the onset of Phase 3 quiescence at 45 Myr·pc. By 46 Myr·pc (Fig. 9B), there has been significant upward flow of the UHP crust and overlying allochthons. This result demonstrates that the margin upper crust only remains immune to its buoyancy while there is a large basal traction associated with active subduction. As soon as the basal traction is reduced, when the subduction stops, the buoyancy causes the upper crust to decouple. A consequence is that UHP temperatures recorded by the margin crust reach a maximum of only ~570 °C at 3.2 GPa, for the reference particle (Fig. 9B), considerably lower than in model f50 (Fig. 6A) owing to the negligible residence/incubation time at UHP conditions.

In models f10 and f25 (WQ × f = 10, 25), the margin crust remains attached longer during Phase 3, until 50 Myr·pc for model f10 and until 53 Myr·pc for model f25. The longer Phase 3 incubation times for models f10 and f25, 5 and 8 Myr, respectively, allow time for some thermal relaxation before the onset of exhumation, so that these models record maximum temperatures of ~600 °C and ~700 °C for the reference particle at 3.5 GPa (Figs. 9C and 9D), before heating reduces the strength of the base of the upper crust and it decouples. In model f10, the decoupled UHP crust exhumes rapidly in the form of a 20-km-thick buoyant plume/tongue (Fig. 9C) that intrudes the overlying allochthons adjacent to their suture with the retrolithosphere. The upper part of this plume of (U)HP crust diverges from the remaining attached HP region of the procontinental upper crust. In model f25, there is limited decoupling by the end of Phase 3, but this increases significantly if the quiescent phase is extended for a total of 20 Myr, but the UHP and HP parts of the procontinent crust remain contiguous. In contrast, model f50 displays no significant decoupling of procontinental crust during the 10 Myr duration of Phase 3 (Fig. 6B).

In summary, these results imply that the continental margin upper crust must be significantly stronger than reference wet quartz, i.e., f = 25–50, if it is to remain in place during Phase 3 quiescence. In the models, this is necessary to achieve temperatures of 750–800 °C at UHP conditions before decompression to amphibolite-facies conditions during Phase 4.

**Effect of Melt Weakening of Procontinental Upper Crust**

It has been suggested that widespread partial melting at (U)HP conditions or during decompression may have assisted or even triggered exhumation by weakening the (U)HP Baltic crust in the Western Gneiss Region (Labrousse et al., 2011; Gordon et al., 2013; Ganzhorn et al., 2014). We have tested the potential role of melt weakening in triggering decoupling and exhumation of the upper procontinental crust using models f50-MW1 and f50-MW2 (where MW is “melt weakening”). These models are the same as model f50 except there is a reduction in the effective viscosity of the procontinental upper crust with increasing pressure and temperature, designed as a simplified parametric approximation to the bulk weakening effect of incipient melting (5%–10%). Melt weakening is restricted to the upper procontinental crust. The more mafic, and assumed “dry,” lower crust and upper mantle would not melt at the ambient conditions developed in the models.

Melt weakening is implemented by a linear decrease in the material viscosity with increasing temperature above the T-and P-dependent solidus $T_{solidus}(P, T)$ (Fig. 10A). At $T_{solidus}(P, T)$, the effective viscosity, $\eta_{eff}$ is the value calculated by power-law viscous flow (Eq. A5) and decreases linearly to a melt-weakened lower limit, $\eta_{eff}$ of $1 \times 10^{19}$ Pa·s at $T_{MW} = T_{solidus} + 50$ °C. The latter approximates the temperature at which the melt fraction of the chosen bulk composition reaches ~10%, similar to the amount required to produce melt connectivity and consequent bulk rheological weakening in laboratory deformation experiments (Rosenberg and Handy, 2005).

Figure 8. Model f50, showing necking (A) and breakoff (B) of the subducted lithosphere during Phase 3 quiescence at 53 Myr·pc and 55 Myr·pc, respectively. Necking develops at depth in the inner rift basin (IRB) lithosphere following end of convergence at 45 Myr·pc under the effective slab-pull of the subducted lithosphere. Breakoff occurs at 54 Myr·pc.
Figure 9. Comparison of tectonic evolution and representative, equivalent pressure-temperature (P-T) paths (same particles) for models f1, f10, and f25, showing the effect of varying procontinent margin strength. The equivalent P-T path for model f50 is shown for comparison in each case (orange). (A) All models are nearly identical up to the end of Phase 2 margin underthrusting (model f50 shown). Complete P-T paths (right) are alike to end of Phase 2 (orange circle) but later diverge. (B) Model f1, with the weakest procontinent margin crust (f = 1). The onset of Phase 3 quiescence leads to immediate detachment and buoyancy-driven exhumation of the ultrahigh-pressure (UHP) margin crust. The short residence time at depth is reflected in the lack of heating of the margin crust at UHP (particle d, right), which only reaches ~575 °C at 3 GPa, when compared with model f50 (particle a). (C) In model f10, the stronger crust (f = 10) remains coupled to the slab somewhat longer, until 49 Myr-pc (4 Myr into the quiescent phase), and consequently the margin heats to ~650 °C at 3 GPa (right). Like model f1, detachment is followed by buoyancy-driven ascent of the UHP margin crust, which undergoes foreland-directed thrusting over the adjacent margin crust. (D) In model f25 (f = 25), the margin crust remains attached to the slab throughout most of Phase 3 quiescence, resulting in heating to ~750 °C (right). Detachment at 51 Myr-pc is followed by relatively slow, buoyancy-driven ascent of UHP crust over the adjacent margin, owing to the higher viscosity of the margin crust (see discussion). Final exhumation in model f25 is achieved during Phase 4 extension, by a combination of buoyancy-driven thrusting and extension resulting from plate divergence, but dominated by the latter. During exhumation, the margin undergoes near-isothermal decompression (particle f, right), similar to model f50 (particle a).
Figure 10. Comparison of tectonic evolution (left) and representative, equivalent pressure-temperature (P-T) paths from models f50-MW1 and f50-MW2, showing the effects of partial melting (right). See text for explanation of melting model. Gray area in corresponding P-T diagrams (right) represents the region of melt weakening and shows the 50 °C temperature interval over which the viscosity of the margin crust decreases linearly from that determined by power-law flow, \( \eta_{PL} \), to the minimum viscosity of the melt-weakened crust, \( \eta_{MW} = 1 \times 10^{19} \text{ Pa}\cdot\text{s} \) at \( T_{Solidus} + 50 \). Particle “a” from model f50 is shown for reference (orange, right). (A) Model f50-MW1, in which melting of the margin crust is determined by the solidus of Auzanneau et al. (2006). At the end of Phase 3 quiescence, only the very leading edge of the margin crust that is in contact with the sublithospheric mantle has reached the >800 °C required to produce melt at ultrahigh-pressure (UHP) conditions. By comparison, the margin crust within the subduction channel (particles g and h) remains too cool (by ~25–50 °C) and too high pressure. (B) Model f50-MW1: during Phase 4 extension, the margin undergoes isothermal decompression, leading to the onset of partial melting at ~2–3 GPa (particles g and h). The amount of partial melting remains minor, and the overall effect on the style of exhumation compared to model f50 (Fig. 6) appears negligible. (C) Model f50-MW2: where the solidus temperature of Auzanneau et al. (2006) is reduced by 100 °C for all pressures to approximate a higher water content, but not water-saturated conditions. Partial melting starts at UHP conditions during the quiescent phase, and the resulting viscosity reduction leads to detachment and buoyancy-driven flow of margin crust. (D) Model f50-MW2: continued melting during Phase 4 extension leads to more widespread detachment of the margin crust from the procontinent lithosphere and insertion into the overlying orogenic wedge.
For demonstration purposes, we have used two pressure-dependent solidi. The high-temperature one (~770–900 °C), model f50-MW1 (Figs. 10A and 10B), uses the melt-fraction data of Auzanneau et al. (2006) for high-pressure melting of graywacke, which has a similar bulk composition to Baltica gneiss of the Western Gneiss Region. In the Auzanneau et al. (2006) experiments, the graywacke contained water in hydrox phases, and the rocks were water-saturated at (U)HP conditions. We judge this to be the minimum water content that will produce melting in the models at UHP conditions. The corresponding anhydrous solidus, for example, based on the dehydration of phengite, is ~900 °C at 3 GPa (Hermann, 1997), and therefore melting is not possible under model (U)HP conditions. In both cases, there would be substantial decompression melting of the (U)HP continental margin following the onset of exhumation. For example, the tracking particle “g” starts to melt at ~2.5 GPa (Fig. 10B), but other hotter regions will melt at somewhat higher pressures, e.g., particle “h.” Our purpose is to test whether melting at UHP conditions may have triggered exhumation.

The solidus used in model f50-MW2 (Fig. 10C) is 100 °C lower than that in model f50-MW1 and is intended to show approximately what would happen to the same material with a higher water content, but still not fully water-saturated (i.e., water-deficient melting). Unfortunately, there are no graywacke experimental data on which to base this case. We have chosen a solidus that lies well above the granite wet melting solidus (Auzanneau et al., 2006, their Fig. 8) and closer to the wet petite solidus (Labrousse et al., 2011, their Fig. 1).

Model results show that for model f50-MW1, there is minor decoupling of the proponent upper crust during Phase 3 because the temperature exceeds the solidus at UHP conditions. ~850 °C, deeper than the “g” and “h” tracking particles in the subduction zone (Fig. 10A). At this point, the model structure is essentially the same as model f50 with minor melt weakening. However, later during Phase 4 extension, thermal relaxation and radioactive heating have created a significant region of the proponent crust that is above 770 °C (Fig. 10B), which leads to decompression melting and melt weakening in this region as the UHP crust exhumes, because the solidus temperature decreases with pressure to a minimum of 770 °C at ~2 GPa. Melt weakening results in some decoupling of the now-retrogressed UHP margin (Fig. 10B) and therefore more deformation in this region than in model f50 (Fig. 6D).

Model f50-MW2 (Figs. 10C and 10D) undergoes significantly more melt weakening during Phase 3 quiescence, such that by the end of Phase 3 at 55 Myr-pc, the UHP procontinent has decoupled, retrogressed, and is rising as a plume adjacent to the suture between the retrocontinent and the allochthons (Fig. 10C). This happens before the onset of extension but also continues with rapid exhumation by 58 Myr-pc (Fig. 10D) after ~30 km of Phase 4 extension.

In summary, melt weakening is significant only during decompression in model f50-MW1 (Figs. 10A and 10B), using the Auzanneau et al. (2006) solidus. Under these conditions, melt weakening does not trigger exhumation at UHP conditions. Melting at higher water content, as shown in model f50-MW2, can trigger exhumation at UHP conditions and results in plume-style exhumation (Figs. 10C and 10D).

Model Sensitivity to Other Variables

As discussed later herein, model f50 reproduces many of the tectonic and metamorphic observations related to the Scandinavian orogeny of the Western Gneiss Region. However, aspects of model f50 are clearly simplifications, and the sensitivity of the results to poorly constrained aspects of both Western Gneiss Region geology and the models needs to be evaluated. Model f50 is intended to be a generic model for systems like the Western Gneiss Region that late-orogenic (U)HP metamorphism in large and hot orogens, not a simulation of the details of Western Gneiss Region evolution.

Accretion of Allochthons

The models specifically consider the building of a large orogen over an extended period by accretion of allochthons. The models simplify this process by considering the allochthons in terms of only two microcontinents, rather than the complex array of continental fragments and oceanic arcs that probably existed in the Iapetus Ocean (Roberts, 2003). Our opinion is that the (U)HP metamorphism and subsequent exhumation of the Baltic continental margin will not be particularly sensitive to the specific number, thickness, and size of the accreted allochthons. What is required is that accretion produces a large, thick orogenic wedge prior to the underthrusting of the margin, similar to the conceptual diagram of Hacker and Gans (2005, their Fig. 9), but probably thicker. This is supported by metamorphic data from Caledonian allochthons (Hacker and Gans, 2005) and is required to produce the observed peak (U)HP pressure conditions in the model. While clearly a simplification for the Western Gneiss Region, the model represents the equivalent generic processes required. More complex models could be designed to reproduce the Caledonian accretion history more closely, but any successful model will need to produce a thick orogenic wedge. The implication is that the thin allochthons observed today must represent the remnants of much thicker, perhaps composite, allochthons, some of which were attenuated during their emplacement (Northrup, 1996) and subsequently mostly removed tectonically or by erosion.

Rate of Underthrusting of Procontinental Margin

A second poorly constrained variable is the rate of underthrusting of the Baltic margin. In the models, this has to be fairly rapid, ~2.5 cm yr⁻¹, in order to underthrust the margin to the required (U)HP conditions in the 10–25 Myr period allowed by geological constraints (i.e., from 415 to 400 Ma; Fig. 3A). The specific rate depends on the dip of the margin and the size of the orogen. However, it is difficult to produce the broad UHP-HP region observed (Figs. 1 and 2) in models with a steeper dip and/or a smaller orogenic wedge. We further note that matching geological constraints also depends on our choice of alignment of the model time with geological time (i.e., 0 Myr–pc = 425 Ma). For example, choosing 0 Myr–pc = 430 Ma would produce (U)HP metamorphism 5 Myr earlier in the geological sense at the same convergence rate.

We also note that an alternative model, model f50-Vp1, which is the same as model f50 except Vₚ = -1 cm yr⁻¹ during Phase 2 and Phase 3, is also in thermal disequilibrium at the end of the now much slower and therefore longer, 25 Myr underthrusting phase. Even at this low velocity, the peak UHP temperature is lower (650 °C) than observed in the Western Gneiss Region. Slow convergence models of this type imply an older alignment time, of ca. 435 Ma in order to achieve (U)HP metamorphism by 420 Ma geological time.

Amount of Divergence

There is also little orogen-scale information to constrain the rate and amount of plate-scale divergence in the Western Gneiss Region, although the amount of extension in Phase 4 in the model is similar to the 50–100 km of normal-sense displacement estimated for the Nordfjord-Sogn detachment zone (Andersen, 1998; Hacker et al., 2003). In the Western Gneiss Region, this phase was transensional, reflecting oblique sinistral plate motion and strike-slip deformation between Laurentia and Baltica (Krabbendam and Dewey, 1998; Dewey and Strachan, 2003; Fossen, 2010), which cannot be reproduced in the present 2-D model. Moreover, additional protracted extension in the Western Gneiss Region during the rifting that...
eventually formed the North Atlantic (e.g., Fos sen et al., 2014) is not considered in the model. Overall, we find that a slow rate of extension, ~1 cm yr\(^{-1}\) for 10 Myr (100 km total lithospheric extension), coupled with a moderate rate of erosion, reproduces the pervasive structural evidence of top-W normal-sense shear and faulting that accompanied exhumation of the Western Gneiss Region (U)HP terrane to crustal levels (Fig. 7D). This deformation cannot be attributed solely to gravitational spreading, as in Phase 3, because spreading also produces thrusting on the basal detachment, as in Phase 2 convergence. This is inconsistent with the interpretation of Fossen and Dunlap (1998) that thrusting apparently ceased by ca. 408 Ma (Fig. 3A; Fossen and Dunlap, 1998).

**Transitions**

We have chosen to present models with four distinct phases, each having constant velocity boundary conditions, except for Phase 1 (Fig. 3B). However, natural systems like the Western Gneiss Region probably evolve with transitions between phases, e.g., slowing of convergence at the Phase 2-3 transition and a progressive increase in divergence rate during the Phase 3-4 transition, as would be expected for an oblique collision that evolved from transpression to transtension. If these transitions are short (<5 Myr), as was likely in the Western Gneiss Region, the effect of omitting them in the models will be minor. This is because the temperature component of metamorphism in large orogens evolves slowly according to the combined effects of radioactive and strain heating, diffusive equilibration, and advection. The abrupt changes versus gradual transitions in velocity represent a modification to the advective component. However, this difference will only result in subtle effects in metamorphic signatures because the advective signal will be dispersed and attenuated by diffusion, which acts as a low-pass filter. Such subtle differences are not likely to be resolved by data from natural orogens.

**Slab Breakoff**

Lastly, slab breakoff (Fig. 8) occurs during Phase 3 in all of the models discussed here. This is consistent with conceptual arguments that slab breakoff follows once plate convergence ceases or slows substantially, as seen in many numerical and analogue models (Duretz et al., 2012; Bottrill et al., 2014; references therein). The main effect of slab breakoff in our models is to prevent decoupling of the pro- and retrolithospheres at the subduction zone, followed by subduction zone hinge retreat of the pro lithosphere late in the quiescent Phase 3. This retreat, seen in models without slab breakoff, would be accompanied by upwelling and widespread decompression melting of the asthenosphere as it fills the gap between the plates. There are no geological data that definitively confirm or deny whether, and when, the slab broke off beneath the Western Gneiss Region. There is, however, no evidence for the widespread intrusion of basaltic magma expected to result from decompression melting. We therefore infer either that the slab did break off beneath the Western Gneiss Region, as in the models, or that if the slab did not break off immediately, there was no decoupling or hinge retreat of the Baltic lithosphere.

**COMPARISON OF MODEL RESULTS WITH THE WESTERN GNEISS REGION**

Model f50, specifically, and others presented herein reproduce several of the key structural/tectonic features and associated metamorphism of the Western Gneiss Region, shown by numbers in Figures 11 and 12 keyed to the following list. Some of these features developed prior to ca. 395 Ma (see reconstruction in Fig. 2A), but the geometry shown is considered to be representative of their original accreted configuration. To aid in the comparison, we have chosen 425 Ma as the time that corresponds to the start of Phase 2 in both the Western Gneiss Region and the models (Fig. 3B). We then quote model times referenced to this time (Fig. 11).

1. The model orogen develops by growth of a thick orogenic wedge during Phase 1 accretion and stacking of allochthonous terranes, driven by subduction of lithosphere beneath the retrocontinent (Fig. 5). The resulting model orogenic wedge (Fig. 5C) is similar in scale to the inferred thickness of the Caledonian orogenic wedge (Figs. 11A and 11B [1]) based on maximum pressure estimates from the allochthons (e.g., Hacker and Gans, 2005). Underthrusting during this model phase also produces local early (U)HP metamorphism of accreted microcontinent crust (Fig. 11H [1]), consistent with the pre-Scandian (U)HP metamorphism preserved in Caledonian allochthons (Janák et al., 2012, 2013; Root and Corfu, 2012).

2. Subsequent Phase 2 and 3 underthrusting of the procontinental margin beneath the orogenic wedge in model f50 leads to (U)HP metamorphism at P-T conditions of up to ~4 GPa and 825 °C, with P and T both decreasing systematically toward the foreland (Figs. 11A, 11B, 11H, and 12B [2]). The metamorphic conditions recorded by the hinterland portion of the model crust, corresponding to the Western Gneiss Region UHP domains, most closely resemble estimates for the Nordsøyan UHP domain, where metamorphic temperatures reached ~820 °C at ~3.9 GPa (Terry et al., 2000b), but they also overlap the highest-temperature estimates of ~3.0 GPa and 750 °C from Nordfjord (Fig. 12B [2]; Young et al., 2007). The systematic decrease in P-T conditions toward the foreland in the model (Fig. 11H [2]) is consistent with regional decrease in peak metamorphic conditions in the Western Gneiss Region from the UHP domains eastward toward the Jotun detachment (e.g., Cuthbert et al., 2000; Walsh and Hacker, 2004).

3. (U)HP metamorphism in model f50 occurs over an ~16 Myr time span (depending on the part of the margin), ending shortly after the onset of Phase 4 extension at 55 Myr-pc (Fig. 6C). Assuming that the onset of Phase 2 corresponds to ca. 425 Ma (the start of the Scandan orogenic phase), then the timing of UHP metamorphism in model f50, from ca. 42.5 to 58.7 Myr-pc, corresponds to ca. 415–399 Ma (Fig. 12C [3]). These results closely resemble the main body of apparent (U)HP metamorphic ages for the Western Gneiss Region obtained from U-Pb zircon (ca. 415–402 Ma; Krogh et al., 2011; Carswell et al., 2003) and Sm-Nd/Lu-Hf garnet geochronology (430–400 Ma; Kylander-Clark et al., 2007, and references therein). As noted already, longer-duration (U)HP metamorphism, e.g., 20–25 Myr, as suggested by Kylander-Clark et al. (2009), could be achieved in the models simply by extending the lengths of Phases 2 and/or 3.

4. The final increment of foreland-directed thrusting and thinning of the allochthons in model f50 during Phase 3, while the margin resides at (U)HP conditions, is consistent with the last foreland-directed thrusting at ca. 408 Ma and the overall absence of late-Scandian thrusting in the foreland (Figs. 6, 11A, and 11B [4]; Fossen, 2000).

5. Phase 4 exhumation of the (U)HP procontinental margin is achieved by top-retroward shearing at amphibolite-facies conditions along the contact with the overlying allochthons during limited lithospheric divergence (Figs. 7C, 7D, 7A, 11B, 11H, and 11F [5]). The kinematics of deformation are similar to those associated with the Nordfjord-Sogn detachment zone and the local “backsliding” of the orogenic wedge (Fig. 11A [5]) inferred from top-W fabrics within the Jotun detachment (Fossen and Dunlap, 1998). Vertical thinning of the margin crust is also consistent with amphibolite-facies subhorizontal E-W stretching lineations throughout the central Western Gneiss Region.

6. P-T paths from the leading edge of the procontinental margin in model f50 record broadly isothermal decompression during Phase 4, (U)HP conditions to ~0.6–1 GPa and 750–825 °C (Fig. 12B [6]), consistent with estimated conditions of widespread amphibolite-facies metamorphism of the Western Gneiss Re-
Exhumation of the Western Gneiss Region (ultra)high-pressure terrane, Norway

A. Schematic reconstruction at ~395 Ma

B. Model f50; 60 Myr-pc / ca. 397 Ma; $\Delta x_{\text{Ext}} = 50$ km

C. Present-day Sognefjord cross-section

D. Present-day Nordøyane cross-section

E. Model f50; End Phase 4; 65 Myr-pc / ca. 392 Ma; $\Delta x_{\text{Ext}} = 100$ km

Figure 11 (on this and following page).
Figure 11 (continued). Comparison of model f50 with structural and tectonic evolution of the Western Gneiss Region. Numbers 1–13 keyed to features discussed in the text. (A) Schematic reconstruction of the Norwegian Caledonides at ca. 395 Ma (modified from Milnes et al., 1997; Fig. 2A), corresponding to normal-sense ductile shearing along the Nordfjord-Sogn detachment zone (NSDZ; x–x′, panel I) and exhumation of the Western Gneiss Region through amphibolite-facies conditions. See Figure 2 caption for more details. Speculative region of former ultrahigh-pressure (UHP) metamorphism is based on distribution of UHP rocks further north in the Western Gneiss Region. Arrows, here and in other panels, show superimposed thrust-sense (black kinematic indicators) and normal-sense (white kinematic indicators) deformation. (B) Model f50 at 60 Myr-pc, Mid–Phase 4 (397 Ma from Fig. 3), showing exhumation of (U) HP margin crust through amphibolite-facies conditions. (C) Present-day cross section through the southern Western Gneiss Region (Sognefjord) modified from Milnes et al. (1997) (x–x′, panel I). “Former UHP?” region is based on extrapolation of UHP domains further north. (D) Schematic present-day cross section through the Nordøyane UHP domain, modified from Milnes et al. (1997) (y–y′, panel I) to illustrate the location of the UHP domain (Hacker et al., 2010) and a more representative cross section through the northern, wider part of the Western Gneiss Region. (E) Model f50 result at 65 Myr-pc, end Phase 4 (392 Ma from Fig. 3), showing final geometry of the model orogen after 100 km of plate divergence and exhumation. (F) Model f50 showing strain (square root of second invariant of deviatoric strain) accumulated during plate divergence and exhumation from 55 to 65 Myr-pc (402–392 Ma; Fig. 3). (G) Model f50 result (equivalent to panel E) showing total strain (square root of second invariant of deviatoric strain) accumulated during the entire model, i.e., through Phases 1–4. Note that the subducted procontinent margin crust is deformed, but the overall structure remains coherent. (H) Model f50 result at 65 Myr-pc (392 Ma from Fig. 3) showing peak metamorphic pressures (dynamic pressure) achieved by model particles throughout the model calculation. Note the smooth gradient in maximum metamorphic pressures within the procontinent margin beneath the detachment, corresponding to the Baltica basement of the Western Gneiss Region. (I) Schematic map of the Western Gneiss Region (WGR) illustrating key structures and locations of cross sections (panels A, C, and D). JD—Jotun detachment; LGF—Lærdal Gjende fault; HSZ—Hardangerfjord shear zone; MTFZ—Møre Tørendal fault zone; NSDZ—Nordfjord-Sogn detachment zone.
Figure 12. Comparison of model f50 with pressure-temperature-time (P-T-t) data from the Western Gneiss Region (WGR). See text for explanation of numbers 2, 3, 6, 7, and 9. (A) Selected pressure-temperature (P-T) estimates and paths from the Nordøyane (Butler et al., 2013b; Larsen et al., 1998; Root et al., 2005; Terry et al., 2000b), Sorøyane (Carswell et al., 2003; Root et al., 2005; Straume and Austrheim, 1999), and Nordfjord (Labrousse et al., 2004; Peterman et al., 2009; Schärer and Labrousse, 2003) ultrahigh-pressure (UHP) domains. (B) Representative P-T paths from model f50 compared with Western Gneiss Region data, showing a similar spread of peak P-T conditions depending on the position of the particle within the procontinent margin. Peak conditions recorded by the leading edge of the margin (bracketed by particles a and b) are consistent with maximum Western Gneiss Region P-T estimates from the Nordøyane UHP domain. (C) Comparison of model f50 paths with selected pressure-temperature (P-T) estimates for the Nordøyane (Krogh et al., 2011; Root et al., 2005; Terry et al., 2000b; Tucker et al., 2004), Sorøyane (Carswell et al., 2003; Kylander-Clark et al., 2008; Root et al., 2005), and Nordfjord (Root et al., 2004, 2005; Young et al., 2007) UHP domains, showing general agreement on the timing of UHP metamorphism and exhumation based on Figure 3. (D) Comparison of model f50 paths with selected temperature-time (T-t) estimates from the Western Gneiss Region UHP domains, showing similar cooling from peak conditions through ~450–350 °C by ca. 375 Ma for particles near the surface of the procontinent margin. Dashed box shows range of ages and crystallization temperatures of Scandian titanite from Spencer et al. (2013; SP13). Particle b remains buried in the crust and is not exhumed. (E) Comparison of representative P-T paths from other models presented in the text with selected Western Gneiss Region data. Paths from models with weak procontinent crust that detaches prior to Phase 4 extension reach relatively low temperatures at UHP conditions but are within the range of the lowest temperature estimates from Nordfjord. (F) Comparison of P-T paths from other models with Western Gneiss Region data. Models with weaker procontinent crust detach earlier than model f50, during Phase 3 quiescence, leading to exhumation that is too early when compared with the Western Gneiss Region.

(8) As in the Western Gneiss Region, later Phase 4 extension leads to frictional-plastic (i.e., brittle) reactivation of amphibolite-facies structures, including the Nordfjord-Sogn detachment zone–equivalent normal-sense shear zone formed above the exhumed margin crust (Figs. 11C and 11F [8]), although in model f50, this reactivation is limited to the hinterland region above the UHP margin (Fig. 11F [8]).

(9) The model (UHP) margin crust close to the top cools through ~425–350 °C (mica closure) at ca. 380–385 Ma, consistent with ca. 375–370 Ma mica 40Ar/39Ar ages from the UHP domains (Fig. 12D [9]; Root et al., 2005).
The continuity of the procontinent margin crust in model f50 is consistent with the apparent absence of regional-scale structural and metamorphic breaks within the Western Gneiss Region (Figs. 11C, 11F, and 11H [10]).

The distribution of deformation within the procontinent margin in model f50 (Fig. 11G [11]), from relatively undeformed in the foreland to more pervasively deformed at the leading (U)HP end of the margin, is similar to the documented westward increase in the intensity of Scandian deformation within the Western Gneiss Region (Hacker et al., 2010).

Lack of detachment and thrusting of the (U)HP procontinent margin crust onto the procontinent in model f50 (Fig. 11F [12]) is consistent with the lack of evidence for a “basal” thrust beneath the (U)HP domains of the Western Gneiss Region (Fig. 11C [12]), except perhaps in the vicinity of the Stolri thrust (see following).

The “final” width of the model f50 orogen, ~300 km, is similar to the present-day width of the Caledonides, ~350–400 km, from the western edge of the Western Gneiss Region to the toe of the orogenic wedge in the foreland (Fig. 2, 11D and 11E [13]), although prior to erosion, the orogen was wider (Fig. 1B).

Some of the additional models described earlier herein clearly fail to reproduce one or more of these basic constraints. Models f1 and f10, with weaker procontinent margin crust, produce earlier detachment and buoyancy-driven exhumation. Consequently, in both models, the onset of exhumation is too early (Fig. 12F, particles d and e), and the peak temperatures at (U)HP conditions are too low (~575 °C and 600 °C, respectively) by comparison with the Western Gneiss Region data (Fig. 12E, particles d and e). In model f50-MW2, relatively low-T melting leads to early detachment and exhumation of relatively cool UHP margin crust. On the orogenic scale, the buoyancy-driven channel flow exhibited in these models leads to foreland-directed thrusting of UHP margin crust over lower-P margin crust, producing a clear “metamorphic break” (Figs. 9B and 9C) rather than the P-T gradient. We interpret the end of thrusting along the allochton–Western Gneiss Region contact (Jotun detachment) by ca. 408 Ma (the interpretation of Fossen and Dunlap, 1998), while UHP metamorphism continued to ca. 400 Ma (Carnwell et al., 2003), to indicate that the Western Gneiss Region resided at UHP conditions for at least ~8 Myr prior to the onset of exhumation. This is consistent with model f50.

As noted already, model f50-Vp1 represents an alternative slow subduction model, identical to model f50 except for a 1 cm yr⁻¹ convergence rate throughout Phase 2. At 55 Myr-pc in this model, the margin crust at UHP conditions remains colder (~650 °C) than it is in model f50 at the end of Phase 3 (also 55 Myr-pc). This demonstrates that advection dominates over diffusion even at convergence rates as slow as 1 cm yr⁻¹. We prefer model f50 because it results in peak UHP conditions similar to those observed in the Western Gneiss Region and meets the ca. 408 Ma end of thrusting criterion, whereas in model f50-Vp1, thrusting continues to a time equivalent to 400 Ma.

**How Long Was the Western Gneiss Region at Mantle Depths?**

Eclogites within the Western Gneiss Region have yielded a broad range of radiometric ages spanning ca. 420–400 Ma, implying that the Western Gneiss Region underwent slow subduction and/or protracted residence at (U)HP conditions (Kylander-Clark et al., 2009). Model f50 results represent an end-member case in which the Western Gneiss Region was rapidly subducted to, and then stagnated at, UHP conditions when plate convergence stopped. The corresponding prograde P-T path has the leading edge of the continental margin first reaching UHP conditions at temperatures of only 600 °C, reflecting the high Péclet number during underthrusting. Despite the initially cool conditions, metamorphic temperatures exceed 800 °C after only 10 Myr of incubation during Phase 3 quiescence. This reflects thermal diffusion at near-zero Péclet number plus radioactive heating of the deeply buried crust. Similar stagnation and incubation of the Western Gneiss Region are consistent with the span of ages attributed to (U)HP metamorphism and could explain why the Western Gneiss Region reached relatively high temperatures during UHP metamorphism (700–850 °C), by comparison with many other UHP terranes (Hacker, 2006).

Alternatively, subduction may have been slower than the 2.5 cm yr⁻¹ used in the models, or it may have slowed gradually. This would imply a hotter prograde P-T path with temperature increasing with pressure. However, in the absence of conclusive evidence for the prograde temperature path, we interpret the end of thrusting along the allochton–Western Gneiss Region contact (Jotun detachment) by ca. 408 Ma (the interpretation of Fossen and Dunlap, 1998), while UHP metamorphism continued to ca. 400 Ma (Carnwell et al., 2003), to indicate that the Western Gneiss Region resided at UHP conditions for at least ~8 Myr prior to the onset of exhumation. This is consistent with model f50.

Whether or not the Western Gneiss Region was deformed internally during subduction and exhumation remains controversial (Cuthbert et al., 2000; Terry and Robinson, 2003; Hacker et al., 2010; Robinson et al., 2014). The broad similarity in the timing and P-T conditions of metamorphism and cooling across the Western Gneiss Region, as well as the apparent lack of regional-scale shear zones/segmentation, has led to the view that it was exhumed as a coherent body of crust (Hacker et al., 2010). Nonetheless, parts of the Western Gneiss Region, in particular the UHP domains, clearly experienced intense deformation during exhumation...
(Terry and Robinson, 2003, 2004) The degree of deformation decreases considerably toward the east, where much of the Baltic crust preserves pre-Scandian mineral assemblages and fabrics, implying limited reworking (Hacker et al., 2010; Spencer et al., 2013). Does Scandian deformation simply fade eastward, or does the Western Gneiss Region consist of multiple tectonic units that were juxtaposed during subduction and/or exhumation (cf. Cuthbert et al., 2000)? As yet, no obvious thrust-sense shear zones have been found to separate the UHP and HP parts of the Western Gneiss Region (Young et al., 2007), and the common view is that the transition represents decreasing metamorphic grade or perhaps kinetic factors (Cuthbert et al., 2000). However, in the northern Western Gneiss Region, a thin slice of HP Baltic basement is interpreted to have been thrust eastward over lower-pressure crust along the Storli thrust during exhumation (Gee et al., 2013; Robinson et al., 2014). Internal thrusting within the Western Gneiss Region basement could represent an exhumation mechanism like that in model f1 (Fig. 9B), where the UHP margin crust detaches from the slab and exhumes buoyantly, driving thrusting over the adjacent lower-deformed HP part of the margin. However, the timing of thrusting remains poorly constrained. Eclogite-facies metamorphism within the Storli thrust sheet appears to have occurred ca. 425 Ma (Beckman et al., 2014). We therefore speculate that the northern Western Gneiss Region may record an early phase of synorogenic overthrusting and possible exhumation of the Baltic margin that predated the main phase of exhumation during post-Scandian extension.

If the coherent slab interpretation applies to most of the Western Gneiss Region, with possible exceptions locally, it follows from the model results that the Baltic basement orthogneisses that dominate the Western Gneiss Region were exceptionally strong throughout subduction and exhumation. This conclusion is consistent with the study of Spencer et al. (2013), who found that pre-Scandian titanite, and by implication plagioclase, must have remained metastable during the 25–40 Myr. Scandian orogenic event at UHP conditions and temperatures as high as 750 °C. That these minerals remained metastable implies that there was little or no deformation and recrystallization. Similarly, Hacker et al. (2010) showed that much of the eastern Western Gneiss Region experienced only minor Scandian deformation. The Western Gneiss Region dry amphibolite- and granulite-facies gneisses were clearly stronger and less reactive than their lower-grade hydrous counterparts (Krabbendam et al., 2000).

To remain nondeforming in the models, the continental margin requires a WQ × 50 (f = 50) flow law, which is consistent with exceptional strength, resulting in effective viscosities similar to those predicted for much stronger feldspar-dominated rocks, and likely near the upper end of possible viscosities for intermediate to felsic crustal rocks (Fig. 13). What could have made the Baltic margin so strong? Effective rock/mineral viscosities, and therefore ductile deformation rates, are strongly dependent on the presence of intracrystalline water (Hirth et al., 2001; Rutter and Brodie, 2004; Rybacki et al., 2006). Prior to the Scandian orogeny, Baltic crust underwent polyphase Proterozoic (Sveconorwegian) high-T metamorphism (e.g., Bingen et al., 2008; Spencer et al., 2013; Corfu et al., 2014), which likely left it dry and therefore strong. Scattered throughout the Western Gneiss Region are examples of metastable granulite-facies rocks (Austreheim, 1987; Krabbendam et al., 2000; Wain et al., 2001; Peterman et al., 2009; Spencer et al., 2013) that were deformed and transformed to Scandian eclogite- or amphibolite-facies assemblages only where infiltrated by fluid. Clearly, the amphibolite-facies orthogneisses that now dominate the Western Gneiss Region were hydrated at some stage. If the Western Gneiss Region was exhumed as a coherent slab because it was dry and strong, hydration must have occurred relatively late in its history, after exhumation from UHP conditions but prior to or during widespread amphibolite-facies deformation. This is also consistent with regional titanite U-Pb data (Spencer et al., 2013) that indicate that, locally, parts of the Western Gneiss Region remained unreacted and by implication undeformed throughout Scandian orogeny.

Did Partial Melting “Trigger” Exhumation of the Western Gneiss Region?

Recent studies have raised the question whether the onset of partial melting of the Western Gneiss Region at (U)HP conditions may have triggered exhumation by decreasing the viscosity and increasing the buoyancy of the subducted crust (Labrousse et al., 2011; Gordon et al., 2013; Ganzhorn et al., 2014). We cannot determine whether melting at peak UHP conditions affected the Western Gneiss Region, but the models can be used to assess whether melting triggered exhumation. The geological evidence summarized here suggests that the Baltic margin remained at UHP conditions for some time (~10 Myr) and, in addition, that the Baltic crust appears to have been exhumed as a coherent slab except in the vicinity of the Storli thrust. Both of these are consistent with model f50, which requires the Baltic crust to have been very strong (f ~ 50) while at peak UHP conditions to prevent decoupling and buoyant exhumation (Fig. 6B). In models with a much lower f (1, 10), the weaker precontinental upper crust decouples, and the UHP region rises as an independent plume that diverges from the adjacent HP region (Fig. 9). This behavior is also consistent with model f50-MW2, in which melting of partly hydrated crust results in melt weakening at UHP conditions and the development of a similar plume of buoyantly rising UHP crust (Figs. 10C and 10D). Given that there is no evidence of an independent plume of exhumed UHP crust in the Western Gneiss Region, or of coeval late-orogenic thrust- and normal-sense shearing, we conclude that models with f less than 25 or 50 or with significant melt weakening at peak UHP conditions, both of which lead to...
early detachment and buoyancy-driven exhumation, must be rejected.

Given that extensive melting did affect the Western Gneiss Region crust, the solution appears to be that minor melt weakening at near-peak UHP conditions in the deepest part of the crust, as seen in model f50-MW1, was not sufficient to trigger decoupling of the crust and its subsequent exhumation as an independent plume (Fig. 10B). The locally extensive melting recorded in Western Gneiss Region migmatites, in particular, within the UHP domains (Hacker et al., 2010; Labrousse et al., 2011; Butler et al., 2013a; Gordon et al., 2013), likely occurred during isothermal decompression at \( P \leq 2.5 \) GPa (Auzanneau et al., 2006). In model f50-MW1, which predicts decompression melting, exhumation is still dominated by extension driven by plate divergence, and therefore the overall orogen-scale style of deformation in model f50, i.e., normal-sense shearing at the top of the exhumed margin crust and negligible Foreland-directed thrusting, is maintained in model f50-MW1. In summary, the model results suggest that melting may have assisted exhumation of the Western Gneiss Region but that its effect was ultimately subordinate to plate extension, as in model f50-MW1 (Figs. 10A and 10B) and in contrast to model f50-MW2 (Figs. 10C and 10D).

What Caused the Extension?

The conclusion that the Western Gneiss Region was exhumed during postorogenic extension is not new (e.g., Fossen, 1992), although the underlying cause of this extension remains debated. The question is whether the observed extension results from reversed subduction or divergence of the retrolithosphere. Andersen et al. (1991) introduced the term “eduction” (which they modified from the original geological definition of Dixon and Farrar [1980], eduction definition 1) to describe exhumation of the Western Gneiss Region. They described eduction as “footwall uplift associated with extensional straining of the middle and upper crust in the hanging wall” and concluded that “eduction occurs principally by a footwall rolling hinge mechanism” (eduction definition 2, p. 306).

They also estimated that there was 180 km of extension in the Western Gneiss Region to achieve the eduction. Their diagram (Andersen et al., 1991, their Fig. 2) indicates that the Baltic crust remains attached to its underlying lithosphere and that extension was achieved by divergence of the retrolithosphere (upper plate; their Fig. 2B).

Brueckner and van Roermund (2004) also appealed to subduction/eduction cycles to explain multiple (U)HP events in the Scandinavanian Caledonides. However, their definition of “eduction” (eduction definition 3) is equivalent to the detachment and buoyant rise of a large nondeforming slice of crust within the subduction zone, like the Chemenda et al. (1995) mechanism, and requires a basal detachment, which they agreed has not been identified below the Western Gneiss Region.

More recently, the term “eduction” has been used explicitly to imply reversal of subduction (Duretz et al., 2012; eduction definition 4) such that in our models, it would require reversal of the retrolithosphere trajectory. This mechanism was shown to be feasible after slab breakoff has removed the downward slab-pull force and the residual buoyancy of the subducted plate reverses the subduction (Duretz et al., 2012).

To avoid confusion, we adopt the Duretz et al. (2012) definition of eduction, the reversal of subduction of the entire lithosphere driven by buoyancy forces following slab breakoff. Bottrill et al. (2014) also demonstrated the Duretz et al. (2012) style of eduction for both normal and diachronous convergence with plate rotation and applied their three-dimensional numerical model results to the Western Gneiss Region.

Instead of eduction (definition 4), we favor an alternative explanation, plate-scale divergence, the product of sinistral transtension between Baltica and Laurentia (Fossen, 1992; Krabbendam and Dewey, 1998), with an orogen-normal component of ~100 km (Dewey and Strachan, 2003). This is the process we have used in models presented here in which the procontinent lithosphere remains stationary while the retrolithosphere plate moves away slowly \((V_t = -1 \text{ cm yr}^{-1})\) in Phase 4. Divergence of 100 km, coupled with surface erosion, is sufficient to achieve exhumation of the Western Gneiss Region that is consistent with observations. Despite the portrayal of the Andersen et al. (1991) model as eduction (definition 4), our interpretation is that their Figure 2 describes the same mechanism we advocate, i.e., plate-scale divergence.

Deep Crustal Subduction or Tectonic Overpressure?

It has been suggested that pressures recorded by metamorphic rocks do not represent near-lithostatic pressure conditions. Instead, the mean stress (dynamic pressure) may deviate from the lithostatic pressure significantly under certain circumstances (e.g., Petrić and Podladchikov, 2000; Mancktelow, 2008; Li et al., 2010). If so, the observed pressure cannot be equated with depth of burial. For example, tectonic overpressure (mean stress − lithostatic pressure) may result from large compressive stresses during continent-continent collision. If tectonic overpressure exists in UHP rocks, it reduces the requirement for deep burial and makes the interpretation of the observed pressure and corresponding tectonics equivocal.

We cannot determine whether tectonic overpressure exists in Western Gneiss Region basement rocks. However, we can determine pressures in the models. The tectonic overpressure in model f50 at the end of Phase 2 (Fig. 14), when overpressure is close to its maximum, shows regions of both over- and underpressure. Overpressure reaches up to ~0.75 GPa in frictional-plastic regions (Fig. 14B, cross-hatched areas) or where strong viscous regions of the lithosphere are subject to bending during subduction. However, despite the high crustal strength (e.g., \( f = 50 \) upper crust), there are no regions of significant (>0.1 GPa) overpressure in either the upper or lower subducted crust. This is similar to the results of Li et al. (2010), who showed that relatively minor pressure deviations (~0.3 GPa) develop when the crust is relatively weak and the subduction conduit does not taper along its length.

Tectonic overpressure is expected in frictional-plastic regions because the mean stress is given by \( P(z) = f (1 + \sin \phi) \), where \( P(z) \) is lithostatic pressure (Petrić and Podladchikov, 2000). In our models, \( \phi = 15^\circ \), implying maximum mean stress at a given depth in the model will be \( \sim 1.35P_L(z) \), an overpressure of ~0.35\( P_L(z) \). In this respect, our models results agree with those of Schmalholz and Podladchikov (2013) in that overpressure develops in the strong frictional-plastic regions bounding weak shear zones. However, unlike their results, there is no significant overpressure in the “weak” crustal regions in our models, even though the crust is actually quite strong, e.g., model f50.

Our preferred interpretation of the Western Gneiss Region is that crustal pressures were close to lithostatic and are a consequence of deep burial, as seen in the models. However, the Western Gneiss Region also offers an excellent opportunity to test for the possibility of tectonic overpressure in the Baltic crust. Mechanisms considered responsible for tectonic overpressure range from microscale (e.g., Moulas et al., 2014) to regional (e.g., Schmalholz et al., 2014). For example, a local overpressure mechanism has been invoked to explain anomalously high pressure estimates from mantle peridotite bodies in the Nordstjane UHP domain (e.g., Vrijmoed et al., 2009). If similar microscale-to-local overpressure mechanisms are responsible for all of the UHP metamorphism, we would expect each case to record its local circumstance, and therefore overpressures will vary significantly from one locality to another. In contrast, there...
cannot be local variations in burial depth; therefore, burial at near lithostatic pressure will produce a spatially smooth variation in maximum pressure. It follows that a detailed systematic study of pressure variations should distinguish between these two mechanisms. The only case that cannot be distinguished in this manner is a regional overpressure mechanism that mimics the effects of burial, that is, a smooth variation of recorded pressures at the scale of the orogen.

PARADIGM REGAINED: BUOYANCY VS. THE EXHUMATION NUMBER

There is considerable diversity in the sizes (i.e., areal extent and thickness) and exhumation rates of UHP terranes worldwide. At one end, there are relatively small (<10³ km²) terranes, like those in the Western Alps and Himalayas, that record short-duration (~5 Myr) “fast” UHP metamorphism followed by very rapid exhumation (3–6 cm yr⁻¹). The Western Gneiss Region represents the other end member; one of the relatively few examples of extremely large (~30 × 10³ km²) UHP terranes that underwent protracted “slow” metamorphism and exhumation (~10–20 Myr). Kylander-Clark et al. (2012, their Table 1 and Fig. 1) classified these types and attributed the differences to the timing of UHP terrane formation with respect to orogenic stage and subduction angle. They hypothesized that small/fast UHP terranes form earlier in collision when subduction is steep, and that large/slow terranes form later in collision when subduction is shallower owing to the buoyancy of the continental lithosphere. While we agree that many small/fast (U)HP terranes formed during the early stages of collision, our models do not support the interpretation that subduction dip is a major factor. Instead, it appears that high crustal strength is critical, such that large/slow (U)HP terranes can be subducted to UHP conditions and held there for prolonged intervals.

Exhumation of the small/fast Western Alpine and Himalayan UHP terranes has been explained by within-orogen extension during continued convergence in response to detachment and buoyancy-driven flow of (U)HP crust upward along the subduction channel (e.g., Beaumont et al., 2009; Butler et al., 2013a). However, if exhumation is driven by crustal buoyancy, why was the Western Gneiss Region—a large buoyant volume of (U)HP crust—exhumed so slowly? A partial answer, suggested by Kylander-Clark et al. (2012) and supported by our model results, is that the Baltic crust was sufficiently strong that it remained coupled to the underlying slab for the entire duration of subduction and post-Scandian extension.

A more fundamental solution to the problem is that the current paradigm—that exhumation of UHP rocks is controlled by buoyancy—is incomplete and needs to be modified (Hacker and Gerya, 2013). The paradigm can be “regained” by recognizing that exhumation is controlled by the exhumation number, which takes both buoyancy and other factors into account. Following Raimbourg et al. (2007), Warren et al. (2008b), and Butler et al. (2014), for example, the exhumation number, \( E_x \), is defined by the competition between the buoyant up-channel Poiseuille flow of the low-density crust and the down-channel Couette flow resulting from the downward traction exerted on the crust by underlying subducting lithosphere, the numerator and denominator in the following force balance, respectively:

\[
E_x = h^3 (\rho_c - \rho_m) g (\sin \gamma) / \eta_m U,
\]

where \((\rho_c - \rho_m) g (\sin \gamma)\) is the effective down-channel pressure gradient, with \(x\) measured in the downdip direction, \(\gamma\) is the dip, and \(\rho_c\) and \(\rho_m\) are the crust and mantle densities. \(U\) is the subduction velocity of the underlying lithosphere, and \(\eta_m\) is the effective viscosity of the crust in the channel. The term \(h^3\) is a measure of the channel thickness estimated when \(E_x < 1\), i.e., for a stagnant channel; \(E_x > 1\) gives burial, and \(E_x > 1\) gives exhumation. Certainly, exhumation requires that the crust is positively buoyant, but this is not sufficient. The point is that many small/fast (U)HP terranes formed during the early stages of collision, our models do not support the interpretation that subduction dip is a major factor. Instead, it appears that high crustal strength is critical, such that large/slow (U)HP terranes can be subducted to UHP conditions and held there for prolonged intervals.

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These contrasting mechanisms are best distinguished by their characteristic patterns of orogen-scale deformation (Fig. 15). Consider exhuma-
Figure 15. Contrasting exhumation mechanisms and their relationship to the exhumation number, $E_x$. (A) Schematic representation of continental subduction/underthrusting, showing burial of a procontinent margin to ultra-high-pressure (UHP) and high-pressure (HP) conditions beneath an orogenic wedge, corresponding to Phase 2 margin underthrusting (Fig. 6A). The strength of the margin is sufficient to overcome the upward buoyancy force acting on the crust, i.e., $E_x < 1$, and the crust therefore remains coupled to the slab. (B) Exhumation type 1, showing upward, buoyancy-driven flow of a plume of UHP crust (e.g., Beaumont et al., 2009; Butler et al., 2013a). Weakening at depth (e.g., strain, reaction, or melt weakening) has reduced the traction exerted on the crust by the subducting procontinent slab. The upward buoyancy overcomes the traction, and the plume flows upward. Insertion of the plume into the orogenic wedge is accommodated by normal-sense shearing and extension above the plume, and thrusting at its base. Thrusting in the foreland is likely to continue during exhumation provided convergence has not ceased. The contact between the plume and the margin is marked by a metamorphic break. The plume may be small, as in the Alps and Tso Morari, but could be much larger. (C) Exhumation type 2, plate-scale extension, corresponding to model f50. In this case, the margin crust is sufficiently strong to resist detachment, and it remains buried at UHP-HP conditions until it is exhumed by extension resulting from plate divergence (here shown as leftward motion of the retrolithosphere, as in the models). Deformation is dominated by hinterland-dipping normal-sense shear zones. In contrast with type 1, in which extension of the orogenic wedge is localized, here the entire orogenic wedge may undergo extension. (D) Exhumation type 3, corresponding to the Chemenda et al. (1995) mechanism that some have proposed for the Western Gneiss Region. In this conceptual model, the procontinent margin crust detaches at depth and exhumes as a rigid slab of crust along coeval normal- and thrust-sense shear zones. Unlike the plume mechanism (type 2), internal deformation of the exhuming slab is considered to be minor.
mation of crust buried in a subduction channel/ conduit as shown in Figure 15A. Exhumation by the $E > 1$ paradigm, and in the form of a plume, produces thrusting at the base and front of the exhuming plume, coupled with normal-sense shearing at its top, and extension of the overlying crust (Fig. 15B). Plumes can be both larger and smaller than the example shown; they just require $E > 1$. By comparison, if the UHP crust remains coupled to the slab, $E < 1$, there will be no buoyant exhumation of detached crust. However, exhumation of the intact UHP crust can be accomplished completely during plate divergence (Fig. 12C) without shortening in the foreland of the orogen or within the terrane, and there is no basal thrust detachment below the terrane. A third possibility is the Chemenda et al. (1995) mechanism in which a large slice of (U)HP crust achieves $E > 1$, decouples, and exhumes buoyantly by overthrusting (Fig. 15D). This crustal slice remains largely undeformed in the Chemenda et al. (1995) models, but it could be accompanied by deformation. Hacker et al. (2010, their Fig. 13) have made a Chemenda-type interpretation for the Western Gneiss Region, whereas we prefer exhumation by extension. An associated question is whether any of these mechanisms explains other examples of large/slow UHP terranes?

As noted by Kylander-Clark et al. (2012), both the small-fast and large-slow mechanisms may operate in the same orogen at different times. This happens in the models described herein. Evidence of pre-Scandian (U)HP metamorphism in several of the allochthons overlying Baltic basement suggests this may have been the case in the Caledonides (Brueckner and van Roermund, 2004). Early Caledonian UHP metamorphism and exhumation probably accompanied allochthon accretion during continuous subduction, requiring detachment and buoyancy-driven flow of small volumes of UHP crust, and implying $E > 1$. Later subduction resulted in UHP metamorphism of the strong Baltic margin crust, with exhumation only during plate divergence, i.e., $E < 1$.

It is also possible that some large and slow UHP terranes may detach and be exhumed buoyantly, $E > 1$, in the absence of plate divergence or extension. For example, in model f50-MW2, a large volume of melt-weakened UHP margin crust detaches from the slab and exhumes buoyantly. Ascent of the plume is initially rapid, but it slows dramatically as it reaches the base of the orogenic wedge, which acts as a low-density “lid,” as envisaged by the Moho “arrest” conceptual model of Walsh and Hacker (2004) and Kylander-Clark et al. (2009). The resulting time span for burial and exhumation in model f50-MW2 is ~30 Myr, making it a relatively slow and large UHP terrane with $E > 1$ and buoyancy-driven flow. However, the geometry and tectonic style of this type of model are not like the Western Gneiss Region. Exhumation by plate divergence provides a better explanation in this case.

CONCLUSIONS

1. The models presented here reproduce the key features observed in the Western Gneiss Region of the Norwegian Caledonides, including the distribution of $P$-$T$ conditions and duration of metamorphism, and the kinematics of pre-Scandian and later exhumation-related deformation.

2. Formation and exhumation of the Western Gneiss Region can be explained by a four-phase evolution including: (a) allochthonous terrane accretion and growth of a thick orogenic wedge; (b) underthrusting/subduction of the Baltic continental margin to UHP conditions beneath this wedge; (c) cessation of subduction with corresponding quiescence, gravitational spreading, and incubation of crust at UHP conditions; and (d) exhumation accomplished by orogen-wide extension during plate divergence.

3. UHP metamorphism in the models results from burial of the continental margin to depths of ~130 km. Despite the high viscous strength of the crust, overpressures within the margin are negligible. We therefore propose that UHP metamorphism of the Western Gneiss Region documented in widespread UHP-HP eclogites represents regional metamorphism related to burial of the Baltic continental margin beneath Laurentia, and not tectonic overpressure, although local overpressures cannot be ruled out.

4. Exhumation of the Western Gneiss Region by extension during plate divergence is consistent with the apparent lack of late-to-post-Scandian thrusting at the orogen scale and within the Western Gneiss Region, and the associated lack of an identified basal thrust below the Western Gneiss Region. Rather than decoupling from the subducted lithosphere and exhumation as a viscous plume, the Western Gneiss Region appears to have remained coupled to the subducted slab at least until convergence ended.

5. The extended duration of Scandian (U) HP metamorphism can be explained by the high strength of the Baltic margin crust, which allowed it to remain coupled to the subducted lithosphere.

6. The models indicate that plate divergence is a viable mechanism for exhumation of large (U)HP terranes. For the Western Gneiss Region, the models suggest that only ~100 km of divergence is required; in nature, this may have been related to absolute Baltica-Laurentia transtensional divergence. If correct, there is no need to invoke reverse/inverse subduction mechanisms like eduction.

7. We propose that the simple “buoyancy-driven exhumation” paradigm for UHP terranes be replaced by one that attributes control of the system to the exhumation number (ratio of the buoyancy force to the basal traction).

APPENDIX A: MODELING METHODS

A1. SOPALE Nested

We used two-dimensional thermal-mechanical models computed using the Arbitrary Lagrangian-Eulerian software (Aoude et al., 2009) to investigate the dynamics of Caledonian-type orogens at the mantle scale. The methodology is similar to that of Butler et al. (2011, 2014). The models are computed by solving the equations for incompressible creeping (Stokes) flows (Eq. A1 and A2), energy balance (Eq. A3), and viscous-plastic constitutive relations that determine stress and effective viscosity (Eq. A4 and A5), on an Eulerian grid subject to mechanical and thermal boundary conditions:

\[ \sigma_{ij} = 0, \quad j = 1, 2. \]  
\[ \frac{\partial T}{\partial t} = c_p \rho \frac{\partial}{\partial x_i} \left( \frac{\partial T}{\partial x_i} + A_T + \nu \frac{\partial T}{\partial x_i} \right) \]  
\[ \frac{\partial}{\partial x_i} \left( \frac{\partial T}{\partial x_i} + A_T + \nu \frac{\partial T}{\partial x_i} \right) = 0, \quad j = 1, 2. \]  

where $c_p$ is the specific heat capacity, $\rho$ is density, $g$ is gravitational acceleration, $\lambda$ is a component of velocity, $T$ is temperature, $t$ is time, $K$ is thermal conductivity, $A_T$ is crustal radiogenic heat production per unit volume, $A_R$ represents shear heating, and $\nu$ is volumetric thermal expansion. Coupling of the mechanical and thermal solutions is achieved through the advection of radioactive crust, shear heating, thermal activation of viscous flow, and buoyancy forces arising from metamorphic phase changes and thermal expansion.

The models use a subgrid approach in which a higher-resolution computational domain is embedded or “nested” within a larger, lower-resolution domain. The velocity and thermal solution is first obtained for the full 4000 km x 1200 km domain. This solution is then interpolated onto the boundaries of the (2 km x 2 km resolution) nested domain encompassing the vicinity of the subduction zone, from 400 to 2000 km and 140 km deep. Coupling of the two solutions is maintained by using a single cloud of Lagrangian tracking particles, which is passed between the two solutions, with the particles within the nested domain always obeying the higher-resolution solution, making this a “two-way” subdomain model.

A2. Model Design

All models have the same geometry (Fig. 4) and are designed to represent a generic Norwegian Caledonian-type subduction-accretion system. The subducting plate consists of oceanic lithosphere, two microcontinents (outer microcontinent and inner microcontinent), and a procontinent with a rifted continental margin, whereas the upper plate is a laterally homogeneous retrocontinent. The outer microcontinent and inner microcontinent are separated by outer and inner rift basins. The composite margin with microcontinents and rift basins is a general representation of the hyperextended margin envisaged by Andersen et al. (2012). For simplicity, we assume subduction and accretion of terranes occur at a single subduction zone at the margin of the retrocontinent. The continental margins and microcontinents each have relatively weaker upper and stronger lower crusts with maximum total thicknesses of 36 km and 30 km, respectively. The crustal units are either capped by a 4- or 6-km-thick weak sediment cover layer for the continental margins and microcon-
frictional-plastic deformation is modeled using the Drucker-Prager yield criterion:
\[ \sigma = (\sigma' / f) + C \sin \phi + C \cos \phi \sin \phi \]
where \( \sigma' \) is the viscosity scaling factor, \( W \) is a strain-weakening factor, \( f \) is the pre-exponential factor converted to the tensor-invariant form, \( \phi \) is the second invariant of the deviatoric stress, \( n \) is the strain-weakening power-law exponent, \( Q \) is activation energy, \( P \) is pressure, \( V \) is the activation volume for power-law creep, \( T \) is absolute temperature, and \( R \) is the universal gas constant. For simplicity, the flow-law parameters for the model materials are based on a small set of laboratory-determined flow laws (Table 1).

Crustal materials and the lithospheric mantle soften/weaken as they respectively deform in the frictional-plastic and viscous regimes. As noted earlier, strain-softening of frictional-plastic materials occurs through a finite decrease in the effective angle of internal friction, \( \phi \), over a specified range \( r \). Viscous strain-weakening occurs through a linear decrease in effective viscosity by the factor \( W \) over a specified range, generally \( W = 10 \) over the range \( r = 5-10 \) (Butler et al., 2013a). Strain-softening/weakening, as implemented in our models, depends on the resolution of the computational grid (2 km x 2 km in the nested region). Given that natural shear zones may consist of an equivalent natural strain-softened/weakening occurs at smaller offsets across shear zones because the shear zones are narrower than the typical 6 km in these models. It follows that the models soften/ weaken more rapidly and have lower effective viscosity than the laboratory flow laws across this range.

Flow laws for rocks that are stronger or weaker than the reference \( WQ \), with no water fugacity effect (dry quartz, DQ) corresponds to \( f = 0.5 \) and \( f = 1 \), respectively (Mackwell et al., 1998). Mante materials have a wet olivine flow law (Karato and Wu, 1993) scaled to represent either "wet" sublithospheric upper mantle and oceanic lithospheric mantle (WOL x \( f = 2, W = 1 \), i.e., no weakening), or relatively dehydrated "dry" continental lithospheric mantle (WOL x \( f = 75, W = 1 \)). Owing to the large range of uncertainties in the values of the effective viscosity by the factor \( (f = 6, 10, 25, and 50) \) respectively. By increasing the strength among models, we test the assumption that it reduces the stress/strain weakening in the continental crust, analogous to Proterozoic granulite-facies basement rocks that dominate the Western Gneiss Region, and is therefore stronger than the reference \( WQ \). The retrocontinental rocks that dominate the Western Gneiss Region, and is therefore stronger than the reference \( WQ \) (Searle, 2008). Therefore, the retrocontinental rocks are analogous to Proterozoic granulite-facies basement rocks that dominate the Western Gneiss Region, and is therefore stronger than the reference \( WQ \).

Crustal materials undergo pressure- and temperature-dependent density changes at the eclogite and coesite-eclogite phase transitions. The densities of the continental crust at low, high, and ultrahigh pressures are 2800, 2900, and 3000 kg/m³, respectively. The density change is driven by the transformation of olivine to ringwoodite in the mantle transition zone (~440–660 km). The fractional volume change accompanying a phase change is small, and its effect on the velocity field is negligible. The fractional volume change corresponding to the eclogite to ringwoodite transformation of olivine to ringwoodite in the mantle transition zone (~440–660 km).

Given that relative ductile flow of different materials in the models is mainly a consequence of their viscosity contrast, the simple scaling guarantees that the viscosity contrast is always given by the scaling factor under the same ambient conditions. This approach simplifies the interpretation of the model results and is the principal reason for choosing it. Instead of having results in which all of the parameters in the power-law creep flow law vary (Eq. A5), only the effective viscosity varies as \( f \) is varied, and it varies by \( f \). We believe that this scaling is an appropriate way to test the sensitivity of the perturbed mantle to wet versus dry conditions or to a moderate change in composition that can be expected given the heterogeneous nature of the lithosphere. The effect of scaling can be seen in Figure 13, where we compare the effective viscosity versus temperature predicted by the reference (dry quartz, DQ) flow laws with those based on laboratory flow laws for a strain rate of 10⁻⁴ s⁻¹. Within the range of uncertainties, the Hirih et al. (2001) quartz flow law with no water fugacity effect (dry quartz, DQ) corresponds to \( f = 0.5 \) for crustal velocities. The other lower mantle boundary conditions, the effect on the velocity field is minor because it only applies at the time of the phase change. The fractional volume change corresponding to the eclogite to ringwoodite transformation of olivine to ringwoodite in the mantle transition zone (~440–660 km).

The fractional volume change corresponding to the eclogite to ringwoodite transformation of olivine to ringwoodite in the mantle transition zone (~440–660 km).

A5. Density, Volume, and Mass Conservation during Phase Transitions

Crustal materials undergo pressure- and temperature-dependent density changes corresponding to the eclogite and coesite-eclogite phase transitions corresponding to the eclogite and coesite-eclogite phase transitions. The latter transformation is determined by a reaction with a Clapeyron slope of ~2.0 MPa/K and results in a density increase (proportional for all mantle materials) of ~6% over a fixed pressure range at a constant temperature. We account for the volume change and its effect on the buoyancy and velocity field. The volume change is calculated as a separate numerical iteration at each time step by applying both the density variations and the velocity field.

A6. Boundary Conditions

The model domain has a stress-free upper surface, and no-slip and free-slip sides and base, respectively. Thermal boundary conditions include a basal heat flux of 21 mW/m², 0 °C at the model surface, and insulated side boundaries (Fig. 4).

Subduction initiates dynamically along a small slice of oceanic crust embedded in the lithosphere at the boundary between the oceanic and continental lithosphere.
The models in the basic set evolve through: Phase 1, oceanic subduction and microcontinent accretion; Phase 2, continental margin subduction; Phase 3, quiescence; and Phase 4, extension. During Phases 1 and 2, the lithospheric boundary velocity is initially $V_r = -5$ cm yr$^{-1}$ (negative velocity indicates motion to the left), decreasing to $V_r = -2.5$ cm yr$^{-1}$ following accretion of a microcontinent with the retrocontinent, while the retrocontinental lithosphere boundary velocity is $V_r = 0$ cm yr$^{-1}$. In Phase 3, both lithospheric boundary velocities are 0 cm yr$^{-1}$. In Phase 4, divergence occurs at $V_r = -1.0$ cm yr$^{-1}$, while $V_r = 0$ cm yr$^{-1}$. The model phases correspond to our working hypothesis that the Western Gneiss Region was subject to both plate convergence and divergence, with a possible intervening quiescent phase.

We also monitor the tectonic forces applied to the model boundaries by the velocity boundary conditions. In model f50, the boundary forces on the prolithosphere (Baltica) range from $-290$ to $-2000$ GN m$^{-2}$, and those acting on the retrocontinua (Laurentia) range from $-500$ to $-2500$ GN m$^{-2}$, where positive forces are tensile. We consider these values to be within the range of expected plate-driving forces in nature, which range from $-2500$ to $-4000$ GN m$^{-2}$ for effective slab pull.

A2 Surface Processes

The models have an upper free surface, and the model topography is a product of the model calculation without any imposed constraints except for water loading where the surface is below sea level. Surface uplift and subsidence are mostly the result of crustal thickening and downward flexural loading near the subduction zone, and the density changes associated with the phase changes have relatively little influence on the topography. Slope-dependent surface erosion acts on the local subduction topography. It varies spatially according to local slope of the model surface, and it is defined by a maximum erosion rate, $E_m$, that operates on a slope of 45° and is scaled down linearly to the local slope of the surface. It results in nominal erosion rates of $-0.5$–$1$ mm yr$^{-1}$ during Phases 1 and 2. During Phases 3 and 4, slope-dependent erosion rates increase somewhat, but the maximum is limited to $-2$ mm yr$^{-1}$. This slope-dependent erosion is a simple proxy for erosion rates increase somewhat, but the maximum is limited to $-2$ mm yr$^{-1}$. This slope-dependent erosion is a simple proxy for tensional stress: Tectonophysics, v. 152, no. 1–4, p. 251–222, doi:10.1016/S0040-1951(98)00518-0.

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