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Quantifying the Exhumation of UHP-Rocks in the Western Gneiss Region, S. W. Norway: a Branch-line – Balanced Cross-section Model

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KEYWORDS

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Abstract

Balanced cross-sections, combined with stratigraphic and seismic data, show that the Western Gneiss Region in the Scandinavian Caledonides is allochthonous (Window Allochthon), with a minimum SE- to SSE-directed displacement of ~215 km and probably of ~320 km. In Neoproterozoic times, it formed a topographic high (Jostedal Plateau) NW of the Hedmark Basin, now imbricated into the internal part of the Osen-Røa Nappe Complex. The Valdres Nappe restores to NW of the Jostedal Plateau. The Jotun Nappe (and coastal equivalents) restore to a microcontinent (Jotunøya) lying outboard of a minor ocean (Fjordane Sea). Based on this palaeogeography, a new model for the evolution of the Western Gneiss Region is proposed. Westward subduction of the Fjordane Sea and then the Jostedal Plateau (425-410 Ma) under Jotunøya caused peak ultra-high pressure metamorphism (~120 km depth) at ~405 Ma. The upper part of the Jostedal Plateau (Western Gneiss Region) decoupled from the lower continental lithosphere and exhumed at ~10 mm yr^{-1}, along the subduction channel, intruding rapidly (~14 mm yr^{-1} horizontally) under the orogenic weage moving relatively slowly towards the SE from Jotunøya, due to the continuing collision with Laurentia. The rapid emplacement of the WGR caused top-to-hinterland faulting at the roof of the Western Gneiss Region, coeval with thrusting in the underlying Hedmark basin, forming the Osen-Røa Nappe Complex. This faulting was accompanied by sinistral strike-slip movement on the Møre-Trondelag Fault Zone, giving local transtension at the roof of the Western Gneiss Region during regional transpression. SSE-directed thrusting in the Osen-Røa Nappe Complex continued after exhumation and ceased at between ~388 to 375 Ma.

Die Kombination von bilanzierten Profilen mit stratigraphischen und seismischen Daten zeigt, daß die Westliche Gneis Region (WGR) der skandinavischen Kaledoniden als allochthon anzusehen ist (Fenster-Allochthon). Der SE- bis SSE-gerichtete Versatz beträgt mindestens 215 km, möglicherweise bis zu 320 km. Die WGR bildete im Neoproterozoikum eine geographisches Hoch (Jostedal Plateau) in einer Position NW des Hedmark Beckens, welches heute in den Osen-Røa Deckenkomplex eingeschuppt ist. Die Valdres Decke befand sich ursprünglich NW des Jostedal Plateaus. Die Jotun Decke und Küstenäquivalente waren Teil eines Mikrokontinents (Jotunøya) am Rand eines kleinräumigen Ozeans (Fjordane Meer). Aufgrund dieser Paläogeographie wird ein neues Modell für die Entwicklung der WGR postuliert: Die westgerichtete Subduktion des Fjordane Meer sowie des Jostedal Plateaus (425-410 Ma) unter Jotunøya verursachte eine Ultra-Hochdruck-Metamorphose (120 km Tiefe) um 405 Ma. Der höhere Teil des Jostedal Plateaus (WGR) löste sich von tieferen Teilen der kontinentalen Lithosphäre ab, exhumierte entlang des Subduktionskanals mit einer vertikalen Rate von 10 mm yr-1 und intrudierte mit hohen horizontalen Bewegunsraten von 14 mm yr^{-1} unter das Orogen. Dieses bewegte sich aufgrund der Kollision mit Laurentia langsam von Jotunøya nach SE. Die rasche Platznahme der WGR führte zu zeitgleichen Hinterland-gerichteten Abschiebungen im Hangenden der WGR sowie zu Überschiebungen im unterlagernden Hedmark Becken, welches so den Osen-Røa Deckenkomplex bildete. Diese Störungen wurden von sinistralen Seitenverschiebungen entlang der Møre-Trondelag Störungszone begleitet und indizieren lokale Transtension im Hangenden und in überlagernden Decken der WGR während regionaler Transpression. Die SSE-gerichteten Überschiebungen im Osen-Røa Deckenkomplex dauerten nach der Exhumation an und endeten erst zwischen 388-375 Ma.

1. Introduction

In many orogens, top-to-hinterland normal faulting occurred at the same time as plate collision, with the block in the footwall of the top-to-hinterland fault overlying active imbrication (cf. Wheeler 1991; Grujic et al., 1996; Ring et al., 1998; Grasemann et al., 1999; Eide and Liou, 2000; Leech and Ernst, 2000; Nowlan et al., 2000). Frequently, this fully allochthonous exhumed block includes (ultra)-high pressure rocks. These are, therefore, exhuming (or have exhumed) contemporaneously with bulk orogenic shortening and are fully allochthonous. In contrast, nearly all models of top-to-hinterland normal faulting and ultra-high pressure exhumation in the SW Norwegian Caledonides invoke a (par-) autochthonous ultra-high pressure block (the Western Gneiss Region; Figs. 1, 2) and some have extension post-dating collision (Chauvet and Dallmeyer, 1992; Chauvet and Séranne, 1994; Fossen, 1992, 1993, 2000; Fossen and Holst, 1995; Fossen and Dunlap, 1998; Fossen and Dallmeyer, 1998; Séranne, 1992; Andersen, 1993, 1998; Andersen and Jamtveit, 1990; Andersen et al., 1991, 1994; Dewey et al., 1993; Wilks and Cuthbert, 1994; Milnes et al., 1997; Krabbendam and Dewey, 1998; Terry and Robinson, 2004a, b; Tucker et al., 2004). Those models which do implicitly invoke an allochthonous Western Gneiss Region (Breuckner, 1998; Breuckner and van Roermund, 2004; Hacker et al., 2003; Root et al., 2005) have not constrained the displacement. Essentially, enormous effort has been expended to constrain the P-T-t-D evolution of the Western Gneiss Region, without a similar effort having being made to investigate the constraints imposed on tectonic models by the regional-scale deformation. Indeed, most discussion has focussed on exhumation, with many articles evading the problem of how the rocks reached ultra-high pressure conditions initially.

This paper presents a model in which the Western Gneiss Region, with its ultra-high pressure rocks, has been displaced several hundred kilometres southeastwards relative to the Baltica craton during and after exhumation (cf. Rice, 1999, 2001a). This is based on a pre-orogenic palaeogeography similar to other parts of the Scandinavian Caledo-nides where the displacement has been constrained by balanced and restored cross-sections (see below). A particular concern has been to ensure that the process rates inferred (subduction, exhumation, thrusting) are both comparable to present day rates and consistent within the model.

2. Scandinavian Caledonides

The Scandinavian Caledonides (Fig. 1) comprise nappes emplaced during the typically SE- to ESE-directed early Palaeozoic collision between Laurentia and Baltica, with sinistral transpression in the latter stages (Harris and Fettes, 1988; Soper et al., 1992). In the south, shortening changed from SE- to SSE-directed. These nappes have been grouped into the Lower, Middle, Upper (with Seve and overlying Köli nappe units) and Uppermost Allochthons (cf. Gee et al., 1985).

Basement rocks, together with an unconformably overlying Neoproterozoic-Ordovician cover, exposed in the cores of tectonic

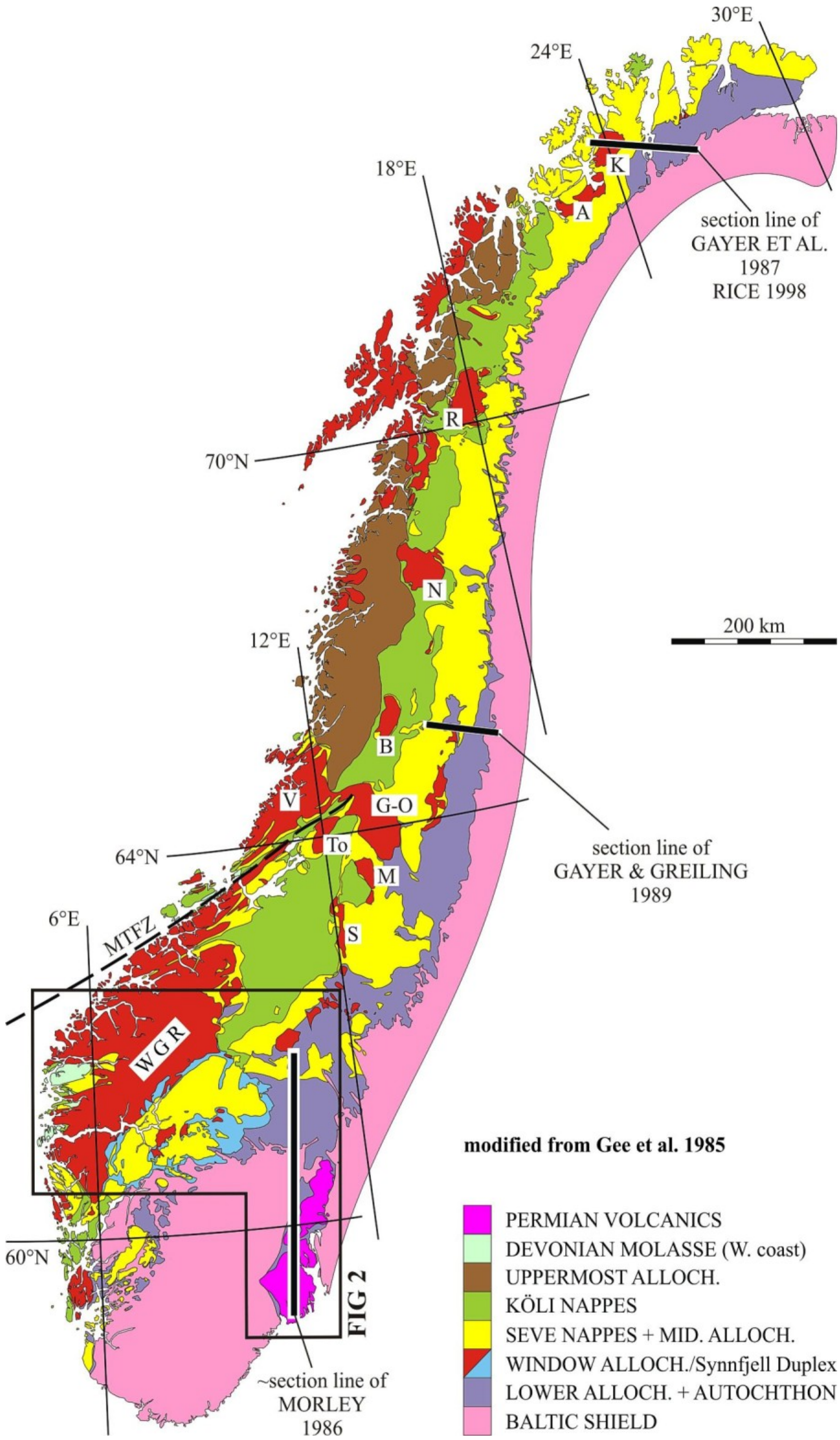

Figure 1: Simplified geological map of the Scandinavian Caledonides showing the distribution of the main structural elements. Section lines through the Lower Allochthon used to show that the basement rocks in the cores of the tectonic windows are allochthonous (Window Allochthon) are shown for Gayer et al. 1987, Rice 1998, Gayer and Greiling 1989, Morley 1986. MTFZ - Møre-Trøndelag Fault Zone (Grønlie and Roberts 1989). Main tectonic windows: K – Komagfjord; A – Alta-Kvænangen; R – Rombak; N – Nasafjäll; B – Børgefjellet; G-O – Grong-Olden; To – Tømmerås; M – Mullfjället; S – Sylarna V - Vestranden; see Figure 2 for minor windows further south.

windows form the Parautochthon and parts of the Lower Allochthon of Gee et al. (1985; Fig. 1). Stratigraphic overlaps revealed by restored balanced-sections through the Lower Allochthon indicate that the rocks coring the Børgefjellet and Komagfjord tectonic windows (Fig. 1) must be restored to the NW of the undeformed Lower Allochthon, by 45 km and 175 km, respectively (Gayer et al., 1987; Gayer and Greiling, 1989; Rice, 1998). Extensive erosion of the Lower Allochthon (Hossack and Cooper, 1986), the shortening in which is critical for estimating the displacement of the Window Allochthon, has resulted in the former distance being relatively small. Similarly, basement/cover rocks in the core of the Rombak and Mullfjället tectonic windows and the Skardöra Antiform (Fig. 1) are allochthonous (Anderson, 1989; Palm et al., 1991). Hereafter, all these window-coring rocks, from both the Lower Allochthon and Parautochthon (cf. Rice, 2001b), are termed the Window Allochthon.

The resulting palaeogeography comprises an internal, relatively shallow basin, now the Lower Allochthon, and an external, deeper basin, now the Middle Allochthon. These were separated by a topographic high, the restored Window Allochthon, which submerged in Ediacaran or later times (Gayer and Roberts, 1973; Gayer and Rice, 1989; Gayer and Greiling, 1989). Outboard of the Middle Allochthon, the transition to Iapetus oceanic crust has been preserved in the Seve nappes. These units (Autochthon, Lower, Window and Middle Allochthons and Seve nappes) thus form the Iapetus Baltoscandian passive continental margin. The overlying Köli nappes represent intra-Iapetus settings and the Uppermost Allochthon comprises rocks derived from the Laurentian side of Iapetus.

Isotopic data suggest that deformation occurred in two main phases: the Finnmarkian (~520-490 Ma) and Scandian (~420-380 Ma) events (cf. Harris and Fettes, 1988). Recently, Breuckner and van Roermund (2004) have proposed a Jämtlandian orogeny (455-445 Ma), incorporating much of the evidence of 'inter-orogenic' deformation throughout the Scandinavian Caledonides (e.g. Andersen et al., 1998; Stephens et al., 1993; Rice and Frank, 2003; Hacker and Gans, 2005).

2.1 Western Gneiss Region and Associated Areas

Only material lying on a transect from Langesund to the Solund-Stadlandet region, passing over Moelv, Jotunheimen and Døvrefjell, has been included (Fig. 2). The Bergen Arc area (Boundy et al., 1996; Bingen et al., 1998a; Kühn et al., 2000; Fig. 2) has been deliberately ignored, although acknowledged to be important.

2.1.1 A Lower Allochthon/Autochthon

South of Langesund, a 1.5 km thick mid-Cambrian to Devonian Autochthon crops out (Figs. 2 and 3; Kumpulainen and Nystuen, 1985), in which the Ringerike Sandstone comprises mid- to Upper Silurian and younger molasse deposits, but a precise depositional age range has not been established. The overlying Osen-Røa Nappe Complex (Lower Allochthon) has been divided into three large-scale tectonic flats (Morley, 1986); northwards, these lie at stratigraphically deeper levels, and are connected by short ramps. In the Oslo Graben, the basal décollement lies in the mid-Cambrian Alum Shales (1st flat), underlain by the autochthonous Lower Cambrian Holmia Series (<50 m) and the Baltic Shield (Fig.

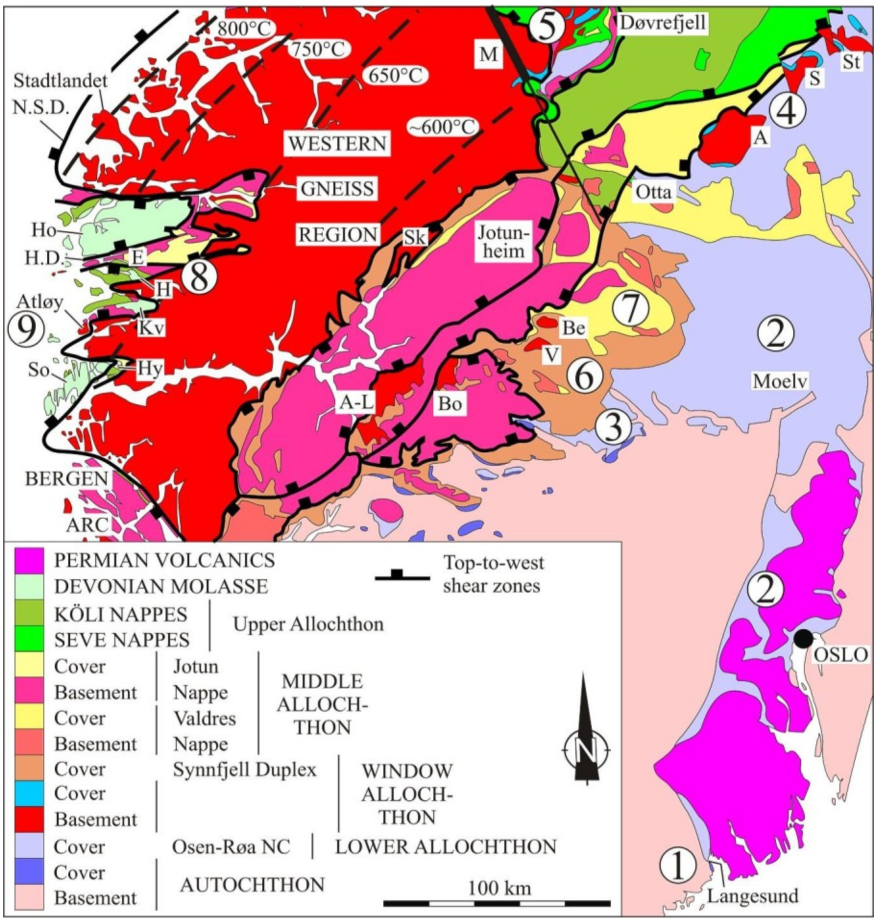

Figure 2: Geological map of the SW part of the Norwegian Caledonides, showing the disposition of the main units discussed in the paper. Modified from Gee et al. (1985). Seismic line of Mykkeltveit et al. (1980) shown, with solid, thicker part (M) indicating position of low velocity layer. Peak metamorphic temperatures shown in the Western Gneiss Region from Griffin et al. (1985) and Wilks and Cuthbert (1994). Distribution of top-to-hinterland (ie top-to-NW/WNW) faults from Fossen (1998) and Andersen (2000). 1-9 refer to stratigraphic columns in Figure 3. NSD – Nordfjord-Sogn Detachment; HD – Hornelen Detachment; Devonian molasse basins Ho – Hornelen; H – Hasteinen; Kv – Kvamshesten; So – Solund; External parts of Window Allochthon A-L – Aurdal-Laerdal; Bo – Borlaug; B – Beito; V – Vang; A – Atnsjøen; S – Spekedalen, St – Steinfjellet. Other localities E – Eikefjord, Hy – Hyllestad, Sk – Skjolden.

3). Along the 1st ramp, the décollement has cut 320 m down-section to the 2nd flat, below the Ediacaran Moelv Tillite and on the 2nd ramp it has cut down ~3 km to the base of the Tonian-Cryogenian Brøttum Formation, giving a total pre-Alum Shale sedimentary thickness of ~3.32 km in the Hedmark Basin (Fig. 3; Kumpulainen and Nystuen, 1985; Morley, 1986).

The Osen-Røa Nappe Complex was formed by epizone grade SE-directed (in the north) to diagenetic zone SSE-directed (in the south) thrusting (Morley, 1986; Robinson and Bevins, pers. comm., 1987). Balanced cross-sections have shown that shortening increases from <20% south of Oslo to ~60% in the Hedmark Basin, with a bulk shortening of ~50 % (Morley 1986), comparable to that in other parts of the Lower Allochthon (Townsend et al., 1986, 1989; Gayer and Greiling, 1989).

West of the Osen-Røa Nappe Complex, the Lower Allochthon comprises the Aurdal Duplex, which has imbricated ~350 m of Ediacaran to Ordovician strata (Fig. 3) and is underlain by ~10 m of autochthonous mid-Cambrian rocks (Hossack et al., 1985). Hossack et al. (1985) estimated shortening in the Aurdal Duplexes at 63% but this value should be regarded as an upper estimate, as the effects of top-to-hinterland faulting were not recognised.

2.1.2 Window allochthon

In the Spekedalen, Atnsjøen, Steinfjellet, Beito, Vang, Borlaug and Aurdal-Laerdal tectonic windows (Fig. 2), an external zone of Window Allochthon has been exposed. Within the Atnsjøen and Spekedalen windows, a <150 m thick cover succession has been correlated with the Moelv tillite (Ediacaran) to Orthoceras Limestone (Ordovician) succession in the Osen-Røa Nappe Complex (Fig. 3, Nystuen and Ilebekk, 1981; Siedlecka and Ilebekk, 1982). Greenschist-facies NW-SE stretching lineations in the cover are parallel to lineations in the Osen-Røa Nappe Complex (Nystuen and Ilebekk, 1981; Siedlecka and Ilebekk, 1982). Similarly, greenschist-facies ductile deformation in the Beito Window Allochthon formed ~NE-SW stretching lineations (Hossack, 1976). Whether these lineations are related to thrusting or to top-to hinterland normal faulting has not been demonstrated, but Andersen (1998) postulated that many tectonic contacts around the Window Allochthon in this area have a top-to-hinterland fault geometry (Fig. 2).

The Western Gneiss Region includes Baltic Shield basement (Gorbatschev, 1985). In the Skjølden area (Fig. 2), the Skjølden Basement Complex has similarities with the basement of the Window Allochthon in the Beito and Vang tectonic windows (Fillefjell-Beito Basement Complex; Milnes and Koestler, 1985). Near Døvrefjell and Trollheimen, the imbricated basement is unconformably overlain by ~200m of Ediacaran to mid-Cambrian sedimentary rocks (Gjevilvatnet Group; Fig. 3; Gee, 1980; Krill, 1980). To the southwest, the Askevatn psammite may have been deposited unconformably on the Western Gneiss Region basement in the Solund area (Fig. 2; cf. Hacker et al., 2003).

The Synnfjell Duplex, which overlies the Aurdal Duplex and imbricates a comparable ~350 m Ediacaran to Ordovician sequence (Fig. 3), has previously been included in the Lower Allochthon (Gee et al., 1985; Hossack et al., 1985; Fig. 1). However, in the model presented below, it is restored to above the Western Gneiss Region and hence is taken to be part of the Window Allochthon (Fig. 2). Hossack et al. (1985) estimated 84% shortening in the Synnfjell Duplex, but this very high value must be regarded as an over-estimate, since the effects of top-to-hinterland faulting were not taken into account. To the northwest, under Jotunheim, the strongly deformed Vang and Fortun Formations have been interpreted as internal Synnfjell Duplex correlatives (Hossack et al., 1985; Milnes and Koestler, 1985);

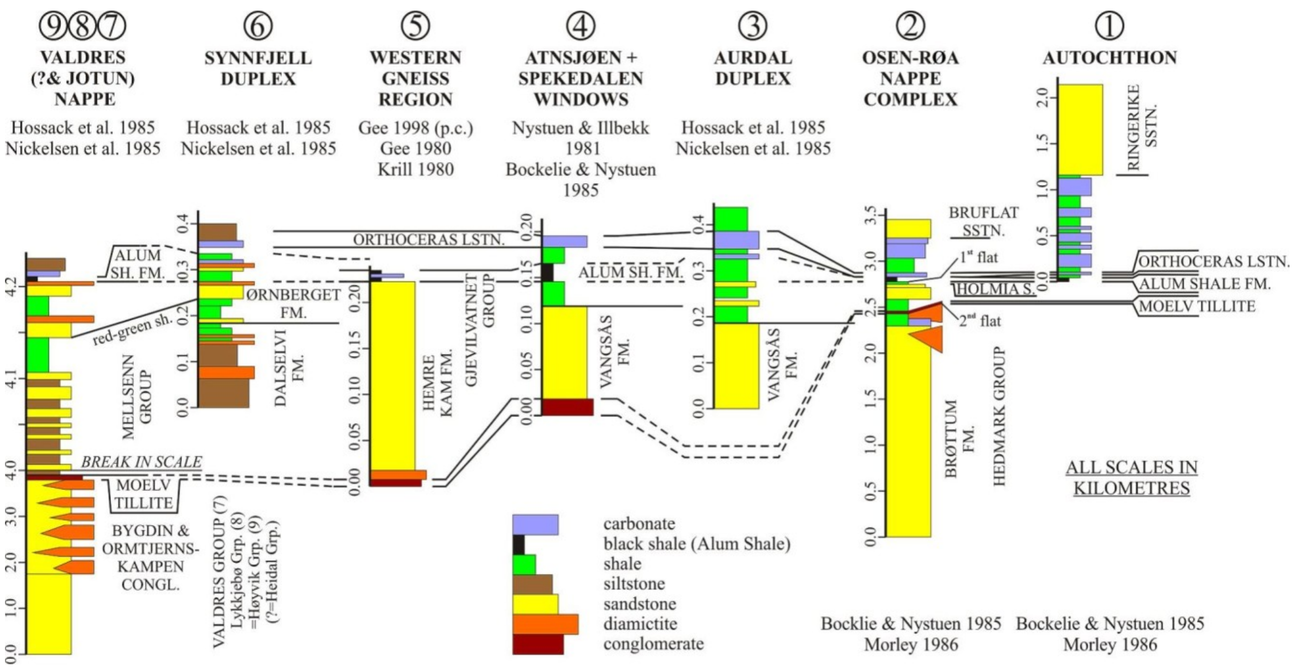

Figure 3: Simplified stratigraphic columns of the main units used in the model proposed here. 1-9 refer to the localities in Figure 2. Note that the vertical scales vary between and within individual columns. For the Osen-Røa Nappe Complex, the base of the stratigraphy preserved in the first and second flats is indicated.

these correspond to the Decollement Phyllites of Fossen and Dunlap (1998) and Fossen and Dallmeyer (1998). The underlying Lower Bergsdalen Nappes are an imbricate of the Western Gneiss Region (Gee et al., 1985) and are here taken as a relict basement plinth to part of the Synfjell Duplex.

Caledonian metamorphism rises from ~15 kbar/600°C in the southeast of the Western Gneiss Region to ultra-high pressure conditions (~120 km/>800°C) in the northwest (Griffin et al.; 1985; Krill, 1985; Dobrzhinetskaya et al., 1995; Austrheim et al., 1997; Wain, 1997; Cuthbert et al., 2000; Terry et al., 2000; Walsh and Hacker, 2004; Root et al., 2005). The Western Gneiss Region underwent severe top-to-hinterland deformation during and after peak metamorphism. This is detailed below.

Seismic profiling (Mykkeltveit et al., 1980) across the Western Gneiss Region revealed a 4 km thick seismic low-velocity layer at ~14 km depth, dipping ~2° W (Fig. 2), interpreted as a segment of oceanic sediments lying between the Baltic Shield and a Laurentia-derived Western Gneiss Region. In the model presented here, this zone has been re-interpreted as a relict of the Hedmark Basin. These data constrain the Western Gneiss Region Window Allochthon to a thickness of ~14 km. As the total crustal thickness in this area has been estimated at ~40 km (Bungham et al. in Mykkeltveit et al., 1980), ~22 km of continental crust below the low-velocity layer is indicated.

2.1.3 Middle Allochthon

The geology of this level of the orogen is relatively unclear, due to uncertainties in its structural development. The lowest unit, the Valdres Nappe, comprises 1.5 km of basement overlain by 4.6 km of Tonian to Cambrian sediments (Valdres/Mellsenn Groups; Fig. 3). The Tonian syn-rifting Bygdin and Ormtjernskampen Conglomerates, both several kilometres thick, accumulated close to actively uplifting basement (Nickelsen et al., 1985); the presence of Ediacaran diamictites lying directly on the basement reflects this topography. The Valdres Nappe was emplaced by SE-directed ductile deformation (Hossack et al., 1985). Marked thinning of the nappe northwestwards under Jotunheim (Hossack et al., 1985; Fig. 2) was caused by later top-to-hinterland deformation (Milnes et al., 1997). The Valdres Nappe has been overthrust by the Jotun Nappe, a thick basement sheet, comparable to Valdres basement, overlain by Valdres Group cover sediments (Hossack et al., 1985; Emmett, 1996; Milnes et al., 1997). Relatively minor Caledonian deformation occurred within the Jotun Nappe, although locally the sequence has been overturned (Milnes et al., 1997). Banham et al. (1979) noted a tectonic melange of serpentinite, gabbro and greenstones below the Jotun thrust; whether these represent a dismembered ophiolite, as originally proposed, and the tectonic level from which it could have been derived, has not been determined (cf. Milnes et al., 1997).

Around Eikefjord (Fig. 2), Bryhni (1989) described deformed sediments of the Lykkjebø Group, comparable with the Valdres Group, infolded with lithologies (Eikefjord Group) similar to basement rocks in the Jotun Nappe. Thus the Lykkjebø and Eikefjord Groups, which have been separated from the Western Gneiss Region by a top-to-hinterland detachment (Andersen and Jamtveit, 1990; Krabbendam and Dewey, 1998; Fig. 2) are part of the Middle Allochthon.

On Atløy (Fig. 2), the Dalsfjord Complex, which also has similarities to Jotun basement (Brekke and Solberg, 1987; Corfu and Andersen, 2002), is overlain by 2.1 km of upper Precambrian sedimentary rocks (Høyvik Group, equivalent to the Valdres Group; Fig. 3). These were deformed prior to the unconformable deposition of the Wenlock Herland Group on both the Dalsfjord Complex and the Høyvik Group (Andersen et al., 1998).

Northeast of Jotunheim, Sturt et al. (1997) defined an Otta Nappe, containing rocks from the Lower, Middle and Upper Allochthons of Gee et al. (1985; the Otta Nappe is not shown in Fig. 2). This includes the pre-Arenig Vågåmo Ophiolite, stitched to a deformed substrate (Høvringen Gneiss basement and Heidal Group cover) by the Arenig/Llanvirn Sel Group (Sturt et al., 1991; Bøe et al., 1993). The basement and cover have similarities with the Jotun basement and Valdres Group, respectively (Emmett, 1996). However, both these correlations and the existence of a Vågåmo Ophiolite have remained contentious (cf. Andersen, 1998). Thus, they have not been directly included in the model presented here, although they were used in previous, similar models (Rice, 1999, 2001a).

2.1.4 Upper Allochthon

Nappes derived from an oceanic realm (Köli nappes) outboard of the Middle Allochthon are represented only in the extreme west of the area, in the hanging-walls of top-to-hinterland normal faults (Fig. 2), where the Sunnfjord Mélange and tectonically overlying Solund-Stavfjord Ophiolite overthrust the Middle Allochthon on Atløy (Andersen et al., 1998). The sediments overlying the ophiolite contain both Precambrian lithologies and early Caledonian arc detritus, but were incorporated into the orogen in the Scandian event (Dunning and Pedersen, 1988, 1993).

2.1.5 Molasse sediments

Apart from the mid- to Upper Silurian molasse in the Autochthon at Langesund, south of Olso (Ringerike Sandstone; Figs. 2 and 3), Devonian-Carboniferous molasse sediments have been mapped in several 'collapse' basins lying on the hanging-walls of top-to-hinterland normal faults along the Norwegian coast (Figs. 1 and 2). In the transect considered here, the basins are separated from the Western Gneiss Region by the Nordfjord-Sogn, Hornelen and Kvamshesten Detachments (Hossack, 1984; Norton, 1987; Osmundsen and Andersen, 1994; Krabbendam and Dewey, 1998). These faults, and the Western Gneiss Region to the north (cf. Root et al., 2005), have been deformed by large-scale upright folds with axes parallel to the extension direction, with sediments preserved in the synclines; such fold structures are typical of extensional terranes (Mancktelow and Pavlis, 1994). Only weak deformation and very-low-grade metamorphism has been reported in the molasse sediments (Roberts, 1983; Torsvik et al., 1986). Although probable Early Devonian fossils (post-416 Ma) have been found in the Solund Basin (Steel et al., 1985), no clasts derived from the Western Gneiss Region

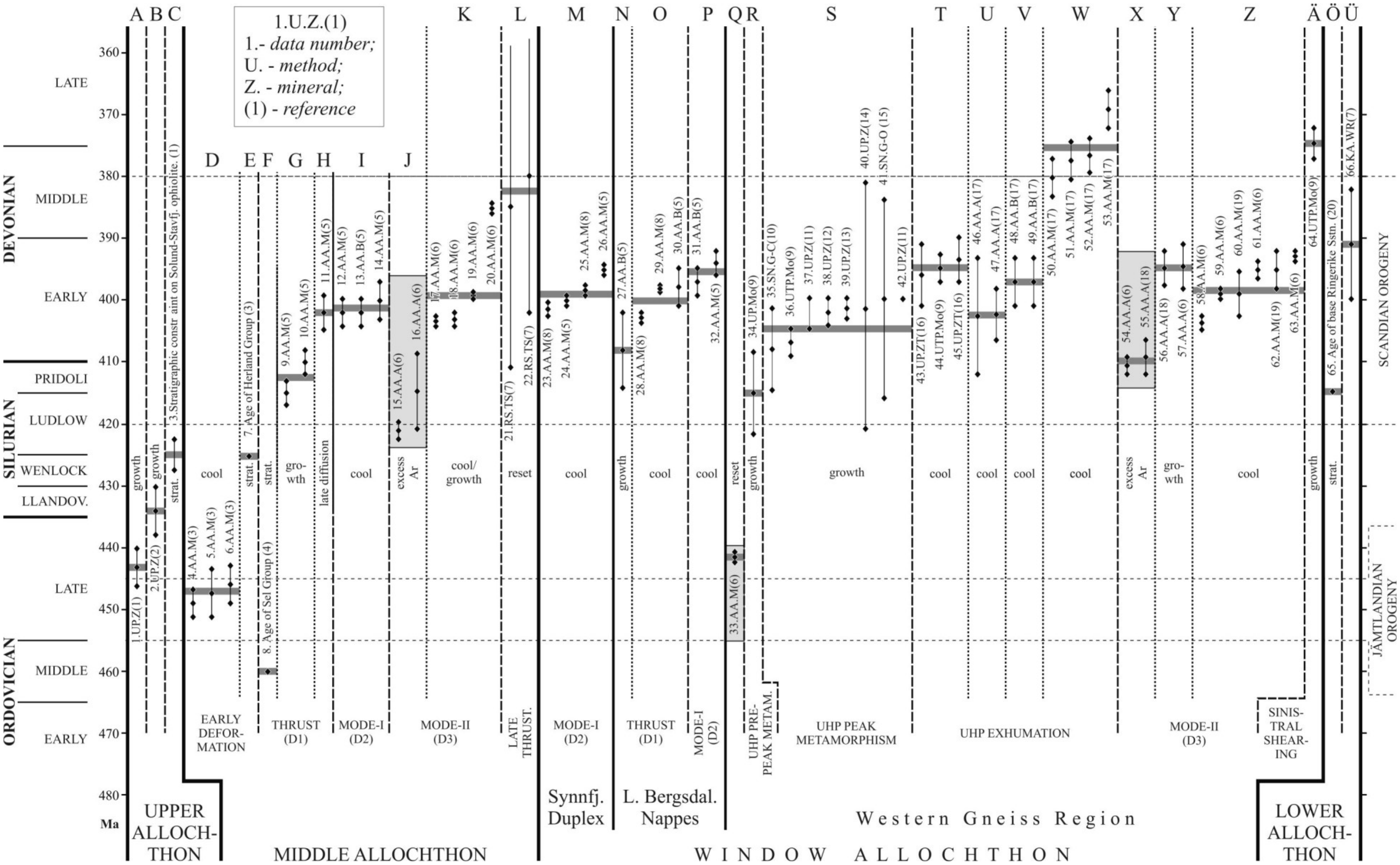

FIGURE 4: Summary of recently published geochronological and stratigraphic constraints on the evolution of the SW Norwegian Caledonides (excluding the Bergen Arc area). Methods used - UP – U-Pb; UT – U-Th-Pb; AA – $^{40}Ar/^{39}Ar$; SN – Samarium-Neodymium; RS – Rb-Sr; KA – K-Ar. Minerals/Fraction used - A – amphibole; B – biotite; C – clinopyroxene; M – muscovite; Mo – monazite; O – omphacite; Z – zircon; ZT – zircon+titanite; WR – whole-rock; TS – thin-slab; STRAT – stratigraphic constraint. Data from - 1 – Dunning and Pedersen (1988); 2 – Hacker et al. (2003); 3 – Andersen et al. (1998); 4 – Sturt et al. (1991, 1997); 5 – Fossen and Dunlap (1998); 6 – Chauvet and Dallmeyer (1992); 7 – Roberts and Sundvoll (1990); 8 – Fossen and Dallmeyer (1998); 9 – Terry et al. (2000); 10 – Carswell et al. (2003a); 11 – Root et al. (2004); 12 – Tucker et al. (2004); 13 – Carswell et al. (2003b); 14 – Krogh et al. (1974); 15 – Mørk and Mearns (1986); 16 – Tucker et al. (1986/7); 17 – Root et al. (2005); 18 – Boundy et al. (1996); 19 – Berry et al. (1994); 20 – Bockelie and Nystuen, (1985). Data in grey boxes have not been used to constrain the model.

have been found in the basins (Cuthbert, 1991), indicating that unroofing of the Western Gneiss Region post-dated deposition of the preserved parts of the basins.

3. Top-to-hinterland Faulting in S. W. Norway

Although complex in detail, only a broad picture of Caledonian top-to-hinterland normal-faulting in southwestern Scandinavia has been outlined below. In Fig. 2, all the mapped or postulated top-to-hinterland normal-sense shear-zones have been shown (cf. Andersen and Jamtveit, 1990; Andersen, 1998; Fossen, 1992, 2000).

Three stages of top-to-hinterland movement have been defined (Modes-I – III; Fossen, 1992, 2000). During Mode-I movement, WNW- to W-directed top-to-hinterland sub-horizontal shearing reactivated thrust contacts. This affected parts of the Jotun, Valdres and Synnfjell units and the top of the Western Gneiss Region (Wilks and Cuthbert, 1994; Fossen and Holst, 1995; Fossen and Dunlap, 1998; Milnes et al., 1997). During this event, the northwestern parts of the Synnfjell Duplex (the Vang and Fortun Formations) and Valdres Nappe lying under the Jotun Nappe were markedly thinned (Hossack et al., 1985, Milnes and Koestler, 1985; Fig. 2).

Mode-II deformation cut Mode-I shear zones along relatively steep W- to NW-dipping normal-sense ductile shear zones up to 6 km thick; Fossen (2000) estimated that the displacement on these structures was in the order of 50 km. The most westerly Mode-II shear zone is the Nordfjord-Sogn Detachment; in several places this has broadened out to separate a relatively low-strain block of the Middle Allochthon and higher units from both the underlying Western Gneiss Region and the overlying syn-Mode-II Devonian molasse (Andersen and Jamtveit, 1990; Chauvet and Séranne, 1994; Fig. 2). During Mode-II deformation, the Mode-I shear-zone was back-rotated and has been generally presumed to have become inactive (Fossen, 1992). However, because the geochronological data show large overlaps in Mode-I and Mode-II movement ages (see below; Fig. 4) they were probably partially coeval. Mode-III deformation was a continuation of Mode-II deformation, but at near-surface, brittle conditions (Fossen, 2000).

Stretching lineations and lineation parallel fold-axes rotate from ~E-W oriented in the bulk of the Western Gneiss Region to ENE-WSW oriented near the MTFZ. Krabbendam and Dewey (1998) suggested that this curvature indicates a component of ~NE-SW oriented sinistral strike-slip displacement during top-to-hinterland movement (transtension). The amount of transcurrent movement increases towards the coastal area, nearer to the NE-SW trending sinistral Møre-Trøndelag Fault Zone (Grønlie and Roberts, 1989) or its southwesterly projection (Fig. 1).

Andersen et al. (1991) proposed a west-directed subduction and eduction model for the Western Gneiss Region, with break-off of the continental lithosphere enabling exhumation of an essentially autochthonous Western Gneiss Region through the overlying crust by faulting. Burial of the crust by the thrust wedge has also been proposed as the burial mechanism (Milnes et al., 1997; Fossen, 2000), with exhumation of a parautochthonous Western Gneiss Region through post-orogenic plate-divergent extension. Krabbendam and Dewey (1998) regarded the Western Gneiss Region as a 'huge metamorphic core-complex', with exhumation resulting from post-collision regional transtension. In contrast, Breuckner (1998), Hacker et al. (2003) and Root et al. (2005) proposed west-directed subduction of the Western Gneiss Region to ultra-high pressure conditions, with subsequent exhumation along the subduction zone, but the palaeogeography and displacements were incompletely constrained in these models.

4. Timing of Deformation and Metamorphism

Fig. 4 summarizes recent geochronological and stratigraphic age constraints for the SW Scandinavian Caledonides. Data from the Bergen Arc have not been included, because its relation to the Western Gneiss Region has not been clarified. The data have been arranged from structural top at left (Upper Allochthon, Köli nappes) to base at right (Lower Allochthon). Within each allochthon, further subdivisions are defined by nappe units and then specified events, younging to the right.

Excluding the stratigraphic age constraints and the age of ocean crust formation in the Upper Allochthon (Fig. 4, cols. A-C) and the phase of Finnmarkian and Jämtlandian deformation within the Middle Allochthon (Fig. 4, cols. D, F) the data show little variation across the diagram; that is, between the Middle and Window Allochthons.

The Solund-Stavjord Ophiolite crust formed at 443±3 Ma, with deformation in the Köli nappes (Upper Allochthon) constrained by the intrusion of the Sogneskollen granite at ~434 Ma and stratigraphically to post ~425 Ma (Dunning and Pedersen, 1988; Hacker et al., 2003; Fig. 4, cols. A, B, C).

In the Middle Allochthon, Ar-Ar muscovite cooling ages from the Høyvik Group on Atløy (Fig. 2) indicate deformation prior to ~447 Ma, probably as part of the Jämtlandian event (Hacker and Gans, 2005; Andersen et al., 1998; Fig. 4, col. D). The age of subsequent deformation on Atløy is constrained only by the Wenlock (430-425 Ma) fossils in the Herland Group that unconformably overlies the Høyvik Group (Andersen et al., 1998; Fig. 4, col. E). Sturt et al. (1997) argued that early deformation in the Heidal Group (Otta Nappe) occurred prior to the Arenig/ Llandovery age of the unconformably overlying Sel Group (Fig. 4, col. F).

Samples containing thrust fabrics (D1) in the Valdres Nappe gave $^{40}Ar/^{39}Ar$ mica ages between 415±2 and 402±3 (Fig. 4, cols. G, H) but the latter age was interpreted by Fossen and Dunlap (1998) to reflect late diffusion, with ~412 Ma reflecting the age of thrusting. Samples with Mode-I top-to-hinterland fabrics gave ages between 402±2 and 400±3 Ma (Fossen and Dunlap, 1998; Fig. 4, col. I), whilst samples from the Middle Allochthon underlying Mode-II shear zones gave ages overlapping this range (403.3±0.7 to 385.1±0.7 Ma; Chauvet and Dallmeyer 1992; Fig. 4, col. K).

Mode-I fabrics in the Synnfjell Duplex gave $^{40}Ar/^{39}Ar$ mica ages of 401.5±1.1 to 395±1 (Fig. 4, col. M; Fossen and Dunlap, 1998, Fossen and Dallmeyer, 1998), comparable to Mode-I ages in the Middle Allochthon. In the Lower Bergsdalen Nappes, a ^{40}Ar/^{39}Ar mica age in rocks with top-to-southeast fabrics gave a slightly younger age (408±6 Ma; Fig. 4, col. N) than in the Valdres Nappe, although the data overlap within error and was interpreted to reflect the age of thrusting (Fossen and Dunlap, 1998). Younger ^{40}Ar/^{39}Ar mica ages in rocks with thrust fabrics were interpreted as cooling ages (Fig. 4, col. O; Fossen and Dunlap, 1998; Fossen and Dallmeyer, 1998). ^{40}Ar/^{39}Ar mica cooling ages of ~395 Ma in rocks with Mode-I fabrics (Fig. 4, col. P) are slightly younger than determined from the Middle Allochthon and Synnfjell Duplex, but overlap within error (Fossen and Dunlap, 1998).

In the Western Gneiss Region, early Sm/Nd and Rb/Sr determinations of the age of ultra-high pressure metamorphism gave results between 400-450 Ma, from which an 'average' age of 425 Ma was often used (Griffin and Breuckner, 1980; cf. Fossen and Dunlap, 1998). Recent U-Pb and U-Th-Pb ages of 407 to 400 Ma (Terry et al., 2000; Root et al., 2004; Tucker et al., 2004) are similar to Sm-Nd ages (408 to 400 Ma; Carswell et al. 2003a; Mørk and Mearns, 1986; Fig. 4, col. S). The 408-400 Ma range is compatible with a 415 Ma prograde metamorphic age (Terry et al. 2000; Fig 4.col. R) and overlaps with 402 Ma and younger top-to-hinterland Mode-I faulting cooling ages (Fig. 4, cols. I, M), which suggests that exhumation may have been active during ongoing burial or peak metamorphism of the internal Western Gneiss Region. In this article, an age of ~405 Ma has been taken as the time of peak metamor-phism; reasons for this are given in the Discussion.

Cooling of the Western Gneiss Region to below the Ar/Ar muscovite closure temperature occurred at ~380 Ma, after which ~E-W oriented folding of the coastal region occurred (Root et al., 2005; Fig. 4, cols. T-W). Compatible data have been determined in Mode-II shear zones (Fig. 4, cols. K, Z) Sinistral shearing on ENE-trending ductile shear zones, possibly a splay of the Møre-Trondelag Fault Zone, occurred at essentially the same time ~374.6±2.7 Ma (Terry et al., 2000; Fig. 4, col. Ä), suggesting a link between exhumation and Møre-Trondelag strike-slip faulting (cf. Mancktelow and Pavlis 1994). The model of Krabbendam and Dewey (1998) suggests that strike-slip faulting was at least in part contemporary with Mode-I top-to-hinterland movement.

The significance of the ^{40}Ar/^{39}Ar muscovite age of 441.6±1 (Fig. 4, col. Q; Chauvet and Dallmeyer, 1992) from basement rocks in the inner Sognefjord area and of the 421-409 Ma ^{40}Ar/^{39}Ar amphibole data (Chauvet and Dallmeyer, 1992; Boundy et al., 1996; Fig. 4, cols. Q, J, X) are unclear. They might be partially reset pre-Caledonian ages or a result of excess argon (Chauvet and Dallmeyer, 1992). Some amphibole ages from Mode-II shear zones are younger than muscovite ages (~394 Ma; Fig. 4, col. Y). More data of comparable ages are required before too great reliance is placed upon such seemingly anomalous results. Consequently, these data are not considered further here.

Difficult to constrain, but critically important, is the timing of deformation in the Lower Allochthon (Aurdal Duplex and Osen-Røa Nappe Complex). Deposition of the Ringerike Sandstone

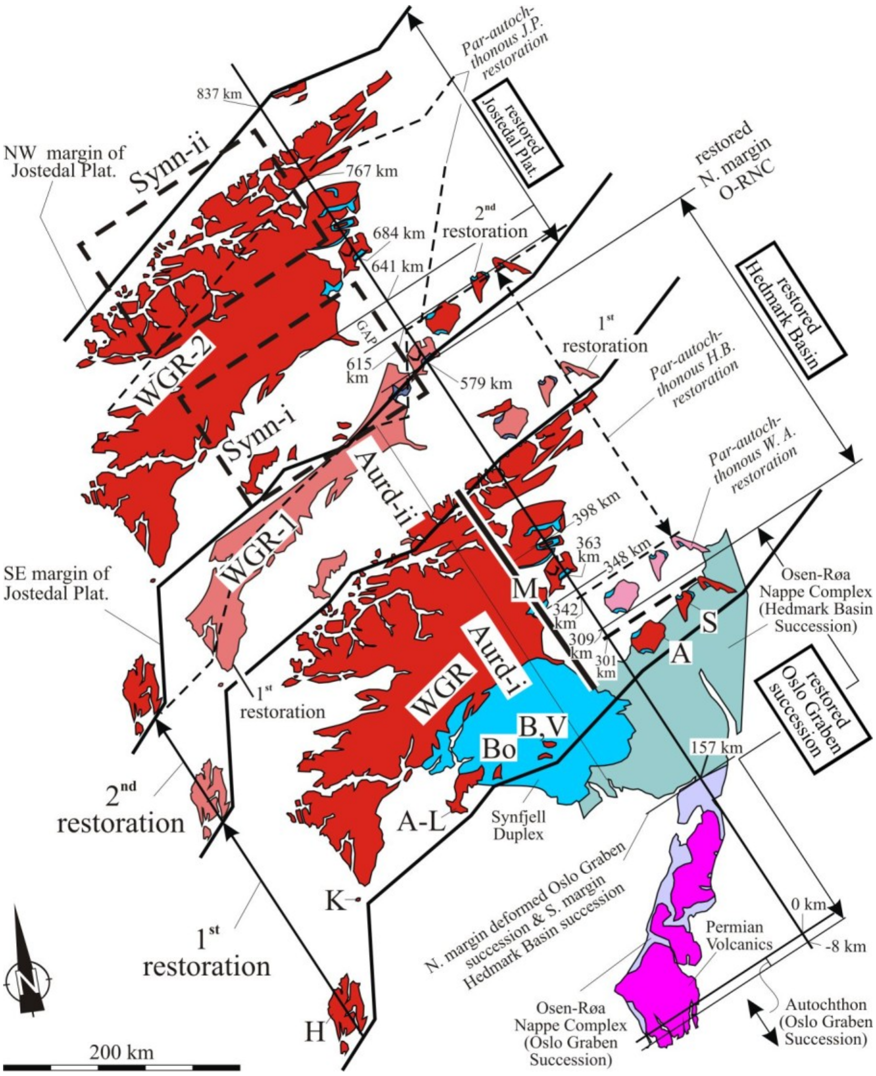

Figure 5: Branch-line restoration of the Scandinavian Caledonides in southern Norway, based on the balanced cross-sections of Morley (1986). M – Mykkeltveit et al. seismic line. The thicker bar is where a low velocity layer was identified. External zone of Window Allochthon shown in tectonic windows at - A – Atnsjøen; A-L – Aurdal-Laerdal; B – Beito; Bo – Borlaug H – Haugesund; K – Kikedalen; S – Spekedalen; V – Vang. S. D. – Synnfjell Duplex; A.D. – Aurdal Duplex; WGR – Western Gneiss Region. WGR-1, WGR-2 – successive restorations of the Western Gneiss Region; A.D. – restored Aurdal Duplex; Synn-i and Synn ii – restored positions of the Synnfjell Duplex. See text for further details.

commenced in late Wenlock to early Ludlow times (Bockelie and Nystuen, 1985; Fig. 4, col. Ö) but there are no constraints on the youngest age of sedimentation and thus no constraint on the age of deformation in the foreland. Fossen (2000) proposed that thrusting in the Ringerike Sandstone occurred at ~410 Ma, contemporary with thrusting in the Jotun Nappe, despite the large distance separating the trailing edge of the Osen-Røa Nappe Complex from the Ringerike Sandstone when deformation in the complex began. In contrast, Dewey et al. (1993) indicated that thrusting in the Oslo area occurred at ~380 Ma, although no grounds for this inference were given. However, ages of ~390 Ma have been obtained from the Caledonian basal décollement elsewhere in the Scandinavian Caledonides, with similar, although very imprecise ages determined from shear bands at the base of the Middle Allochthon (Roberts and Sundvoll, 1990; Fig. 4, cols. L and Ü). For reasons outlined in the Discussion, the end of thrust-related deformation has been estimated at ~388 Ma, prior to the end of Mode-II top-to-hinterland shearing.

5. Palinspastic Restoration

Before a model for the burial of the Western Gneiss Region to ultra-high pressure conditions and its subsequent exhumation can be proposed with any degree of certainty, the pre-orogenic palaeogeography of both the Western Gneiss Region and the adjacent nappes has to be established. Evidence to constrain this is given in the following section.

Restoration of balanced cross-sections through the Osen-Røa Nappe Complex places its trailing edge NW of the Norwegian coast (Morley, 1986; Fig. 5, 'restored N margin O-RNC'). As a result, the Tonian Brøttum Formation in the restored Hedmark Basin overlies the Ediacaran to Cambrian Gjevilvatnet Group, which rests unconformably on the Western Gneiss Region in the Døvrefjell-Trollheimen area (Gee, 1980; Krill, 1980; Figs. 2 and 3). This stratigraphic repetition can be resolved either by restoring the Hedmark Basin to the northwest of the Western Gneiss Region, whilst leaving the Olso Graben succession southeast of it (cf. Nystuen, 1987) or by displacing the Western Gneiss Region ~215 km northwestwards, to lie NNW of the restored Hedmark Basin (Fig. 5, '1st restoration', 'WGR-1'). The former model has been rejected as it divides the Osen-Røa Nappe Complex into two parts, for which there is neither structural nor sedimentological evidence. The Western Gneiss Region has thus been interpreted as having been part of a basement high, here termed the Jostedal Plateau (broadly comparable to the present day Rockall Plateau in the N. Atlantic) lying outboard of the Hedmark Basin (Fig. 6).

In restoring the Oslo Graben succession (1st and 2nd flats) of the Osen-Røa Nappe Complex, mid- to Upper Cambrian Alum Shales come to rest above the Ordovician Orthoceras Limestones of the Window Allochthon in the Atnsjøen-Spekedalen tectonic windows (Fig. 5). This indicates that the Alum Shales might be continuous under the basal décollement from Oslo to the Atnsjøen-Spekedalen area, with the external Window Allochthon units being parautochthonous (~47 km displacement; Fig. 5, 'Parautochthonous WA restoration'). If so, the Hedmark Basin must be restored to NW of the parautochthonous external units of the Window Allochthon, with the Western Gneiss Region also restored further NW (Fig. 5, 'Par-autochthonous H.B. restoration' and 'Par-autochthonous J.P. restoration'). However, since this model brings the external Window Allochthon units (and thus their thin basin margin-fill cover successions) very close (~6 km) to the low velocity layer of Mykkeltveit et al. (1980), the determined length of which was constrained only by the overall length of the seismic experiment, this restoration seems unlikely. For this reason, the basement rocks of the Atnsjøen-Spekedalen tectonic windows (and the other external units of Window Allochthon) are included as the southeasternmost part of the Jostedal Plateau (Fig. 6). The similarity of the Skjølden Basement Complex in the Western Gneiss Region and the Fillefjell-Beito Basement Complex in the Beito and Vang tectonic windows (Fig. 2; Milnes and Koestler, 1985) supports this restoration.

The gap between the presently observed trailing edge of the Atnsjøen-Spekedalen Windows and the leading edge of the Western Gneiss Region has been included in the Jostedal Plateau, giving a total plateau width of ~258 km along the section line. Restoring these segments to northwest of the Hedmark Basin (restored only to the northwest of the undeformed Oslo Graben succession) gives a further displacement of 105 km (Fig. 5, '2nd restoration', 'WGR-2'). There is a 70 km difference in total displacement for the Western Gneiss Region between the parautochthonous and the fully allochthonous models for the external units of the Window Allochthon (Fig. 5).

The unconstrained displacement of the top-to-hinterland normal faulting in the Vang and Fortun Formations (Fig. 2) precludes a balanced restoration. Such deformation may also have affected the Synnfjell Duplex itself. Assuming that the

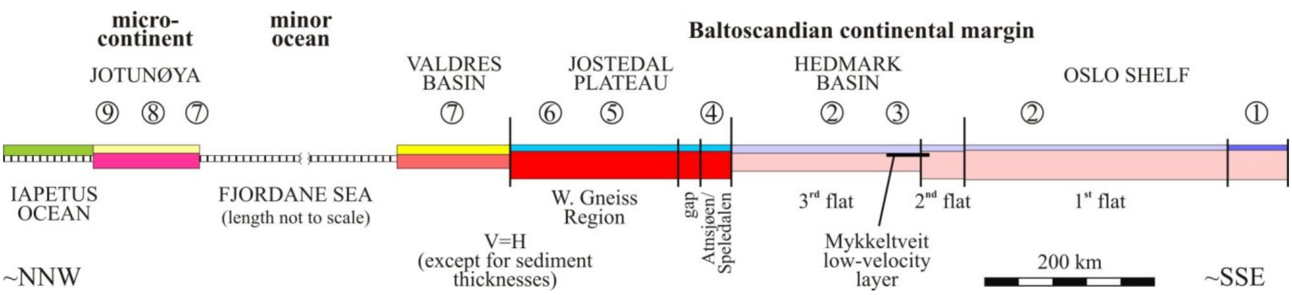

Figure 6: Cross-section showing the restored palaeogeography. See text for details. Horizontal dimensions of oceans are not constrained. Simple isostatic equilibrium based on filled pre-Ediacaran-glaciation basins has been assumed. 1-9 refer to the localities in Figure 2 and the stratigraphic columns in Figure 3.

original thrust related shortening was ~50%, typical for orogens (e.g. Morley, 1986; Townsend et al., 1986), and that this was essentially cancelled by later top-to-hinterland deformation, allows the present day dimensions of the Synnfjell-Vang-Fortun unit (~110 km) to be used as a very rough proxy of the unstrained palaeogeographic width. The restored length of the Aurdal Duplex, here presumed to be unaffected by top-to-hinterland deformation, was estimated at 89 km by Hossack et al., (1985). Thus the combined restored length of the duplexes and their internal equivalents was ~200 km. Although their palaeogeography is uncertain, both units chronostratigraphically overlap with the Mellsenn and Gjevilvatnet Groups (Gee, 1980; Krill, 1980, Nickelsen et al., 1985; Fig. 3). Their restored position must lie northwest of the Autochthon (Hossack et al., 1985), possibly overlying the margin of the Hedmark Basin and the Jostedal Plateau, to the southwest of the Gjevilvatnet Group (Fig. 5, 'Synn-i', 'Aurd i'). However, if the Aurdal Duplex continues further to the NW, under the Synnfjell Duplex, with the same shortening, its total restored length would be ~240 km and could have covered much of the Hedmark Basin (Fig. 5, 'Aurd-ii'). Re-storation of the Synnfjell Duplex only to northwest of the Gjevilvatnet Group (Fig. 5, 'Synn-ii') results in a possible overlap with the Mellsenn Group, but the similarity of the purple-green shales below the inferred base of the Cambrian in the Mellsenn Group and Ørneberget Formation (Fig. 3; Nickelsen et al., 1985) does suggests a close palinspastic proximity. Essentially, since the leading edge of both duplexes has been eroded, it seems likely that originally they imbricated the Ediacaran to Ordovician sequence of the entire Jostedal Plateau to Autochthon segment along the section-line.

Hossack et al. (1985) estimated the restored widths of the Valdres and Jotun Nappes at 132 and 124 km respectively. Although these values also ignored the top-to-hinterland deformation, they are used here; Milnes et al. (1997) noted that much of the Jotun Nappe was only slightly affected by Caledonian deformation. The Valdres Nappe has been restored to directly northwest of the Western Gneiss Region, with a thinned continental crust. The presence of Ediacaran diamictites locally lying directly on the basement in the Valdres Nappe (Nickelsen et al., 1985) is taken to reflect thickening of the crust adjacent to the NW-margin of the Jostedal Plateau (Fig. 6). This restoration requires a ~600 km displacement for the Valdres Nappe, considerably more than Hossack et al. (1985) proposed, due to the allochthonous status of the Western Gneiss Region, on which the Valdres Nappe was carried piggyback much of the time.

Restoring the Jotun Nappe presents difficulties, because although it is lithologically similar to the Valdres Nappe (Hossack et al., 1985; Emmett, 1996), its root was not subducted, as that of the Valdres Nappe has been inferred to have been in the model outlined below. Emmett (1996) speculated that the Jotun Nappe represented a microcontinent rifted from Baltica, with an intervening ocean along the section line; this concept has been adopted here. The tectonic melange with mafic and ultramafic detritus under the Jotun Nappe (Banham et al., 1979) might be evidence for this ocean, although an ophiolitic origin for the melange is not certain (Milnes et al., 1997). The rifting age was probably similar to that of the Iapetus rift-drift transition at ~650-550 Ma (Claesson and Roddick 1985; Zwaan and van Roermund, 1990; Bingen et al., 1998b; Svenningsen, 2001; Rice et al., 2004). The microcontinent and minor ocean are here called Jotunøya and the Fjordane Sea, respectively (Fig. 6).

6. Model for ultra-high Pressure Metamorphism and Exhumation

The foregoing sections have placed the regional geology into a palinspastic framework and summarised the constraints on the timing of deformation and metamorphism. In the following section, a new tectonic model has been proposed for the development of ultra-high pressure rocks in the Western Gneiss Region and for their subsequent exhumation, using published lengths and stratigraphic and crustal thicknesses. The model has assumed plane strain and has been 'area-balanced' with a ramp-flat type geometry. That these assumptions are not valid in rocks which underwent intense ductile deformation (Meakin, 1983; Krill, 1985; Vollmer, 1988; Robinson, 1995; Terry and Robinson, 2004a, b) is fully recognised, but it is thought to be a better approach than one in which material is gained or lost from the profile without justification or control. Smoothing out the ramp-flat geometry would not affect the model presented, which is essentially a first-order approximation. Since the late- to post-Caledonian sinistral strike-slip movement, partly or wholly on the Møre-Trondelag Fault Zone, was essentially orthogonal to the modelled section-line (cf. Krabbendam and Dewey, 1998; Terry et al., 2004a, b; Fig. 1), it has been assumed that material moved into the modelled section line at the same rate as equivalent material moved out. The cross-sections have all been drawn relative to a datum line (d.l., Fig. 7) at the base of the mid-Cambrian Alum Shale in the Caledonian foreland. Inferred process rates are given in the Discussion.

The early model stages have been covered very simplistically, as they relate to rocks only locally exposed and relatively poorly understood in this transect (cf. Breuckner and van Roermund, 2004; Hacker and Gans, 2005). Collision at the NW-margin of Jotunøya occurred pre-447 Ma, probably as part of the Jämtlandian Orogeny, deforming the Heidal Group and the Dalsfjord Suite (Fig. 7A; Andersen et al., 1998). Southeast-directed obduction of back-arc basin oceanic crust (formed at 443 Ma), occurred after 425 Ma, seen in the Solund-Stavfjord Ophiolite, together with the other Köli nappes (Fig. 7B). This deformation, which was broadly a consequence of the collision of the west side of Jotunøya with Laurentia, was followed by the initial development of the Jotun Nappe on Jotunøya, in the footwall of the Köli nappes.

Subduction of the Fjordane Sea under the SE-margin of Jotunøya commenced at an early stage, possibly in Finnmarkian times. During the subsequent Baltica-Jotunøya collision, some time after ~425 Ma (Fig. 7), the basement of the former began to subduct, whilst the Valdres Group was imbricated and remained near the surface. The top of the Jostedal Plateau, together with the cover sediments forming the Vang and Fortun

Formations and the Synnfjell Duplex, were then excised by footwall shortcut thrusting, thinning the crust (footwall part of the Jostedal Plateau) as it subducted (Fig. 7C). At the same time, the leading edge of the orogenic wedge, driven by the collision with Laurentia and floored by the Jotun Nappe, began to move from Jotunøya onto the developing Valdres Nappe. This occurred at ~410 Ma, since $^{40}Ar/^{39}Ar$ muscovite data give D1 (thrusting) ages of ~412 and 408 Ma in the Valdres Nappe and Synnfjell Duplex. During this period, the Jotun Nappe was shortened by 41%, from a pre-orogenic length of ~124 km to a modelled 73 km.

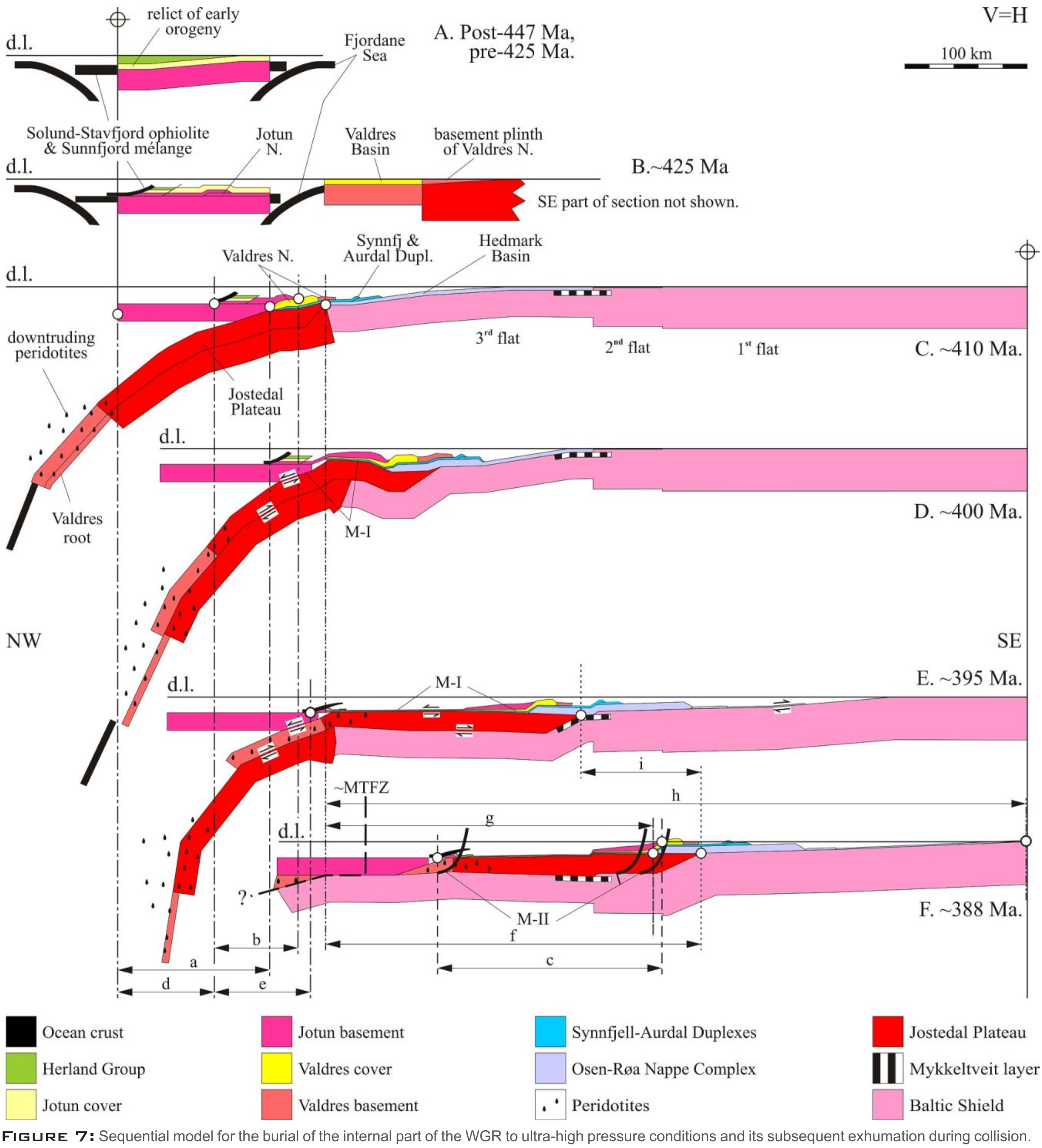

FIGURE 7: Sequential model for the burial of the internal part of the WGR to ultra-high pressure conditions and its subsequent exhumation during collision. Note that the vertical and horizontal scales are the same and that the profiles are area balanced. Subduction zone geometry based on Chemenda et al. (1986; fig. 7a, b). The sections are all drawn relative to a datum line (d.l.) taken as the base of the Alum Shale at the start of its deposition. Preserved overlying sequences are generally thin. The SE part of the 410 Ma profile has not been drawn, but is the same palaeogeographically as for the 425 Ma profile. See text for further details.
a: 124 km - undeformed length of Jotun Nappe. b: 73 km - thrust-shortened length of Jotun Nappe. c: 182 km - top-to-hinterland stretched length of Jotun Nappe.d: 79 km - displacement of trailing edge of Jotun Nappe between ~425-410 Ma. e: 78 km - displacement of trailing edge of Jotun Nappe between ~410-395 Ma. f: 306 km - displacement of leading branch-line of Jostedal Plateau. g: 266 km - displacement of trailing branch-line of Osen-Røa NC. h: 572 km - propagation of deformation through Osen-Røa NC (i.e. undeformed length of the ORNC). i: 98 km - displacement of Jostedal Plateau after exhumation. M-I - mode I shear zone. M-II - mode II shear zone

At ~410 Ma, convergence between Baltica and Jotunøya (subduction) ceased, although collision between Laurentia and Jotunøya-Baltica continued, driving thrusting. Peak ultra-high pressure conditions (408-400 Ma; Fig 4. col. S) were contemporary with nearer-surface Mode-I deformation, defined by cooling ages of 403 Ma and younger in the nappes (Fig. 4, cols. I, M). Exhumation of the leading edge of the Window Allochthon contemporary with ongoing burial and peak metamorphism of the trailing edge can be modelled with subduction zone steepening (cf. Chemenda et al., 1996). In Fig. 8, the Baltoscandian continental margin has been superimposed on the subduction curves of Chemenda et al. (1996), with the southeast end of the Jostedal Plateau (leading edge of the Window Alloch-thon) at its peak metamorphic ~mid-greenschist facies depth. In curve A, this brings the leading edge of the Jostedal Plateau to a depth of ~95 km. Steepening of the subduction zone to curve B takes this point to ~142 km depth (line X, Fig. 8), deeper than the ultra-high pressure metamorphism recorded in the Western Gneiss Region. However, if the upper part of the Jostedal Plateau separates from its substrate (forming the Window Allochthon), the lower, high-density part (mantle plus lower crust) could sink rapidly, steepening the subduction zone. With a high steepening rate, the buoyant Window Allochthon could have started to move 'upwards' relative to its footwall, whilst effectively still sinking. Since the length required for a subduction zone to reach a given depth decreases as the subduction angle steepens, this would have caused the shallow part of the Window Allochthon to start exhuming (compare lines Y and Z, Fig. 8). Contemporary with the subduction of the leading edge of Baltica, downtrusion of garnet peridotites from under Jotunøya (Baltica-type lithosphere) occurred (Fig. 7C and D; Breuckner, 1998).

During the inferred subduction zone steepening, the orogenic wedge moved farther southeastwards onto Baltica. By ~400 Ma, the exhuming Window Allochthon had significantly intruded under the orogenic wedge, below the Synnfjell Duplex. The rapid vertical exhumation of the ultra-high pressure rocks resulted in a horizontal displacement of the leading edge of the Window Allochthon that was greater than the continental collision rate. Hence, those parts of the Middle Allochthon and Synnfjell Duplex lying on the leading edge of the Jostedal Plateau were carried piggy-back towards the foreland faster than parts still resting on Jotunøya, forming the Mode-I top-to hinterland detachments and stretching and thinning the nappes. During this deformation, the Jotun Nappe was stretched from its initial thrust-shortened length of ~73 km to ~182 km, an extension of 149 %, and an extension of 47 % from it initial unstrained modelled sedimentary length (124 km).

Initially, the Window Allochthon intruded under the Hedmark Basin, whilst the nappes, perhaps controlled by the pre-existing décollement at the base of the Synnfjell Duplex, overrode the basin. After ~40 km of underthrusting, shortening in the Hedmark Basin started, forming the Osen-Røa Nappe Complex. This scenario is necessitated by the exposure of the Osen-Røa Nappe Complex above the Spekedalen, Atnsjøen, and Steinfjellet Window Allochthons (Fig. 2).

In the model, exhumation of the Window Allochthon was essentially completed by ~395 Ma (Fig. 7E); resistance to further exhumation outgrew the body force exerted by the root of the Valdres Nappe. At this stage, the 'Laurentian' orogenic wedge lay on the margin of Jotunøya, whilst the leading edge of the Window Allochthon, together with the external part of the Middle Allochthon, overlay the buried sedimentary layer identified by Mykkeltveit et al. (1980), indicating a considerable stretching of the Middle Allochthon and Synnfjell Duplex. The leading edge of the developing Osen-Røa Nappe Complex lay further towards the foreland, having climbed up onto the first flat. By ~388 Ma, collision had ceased; reasoning for this date is given in the Discussion.

Mode-II extension might also have started as early as ~405 Ma (Fig. 4, cols. K, M, Z), contemporary with Mode-I shearing and the latter part of exhumation, and continued beyond the ~388 Ma inferred end of thrusting.

7. Discussion

7.1 Palaeogeography

The dimensions used in the model have been taken from balanced cross-sections, log-sections or direct measurements of geological maps, or are assumptions discussed in the text. These values will change as more work is done; the undeformed lengths of the Jotun and Valdres Nappes and Synnfjell Duplex given in Hossack et al. (1985) are probably erroneous, since top-to-hinterland movements were not recognised at that time. However, changing the length of the Valdres Nappe would not significantly affect the model, although it would alter the maximum depth to which Baltica continental crust descended.

The basin geometry proposed, with the Window Allochthon and parts of the Valdres Nappe forming a topographic high (Jostedal Plateau) northwest of the restored Lower Allochthon (Fig. 6) is similar to that in central and northern Scandinavia (Gayer and Roberts, 1973; Gayer and Greiling, 1989; Gayer and Rice, 1989). In southern Norway, Neoproterozoic extension produced a more complex geometry, as Jotunøya rifted away from Baltica in the section drawn (cf. Emmett, 1996). The resulting continental margin was probably comparable to the early stages of Atlantic rifting, when continental extension developed as a series of linked ridge-ridge-ridge junctions (Burke, 1976). Although the width of the Fjordane Sea is unconstrained, it is envisaged to be tectonically and palaeogeographically similar to the present-day Labrador Sea (~1000 km wide).

The microcontinental model for the Jotun Nappe requires a subduction zone to close the Fjordane Sea during Caledonian orogenesis. Evidence for this ocean is lacking, although possible ophiolitic material occurs at the base of the Jotun Nappe (Banham et al., 1979) and in the pre-Arenig Vågåmo Ophiolite (Sturt et al., 1991, 1997). Chemenda et al. (1995, 1996) showed that, in their models, continental subduction can be induced by a pull from a pre-existing oceanic subduction zone and Breuckner (1998) noted that something must have pulled the WGR down; in the model, this is the subducting Fjordane Sea.

7.2 Model choice

The model proposed is broadly based on the analogue models of Chemenda et al. (1995, 1996). For exhumation of ultra-high pressure rocks to the surface, Chemenda et al. (1996) proposed a high pull-force model, with the leading-edge of the exhuming block at the base of the overlying lithosphere; thus ~120 km burial conditions should be the lowest pressures recorded in the exhumed block. The requirements for the high pull-force model occurred in the Baltoscandian margin: Crustal thinning occurred around the Jostedal Plateau, forming the Valdres Basin, with consequent thickening of the lithospheric mantle during thermal relaxation, increasing the bulk lithospheric density. Excision of both the sedimentary cover and the upper part of the Jostedal Plateau, as the basement plinth of the Valdres Nappe would also have increased the bulk density, as would the locally common eclogites and peridotites in the Western Gneiss Region (Griffin et al., 1985; Wain, 1997). Despite this, a leading edge close to the surface is clearly indicated by the greenschist facies peak metamorphic conditions in the external part of the Window Allochthon. This implies a low-pull force model (Chemenda et al., 1995), in which ultra-high pressure rocks are not brought to the surface. Essentially, it is here thought that the Chemenda et al. (1995, 1996) models show that continental crust can be subducted to great depths and return to the surface; detailed regional variations, beyond the control of analogue models will likely determine the precise geometry of both burial and exhumation.

Exhumation of ultra-high pressure rocks by channel-flow (Mancktelow, 1995) is conceptually very attractive and it is thought to have probably operated to an unknown degree. However, incorporating it in the model proved to be difficult, although complex ductile deformation occurred in the Western Gneiss Region, both shortening and extending it and incorporating the overlying nappes (Meakin, 1983; Krill, 1985; Vollmer, 1988; Robinson, 1995; Terry et al., 2004a, b). Since the present length of the Window Allochthon is sufficient to account for burial to ultra-high pressure conditions along subduction zones dipping at reasonable angles, any internal deformation within it, due to channel-flow and other processes (cf. Boutelier et al., 2002), seem to have balanced out, so far as the model presented here is able to suggest.

7.3 Movement rates

The proposed model for ultra-high pressure metamorphism in the Western Gneiss Region requires that ~250 km of Baltoscandian margin was subducted below Jotunøya (see Fig. 7C) along a curved subduction zone. If the Fjordane Sea had sub-ducted at a rate of ~5 cm yr^{-1}, and had had an original width of ~1000 km, the subduction initiation could be linked to the inter-Finnmarkian-Scandian event noted above. Alternatively, taking a rate of 1.7 cm yr^{-1}, very slow for an oceanic subduction zone, subduction would have started at 485 Ma, perhaps accounting for the Vågåmo Ophiolite (Sturt et al., 1991). Considering the lack of constraints, the Fjordane Sea is shown to have already started subduction in Fig. 7A with the Jostedal Plateau about to subduct in Fig. 7B.

Exhumation of the Western Gneiss Region brought it up from a maximum of ~120 km depth to lie under the nappes, at a depth here taken to be ~20 km, (cf. Root et al. 2005; Walsh and Hacker, 2005), giving an exhumation of ~100 km. After this, exhumation was driven by a combination of erosion and tectonic unroofing through Modes-II and III faulting. U-Pb cooling ages (cessation of Pb-loss) within the Western Gneiss Region show a narrow range, with U-Th-Pb data from monazite giving ages in a bracket between 396.1±4.9 to 394±2.3 (Fig. 4, col. T); Tucker et al. (2004) give a combined age of 395±3 Ma, based on a larger data set. $^{40}Ar/^{39}Ar$ muscovite cooling ages go down to ~370 Ma (Fig. 4, col. W). If an ultra-high pressure metamorphism age of ~405 Ma is taken, in the middle of the available age data (408-400 Ma, Fig. 4 col. N) an exhumation rate of 10 mm yr^{-1} (100 km in 10 Ma) is inferred. This is comparable to the fastest active exhumation rates recorded (Blythe, 1998) and similar to those determined by Terry et al. (2000a;10.9 mm yr^{-1}), Carswell et al. (2003a; 10 mm yr^{-1}) and Hacker et al. (2004; >2.5 – 8.5 mm yr^{-1}).

By taking an average 45° dip angle, for the exhumation channel, the 10 mm yr^{-1} rate of vertical exhumation of the Western Gneiss Region translates into a horizontal displacement of ~14 mm yr^{-1}, defining the lateral displacement rate of the leading edge of the Window Allochthon in Fig. 7. In contrast, the collision rate of Laurentia in the model is constrained by lengths d and e in Fig. 7, both of which indicate a displacement of ~78 km in ~15 Ma, giving a shortening rate of ~5.2 mm yr^{-1}, considerably slower than the lateral emplacement rate of the Window Allochthon.

Initially, the exhuming block underthrust the Hedmark basin for ~40 km before imbrication of the latter into the Osen-Røa Nappe Complex commenced. The trailing branch line of the complex was then displaced towards the foreland at the same rate as the Western Gneiss Region (14 mm yr-1). By assuming that the thrusting rate remained the same after exhumation

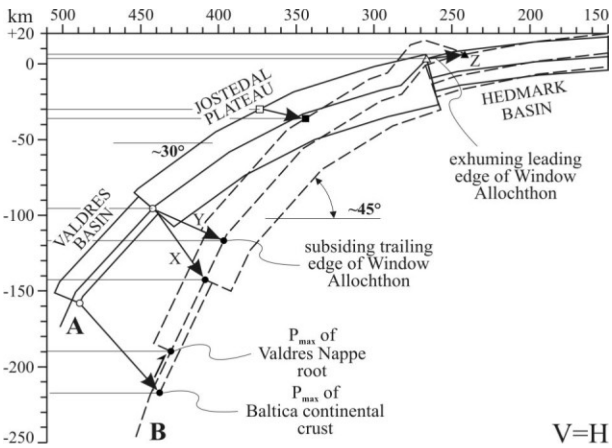

Figure 8: This shows the effects of superimposing the palaeogeography of the Jostedal Plateau and Valdres Basin (see Fig. 7) onto the subduction curves of Chemenda et al. (1996; fig. 7a, b). The external part of the Jostedal Plateau, formed of the Atnsjøen and Spekedalen Window Allochthons was placed at ~mid-greenschist facies depths, with the WGR lying at greater depths, using the lengths currently observed in SW Norway. Line X reflects shows burial path for rocks which did not detach from the sinking lithosphere, Line Y shows the path taken if the Jostedal Plateau detached as soon as subduction zone steepening began.

ceased (i.e. for displacement g, Fig. 7), final deformation in the external part of the Osen-Røa Nappe Complex can be inferred at 388 Ma, close to the age determined for final thrusting in Finnmark (390 Ma, Fig. 4, cols. L, Ü; Roberts and Sundvoll, 1990). If, however, the rate slowed to the thrusting rate in the hanging-wall of the Western Gneiss Region (5.2 mm yr-1), thrusting in the Lower Allochthon would have ceased at ~375 Ma.

7.4 Fault Geometries

For thrust faults, a ramp-flat geometry has been used. That this is not valid is acknowledged, but the advantage in ensuring that the model is essentially plane strain outweighs the disadvantages; it is altogether far too easy to draw models in which material is gained or lost without proper control. Such losses and gains can make models spuriously appear plausible. The unrealisticappearing buckle in the footwall of the exhuming Western Gneiss Region (Fig. 7D) is an obvious result of this geometry.

Although sinistral strike-slip motions during Caledonian collision have been proposed (Soper et al., 1992; Krabbendam and Dewey, 1998; Terry et al., 2004a, b) the assumption made here is that material moved into the plane of the section at the same rate as comparable material moved out, giving the effect of plane-strain in a 2-D vertical profile. So long as the strike-slip component is relatively small, this assumption is not unreasonable.

An important point in the model is that no absolute extension occurred across the orogen (Laurentian foreland to Baltica foreland) during exhumation of the Western Gneiss Region (Mode-I movement). Top-to-hinterland structures formed because the Window Allochthon moved more rapidly (~14 mm yr^{-1}) towards the southeast than the overlying Laurentian driven nappe pile (~5.2 mm yr^{-1}), resulting in significant stretching of the nappes draped across the junction between Jotunøya (Laurentia driven) and the leading edge of the Window Allochthon. Using this logic, the regional, plate-scale deformation pattern during Mode-I faulting was transpressional rather than transtensional as suggested by Krabbendam and Dewey (1998), although at the roof of the Western Gneiss Region, transtension did occur.

Isotopic data suggest that Mode-I faults may have been active during peak ultra-high pressure metamorphism and thus before exhumation had started. Subduction-zone steepening would allow contemporary exhumation of material at the upper end of the subduction channel at the same time as the lower part was still moving to greater depths (Fig. 8). The shallower the initial subduction angle, the more pronounced this effect could become.

Mode-II faults also moved at least partly synchronously with Mode-I faults (Fig. 4). Since, as argued above, Mode-I was syncollisional, Mode-II must have also been, at least initially, syncollisional. The rapid intrusion of the 14 km thick Window Allochthon crust may have induced gravitational collapse by lifting the orogenic lid, with the formation of the Mode-II faults. In this sense, the gravitational collapse model of Terry et al. (2004a, b) is compatible with the model presented here. The geometry of the Mode-II faults at depth remains uncertain; previously, it has been presumed that they either cut through the crust or die out within it. Here, it is proposed that they link with the thrust detachment at the base of the WGR, implying that the late top-to west movement (Modes-II and III) did not affect Baltica itself. This geometry seems more compatible with a regime in which there was shortening, at least initially, between Baltica and Laurentia

The displacements on the Mode-II faults have not been included in the model. Although Fossen (2000) suggested that this might be as much as 50 km on each fault, the distribution of Mode-II structures is as yet uncertain and hence the total amount of displacement is unknown. When the distribution and displacements are better known, they can be incorporated into the model; it is likely that this will result in a considerable change in the length of the Western Gneiss Region, and thus the Jostedal Plateau.

8. Conclusions

Shortening in the Lower Allochthon, averaging ~50%, combined with branch-lines, indicates that the minimum restored position for the basement rocks in the tectonic windows, here termed the Window Allochthon (Western Gneiss Region plus the Atnsjøen, Spekedalen, Vang, Beito, Aurdal-Lærdal, etc. Window Allochthon) must lie west of the Norwegian coast. These units formed a topographic high (Jostedal Plateau), overlain by a thin Ediacaran and younger sequence, in comparison to the thick middle to upper Neoproterozoic 'sparagmite' successions of the Hedmark and Valdres Basins. Conceptually, this palaeogeographic model follows that proposed by Gayer et al. (1987), Gayer and Greiling (1989) and Rice (2002) for the evolution of the basement rocks exposed within the major tectonic windows throughout the Scandinavian Caledonides.

This estimated displacement of the Western Gneiss Region (~320 km) has to be included in all models for the Western Gneiss Region, irrespective of the burial and exhumation mechanisms proposed for the ultra-high pressure rocks. If this value should seem large, it is worth recalling that in the Himalaya, the best studied modern analogue of the Caledonides, continental shortening has been constrained to 2,850 km, based on an India-Asia collision at ~57 Ma and a subsequent collision rate of 5 cm yr^{-1} (Leech et al., 2005).

In the model proposed, the NW-margin of the Jostedal Plateau (coastal Western Gneiss Region) was subducted to depths of ~95 km by ~410 Ma, after which steepening of the subduction zone resulted in further burial to ~120 km, whilst contemporary exhumation of the SE-margin of the Jostedal Plateau began (external Window Allochthon). If peak ultra-high pressure conditions were reached at ~405 Ma and followed by exhumation to near surface levels at a ~10 mm yr^{-1} vertical displacement, as modelled, this would have caused a ~14 mm yr^{-1} horizontal displacement rate and imbrication of the Osen-Røa Nappe Complex. This rate is faster than the modelled emplacement rate of the nappes above the Western Gneiss Region (~5.2 mm yr^{-1}), which were consequently strongly stretched (~149 %). After exhumation, shortening in the Osen-Røa Nappe Complex continued to at least ~388 Ma and possibly down to 375 Ma.

By modelling the lateral displacement of the exhuming continental crust to be faster than the lateral movement of the Laurentian driven nappes, the conflict between large-scale Caledonian sinistral oblique collisional, proposed by Soper et al. (1992), and the local transtensional regime observed in the Western Gneiss Region (Krabbendam and Dewey, 1998) is resolved.

The Western Gneiss Region is one of the world's major areas for studying the formation and exhumation of ultra-high pressure rocks. Consequently, it is critical that all the available geological constraints are incorporated into models accounting for the burial and exhumation of the ultra-high pressure rocks, to ensure valid applicability to other areas. Models based on only limited regional data are likely to be incorrect, yet have a strong influence on studies elsewhere, both field and laboratory based. In this respect, the Lower Allochthon, an area which represents a very considerable constrained shortening (Morley, 1986), must be regarded as a critical component in the understanding of the evolution of the Western Gneiss Region.

Despite its acknowledged limitations, the geometrical model presented here is thought to give a reasonable, if much simplified account of the broad-scale development of the region. Further, it shows that regional palaeogeographical constraints, however uncertain, can be usefully applied to such areas. Hopefully, this will provide a firmer basis for subsequent, more detailed models of the region.

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