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Subduction of continental crust, the origin of post-orogenic granitoids (and anorthosites?) and the evolution of Fennoscandia

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Abstract: Slices of silica-rich continental crust subducted into the mantle during collision may undergo metamorphism and exhumation towards the surface as coherent high-pressure or ultrahigh-pressure (HP or UHP) terranes or, if stalled in the mantle, melting and return towards the surface as magmas, or a combination of these two processes. Some exposed HP or UHP terranes contain anatectic granitoids demonstrating that melting does occur during exhumation. Therefore crust trapped in the mantle should also melt when radioactive heating and/or conductive heating raise temperatures to the appropriate solidus. Terranes with hydrous phases will melt readily through hydrate-breakdown reactions. Terranes lacking hydrous phases may require adiabatic decompression to melt, possibly as heated quartz-rich crust becomes ductile and rises diapirically. The magmas generated will intrude the overlying plate to form late-, post- and possibly anorogenic granitoids, depending on the time required to reach solidus temperatures. Geochemical characteristics will depend on P–T conditions, the chemistry and mineralogy of the subducted terrane (especially the presence of hydrous phases), and the amount of melt interaction with the mantle. The removal of sialic upper crust may strand the denser mafic lower crust, which subsequently could melt to generate anorogenic anorthosites, Fe–K granitoids and related rocks. The evolution of the Fennoscandian Shield documents a change from slab melting in the Mesoproterozoic to combined melting and exhumation in the Neoproterozoic to intact exhumation without significant melting in the Palaeozoic.

This paper investigates the simple idea that quartz-rich continental crust subducted into the mantle can become trapped and, ultimately, melt. The idea that continental crust is subducted into the mantle to depths of at least 60–150 km is becoming accepted by the geological community as a mechanism for explaining the presence of high-pressure (HP) and ultrahigh-pressure (UHP) eclogite terranes in many collisional mountain systems (see reviews by Chopin 2003; Ernst 2005; Brueckner 2006). It also provides a simple model for the presence of mantle-derived ‘orogenic’ peridotite lens in many HP or UHP terranes, as they could be derived from the mantle wedge that overlies the subducted crust (Brueckner & Medaris 2000). Exposed HP or UHP terranes with well-preserved eclogite-facies assemblages are generally assumed to have returned buoyantly towards the surface as coherent slabs because of their low density relative to the enclosing mantle (Chemenda et al. 1996; Ernst et al. 1997).

Slices of continental crust that do not return as coherent slabs but instead stall indefinitely within the mantle will undergo a different history. Initially, they will undergo metamorphic recrystallization as pressure and temperature increase during subduction. However, only temperature will continue to rise when they are trapped in the mantle, presumably to levels at which partial melting occurs. Limited melting may not destroy the integrity of a terrane, allowing it to exhumate eventually as a coherent mass, possibly as a diapir or plume (Gerya et al. 2006). Higher degrees of melting, however, would probably preclude the terrane from ever exhuming as a solid. Instead, granitic magmas will rise towards the surface leaving behind a dense residue that may linger indefinitely in the mantle.

A growing body of experimental and field evidence exists for melting of continental crust within the mantle during subduction and exhumation (Vrána 1989; Patiño-Douce & McCarthy 1998; Hermann & Green 2001; Parkinson & Kohn 2002; Schmidt et al. 2004; Patiño-Douce 2005; Wallis et al. 2005; Auzanneau et al. 2006; Zhao et al. 2007). Crustal xenoliths within Miocene ultrapotassic lavas from the Pamir provide direct textural and compositional evidence of dehydration-driven partial melting while under HP–UHP conditions in the mantle (Hacker et al. 2005). Most studies emphasize the role of pressure changes to induce melting (i.e. adiabatic decompression). Here the emphasis is on terranes that stay at depth for considerable periods of geological time. This static configuration will, over time, cause a terrane to heat isobarically to temperatures that are not attained in exhumed terranes, resulting, in suitable circumstances, in melting. Although this model reduces the importance of decompression to induce melting for many situations, it will become clear that decreasing pressure may still be required to explain the melting of some terranes, probably through the buoyant upwelling of sialic diapirs (Whitney et al. 2004).

Factors that might stall slices of continental crust in the mantle include the following: (1) eclogitization and/or early melt extraction, which results in the formation of mineral assemblages that are close to the density of the enclosing mantle and thus have reduced buoyancy (Massonne et al. 2007); (2) slices that are so thin that the body forces within them cannot overcome frictional resistance along the lower and upper boundaries; (3) a low initial subduction angle so that subsequent eduction is prevented by too much friction along the upper fault boundary for reverse motion;
(4) strong compression in the overlying mountain system that prevents the slabs from inserting themselves into high levels of the orogen.

The ultimate melting of trapped continental crust may explain the origin of late orogenic and post-orogenic igneous complexes in several Proterozoic and Phanerozoic mountain systems, and may even apply to ‘A-type’ or anorogenic granite-rich systems with no obvious relationship to orogeny. If a subducted HP–UHP terrane melts instead of being exhumed as a solid, the mountain system could end up losing the evidence for HP–UHP metamorphism (Patino-Douce & McCarthy 1998) and instead would be characterized by voluminous igneous intrusions after subduction had ceased. Thus the apparent absence of HP–UHP terranes in some Proterozoic or Phanerozoic mountain systems does not preclude subduction of continental crust into the mantle during their evolution.

The term ‘granitoid’ used throughout this paper should be understood as shorthand for suites that are dominated by granites, but can include a variety of other high-silica rock types. Late orogenic, post-orogenic and anorogenic granitoids (all henceforth referred to as ‘PO granitoids’ unless specificity is required) are mineralogically, chemically and petrogenetically complex. Space limitations preclude an extended discussion of these complexities. Instead, the emphasis is on a general tectonic framework for how PO granitoids might be generated.

**Exhumed HP–UHP terranes**

All exposed HP–UHP terranes show evidence of at least some melting during exhumation, but the surface area of exposed granitoids ranges from relatively inconsequential (<5–10% granitoids; e.g. the Western Gneiss Region, Norway; the Dora Maira Massif, Italy and the North East Greenland Eclogite Province), to extensive (25–50% granitoids; e.g. Liverpool Land, Greenland; the Bohemian Massif, central Europe). Melting is invariably associated with a high-temperature–low-pressure (HT–LP) overprint suggesting that melting occurs during exhumation (Gerdes et al. 2000). The presence of a hydrous vapour phase would induce water-saturated melting at temperatures as low as 650–700 °C (Fig. 1). However, free H₂O-rich fluids are probably absent from subducted continental crust (Yardley & Valley 1997), so dehydration melting (also known as hydrate-breakdown melting) is probably the most common melting mechanism when hydrous phases are present (Thompson 1982). Muscovite, biotite and amphibole at low to moderate pressures and phengite and zoisite at high and ultrahigh pressures (Fig. 1) are all capable of generating melts as a result of increased temperature or decreased pressure at constant temperature (Huang & Wyllie 1973; Thompson & Algor 1977; Patiño Douce & McCarthy 1998; Hermann & Green 2001; Hermann 2002; Schmidt et al. 2004; Patiño-Douce 2005; Auzanneau et al. 2006). Hence the abundance of hydrous minerals within a subducted terrane becomes a critical variable (Vielzeuf & Holloway 1988; Montel & Vielzeuf 1997). Terranes lacking hydrated minerals, such as granulite-facies crystalline basement complexes, cannot generate magma through dehydration reactions whereas terranes containing mica-rich schists and gneiss or amphibole-bearing metasedimentary rocks should melt readily, with the degree of melting dependent on the amount of hydrous phases present in the rock.

The other important variable is the $P$–$T$ path taken as a terrane subducts and then educts. Terranes that do so at fast rates (>10 mm a⁻¹) will have less time to heat and should have ‘skinny hairpin’ $P$–$T$ paths (‘S’ in Fig. 1) that are less likely to intersect hydrate-breakdown melting curves as they returns towards the surface (Patiño-Douce 2005). Presumably these terranes will return as intact slabs without a significant HT–LP overprint and without melting. Terranes that subduct and educt at slow rates (<1 mm a⁻¹) are more likely to warm up as heat is conducted in from the enclosing mantle and/or generated by radioactive decay and would undergo a ‘fat hairpin’ $P$–$T$ evolution (arrow ‘F’, Fig. 1) resulting in high temperatures during exhumation. These terranes will be characterized by recrystallization of the HP–UHP assemblages to HT–LP amphibolite or lower facies assemblages and are likely to partially melt as their $P$–$T$ path crosses dehydration melting curves (Fig. 1) potentially producing volumetrically significant granitoid igneous complexes. These granitoids will be largely ‘syntectonic’ if they do not leave the slab during exhumation; that is, they will develop fabrics (foliations, lineations, boudins) during and after crystallization associated with the deformation that accompanies exhumation. However, magmas may also be sheared out of the educting slab to rise through the upper shear boundary into the overlying plate (Brown & Solar 1998). If they enter an upper plate undergoing little or no deformation they will develop few structural elements and will appear to be PO granitoids even as isochronous granites in the exhuming slab will appear to be synorogenic. If, however, the overlying plate is actively short-
Trapped terranes with hydrous assemblages

A subducted HP–UHP terrane may become trapped in the mantle at its deepest penetration or at some intermediate level during exhumation. Even terranes that rise to the top of the mantle may stall at or near the Moho (Walsh & Hacker 2004; Schneider et al. 2007). Given sufficient time, these stalled terranes will become heated to the ambient temperature as depressed isotherms slowly but inevitably relax. Melting is a function of temperature if the terrane is truly stalled in place: adiabatic decompression would play no role. Ignoring heat generated by radioactive decay, the attained temperature in a trapped slab will be a function of depth and the regional geothermal gradient that develops after active subduction has ceased. A simplified case assuming a relatively low post-orogenic geothermal gradient (i.e. a ‘shield’ gradient) is shown by arrow ‘A’ in Fig. 1 (adapted from fig. 5.3 of Hacker & Peacock 1995), which shows isobaric heating towards a linear thermal gradient of 10 °C km⁻¹ so that a terrane at c. 100 km depth will achieve a temperature of c. 1000 °C. This temperature is not high enough to reach the ‘dry’ solids for granite, but is close to or above the dehydration melting curves of zoisite and phengite-bearing assemblages (curves 7 and 8). Subducted crust containing a significant volume of shale, greywacke or other mica-rich rocks will form phengite and zoisite at high pressures (>20 kbar) and therefore will initiate dehydration melting reactions deep in the mantle. Partial melting becomes more likely at higher pressures as the geotherm temperature exceeds the stability fields of phengite and zoisite by increasing amounts. In short, deeply trapped terranes with hydrous minerals could generate granitoid melts by conductive heating alone.

Trapped terranes without hydrous minerals

Subducted terranes that lack hydrous phases are harder to melt and therefore more likely to return towards the surface as coherent HP–UHP terranes. Indeed, many now exposed UHP terranes were largely old, crystalline, granitic, basement complexes that lacked significant amounts of hydrous phases at the time they were subducted into the mantle (Ernst 2005; Brueckner 2006). Subsequent retrograde metamorphism during exhumation turned many of these terranes into gneisses with hydrous minerals, making it difficult to assess how ‘wet’ or ‘dry’ these terranes originally were. Regardless, even anhydrous terranes should melt if they stall in the mantle, even at depths where the geotherm does not intersect the dry solidus.

The depressed isotherms typical of actively subducting slabs (i.e. Peacock 1990) will migrate ‘up slab’ with time when subduction ends, the mantle becomes static, and heat begins to be introduced conductively from above and below (Fig. 2a). Uncertainties in the thickness and initial temperature of the subducted continental crust, the angle and depth of subduction, and the temperature gradients below and above the initially cold crust complicate detailed modelling of the amount of time it would take to heat the slab to melting temperatures through conduction. However, if the ‘cold subduction’ model of Peacock & Wang (1999) is taken as analogous to a recently stalled slab of subducted continental crust, mantle at a temperature of c. 1250 °C will be 10–20 km above the coldest part (c. 250 °C) of the subducted slab at a depth of c. 100 km. A thermal diffusivity of the mantle of 10⁻６ m² s⁻¹ will cause the temperature of the slab to reach 1000 °C in 20–60 Ma, depending on distance (equation (4-113) of Turcotte & Schubert 1982). These times are minimum values, as the calculation assumes that the hot end of the thermal gradient maintains a constant temperature, whereas in a static mantle it presumably cools as heat migrates away. In any case, conductive heating alone (arrow ‘A’ in Fig. 1) is not sufficient to elevate temperatures to the dry granite solidus if a relatively low geothermal gradient is established after subduction ends.

However, additional heat will be generated, after a suitable ‘incubation period’ (Sylvester 1998), by radioactive decay. The heat generated through the decay of K, U and Th is generally discounted as a thermal source for deep-level melting because the parent isotopes are incompatible elements that are concentrated in the upper crust. The thinness of the crust and the high temperature gradient within it results in most of the generated heat being released into the atmosphere rather than remaining trapped within the crust. Suggested mechanisms for trapping more heat include thickening of the continental crust through thrusting (e.g. England & Thompson 1984; Chamberlain & Sonder 1990; Goffé et al. 2003) or by thrusting and subsequent thinning (Gerdes et al. 2000). These mechanisms theoretically can induce melting at the base of the crust, but they are unnecessary if a slice of crust becomes embedded deep within...
the mantle through subduction. In effect, the crust will be within a thermal pocket or sheath with relatively hot mantle both above and below. This configuration reduces or even eliminates the loss of heat from the crustal slab. Radioactive heat production will cause the initially depressed isotherms within the slab to migrate further up slab (Fig. 2a) as time passes, with the ultimate result that the slab will be hotter than the ambient geotherm of the mantle alone (arrow ‘B’ in Fig. 1). A rise of c. 100–300 °C would be enough to induce melting in even the driest terrane imaginable, depending on depth.

The typical granite generates $3.4 \times 10^{-5} \text{ J g}^{-1} \text{ a}^{-1}$ by radioactive decay. A granite requires 0.8 J g$^{-1}$ to raise its temperature by 1 °C, which would occur in c. 23 ka. Thus roughly 10 Ma are required to raise the temperature of 1 g of granite by 500 °C. Melting requires even more heat; c. 340 J g$^{-1}$, which means that another 10 Ma or so would be required to melt 1 g of rock. As was the case with conduction, the surprisingly short time (c. 20 Ma) needed to heat and melt a rock should be regarded as a minimum, as subducted terranes are not simple slices of granite. Also, early melting will partition U, Th and K into the melt and remove it from the system, leaving behind a residue that will generate less heat per unit time.

These processes do not necessarily eliminate the traditional process of adding heat to the crust by the upwelling of the hot asthenosphere (>1200 °C). This mechanism is usually invoked to melt the base of overthickened continental crust (e.g. Sylvester 1998; Duchesne et al. 1999). However, the subducted slab mechanism works even better, as slab breakoff (Brueckner & Medaris 2000; Parkinson & Kohn 2002) provides a straightforward mechanism for allowing the asthenosphere to well up into the mantle wedge and adjacent to the subducted slab of continental crust. The introduction of heat through this mechanism could be relatively rapid, depending on how much time has passed before breakoff occurs.

It seems inevitable that introduction of heat by conduction, radioactive decay, upwelling asthenosphere, or a combination of these processes will raise the temperature of a slice of continental crust trapped deep in the mantle to temperatures close to the dry granite solidus. The calculations presented above suggest that these mechanisms will take tens of millions of years after collision, providing the necessary delay to explain why some PO granitoids intrude long after compressional orogeny has ceased.

**The diapir model**

Melting reactions are endothermic, so a terrane that gradually heats to the dry granite solidus will undergo only limited melting, initially. Adiabatic decompression may still be required to induce larger-scale melting. Relatively rapid decompression can occur through the upward movement of diapirs (Whitney et al. 2004; Wallis et al. 2005). Rocks become ductile at temperatures near their melting solidus (i.e. at high homologous temperatures). Even small degrees of melting will further increase the ductility of the diapir dramatically (Brown & Solar 1998; Whitney et al. 2004). The density contrast between quartz-rich granitic crust and the overlying, static mantle wedge should allow crustal rocks to move upward in a fashion analogous to salt or gneiss domes (Fig. 2b) or perhaps as plumes (Gerya et al. 2006). Ascent in a largely solid state would limit the interaction of the diapirs with the enclosing mantle peridotite so that the bulk chemistry of the ascending mass should not change significantly. Upwelling rates are not necessarily rapid if the granitic rocks have been eclogitized and hence are only slightly less dense than the peridotite of the mantle (i.e. Massonne et al. 2007). An essentially spherical diapir will lose heat more slowly than a tabular slab. There will be interplay between heat loss to the surrounding mantle as the diapir moves upward, new heat produced by radioactive decay, and the limited melting produced by this heat (e.g. Whitney et al. 2004, figs. 4 and 5). The essential question becomes, where will the diapir melt extensively; in the mantle or in the crust?

**Post-orogenic granitoid chemistry and other features**

The major and trace element chemical and isotopic characteristics of PO granitoids vary considerably (see, for example, Küster & Harms 1998, and other papers in Lithos, 1998, Volume 45). Some recurring features include high silica content (i.e. monzonite, tonalite, granodiorite, granite, high-silica rhyolite), high alkali content, particularly K (i.e. peralkaline, high-K calc-alkaline), but low alkaline-earth contents (Ca, Ba, Sr), variable but generally low Al content (metaluminous), high halogen content, and high incompatible element content (i.e. Rb, Cs, Th, REE; Frost et al. 2001; Vander Auwera 2003; Best 2003). Some PO granitoids contain zircons with older cores of metamorphic origin (Paterson et al. 1992; Zhao et al. 2007). Most of these features are consistent with the melting of old sialic continental crust. However Sr, Nd and O isotopic ratios and trace element discriminators vary considerably, commonly displaying both ‘mantle’ and ‘crustal’ signatures (i.e. Best 2003, p. 389; Hess 1989, pp. 214 and 228; Vigneresse 2004, fig. 8; Andersen et al. 2001; Bolle et al. 2003). Major element chemistry also indicates that many granitoid suites, including the alumina-deficient ‘A-type’ granites considered here, are ‘hybrid’ rocks with both mantle and crustal components (Patiño-Douce 1999).

This interaction between crust and mantle to produce hybrid magmas is usually modelled as the upwelling of mafic magmas from the mantle into the base of the crust where the resultant high temperatures result in crustal melting (e.g. Liankun & Kuirong 1991, fig. 5). The model presented here turns this relationship on its head: sialic melts generated from slices, diapirs or plumes of continental crust within the mantle interact with the ultramafic mantle as they rise through it. This mechanism avoids several problems. With some exceptions, the lower continental crust is richer in Fe and Mg and lower in Si, K, and Na than the upper crust, making it more difficult to generate the essentially upper crust characteristics (i.e. high $^{87}\text{Sr}/^{86}\text{Sr}$, high large ion lithophile element (LILE) concentrations, etc.) of many granitoid suites by melting the bottom of the lower crust alone. The rise in temperature required to melt the lower crust requires either thickening of the continental crust or the introduction of heat by mantle magmatism and/or upwelling of the asthenosphere. The subduction model eliminates these difficulties, as it introduces the upper crust into the high-temperature realm of the upper mantle, creating an ideal mechanism for melting this crust and obviating the need to generate mantle magmas. Furthermore, these magmas can interact with the mantle to produce hybrid characteristics. Finally, the subducted crust model allows melting to occur at higher pressures than at the base of the crust, and, generally speaking, increasing pressure results in melts that are less ferromagnesian and richer in alkalis (Patiño-Douce & McCarthy 1998; Hermann & Green 2001). If the subduction model is correct, the degree to which the chemistry and isotopic patterns of granitoid melts are hybridized by interaction with the mantle depends on whether they initially rise through the mantle as melts or solids.
Mantle interactions

Granitoid melts should react extensively with the mantle. They will not cause mantle melting and mixing as a result of the much higher melting temperatures of peridotite relative to granite. However, magma bodies will undergo exchange reactions at their margins with the host peridotite to form pyroxenite and orthopyroxene-rich peridotite (Kelemen et al. 1998). Pyroxene crystallization requires the transfer Mg, Cr and Fe(?l) from the peridotite into the magma and Si, Al and Ca from the magma into the mantle. Extensive melt–mantle interactions probably produced the Mg-rich hybrid magmas (melagranites, quartz monzonites, Gerdes et al. 2000) that occur within some PO suites. K and LILE such as Rb and Cs are not taken up by pyroxenites or other anhydrous mantle phases, so these incompatible elements will remain in the granite melt, provided that H2O is absent. If water is dissolved in the magma, K-rich phlogopite or amphibole will crystallize and change fractionation trends significantly by depleting the magma in K and the LILE. Therefore the amount of H2O dissolved in the magma is a significant variable. Isotopic patterns will be modified if the mantle wedge contains significant clinopyroxene. Exchange reactions between mantle clinopyroxenes and magmas would lower the 87Sr/86Sr and raise the 143Nd/144Nd ratios of the magma so that the enriched isotopic signature of melted continental crust will become progressively more mantle-like or depleted.

A significant geochemical signature of PO granitoids is the relatively low FeO/(FeO + MgO) of post-orogenic granites (‘magnesian’ granitoids; Frost et al. 2001) compared with the significantly higher ratios of ‘A-type’ or anorogenic granites (‘ferroan’ granitoids). Other differences are that anorogenic granitoids form much later than PO granitoids, tend to melt at higher temperatures (>950 °C), have higher concentrations of high-field strength elements (HFSE) such as Zr and Nb, and tend to lack hydrous minerals, relative to post-orogenic granites. These chemical differences suggest that some hydrous phases existed in the source rock for the post-orogenic granites whereas the source rock for anorogenic granites lacked hydrous minerals. The amount of hydrous minerals in the source terrane may also influence the degree of hybridization with the mantle. A ‘some-what wet’ scenario (some hydrous minerals) will feature relatively low-temperature dehydration melting (relative to the dry solidus) early in the heating history of the trapped slab. For example, phengite dehydrates at a temperature of as low as 850 °C at 100 km depth in the presence of trace amounts of H2O (Fig. 1; Hermann & Green 2001; Auzanneau et al. 2006). The small melt volumes generated should initiate early ductile behaviour and the early formation of ascending diapirs (Fig. 2). The resultant decompression melting is likely to occur while the diapir is still within the mantle (Fig. 2b), increasing the likelihood that interaction of the melt with the mantle will form relatively early ‘magnesian’ granitoids and other granitoids with ‘mantle’-like signatures, as discussed above.

The higher temperatures required to melt a (‘dry’) terrane that is free of hydrous phases requires that extended time must pass before conductive and/or radioactive heating can initiate ductile upwelling. A solid diapir will not react extensively with the surrounding mantle as it rises. Decompression melting will also be delayed and may not occur until the diapir has reached relatively shallow levels (arrow ‘C’ in Fig. 1), which can be close to the Moho, depending on the thickness of the crust (Figs 1 and 2b). Melts generated at shallow levels will not react as extensively with the mantle as those generated at deeper levels, although some interaction presumably must occur before they leave the mantle. Afterwards, they will rise more or less isochemically through the crust to form large-scale post-orogenic intrusion complexes retaining strong ‘crustal’ geochemical signatures. Melting near the Moho will occur within the plagioclase stability field. The geochemical signatures of anorogenic granites are low Al2O3, CaO, Sr and Eu concentrations (Patino-Douce 1999), requiring melting within the plagioclase stability field (Patino-Douce 1997).

A possible problem with the subducted crust melting model is that extensive interaction of granitoid melts with the mantle could result in the complete crystallization of the melt in the mantle, preventing the magmas from ever rising into the crust. However, a pyroxenite carapace is likely to form between the ascending melt and the enclosing mantle, which would armour the melt from further contact with the mantle. This shield may remain in place rather than disintegrating, because the mantle would be essentially static as a result of the termination of subduction. Dismembered fragments of this carapace might be produced if there is residual flow in the mantle. These fragments could be the source of some of the mafic xenoliths or enclaves that characterize many PO granitoids (Didier & Barbarin 1991); for example, the pyrope-bearing peridotite xenoliths in perpotassic granites in the Bohemian Massif (Vrana 1989). Mantle enclaves can react extensively with the host melt and will not necessarily be distinguishable from mafic enclaves of crustal origin, as both hydration and dehydration processes accompany disintegration and mixing (Beard et al. 2005). For example, gabbronitite and hornblende xenoliths from the ‘Big Jim Complex’ in the Cascade Range of Washington State would be difficult to distinguish from lower crustal enclaves even though they formed through the interaction of a large peridotite body with a diorite magma (Kelemen & Ghiorso 1986). Similarly, mafic and intermediate granulite xenoliths within the Hannuoba basalts of North China could be mislabelled as fragments of the lower crust event though they formed through silicic melt–peridotite reactions in the mantle (Liu et al. 2005).

Orogenic cycles within the Fennoscandian Shield

The evolution of the Fennoscandian (or Baltic) Shield and the Caledonian system that occurs along the shield’s western margin in Scandinavia provides several possible examples of subducted continental crust that melted during the Proterozoic. The connection between orogeny and the intrusion of ‘anorogenic’ rapakivi granites and rhyolites in this shield is not a new one and has been proposed previously for both the Svecofennian (Windley 1993) and Gothian (A˚hall et al. 2000) orogenies. New discoveries related to the Sveconorwegian orogeny have strengthened evidence for this connection.

The Svecofennian orogeny occurred between 1.91 and 1.87 Ga, overlapping the intrusion of anatectic granites (the Central Finland Granitoid Complex) at 1.89–1.88 Ga, followed by the intrusion of anatectic potassium granites at 1.84–1.82 Ga, then by the Transscandinavian Igneous Belt at 1.81–1.65 Ga (Nironen 1997). The subsequent emplacement of Rapakivi Granite Suites at 1.65–1.50 Ga may also be related to Svecofennian subduction (Windley 1993); however, recent evidence suggests that rapakivi emplacement maybe related to the enigmatic Gothian orogeny (A˚hall et al. 2000). The tectonic scenarios for Fennoscandian evolution are complex but generally involve the accretion of arcs and microcontinents to the southwestern margin of an Archaean nucleus (present coordinates) along a NW–SE-trending suture (the Luleá–Jääkkömm and Ladoga–Bothnian Bay Zones, Nironen 1997, fig. 1) during the early phases of this evolution.
(Windley 1993; Nironen 1997). Structural, sedimentological and geochemical criteria suggest that the Archaean Shield was subducted to the SW beneath the arcs. The above-mentioned plutonic suites occur to the SW of the suture in southern Finland and central and southern Sweden. The SW edge of the Archaean Shield (i.e. the Karelian Province) underwent high-temperature but low-pressure metamorphism, locally into the low-P granulite facies (Nironen 1997), suggesting that only the shallowest portion of the subducted Archaean slab was exhumed and that the more deeply subducted portions were trapped in the mantle and ultimately melted to generate the numerous PO granitoid suites, most of which have geochemical and mineralogical characteristics of magmas generated from older continental crust and mixed with a mantle component (Nironen 1997; Elliot 2003). The 1.89–1.82 Ga ages of the two earlier episodes of anatectic melting, as well as the older c. 1.81 Ga from the Transscandinavian Igneous Belt overlap with or are up to 80 Ma younger than the Svecofennian orogeny, consistent with the tens of millions of years required for subducted crust to heat through a combination of radioactive decay and conductive heating. The 1.65 Ga Transscandinavian Igneous Belt is more than 200 Ma younger than Svecofennian collision, which may be too long to relate them to the collision as proposed by Windley (1993).

It is more likely that these younger plutons are related to the subsequent Gothian orogeny. Åhäll et al. (2000) directly related three distinct convergent events during the Gothian orogenic cycle at 1.69–1.65, 1.62–1.58 and 1.56–1.55 Ga with subsequent phases of rapakivi magmatism (locally associated with anorthosites) at 1.65–1.62, 1.58–1.56 and 1.55–1.50 Ga, respectively. Both the collisional belts and the resultant rapakivi belts trend north–south, and both belts migrated from east to west as three terranes docked successively outboard of the southwestern margin of Fennoscandia. A problem with this model is that the rapakivi belts occur significantly further east (300–600 km) than the collisional belts. Possibly the subducted Gothian crust was carried eastward within the asthenosphere after collision. More significantly, the two initial convergent events involved the collision of island arc systems with Fennoscandia, which implies that subduction was outboard, beneath the arcs, instead of inboard, beneath Fennoscandia. Outboard subduction could not have emplaced continental crust beneath Fennoscandia as proposed in the model presented here. The third event, however, involved the development of a continental arc on the Fennoscandian margin followed by collision with a microcontinent (‘proto-southwest Norway’; Åhäll et al. 2000). Eastward subduction of oceanic lithosphere beneath the western margin of Fennoscandia during the evolution of the arc would have been followed by the subduction of the leading edge of the microcontinent into the mantle, which would be entirely consistent with the model presented here.

### The Sveconorwegian orogeny

Subsequently, in southern Norway and southwestern Sweden, the Sveconorwegian (Grenville) orogeny occurred at 1.13–0.96 Ga (Bingen et al. 2008c) followed, with some overlap, by the intrusion of post-orogenic granites at 1.0–0.93 Ga and the famous anorthosite–mangerite–charnockite Rogaland Complex (also known as the Egersund Igneous Complex) of southern Norway at c. 0.93 Ga (Andersen et al. 2001; Bingen et al. 2008a–c). The Sveconorwegian orogeny involved the successive accretion of terranes with metamorphic ages that ‘young’ from west to east (Söderlund et al. 2008; Bingen et al. 2008b,c); namely, the 1.14–1.08 Ga Bamble–Kongsberg terranes, the 1.05–1.025 Ga Idefjorden terrane and the 0.98–0.96 Ga ‘Eastern Segment’ (Fig. 3). A younger metamorphism between 0.97 and 0.90 Ga in the westernmost Telemarkian terrane (the ‘Dalan phase’ of Bingen et al. 2008b) appears to violate this general pattern, but the metamorphism not necessarily date an accretionary event but rather may date metamorphism related to the intrusion of PO granitoids discussed below.

The tectonic evolution of the easternmost Eastern Segment (Andersson et al. 2008; the ‘Falkenber phase of Bingen et al. 2008b) fits very well with the proposed crustal subduction–melting model. It is a north–south-trending (Fig. 3) belt of eclogite-bearing high-pressure granulite-facies rocks in southwestern Sweden (Möller 1998; Andersson et al. 2008). It is separated to the east from relatively unmetamorphosed largely magmatic rocks of the Transscandinavian Igneous Belt (discussed above) by a large-scale tectonic boundary, the Sveconorwegian Frontal Deformation Zone, which extends northward all the way to the west coast of the Western Gneiss Region. It is separated to the west by a major mylonite zone (Andersson et al. 2002, 2008; Möller et al. 2007) from the Idefjorden terrane, a c. 1.6–1.53 Ga Gothian volcanic arc sequence (Brewer et al. 1998, 2004) that was metamorphosed at high pressures 50 Ma before the Eastern Segment (Söderlund et al. 2008; the ‘Agder phase of Bingen et al. 2008b). All of the voluminous Sveconorwegian PO granitoids occur west of this mylonite zone, within the Idefjorden and Bamble–Kongsberg terranes and, further to the west, within the Mesoproterozoic Telemarkian terrane (Fig. 3).

The eastern and western boundaries of the Eastern Segment terrane can be interpreted (Fig. 3) as the lower thrust and upper low-angle normal detachment, respectively, of an exhumed HP slab (Hegardt et al. 2003). East–west stretching lineations (Andersson et al. 2008) with a top-to-the-west movement along the mylonite zone (Andersson et al. 2002) provide evidence that the mylonite zone was a low-angle normal fault during exhumation. If subduction was in the opposite direction (i.e. slab subduction to the west) a reasonable tectonic model would have the western edge of Fennoscandia subducted beneath arcs (i.e. the Idenfjorden and Bamble–Kongsberg terranes), microcontinents (i.e. ‘Telemarkia’, Bingen et al. 2008c) and perhaps even beneath the eastern edge of a major exotic craton (Amazonia?, Bingen et al. 2008b). The subducted western edge of Fennoscandia at the time was probably crust formed during the earlier Gothian orogeny (i.e. the Transscandinavian Igneous Belt), an inference supported by orthogneisses within the Eastern Segment that give U–Pb zircon crystallization ages of 1.7–1.68 Ga (Söderlund et al. 2002).

Eclogite-facies metamorphism of the ‘Eastern Segment’ occurred between 0.98 and 0.96 Ga (Johansson et al. 2001; Hegardt et al. 2005; Möller et al. 2007). Here it is proposed (see Fig. 4) that the subducted continental crust separated (boudined?) into two slabs: a shallower one that was exhumed essentially intact to expose the HP eclogites of the eastern segment, and a deeper segment that remained trapped within the mantle (Fig. 4). Evidence for a granulite-facies overprint accompanied by migmatization (Möller et al. 2007) suggests that exhumation of the HP eastern segment was slow enough to cause heating and partial melting of the slab during exhumation (Hegardt et al. 2005). It seems likely that a stranded more deeply subducted terrane further to the west also melted and intruded the overlying complex of arcs and microcontinents to form the 0.97–0.92 Ga PO granitoids of southern Norway. These intrusions are associated with the Dalan phase of Sveconorwegian tectonism (Bingen et al. 2008a, c), which was characterized by the develop-
ment of large-scale gneiss domes similar to those described by Whitney et al. (2004), suggesting that the Dalan phase was dominated by thermally driven vertical tectonics. If the end of eclogite-facies metamorphism is taken to be between 0.98 and 0.96 Ga, then the intrusion of the youngest plutons dated at 0.93 and 0.92 Ga (Andersen et al. 2001; Bingen et al. 2008a, b) suggests that 30–60 Ma were required to heat the subducted portion of the Eastern Segment to melting temperatures. Isotope and trace element studies of these plutons show an increasing component of ‘mantle derived material in the granite sources’ (Andersen et al. 2001) from east to west, as would be predicted in a tapered mantle wedge that increases in thickness from east to west (i.e. Fig. 4).

Finally, the Caledonian orogenic cycle involved the closure of Iapetus and the collision of the west side of composite Fennoscandia with Laurentia during the c. 400 Ma Svecadian orogeny. The HP–UHP terrane known as the Western Gneiss Region was subducted and exhumed without significant amounts of melting (Brueckner & Van Roermund 2004). The change in behaviour from very high degree of melting of subducted continental crust during the Svecofennian and perhaps Gothian orogenies to exhumation plus melting during the Sveconorwegian orogeny to exhumation without melting during the Caledonian orogeny may signal a change in the behaviour in subducted continental crust as a result of the secular cooling of the mantle through time.

Sveconorwegian anorthosites, mangerites, charnockites and ferro-potassic granitoids

The anorthosite–mangerite–charnockite Rogaland Complex in southernmost Norway (Fig. 3) is characteristically composed of anhydrous minerals. East of the Rogaland Complex is a 300 km long north–south belt intruded by ferro-potassic granitoids (also known as hornblende–biotite granitoids). The two associations
intruded contemporaneously at 0.93–0.92 Ga (Schärer et al. 1996; Vander Auwera et al. 2003; Bingen et al. 2006, 2008a), roughly 30–60 Ma after the eclogite-facies metamorphism of the Eastern Segment. The apparent lack of relationship of the anorthosite–mangerite–charnockite and hornblende–biotite granitoid complexes with orogeny earned them the sobriquet of ‘anorogenic’ or A-type granitoids (Schärer et al. 1996; Vander Auwera et al. 2003), but here it is argued that the source for these intrusions was put in place during the Sveconorwegian subduction of the Eastern Segment.

Both rock suites required very high melting temperatures; 1200–1300 °C at depths of c. 40–50 km for the anorthosites (Duchesne et al. 1999; Longhi et al. 1999) and 1000–1050 °C for the biotite–hornblende granitoids (Vander Auwera et al. 2003). Both suites were generated by melting rocks of broadly basaltic composition, but the anorthosite–mangerite–charnockite suite requires an anhydrous source (gabbro-norite or mafic granulite) whereas the hornblende–biotite granitoid suite requires a hydrous parent (amphibolite). Duchesne et al. (1999) proposed that the source rocks for the Rogaland anorthosite–mangerite–charnockite suite were mafic ‘tongues’ that projected into the mantle from the Moho. The subducted crust model presented here provides the perfect mechanism for generating these ‘tongues’ (Fig. 4), assuming the upper part of subducted crust is primarily sialic whereas the lower crust is mafic. The upper quartz-rich crust could be removed either by diapiric upwelling or delamination and exhumation along the boundary between lower and upper crust (i.e. the ‘Conrad’ discontinuity). Both processes apparently operated during the evolution of the Sveconorwegian orogenic belt; that is, melting resulted in the intrusion of the 0.96–0.92 PO granitoids of southern Norway and delamination and eduction resulted in the exhumation of the HP ‘Eastern Segment’ of southwestern Sweden.

The denser and more refractory lower crust left behind would have been exposed to the introduction of heat not only from below, but also from above. The eventual establishment of a late Proterozoic sub-continental geothermal gradient and the heat generated by radioactive decay may not have been sufficient to heat the mafic slab to solidus temperatures (>1000 °C). Upwelling asthenosphere may still have been required to raise temperatures to the melting solidus (Ashwal 1993). The subducted slab model offers a simple mechanism for initiating upwelling by converting the deepest levels of subducted mafic crust to dense eclogite, which would cause it to break off and sink to deeper levels of the mantle, the same fate that occurs to subducted oceanic lithosphere. This ‘slab breakoff’ would create a gap allowing upwelling of the asthenosphere (Fig. 4). The shallower (<40–50 km) mafic crust would recrystallize to mafic garnet granulites, or, at still shallower levels, and provided hydrous fluids were available, amphibolites. The lower density of these plagioclase-bearing assemblages makes it unlikely that they would sink. Instead, they would remain in place to form the ‘tongues’ of Duchesne et al. (1999). The upwelling of the hot asthenosphere through the breakoff gap (Fig. 4) would engulf these mafic slices, causing them to melt and to generate, respectively, the anhydrous magmas of the anorthosite–mangerite–charnockite suite and the hydrous magmas of the hornblende–biotite granitoid suite. This model is consistent with low temperatures and shallower melting levels that lead to the hornblende–biotite granitoid suite to the east and higher temperatures and deeper melting levels that led to the anorthosite–mangerite–charnockite to the west as would be expected for a slab that dipped westward during Sveconorwegian subduction.

Conclusions

Slices of subducted continental crust may melt and return towards the surface as magmas or recrystallize and return towards the surface as relatively coherent HP–UHP terranes or, as elegantly demonstrated by the evolution of the Sveconorwe-
gian orogenic belt, exhibit some combination of these two processes. The evolution of the Fennoscandian Shield suggests that there was a shift from the former towards the latter as the Earth cooled and the nature of metamorphism changed. HP metamorphism is not unknown in the Precambrian, but most UHP eclogite-facies belts formed during the Phanerozoic whereas large volumes of PO granitoids were generated largely during the Proterozoic. Solid exhumation returns continental crust to the outer layer of the solid Earth as more or less intact and chemically unmodified terranes, whereas melting returns to the surface igneous rocks that are more silica, aluminum and alkali rich than the rocks that were originally subducted. Thus crustal subduction and melting may be an important distillation mechanism for creating ever more silicic continental crust.

Simultaneously, deep-level crustal melting would leave behind residues of more mafic rocks (Patiño-Douce & McCarthy 1998; Patiño-Douce 2005), which would remain in the mantle as a result of their high density. In addition, deeply generated granitoid magmas will enrich the overlying peridotite as they pass through the mantle wedge. Both mechanisms can generate enriched geochemical reservoirs that would play an important role in subsequent mantle melting processes (Hermann & Green 2001; Schmidt et al. 2004; Patiño-Douce 2005).

Many mountain systems that appear to lack evidence of HP–UHP metamorphism, but are instead characterized by voluminous syntectonic and PO granitoids, may nevertheless have evolved through the subduction of continental crust into the mantle and the formation of eclogite-facies assemblages. Subtle clues for this metamorphism, such as coesite in zircons within the upper mantle and the formation of eclogite-facies assemblages. Subtle clues for this metamorphism, such as coesite in zircons within granitoid magmas, should be sought to either prove or disprove the subducted crust model.

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References


