Chapter 11

Was the Precambrian Basement of Western Troms and Lofoten-Vesterålen in Northern Norway Linked to the Lewisian of Scotland? A Comparison of Crustal Components, Tectonic Evolution and Amalgamation History

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1. Introduction

Temporal and spatial linkage of Archaean and Palaeoproterozoic crustal provinces in the North Atlantic realm requires a well-established geological and geodynamic framework. Such a framework is well established for the Fennoscandian Shield of Finland, Sweden and northwestern Russia [1-4], for Greenland/Laurentia [5, 6] and for the Lewisian of NW Scotland [7, 8], but not yet for the Precambrian crystalline rocks within and west of the Scandinavian Caledonides in North Norway (Figure 1).

In western Troms (Figure 2) and in the Lofoten-Vesterålen areas of North Norway (Figure 3) Neoarchaean and Palaeoproterozoic continental crust (2.9-1.67 Ga) is preserved as an emerged basement horst bounded to the east by thrust nappes of the Scandinavian Caledonides (Figure 1b) [9-11] and to the west by offshore Mesozoic basins [12]. These basement outliers are believed to be part of the Archean-Palaeoproterozoic Fennoscandian Shield [1, 11, 13] that stretches from NW Russia, through Finland and Sweden (Figure 1a). Similarly, a pronounced magmatic suite in the Lofoten area [9, 10] corresponds in age (1.80-1.78 Ga) and structural position with the NNW-trending Transscandinavian igneous belt of Sweden [14, 15]. In spite of the internal position relative to the Caledonides, and in great contrast to the basement inliers in southern Norway where Caledonian high-grade metamorphic reworking is widespread, the geotransect in western Troms and Lofoten displays only modest Caledonian reworking and, thus, provides a reliable framework for regional correlation of Neoarchaean and Palaeoproterozoic crust [9,10, 16,17].

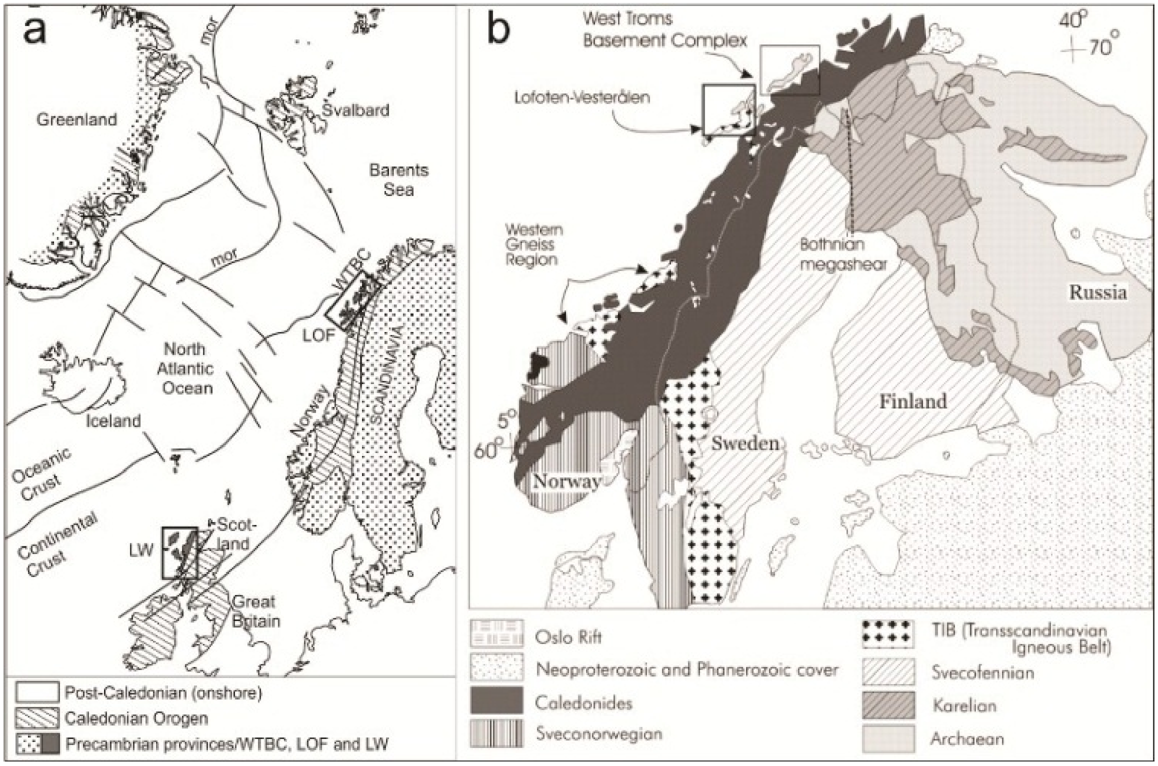

Figure 1. (a) Location of Precambrian basement outliers in western Troms (WTBC), Lofoten-Vesterålen (LOF) and NW Scotland (LW) and related basement provinces within today's plate setting of the North Atlantic Ocean. (b) Geologic map of Fennoscandia with location of the West Troms Basement Complex and the Lofoten-Vesterålen province west of the Scandinavian Caledonides [16].

In NW Scotland, the Lewisian Complex is situated to the west of the Caledonian Moine thrust and covers the Outer Hebrides, the NW Scottish mainland and part of the Inner Hebrides (Figures 1a and 4). The Lewisian rocks also form inliers in the Caledonian orogenic belt, possibly continuing under the Moine Group up to the Great Glenn fault in the southeast, while Mesozoic to early Cenozoic extensional basins offshore Scotland largely disrupted the continuity of the Lewisian outcrops [18, 19, 20].

A possible linkage of the western Troms and Lofoten-Vesterålen basement rocks with the Lewisian basement inliers of the Caledonides in NW Scotland (Figure 1a) and with Laurentia-Greenland has been raised [3, 4, 8, 21, 22], but excact correlation of these cratonic-marginal provinces in the North Atlantic realm, their role during assembly of Fennoscandia and Laurentia in the Neoarchaean [23, 24], and the situation prior to Palaeoproterozoic orogenies [3, 25, 26], still remain enigmatic.

This paper reviews the current knowledge of the crustal components, tectono-magmatic evolution and amalgamation history of the basement rocks in western Troms and Lofoten-Vesterålen, North Norway, and compares them with the Lewisian of Scotland (Figures 2-4, Table 1). New and focused structural and geochronological work in the West Troms

Basement Complex [16, 17, 27-32] has sparked off new interest in these provinces. Questions specifically raised for these basement suites concern the age and nature of supracrustal units, and the character of crustal-scale ductile shear zones, either as potential terrane boundaries between assembled older crustal blocks or just reflecting episodes of basin formation and later reworking. Such boundaries can, in general, help to restore the outline and correlation of each craton and the cratonic margin characteristics and to unravel cycles of tectono-magmatic events [33].

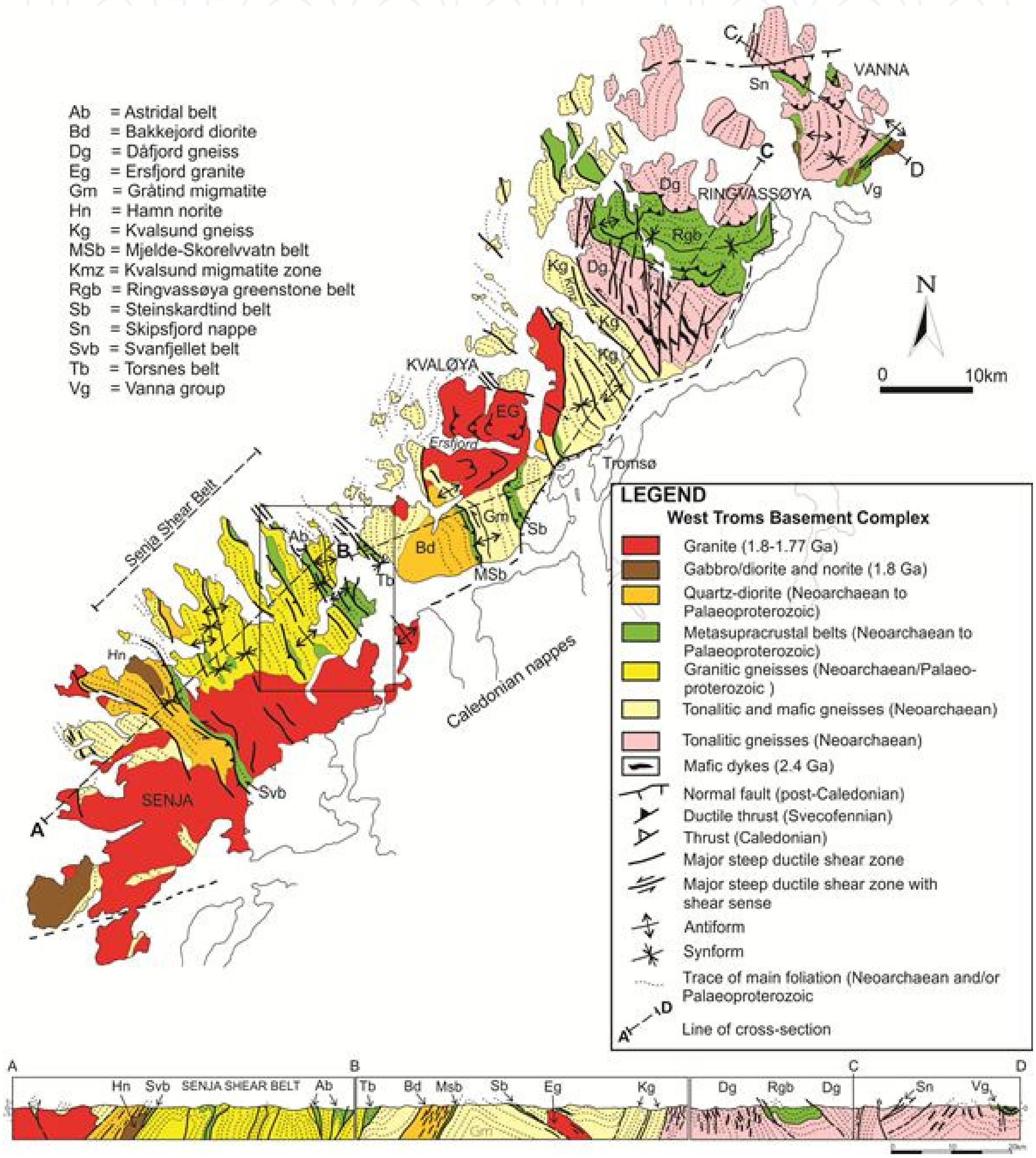

Figure 2. Regional map of the West Troms Basement Complex, North Norway, that shows the main crustal components and tectonic features, with a generalized cross-section. Frame shows location of Figure 10a. The map is revised after [17, 34].

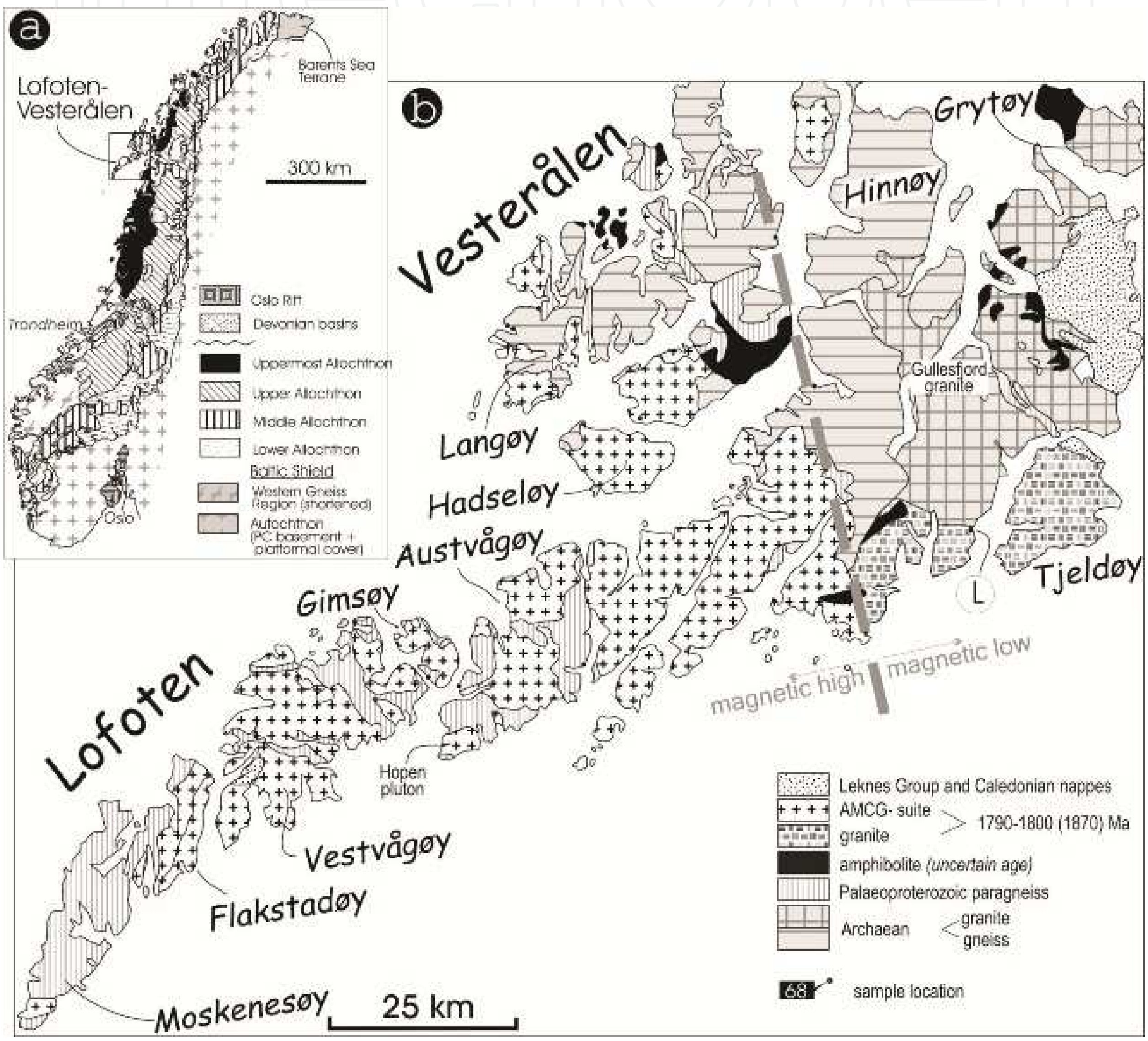

Figure 3. Geological map of the Lofoten-Vesterålen province, showing the Neoarchaean-Palaeoproterozoic basement rocks and the anorthosite-mangerite-charnochite-granite igneous suite [10, 35, 36]

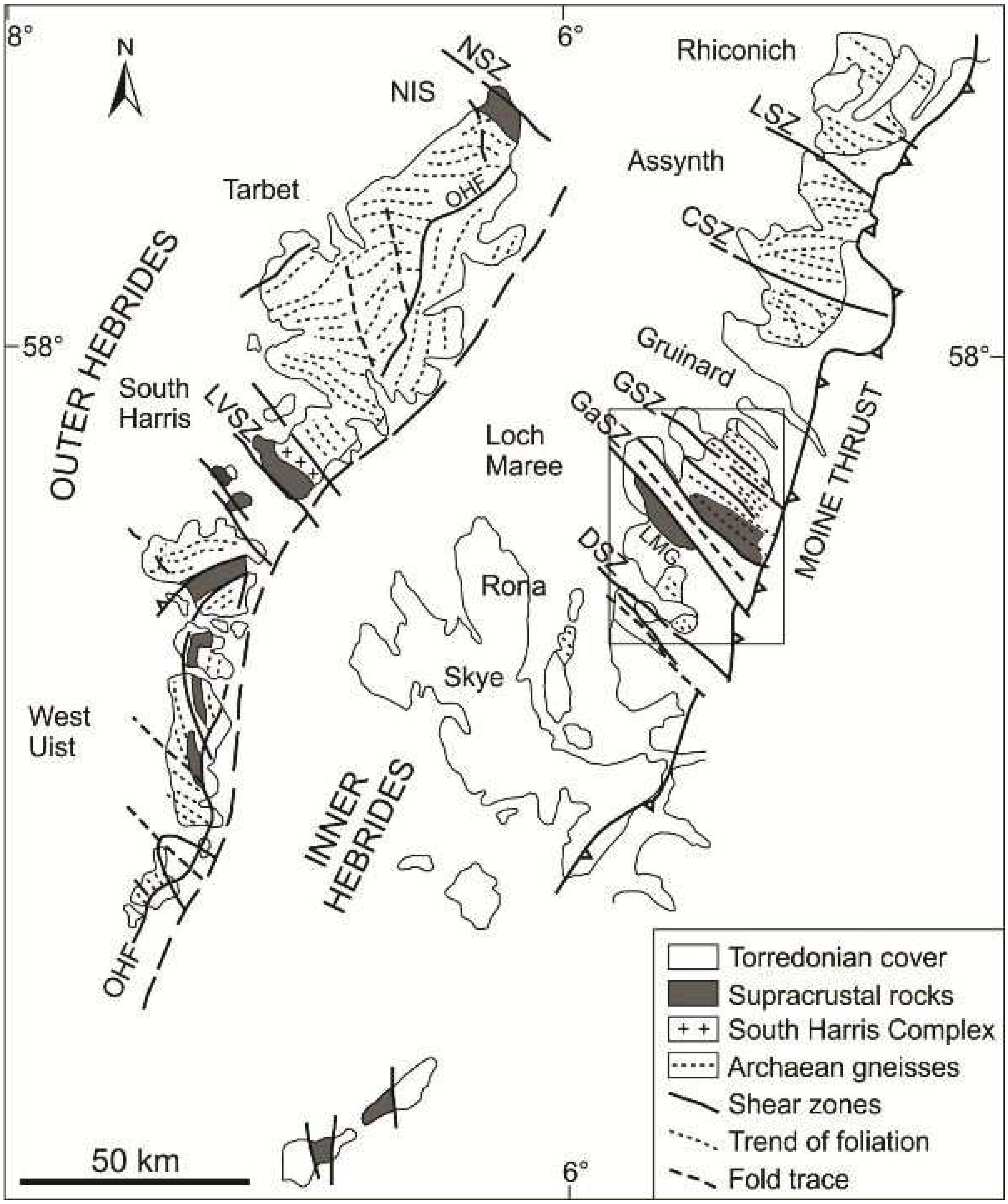

Figure 4. Simplified geologic and structural map of the Lewisian Complex in northwest Scotland showing the main rock units and the overall subdivision of the mainland into multiple regions (or terranes) separated by main Palaeoproterozoic ductile shear zones. The map is modified from from [20, 37]. Abbreviations: LSZ=Laxford shear zone, CSZ=Canisp shear zone, GaSz=Gairloch shear zone, DSZ=Diabaig shear zone. Frame shows outline of figure 18.

West Troms Basement Complex		**Lofoten-Vesterålen Province**		**Lewisian Complex**	
Age (Ga)	Components and events	Age (Ga)	Components and events	Age (Ga)	Components and events
2.92-2.8 Ga 2.85-2.83 Ga 2.75-2.6 Ga 2.75-2.7 Ga 2.7-2.67 Ga 2.69-2.56 Ga	**Neoarchaean cratonization**: -Tonalite crystallization (*Dåfjord & Kvalsund gneisses)* Volcanism and sedimentation: - *Ringvassøya greenstone belt* **Continued Neoarchaean cratonization** -Mafic plutonism (*Bakkejord diorite*) in the southwest Neoarchaean deformation and metamorphism: -Magmatism, migmatization (*Gråtind migmatite*) and ductile shearing (in *Dåfjord* and *Kvalsund* gneisses) - Main gneiss foliation (initially flat-lying), ductile shear zones, tight folds and dip-slip stretching lineation. - Medium/high-grade metamorphism, ENE-WSW crustal contraction and thickening by accretion and/or underplating -High-grade metamorphism and resetting	2.85-2.7 Ga 2.75-2.68 Ga 2.72-2.66 Ga 2.64 Ga	**Neoarchaean cratonization:** Accretion, convergence and crustal thickening -Tonalite magmatism **Neoarchaean deformation and metamorphism:** - Various orthogneisses (e.g. *Bremnes gneiss*) formed by crustal shortening - Emplacement of tonalities, followed by high-grade metamorphism and localized migmatization and ductile crustal shearing (*Sigerfjord migmatite, Ryggedalen granulite*)	3.145-2.75 Ga 2.9-2.7 Ga 2.8 Ga? 2.7-2.6 Ga 2.49-2.4 Ga	**Neoarchaean cratonization**: -Various types of *TTG*-gneisses, protoliths of tonalite (*Scourian Gneiss*), granites (*Laxfordian Gneiss*), Regional high-grade metamorphism, crustal accretion (?) and thickening **Volcanism and sedimentation**: - Mafic/ultramafic volcanics and supracrustal rocks (*Eanruig paragneiss, Claisfearn supracrustals*). Neoarchaean deformation and metamorphism: - Subhorizontal foliation & tight folding and thrusting, regional granulite facies metamorphism. NE-SW shortening -Open macro-folds and axial-planar dextral transpressive shear zones (*Canisp* and *Laxford*? shear zones). - Reworking and retrogression to amphibolite facies
2.40 Ga	**Crustal extension and intrusion** of the *Ringvassøya* mafic dyke swarm	?		2.4-2.0 Ga	**Crustal extension and intrusion** of the *Scourie mafic dyke swarm* and Na-rich pegmatites. Dextral transtensional setting.
2.4-2.2 Ga	**Deposition** of *Vanna group* clastic sediments in a marine subsiding basin	?	**Deposition** of supracrustal units, heterogeneous mafic gneisses, banded iron formations, quartzite, marbles, graphite schists	2.2-1.9 Ga	**Deposition** of the *Loch Maree Group* clastic and volcanic succession in marine extensional and/or arc-settings -Marine mudstones (*Flowerdale schists, Aundrary amphibolites*)
2.22 Ga	**Intrusion** of *Vanna diorite* sill	?		?	
2.2-1.9 Ga	**Deposition** of *Mjelde-Skorelvvatn*, *Torsnes* and possibly, the *Astridal* supracrustal belts	1.87-1.86 Ga	Precursory stage magmatism, AMCG-suite, *Lødingen granite*	1.9-1.8 Ga	**Intrusion** of early-stage *granites*

West Troms Basement Complex		**Lofoten-Vesterålen Province**		**Lewisian Complex**	
1.993 Ga	**Intrusion/volcanism** in the *Mjelde-Skorelvvatn belt*	1.8-1.79 Ga	Main stage intrusion of AMCG-suite plutonic rocks and the *Tysfjord Granite*	1.9-1.87 Ga	**Kalk-alkaline volcanism**: *Ard kalk-alkaline gneisses* Magmatism: *South Harris Igneous Complex* (Hebrides)
1.80 Ga	**Magmatism/int**rusion of *granites* and *norite* in Senja				
1.79 Ga	**Magmatism/intrusion** of *Ersfjord Granite* in Kvaløya	1.77 Ga	Intrusion of granite pegmatite dykes		
				1.7-1.68 Ga	**Intrusion** of late-stage granites
c. 1.9-1.7 Ga	**Palaeoproterozoic deformation**:		**Palaeoproterozoic deformation**	1.97-1.67 Ga	**Palaeoproterozoic deformation**:
1.9-1.8 Ga?	**Early**: -Mylonitic foliation (initially flat-lying), NW-SE trending gently-plunging isoclinal folds, NE-directed ductile thrusts with dip-slip stretching lineation. Prograde medium/high-grade metamorphism in the southwest. NE-SW orthogonal shortening, NE-directed thrusting/accretion	1.87-1.79 Ga	Strong ductile deformation and high-grade (granulite facies) metamorphism and reworking	1.8-1.85 Ga?	**Early stage** -Deep thrusting (*Gairloch* shear zone) and amphibolite facies metamorphism, accretion of the Loch Maree Group onto the continental crust, in subduction or arc-setting.
		I.83 Ga	Major ductile shear zone (suture)? Arc-related and/or collisional setting		
1.78-1.77Ga	**Mid**: -Regional open upright folding, NW-SE trend, flat-lying hinges and steep limbs. Medium/low grade metamorphism. Continued NE-SW orthogonal crustal shortening	1.78-1.76 Ga	Retrogressive metamorphism of AMCG-suite rocks	1.85 Ga	**Mid/main stage** - Isoclinal folding and thrusting, upright macro-folding and transpressive shear zones (*Canisp*, *Laxford* and *Shieldaig* shear zones), Accretionary tectonic event
c. 1.75 Ga?	**Late**: -Regional steep N-plunging folds, NW-SE striking, steep ductile shear zones (strike-slip). -Retrogressive low grade metamorphism.			1.70-1.67 Ga	**Main/late stage** - Partitioned deformation, thrust and steep dextral-oblique strike-slip shear zones on a flat detachment (on *Laxford* and *Canisp* shear zones), Amphibolite facies metamorphism, likely collisional event
1.7-1.67 Ga?	**Latest**: -NE-SW trending upright folds of the *Vanna group* and SE-directed thrusts, steep semi-ductile strike-slip shear zones. Retrogressive low grade metamorphism Partitioned NE-SW shortening and orogen-parallel (NW-SE) strike-slip shearing				**Late-stage** NW-SE steep transpressive shear zones, retrogression to greenschist facies, crustal rejuvenatiion.
1.57 Ga	**Intrusion** of felsic pegmatites and retrogression			1.6 Ga?	Crustal uplift, retrograde metamorphism, cooling

Table 1. Summary and comparison of tectono-magmatic components and events in the West Troms Basement Complex, the Lofoten- Vesterålen area and the Lewisian of Scotland. The data is based on references listed and discussed in the text.

2. Geological features of western Troms and Lofoten

Archaean and Palaeoproterozoic crust underlies much of the northeastern part of the Fennoscandian Shield [1, 3], including also basement outliers west of the much younger, Palaeozoic Scandinavian Caledonides (Figure 1). Here, the West Troms Basement Complex [17] and basement in Lofoten-Vesterålen [9, 10] emerge as a c. 300 km long horst, separated from the Caledonian nappes by Mesozoic rift-normal faults [11, 12]. The West Troms Basement Complex (Figure 2) is composed of various Mesoarchaean to Palaeoproterozoic plutonic rocks and orthogneisses (2.9-1.7 Ga), metasupracrustal rocks, mafic dyke swarms, and networks of ductile shear zones [17, 38]. The basement of Lofoten and Vesterålen (Figure 3) consists of similar metamorphic Neoarchaean rocks intruded by a very extensive suite of 1.80-1.78 Ga plutonic rocks of the anorthosite-mangerite-charnockite-granite (AMCG) suite [9, 10] that appears to link up with the Transscandinavian Igneous Belt of southern Sweden (Figure 1a). Both areas display dominant NW-SE structural trends parallel with Archaean and Palaeoproterozoic orogenic belts of the Fennoscandian Shield that stretch from Russia, through Finland and northern Sweden into the Bothnian basin of central Scandinavia [1].

2.1. The West Troms Basement Complex

The West Troms Basement Complex (Figure 2) is underlain by Meso to Neoarchaean gneisses, various generations of metasupracrustal rocks and mafic dyke swarms that were later intruded by felsic and mafic plutonic suites and variably reworked, deformed and metamorphosed during the main Palaeoproterozoic (Svecofennian) orogeny [17].

2.1.1. Archaean crust

The Meso-Neoarchaean rocks of the West Troms Basement Complex (Figure 2) consist of tonalite-trondhjemite and anorthositic gneisses with mafic and ultramafic layers and banded intercalations (Figure 5a) and are overlain by the Neoarchaean Ringvassøya greenstone belt. These rocks were deformed and metamorphosed up to granulite/migmatite facies prior to deposition of Palaeoproterozoic cover units and intrusion of a 2.4 Ga mafic dyke swarm [17, 39]. A steep NW-SE trending transposed gneiss foliation with dip-slip stretching lineations (Figure 5b) and tight ENE-vergent intrafolial folds (Figure 5c) attests for WSW-ENE contraction and thrusting during the Meso/Neoarchaean [17]. Prominent high-grade migmatite zones interpreted as a ductile shear zone (Figure 5d) separate compositionally different gneisses [17], e.g. the Kvalsund migmatite zone separating the Dåfjord and Kvalsund gneisses on Ringvassøya [36] and similar zones within the Senja Shear Belt [38, 40]. Polyphase refolding and thrusting is common and suggests protracted Neoarchaean deformation [17]. In Ringvassøya, tonalitic orthogneisses and granitoids (Dåfjord gneiss) reveal U-Pb zircon crystallization ages of 2.92-2.77 Ga (Figure. 6) [36, 41], and these rocks are considered to be related to tonalites on the island of Vanna farther north, where a U-Pb crystallization age of 2885 ± 20 Ma has been obtained [33]. This Mesoarchaean basement was also intruded by the 2695 ±15 Ma Mikkelvik alkaline stock [42]. The overlying Ringvassøya Greenstone belt [43] comprises arc-related meta-volcanic rocks with MORB-transitional, tholeiitic to calk-alkaline affinity. Two meta-volcanic rocks of the Ringvassøy greenstone belt yield ages of c. 2.85 Ga [44].

Figure 5. Outcrop features of Meso-Neoarchaean tonalitic gneisses in the West Troms Basement Complex: (a) Foliated tonalitic and mafic gneisses (Dåfjord gneiss) in central part of Ringvassøya. (b) Steep SW-dipping foliation in tonalitic Dåfjord gneiss with dip-slip stretching lineation. (c) Banded tonalitic gneiss with tightly folded mafic inclusions, cut by granitic pegmatite veins presumed to be related to the 1.79 Ga Ersfjord granite. (d) Zone of major migmatized Kvalsund gneiss in southwestern part of Ringvassøya. The zone is cut by a mafic dyke, which is part of the Ringvassøya dyke swarm dated at 2.4 Ga [39].

By contrast, all the dated metaplutonic rocks on the islands of Kvaløya and Senja farther south (Figure 2) are Neoarchaean. The Bakkejord pluton, neosome in the Kattfjord gneiss, and granodiorite units bordering the Torsnes belt on Kvaløya, as well as several major intrusive bodies on Senja all yield ages between 2.72 and 2.68 Ga [16, 32]. A somewhat younger element at 2.67 Ga is shown by mafic dykes that cut the Bakkejord pluton on Kvaløya [32]. The only potentially younger Archaean event is the formation of migmatites in southern Senja, where zircon in a neosome suggests crystallization at c. 2.6 Ga [32]. The Kvalsund migmatite zone in southwestern Ringvassøya (Figure 2) appears to represent a boundary separating the Mesoarchean crust to the north from Neoarchean crust to the south. The time of deformation is not yet dated, but dynamic melting structures in the migmatite are cut by mafic dykes considered to belong to the 2.4 Ga swarm, hence indicating an Archean age of shearing. One sample of neosome has a primary age of about 2.7 Ga indicated by zircons, which, however, also records an event ≤ 2.55 Ga possibly reflecting the time of deformation [32].

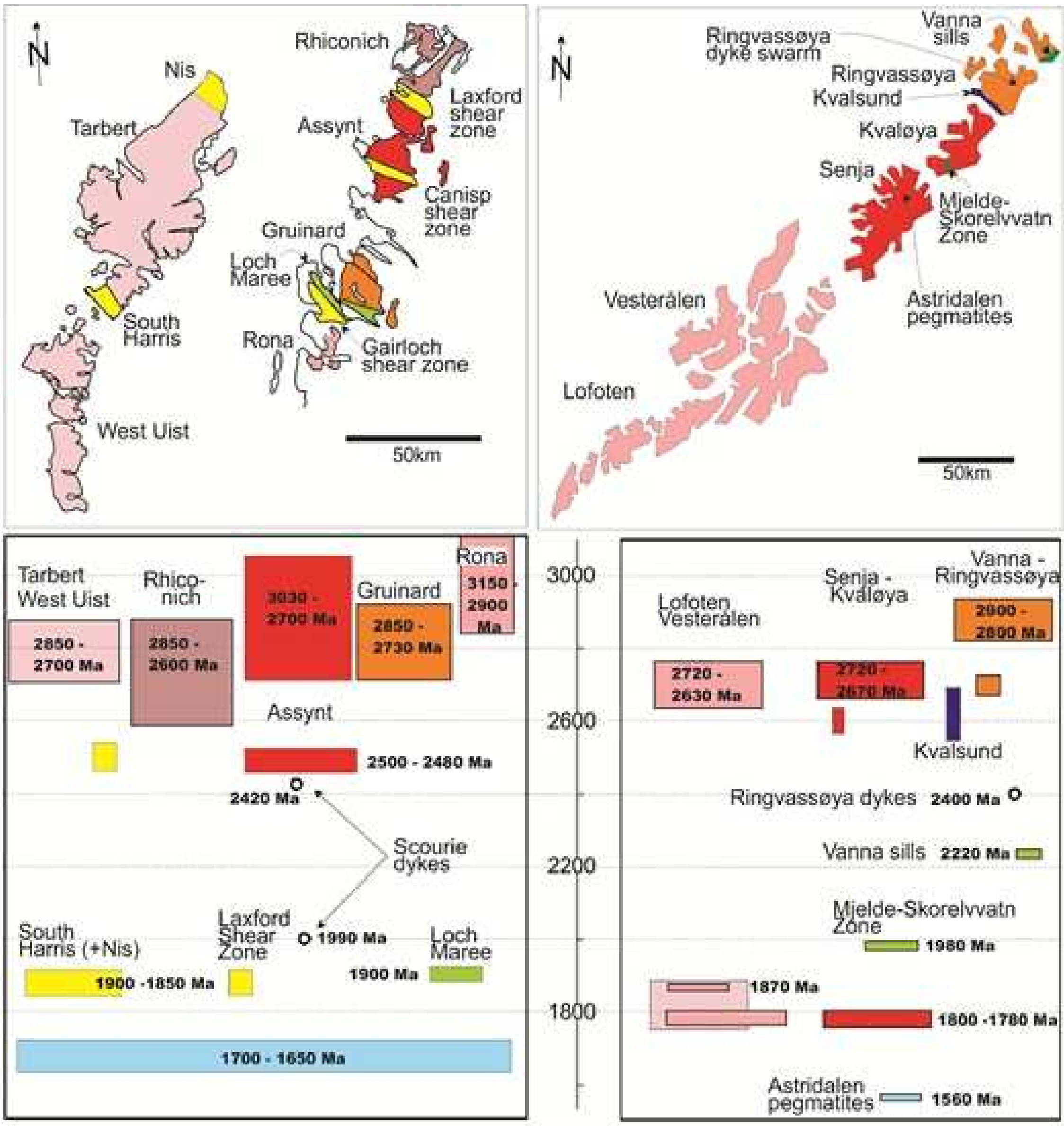

Figure 6. Summary of the main geochronological features in the Lewisian Complex (a) and the West-Troms Basement Complex and Loften province (b). The age compilation is based principally on U-Pb data. The references are listed and discussed in the text.

2.1.2. Ringvassøya mafic dyke swarm

The Neoarchaean gneisses and the oldest metasupracrustal belts of the West Troms Basement Complex have been intruded by a huge mafic, plagioclase phyric and gabbronoritic dyke swarm (Figure 7), the Ringvassøya dykes [39]. These dykes are widely distributed and display offset, shearing and reworking, thus providing a good time-marker for resolving the subsequent Svecofennian deformation [17, 27]. Zircon and baddeleyite from a dyke swarm on Ringvassøya provides a crystallization age of 2403 ± 3 Ma, and the dykes are classified as transitional between MORB and within-plate basalts with an affinity to continental tholeiites [39].

Figure 7. Outcrop features of the 2.4 Ga mafic dyke swarm that intruded Neoarchaean massive tonalites in Ringvassøya [39]. Note irregular and varied dyke orientations (a, b) and that the dykes truncate the main tonalitic gneiss fabric (c).

2.1.3. Palaeoproterozoic supracrustal rocks

The metasedimentary Vanna group represents unconformable continental deposits [29, 45]. Common rock types include layered meta-psammites locally exhibiting pronounced cross-bedding (Figure 8a, b). The age of deposition is constrained between 2403 Ma, the age of the underlying Ringvassøya dykes, and 2221 ± 3 Ma, the age of a diorite sill in the supracrustal rocks [29, 39].

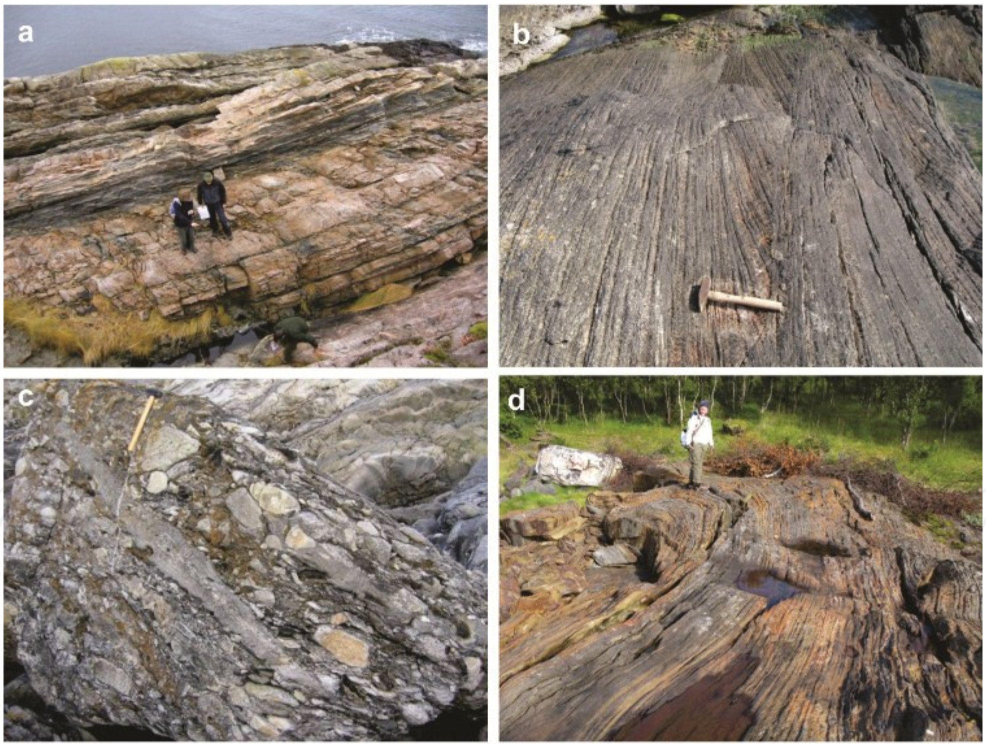

Figure 8. Outcrop features of meta-supracrustal rocks in the West Troms Basement Complex: (a) Layered meta-psammites with interbedded mudstones/mica-schists of the Vanna group [29, 45]. (b) Subvertical thinly bedded meta-sandstone of the Vanna group, with pronounced trough and planar cross-bedding in internal lenses. Up is to the right. (c) Basal meta-conglomerate of the Torsnes supracrustal belt. Note that the dominant clast-type is tonalite, tonalitic gneiss and granitoid gneisses (d) Rythmical laminated quartz-feldspatic meta-psammite with enrichment of iron-hydroxide staining from the Astridal belt, Senja. The beds are steeply dipping and subvertically folded. The fold hinge is located near person.

Arc-related meta-volcanic rocks with MORB-transitional, tholeiitic to calk-alkaline affinity occur in the central and southwestern parts, e.g. in the Astridal and Torsnes belts [17, 27, 31]. The Mjelde-Skorelvvatn belt (Figure 2) is dominated by metabasaltic rocks together with ultramafic rocks, meta-psammites, marble and calc-silicate gneisses. A differentiated gabbroic portion of the meta-basaltic pile yields an age of 1992 ± 2 Ma [31]. The Torsnes belt comprises a basal conglomerate (Figure 8c) overlain by meta-psammite and a thick sequence of mafic metavolcanic rocks. Detrital zircons indicate a maximum-age of deposition of 1970 ±14 Ma [31]. The Astridal belt is compositionally similar to the Mjelde-Skorelvvatn belt, but there are no direct radiometric dates yet. It contains abundant mica-schists, local graphite schists, mafic meta-volcanic rocks and widespread sulphide ore deposits (Figure 8d).

2.1.4. Late palaeoproterozoic igneous suites

The Neoarchaean crust in Kvaløya and Senja was intruded by an extensive suite of felsic and mafic plutonic rocks [17]. The most prominent are the Ersfjord granite (Figure 9a, b) on Kvaløya [46] and large granites and mafic plutons (Hamn norite) on Senja (Figure 2). The Ersfjord granite has a U-Pb zircon crystallization age of 1792 ± 5 Ma [16] and the Hamn norite 1802.3 ± 0.7 Ma [28], whereas the granitic masses farther south in Senja give Rb-Sr [47] and zircon-titanite ages of 1805 ± 2.5 Ma [16]. Metamorphic overprints of the Ersfjord granite are recorded by U-Pb titanite ages of 1769 ± 3 Ma and 1756 ± 3 Ma [16]. Pronounced and widespread granite pegmatite dykes (Figure 9b, c) formed syn-tectonically with shear zones in the metasupracrustal belts at c. 1768 ± 4 Ma, probably genetically related to the main intrusive activity [16]. All these ages are within the interval when most known Precambrian juvenile crust generated by arc-related magmatism [48].

Figure 9. (a) Aerial view of the Ersfjord granite with its rugged mountains and presence of both steep and gently dipping planar fabrics. (b) Ersfjord granite pegmatite dykes and veins cutting Neoarchaean tonalitic gneiss foliation in southwestern ringvassøya, and later boudinaged during Palaeoproterozoic tectonism. (c) Ersfjord granite pegmatite dykes cutting dioritic gneisses near its western boundary against the Kattfjord gneiss.

2.1.5. Palaeoproterozoic deformation and metamorphism

Strong deformation and metamorphism at 1.8-1.76 Ga produced mega-blocks or segments delineated by NW-SE trending, variously mylonitized metasuprascrustal belts and lens-shaped ductile shear zones as outlined in Figure 10 [17, 38]. This shear belt deformation was characterized by high-strain, complex and multiphase deformation and up to amphibolite facies metamorphism and reworking. The most distinct one, the Senja Shear Belt is c. 30 km wide (Figures 2 and 10) and is delimited by the Svanfjellet belt to the south [38, 49] and the Torsnes belt in the north [17]. This linear crustal structure is thought to continue beneath the Caledonian nappes into parallelism with the Bothnian-Senja shear zone of the Swedish part of the Fennoscandian shield, but gravity and magnetic anomaly patterns do not uniquely confirm such a correlation [11, 13, 50].

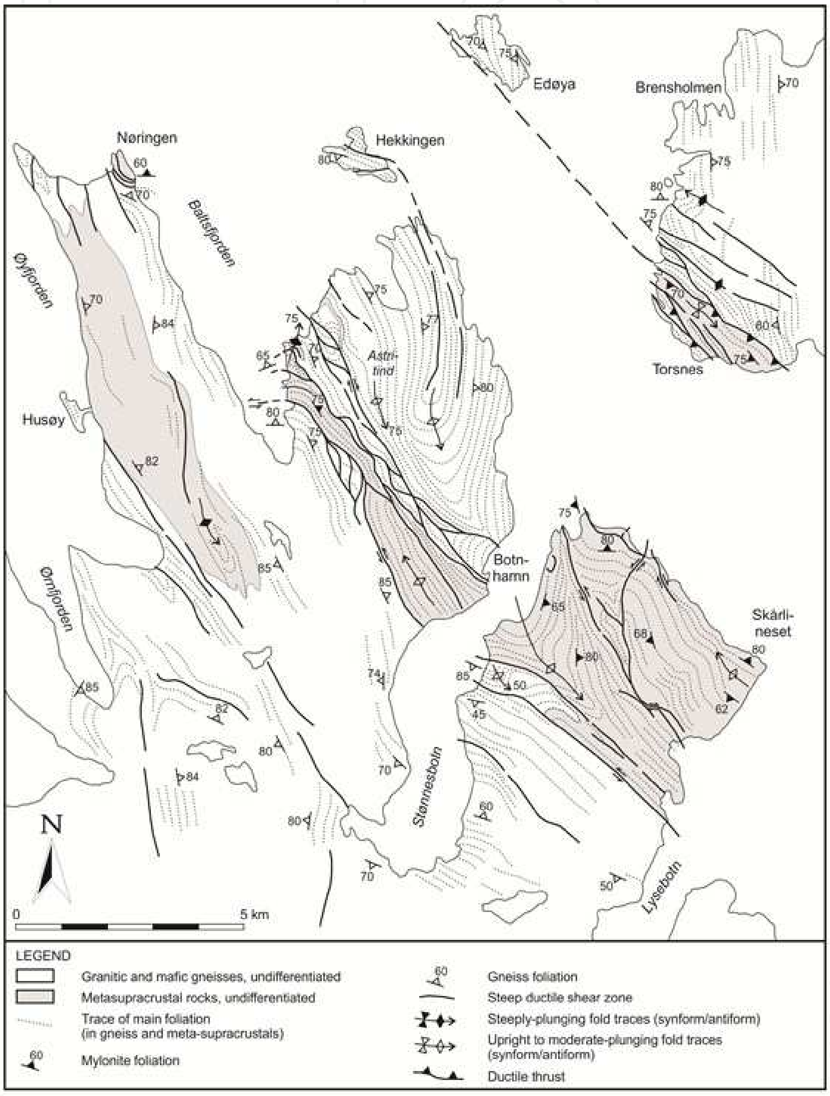

Figure 10. Tectonic map of the Senja Shear Belt in central part of the West Troms Basement Complex (for location see figure 2), illustrating the lens-shaped architecture of the metasupracrustal belts. Note macro-scale, steep-plunging folds of the belt and adjacent tonalitic gneisses, where fold hinges are bent into parallelism with the trace of the Astridal belt. Note also the major shear zone boundaries with thrust and sinistral strike-slip characters. The map is from [17].

The Palaeoproterozoic deformation structures of the West Troms Basement Complex (Table 1) include a main NNW-SSE striking, mylonitic foliation mostly present in the meta-supracrustal belts (Figure 11a), that formed axial-planar to early-stage isoclinal folds (Figure 11b) with gently-plunging axes, at amphibolite facies conditions. The foliation has a steep

WSW dip and exhibits a dip-slip, west-plunging stretching lineation. Macroscopic NW-SE-trending and mostly upright antiform-synform folds (Figure 11c) are widespread, causing the steep tilt and apparent repetition of most of the supracrustal belts in synformal troughs [17]. Corresponding upright folds also occur in the adjacent gneisses. Younger (Late Palaeoproterozoic) superimposed structures include tight to vertical folds on all scales (Figure 11d, e) with axial-planar cleavages coeval with an anastomosing network of steeply-dipping, NW-SE to N-S trending sinistral and dextral strike-slip shear zones (Figure 10). The latter zones are mylonitic and retrogressed into greenschist facies. These shear zones caused subvertical drag-folding of the surrounding gneisses and became boudinaged and masked by quartz precipitates along the main foliation. Later on the foliation was folded by steeply north-plunging shear folds and cut by oblique-slip crenulation cleavages, sigmoidal clasts and multiple shear bands, all supportive of strike-slip displacements (Figure 11d, e). The youngest set of structures occurs in the northeastern parts of the West Troms Basement Complex, as gently-dipping, SE-directed phyllonitic shear zones (thrusts) with abundant SE-verging folds and thrusts [17].

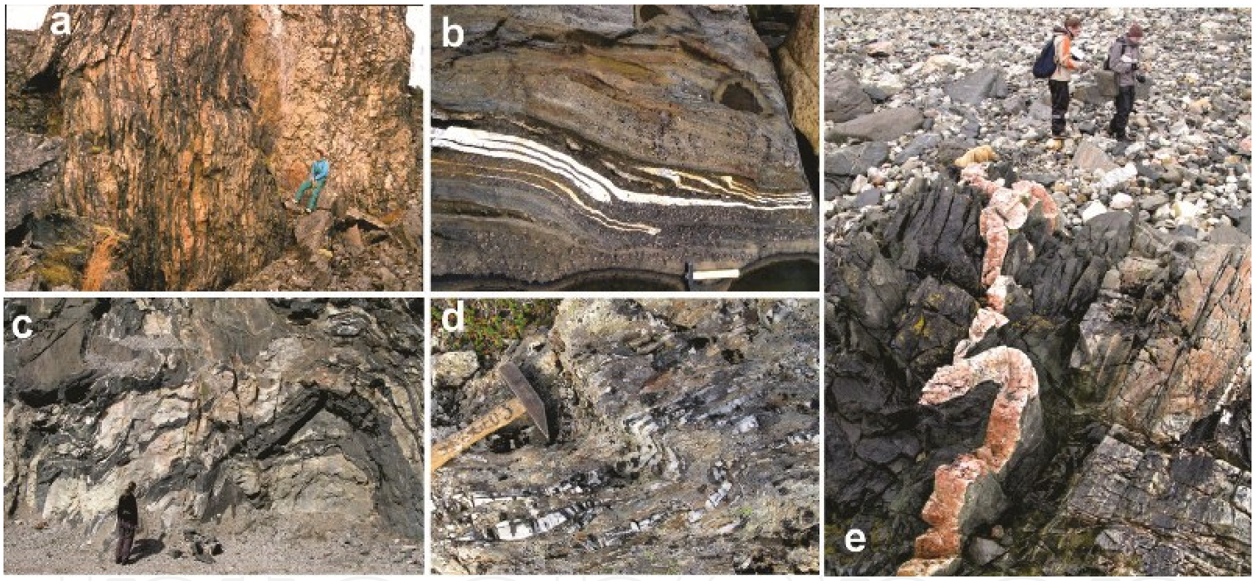

Figure 11. Outcrop illustrations of Palaeoproterozoic deformation structures in the West Troms Basement Complex: (a) Subvertical high-strain mylonites along the eastern contact between the Torsnes belt and the granitoid Kattfjord gneiss (right). (b) Foliation-parallel felsic vein in garnet-mica-schists of the Astridal bel that is isoclinally folded and sinistrally sheared, causing multiple repeatitions of the vein. View is to the NE on near-horizontal surface. (c) Upright asymmetrical NE-verging fold that refolds isoclinal folds in meta-volcanic and siliciclastic rocks of the Svanfjellet belt, Senja. (d) Subvertical phyllonitic shear fabric with quartz precipitates along the main foliation.Note dextral subvertical folding of the main fabric and quartz veins. (e) Ersfjord granite pegmatite dyke in mafic Kattfjord gneiss northeast of the Torsnes belt that is folded by subvertical sinistral folds.

The early and middle stages of deformation were associated with prograde metamorphism varying from low grade in the northeast to amphibolite and granulite facies in the central and southern parts and terminated with late stage retrogressive greenschist facies reworking [27]. A migmatitic shear zone in meta-psammites southwest of Ringvassøya displays a garnet granulite facies assemblage succeeded by a two-pyroxene granulite facies assemblage and is dated with the zircon U-Pb method to 1777 ± 12 Ma, the same age obtained for a granitic dike (1776.6 ± 1.1 Ma) cutting the lens [51]. Other critical radiometric ages for Palaeoproterozoic high-strain deformation include metamorphic overprints of Ersfjord granite pegmatite dykes at c. 1.77-1.76 Ga, and an interval of 1774, 1768 ± 4 Ma and 1751 ± 8 Ma for a granitoid pegmatite dyke formed syn-kinematically with late-stage strike slip shear zones in the Astridal belt [16].

2.1.6. Mesoproterozoic reactivation

Late-tectonically deformed granitic pegmatites in the Astridal belt of the Senja Shear Belt yield U-Pb ages of 1725 ±22 Ma and 1562 ± 2 Ma, indicating that formation of the pegmatites occurred after termination of the main orogenic events (Corfu et al. in review). These occurrences are attributed to 'anorogenic' far-field effects reflecting intracratonic strain, possibly caused by the emplacement of A-type massifs in the core of the Fennoscandian Shield.

2.2. The Lofoten-Vesterålen province

This province comprises gneisses and major plutonic suites of Precambrian age [52] that suffered major tectono-magmatic events at 2.8 and 1.8 Ga [53]. The basement complex in western parts of Lofoten and Vesterålen (Figure 3) comprises granulite facies rocks, whose distribution also coincides with a major magnetic and gravity high caused by the presence of dense rocks in the crust and an elevated Moho discontinuity [9, 11, 54, 55]. The latter is the result of differential uplift and extension in the aftermath of the Caledonian orogeny and the subsequent Late Palaeozoic and Mesozoic processes that led to the opening of the North Atlantic. The eastern part of the region consists of various amphibolite facies gneisses, migmatites, greenstone belts and granitic plutons. After the initial geochronological studies more work followed using Pb-Pb, Rb-Sr and Sm-Nd dating [9, 36, 56-58]. The chronology of the basement complex and the anorthosite-mangerite-charnockite-granite suite (AMCG) has now been refined by modern U-Pb geochronology [10].

2.2.1. Neoarchaean basement gneisses

Neoarchaean crust occupies large parts of the islands of Langøya in Vesterålen and Hinnøya farther east (Figure 3), and small remnants are also present at the southwestern tip of Austvågøya [58]. The Neoarchaean rocks of Langøya consist mainly of high-grade gneisses interpreted to represent metasupracrustal rocks of intermediate composition, while Neoarchaean gneisses on Hinnøya define an amphibolite facies metamorphic domain that was migmatized in the Neoarchaean and subsequently deformed and metamorphosed at granulite-facies conditions in the Palaeoproterozoic [9]. The appearance of orthopyroxene to the west of the amphibolite-facies domain on Hinnøya has been interpreted as either a prograde metamorphic transition [9, 54], or an abrupt transition marked by a crustal scale ductile shear zone of presumed Neoarchaean age. The zone east of the metamorphic boundary (Figure 3) is dominated by intrusive rocks of tonalitic to granitic composition, migmatitic domains, and local greenstone belt remnants, considered to be Neoarchaean in age [9, 59].

2.2.2. Palaeoproterozoic supracrustal rocks, deformation and metamorphism

A younger sequence of heterogenous mafic gneisses in Lofoten has been interpreted as a Palaeoproterozoic supracrustal succession of volcanogenic derivation as deduced from geochemical compositions [9]. Metasedimentary rocks consist of fine-grained gneisses, locally quartzitic (Figure 12a), with subordinate graphite schist, banded iron formation and marble. These gneisses have been overprinted by the same granulite-facies metamorphism as the Neoarchaean rocks, and the boundary between the two metamorphic domains corresponds to the eastern limit of the magnetic high and the first appearance of orthopyroxene [54] in eastern Langøya (Figure 3). In the south this boundary is considered a major thrust that juxtaposes the Eidsfjord anorthosite and deformed intrusive mangerite [54, 60]. Some studies [9] suggest that the orthopyroxene isograde is folded and continues southeastward across Hinnøya (Figure 3). The granulite facies gneisses tend to be rather massive, with faint banding, and they equilibrated at 3 to 4 kb and 750 to 780 °C [61]. They are considered to be orthogneisses [59] or migmatized supracrustal rocks of intermediate composition [9].

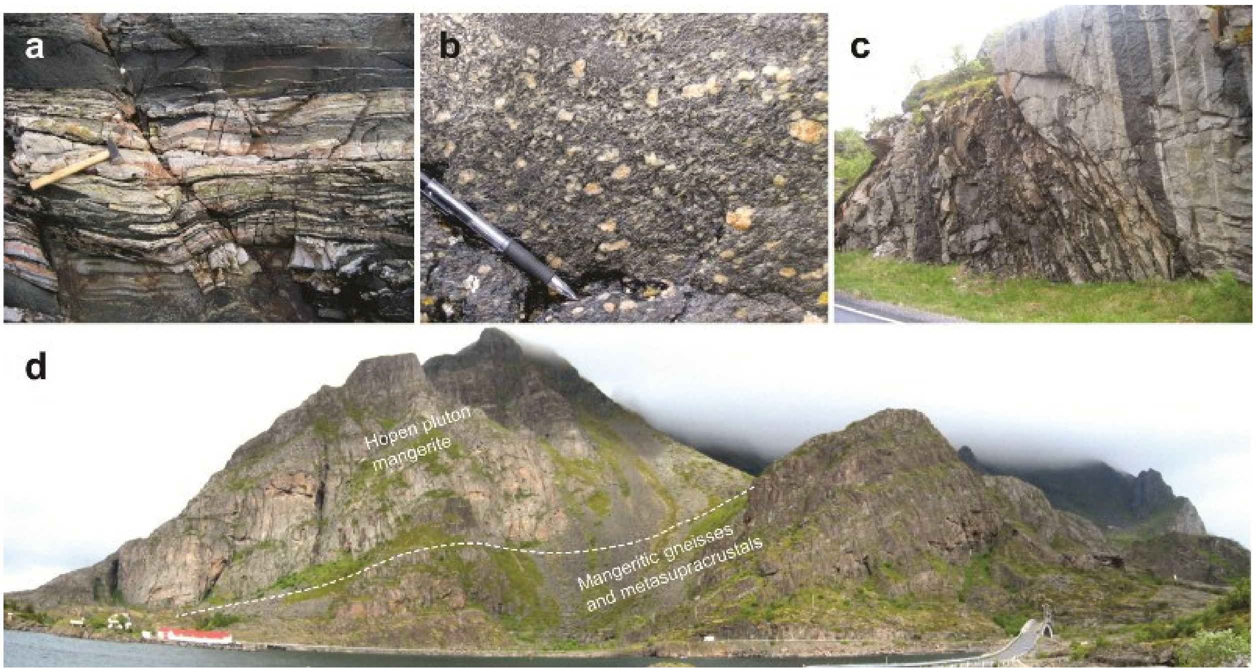

Figure 12. Outcrop photographs of the AMCG suite and metasupracrustal rocks in western parts of Lofoten (Figure 3)(a) Foliated paragneisses composed of alternating quartz-rich and mafic metavolcanic rocks. (b) Detail view of mangerite (hypersthene-bearing monzonite) which is the dominant rock type of the Lofoten igneous province. Note phenocrysts of plagioclase and orthoclase. (c) Mangerites with mafic intercalations and dykes aligned parallel to a weak magmatic foliation. From a road cut in eastern Lofoten (Austvågøya). (d) Panorama view of the Hopen pluton mangerite and its boundary to migmatized gneisses and altered mangerites. View is toward the north.

2.2.3. Palaeoproterozoic magmatic rocks, deformation and metamorphism

Plutons of the AMCG suite occupy about 50 % of the Lofoten islands [62]. The suite (Figure 3) is dominated by mangerite and charnockite and their metamorphosed equivalents (Figure 12b-d), with local but important occurrences of gabbro, anorthosite, and granite and associated mafic dykes (Figure 12c). The mafic and felsic phases locally grade into, or mutually cross-cut each other indicating that some of the intrusions are genetically related. In addition, a widespread network of feldspathic veins is present throughout the granulite facies domain [9]. Mangerite and charnockites are widespread throughout the region, in the form of large plutons (Figure 12d) such as in the southwestern Lofoten islands [9, 62, 63]. Anorthosite and gabbro forms plutons in most of the islands[54, 64, 65]. There is one distinctive body of granite-syenite on Langøya [54, 66]. To the east, on Hinnøya, Neoarchaean gneisses are cut by Palaeoproterozoic plutons such as the Lødingen granite with a zircon and titanite age of 1870 Ma [10, 36], whereas the Tysford granite covers a large area on the mainland and has given U-Pb ages between 1.8 and 1.7 Ga [67].

The mangerites and charnockites were shown to have intruded at about 4 kb and >925 to 850 °C, whereas the anorthosite on Flakstadøy records a polybaric crystallization history from 9 kb to about 4 kb at 1180 to 1120 °C. The data imply that anorthosites, mangerites and charnockites in Lofoten were emplaced at the same depth of about 12 km [64]. Later on in the same time span, the Palaeoproterozoic plutonic rocks and surrounding Neoarchaean gneisses were deformed and metamorphosed up to granulite facies conditions, and portions of the plutonic rocks, e.g. anorthosite bodies, were thrusted over mangeritic gneisses (Figure 13a, b) [54]. Structural fabrics include transposed foliations (Figure 13c), intrafolial isoclinal folds (Figure 13d) and local migmatization structures, and upright folds that refolded the earlier shear fabrics (Figure 13e). A younger, very extensive but narrow ductile shear zone network is characterized by greenschist facie retrogressive shear zones that truncate all other structures (Figure 13f) and display both low- and high-angle attitudes. The exact age of the latter is unknown, but presumably, late Palaeoproterozoic.

The structural and isotopic data show that the Neoarchaean crust played an important role in the genesis and deformation of the Lofoten igneous rocks, and both as a source and as a contaminant [58, 68]). The Pb data define a linear trend that may represent mixing between Neoarchaean lower crustal components and late Palaeoproterozoic juvenile additions, whereby the mangerites and charnockites contain more Neoarchaean Pb than the mafic rocks. A multistage evolution with a basaltic parental melt undergoing polybaric crystallization and differentiation to form anorthosites as cumulates and ferrodiorites as residual melts has been proposed [69]. The mangerites and charnockites are inferred to represent feldspar cumulates and residual liquids, respectively, derived from magmas broadly syenitic in composition.

The more recent U-Pb results show that the Lofoten AMCG suite was emplaced during two quite distinct events (Figure 6), the first one at 1.87-1.86 Ga followed by a second and dominant magmatic event at 1.80-1.79 Ga. A concluding period, lasting some 20-30 my, was characterized by local hydration of the dry AMCG rocks, and by the infiltration of pegmatite melts [10]. Local granitic pegmatites belong to a distinct Palaeoproterozoic (ca. 1.77 Ga) generation.

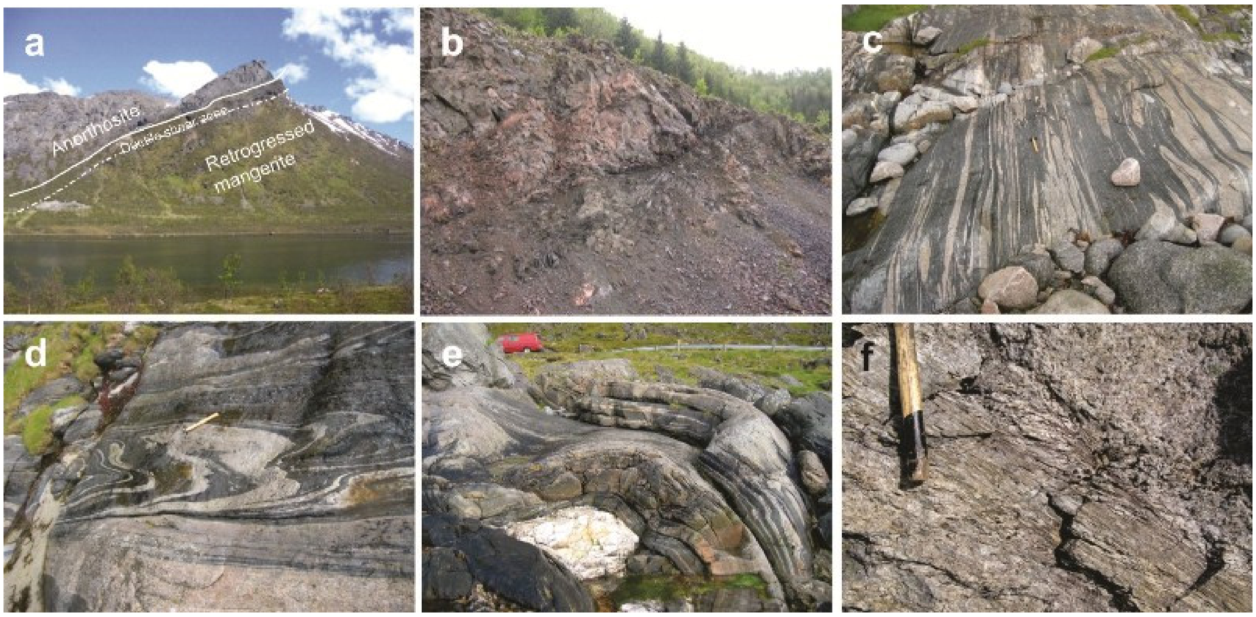

Figure 13. Outcrop features illustrating Palaeoproterozoic deformation fabrics of the Lofoten-Vesterålen province. (a) View of an anorthosite complex in Langøya which is thrusted over granulite facies mangeritic gneisses to the southeast. The ductile thrust zone is c. 50 m thick and made up of mylonitic gneisses [54]. Note the lack of vegetation on the grey-coloured Eidsfjord anorthosite above the shear zone contact. View is toward NE. (b) Thrust in Palaeoproterozoic granitic gneisses on Langøya [54]. (c) Transposed foliation in mangeritic gneiss with granitoid bands and intercalations. Locality: Austvågøya. (d) Ductile shear zone in mangeritic gneiss with felsic intercalations. (e) Open upright fold in mangeritic gneiss. The fold axis trends NW-SE, and view is toward SE. (f) Steep and localized retrogressed ductile shear zone in mangerite in southern part of Vestvågøya.

3. Geological features of the Lewisian Complex of Scotland

The Lewisian Complex in NW Scotland (Figure 1a) is situated to the west of the Caledonian Moine thrust and covers the Outer Hebrides and the NW Scottish mainland of the Inner Hebrides in the south (Figure 4). The Lewisian rocks also form inliers in the Caledonian orogenic belt possibly continuing under the Moine Group up to the Great Glenn Fault in the southeast [18-20], while Mesozoic to early Cenozoic extensional basins bound the Lewisian outcrops offshore Scotland (Figure 1a).

The Lewisian Complex was considered by earlier workers [70, 71] as a continuous crustal block composed of up to three Neoarchaean gneiss domains overlain by Palaeoproterozoic metasedimentary, metavolcanic and intrusive rocks separated by NW-SE trending ductile shear zones. In the last decade it has been proposed that the region consists of distinct terranes [7, 72], although there are disparate views on how these terranes are related [8]. A classical Lewisian nomenclature (Badcallian, Scourian, Inverian, Laxfordian) evolved progressively, originally to designate specific rock forming, deformational or metamorphic events, and eventually becoming linked to specific dates obtained from forthcoming geochronological studies. Some of these terms, however, have now become problematic since they have been and can be used to designate different tectonic expressions and times. To maintain clarity in the following we shall therefore avoid their use.

3.1. Main structures

The Neoarchean rocks of the Lewisian were deformed during an early granulite facies tectonic event involving crustal shearing and isoclinal folding that formed a gently dipping gneiss foliation (Figure 14a, b). During a later retrogressive event the gently dipping fabric was macro-folded into NE and SW-dipping steep attitudes and subjected to steep strike-slip shearing (Figure 14 c, d) in amphibolite facies [73-75]. A key observation is the local truncation of these macrofolds by the c. 2.4 Ga Scourie dykes [76], as outlined later (Figure 15), indicating that these early deformation events are likely Archaean in age.

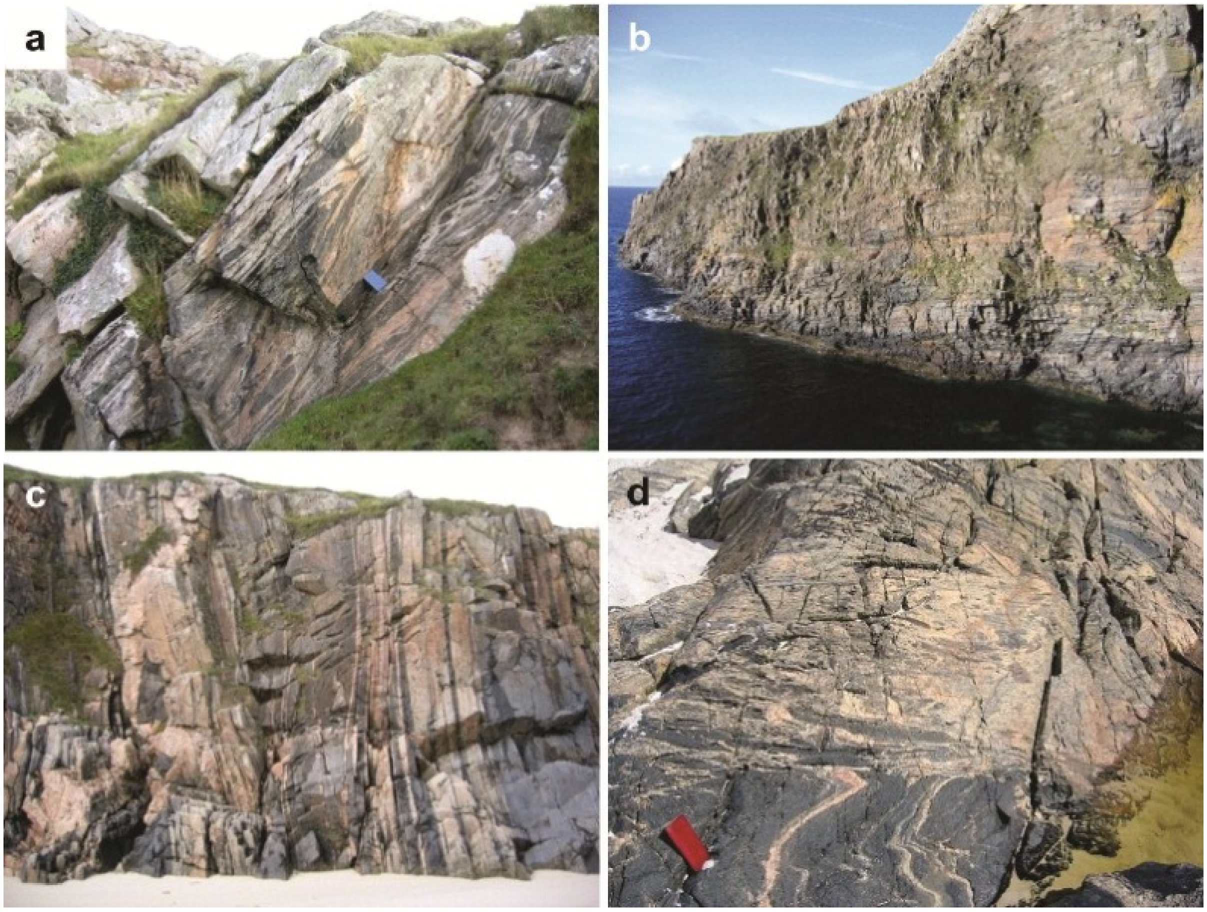

Figure 14. Outcrop features of Neoarchaean tonalitic gneisses in the Lewisian Complex. (a) Tight to isoclinal intrafolial folds in the TTG gneisses with a presumed 2.7 Ga age, northwest of the Canisp shear zone in Assynt terrane. (b) Cliff face made up of Neoarchaean TTG-gneisses with a subhorizontal foliation. Height of the cliff is ca. 50 m. Locality: in Assynt terrane. (c) Steeply dipping, alternating banded tonalitic, granitic and mafic orthogneisses. Locality: north tip of Rhiconich terrane. (d) TTG- gneiss with an older foliation cut and transposed into a steep ductile shear zone of presumed 2.7 Ga age. Locality: north of Canisp shear zone.

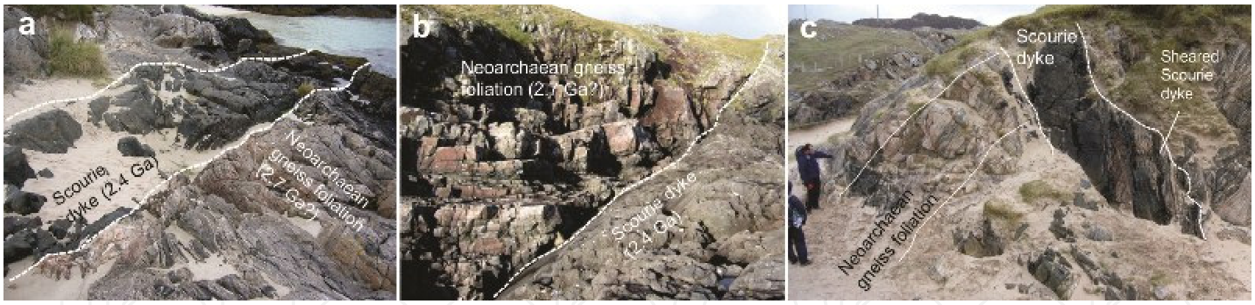

Figure 15. The Scourie dyke swarm in the Assynt terrane (see Figure 4). (a) Steep, Scourie mafic dyke that cuts through Neoarchaean gneiss foliation. Note the very sharp an unaffected intrusive contacts. Dyke thickness is approximately 5 m. Locality, north of Canisp shear zone. (b) Scourie dyke intruded into gently dipping TTG-gneisses. (c) Scourie dyke cutting Neoarchaean gneisses and which is again strongly sheared along steep, presumed 2.49 Ga Palaeoproterozoic shear zones. Locality, near contact to Canisp shear zone.

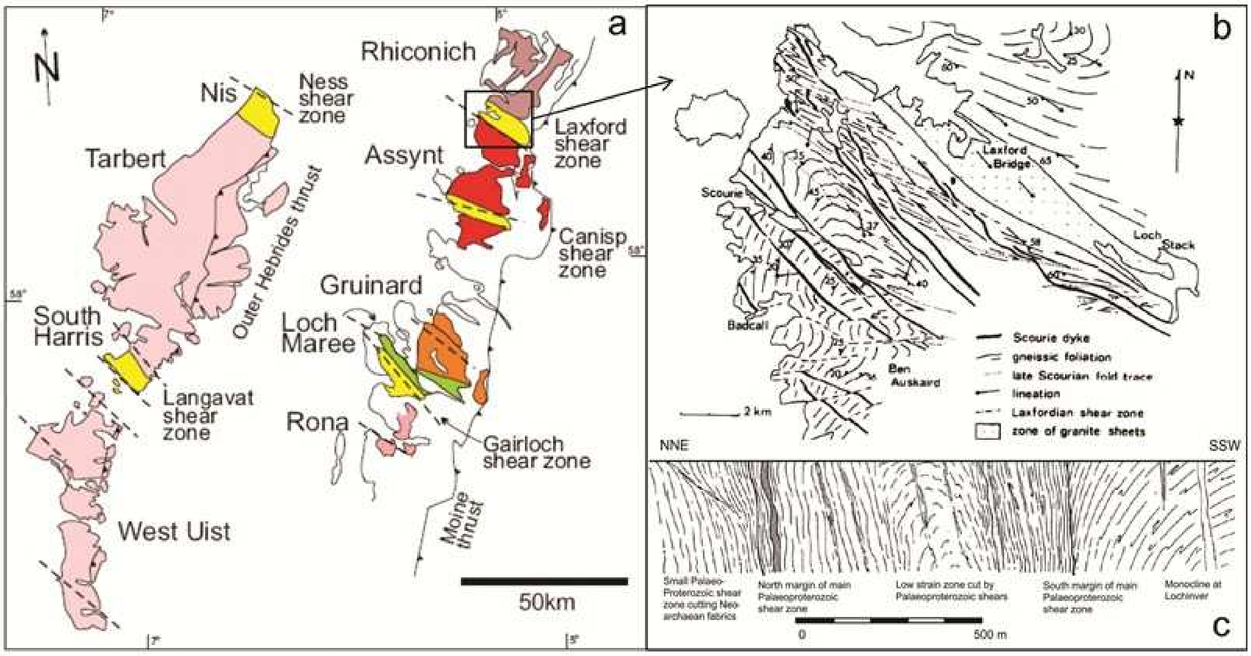

Figure 16. (a) Regional structural map showing major ductile shear zones in the Lewisian Complex, that suffered Palaeoproterozoic deformation and reworking [37]. (b) Map of the Laxford shear zone, with Scourie dykes that cut presumed regional 2.49 Ga folds that are becoming tightened toward the Laxford shear zone [37]. (c) Schematick NNE-SSW cross-section of the Canisp shear zone showing major upright folding of the gneiss foliation and formation of localized, steep, axial-planar shear zones [77].

The main episodes of Palaeoproterozoic crustal deformation in the Lewisian produced major folds and NW-SE striking, dextral-reverse, transpressive shear zones (Figure 16a) that were superimposed on, and largely obliterated the pre-Scourie dykes fabrics, except in some low strain lenses. An early/main stage of deformation involved tight to isoclinal folding of the flat-lying Neoarchaean gneiss foliation and subsequent upright folding leaving the limbs in a steep attitude (Figure 16b, 17a). In addition, localized moderately NE-dipping ductile reverse and dextral oblique shear zones developed by strain partitioning, likely due to reactivation of steep pre-Scourie dyke shear zones [77], and they affected the Scourie dykes

(Figure 17b). This event was associated with upper amphibolite, locally granulite facies metamorphism [37] and intrusion of syn-tectonic pegmatite sheets and veins in e.g. the Laxford shear zone (Figure 4). The major ductile shear zones show complex multiphase strain partitioning, including thrusting and refolding of early-stage subhorizontal shear zones developed parallel to the pre-existing foliation [78] and later development of steep strike-slip shear zones [77]. The late stage of deformation coincides with retrogression to greenschist facies conditions [79] associated with the formation of steep-plunging asymmetric folds and retrogressed cleavage and strike-slip shear zones [37].

From northeast to southwest the main shear zones are the Laxford, Canisp, Gairloch and Diabaig shear zones (Figure 16a). The ca. 8 km wide, SW-dipping *Laxford shear zone* [37] is a major zone of folding (Figure 16b) that reflects thrusting of the gneisses in the Assynt block over the Rhiconich gneisses to the north. This presumed terrane-bounding shear zone [80] evolved from a pre-existing, steep fabric that displayed early stages of sinistral and dextral shearing and subsequent oblique-thrusting and dextral strike-slip movement [81], evidenced by SSE plunging stretching lineations [37]. Numerous granites and pegmatite sheets were injected within this shear zone.

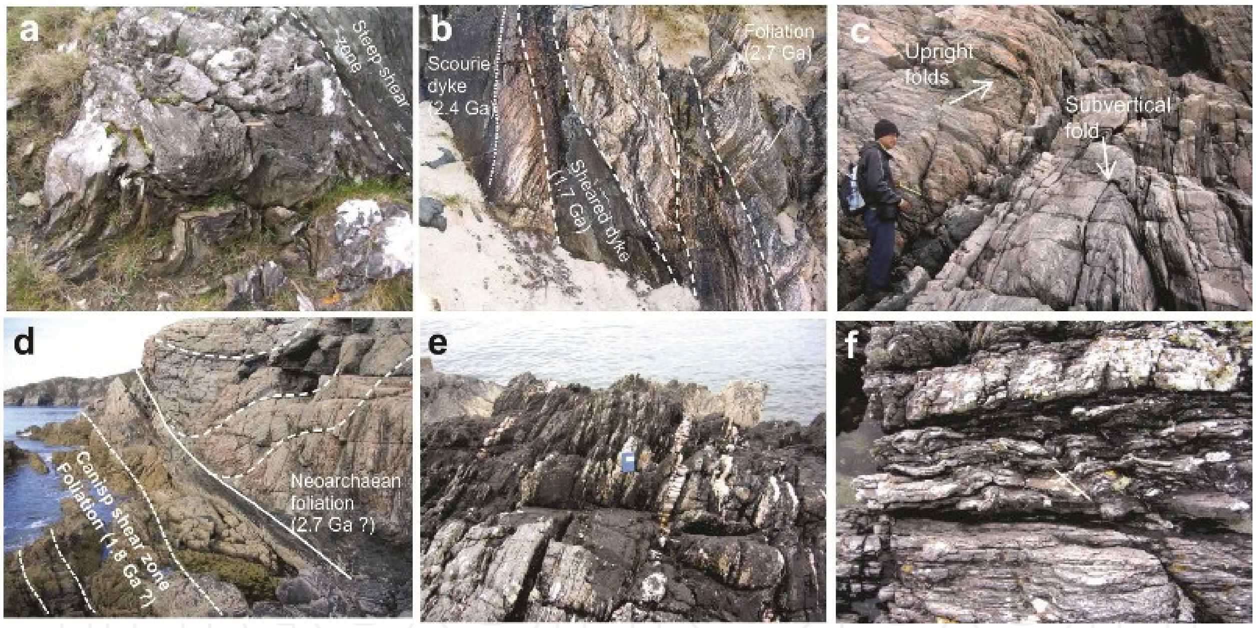

Figure 17. Outcrop features illustrating Palaeoproterozoic (c. 1.9-1.67 Ga) deformation fabrics in the Lewisian Complex. (a) Meso-scale upright folds within low-strain domain of Canisp shear zone. Hammer is parallel to fold axis, ESE-WNW, and the axial-surface dips steeply to the ENE. A steep, SSW-dipping mylonitic shear zone developed on the fold limb (to the right). (b) Steep shear zones that cut and displace a Scourie dyke and the Neoarchaean gneiss foliation near Canisp shear zone. (c) Upright folding of TTG-gneisses outside the Canisp shear zone (left) and refolding by tight sub-vertical folds within the shear zone (central and right). View is toward ESE. (d) Contact between TTG- gneiss and the Canisp shear zone. Note subvertical sinistral drag-folding of the gneisses into the mylonitic shear zone. View is toward WNW. (e) High-strain mylonite in the Canisp shear zone, with retrogressed chlorite-mylonitic schist and quartz veins aligned along the main fabric. View is toward the WNW. (f) Detail from the steep mylonite zone in e, showing asymmetric sinistral folding of the main fabric, including the quartz veins, seen on a horizontal view .

Farther south, the Canisp and Gairloch shear zones define steep oblique crust-internal shear zones of the Assynth and Gruinard terranes (Figure 16a). The *Canisp shear zone* dips steeply SSW [37] and typically truncates the Scourie dykes (Figure 17b) and displaces them into zones of alternating high- and low-strain [77]. Variably dipping high-strain thrust zones evolved in the hinge zone of a major ESE-WNW trending fold (Figure 16c), while steep high-strain mylonitic shear zones truncate the limbs of the macro-fold (Figure 17c, d). Strain partitioning is observed at all scales, and low strain lenses are typically overprinted and transposed into high strain zones. Folds in the high strain zones are asymmetric, tight and have generally steep plunge (Figure 17c). These folds may have been produced by early deformation which partitioned later into the higher strain domain as tighter and steeply plunging folds (Figure 16b), leaving a weak signature in the low strain domain [82]. Fabric-parallel quartz veins are abundant in the high-strain zones, and these veins have been further deformed and folded internally (Figure 17e, f). Stretching lineations in high-strain parts of the *Canisp shear zone* suggest dextral-oblique-reverse movement (south-side up thrust) followed by strike-slip shearing [77, 79, 82]. The *Diabaig shear zone* is thought to be an inclined thrust ramp dipping toward NE. Similar shear zones exist in the Outer Hebrides portion of the Lewisian Complex (see Figures 4 and16a).

3.2. A chronology of events

In the past sixty years the Lewisian Complex has been the subject of very extensive geochronological studies with the application of many different decay systems. Much of the initial work, especially that done with whole rock methods, documented the existence of Archaean and Palaeoproterozoic events, matching the subdivision proposed by [70] based on the pre- and post- Scourie dyke position of the rocks [83-88]. The details of the picture, however, have remained fuzzy due to the complications introduced by the multistage evolution of the rocks. More recent dating of zircon and other minerals with U-Pb has helped to shed more light into the timing and importance of the events in the different domains. This evolution included five major events, interspersed by some minor but tectonically important events (Figure 6): (i) Meso- and Neoarchean orogenic activity, mainly between 3.00 and 2.70 Ga, built the bulk of the Lewisian crust; (ii) an earliest Palaeoproterozoic event at 2.50-2.48 Ga, had a profound influence on the Assynt block, but is not seen elsewhere, except for anorthosite in South Harris; (iii) emplacement of Scourie dyke during at least two episodes at 2418 and 1992 Ma; (iv) deposition of clastic sediments at Loch Maree sometime between 2.0 and 1.9 Ga, leading to, or associated with a localized but intense episode of arc magmatism at 1.90-1.85 Ga at Loch Maree, Laxford Bridge and South Harris; and (v) local migmatization and emplacement of pegmatite dyke swarms at 1.70-1.65 Ga, mainly in South Harris but recorded by sporadic pegmatites and titanite across the entire Lewisian Complex.

i. Meso- and Neoarchaean evolution

The Lewisian Archaean crust is dominated by banded, felsic to intermediate TTG-gneisses of presumed igneous origin. Meso- and Neoarchaean supracrustal rocks composed of semipelites, calc-silicate schists, meta-arkoses and metavolcanic rocks associated with mafic to ultramafic and anorthositic rocks with a tholeiitic composition are also preserved, and when present, may represent the protoliths of some of the gneisses. The metamorphic grade reached granulite facies in the Assynt block, and varies from granulite to amphibolite facies elsewhere. The geochronology of these gneisses tends to be very complex because of the multiple overprints which are recorded in zircon so that in many cases the U-Pb data are scattered and do not entirely resolve the sequence of events. However, the overall picture permits to distinguish a main pattern typical of the Assynt block and distinct from that of the other segments (Figure 6).

The TTG rocks in the Assynt block preserve some of the oldest zircon ages of 3.04-2.96 Ga and it has been suggested that this is the age of formation of the rock [89-91], but it remains uncertain whether the oldest grains may just be xenocrystic and the gneisses actually formed about 2.85 Ga [92]. The early high grade metamorphism occurred at around 2.80-2.75 Ga, and was followed by the intrusion of trondhjemite at 2.72 Ga [93-95] and emplacement of mafic-ultramafic rocks [88].

The other Archaean segments of the Lewisian rocks had the most prominent period of development between 2.9 and 2.0 Ga, with a peak at 2.84-2.82 Ga [7, 90, 91, 93, 96-101]. An exception is the Rona segment where ages between 3135 and 2880 Ma have been reported for tonalitic gneisses [102]. The Gruinard block underwent granitic to trondhjemitic magmatism and high-grade metamorphism at about 2.73 Ga [91, 96, 97] and in the Richonich block there are some indications for further activity as late as ca. 2.6 Ga [90].

ii. Earliest Palaeoproterozoic metamorphic progression: 2.50-2.48 Ga

The Assynt block was affected by a second high-grade metamorphic event at 2.50-2.48 Ga. This event caused strong resetting of U-Pb in zircon due to recrystallization and local new-growth [90-94] and is also recorded by Sm-Nd isochrons obtained from garnet and coexisting minerals in ultramfic pods [104], as well as by U-Pb in titanite and monazite [94, 96]. The dry high temperature metamorphism was concluded by re-hydration which caused local formation of granitic pegmatites and leucosome in migmatites [94] and likely led to retrogressive amphibolite facies metamorphism and deformation, termed the Inverian [73, 74]. The only well constrained temporal analogue to these metamorphic events elsewhere in the Lewision Complex is an anorthosite body present in the younger South Harris igneous complex [105].

iii. Scourie dyke swarms

The Scourie dykes form part of an extensive dyke swarm present throughout the Lewisian Complex [37] but most abundant in the southern region. These dykes display many different geometries and attitudes, mostly steeply dipping, and they cut all Neoarchaean folds and planar fabrics (Figure 15). The composition of the dykes varies from mafic to ultramafic [105], and their trace element geochemistry indicate formation in a marginal continental setting, as island arcs or from a mixtures of crustal and oceanic material [106, 107]. The Scourie dykes intruded during at least two different events at 2418 Ma and 1992 Ma [72, 75, 108, 109]. The youngest of these dykes intruded into hot gneissic and migmatitic rocks syn-kinematically with a late stage of shear zone deformation [79, 107], of the same kind as that shown in Figure 14d.

iv. Deposition of supracrustal rocks (2.00-1.90 Ga) and arc magmatism (1.90 – 1.85 Ga)

Banded iron-formation, marble, chlorite schists and meta-psammites of the Loch Maree Group were deposited on the Neoarchaean crust and are now arranged in two narrow, NW-SE trending synformal belts [110, 111] in the Gairloch area (Figure 18). Deposition of the Loch Maree Group is constrained between 2.0 Ga, the youngest detrital zircons from a metagreywacke in the Gairloch area [112], and 1903 Ma, the age of the syntectonic Ard gneiss intrusion [102, 110]. The latter was emplaced during the early stages of deformation associated with amphibolite to granulite facies metamorphism and interpreted to be related to the development of a subhorizontal shear zone net work. The Ard gneiss is considered to be a product of arc magmatism and the deformation caused by lateral accretion of oceanic plateaus and primitive island arcs [110].

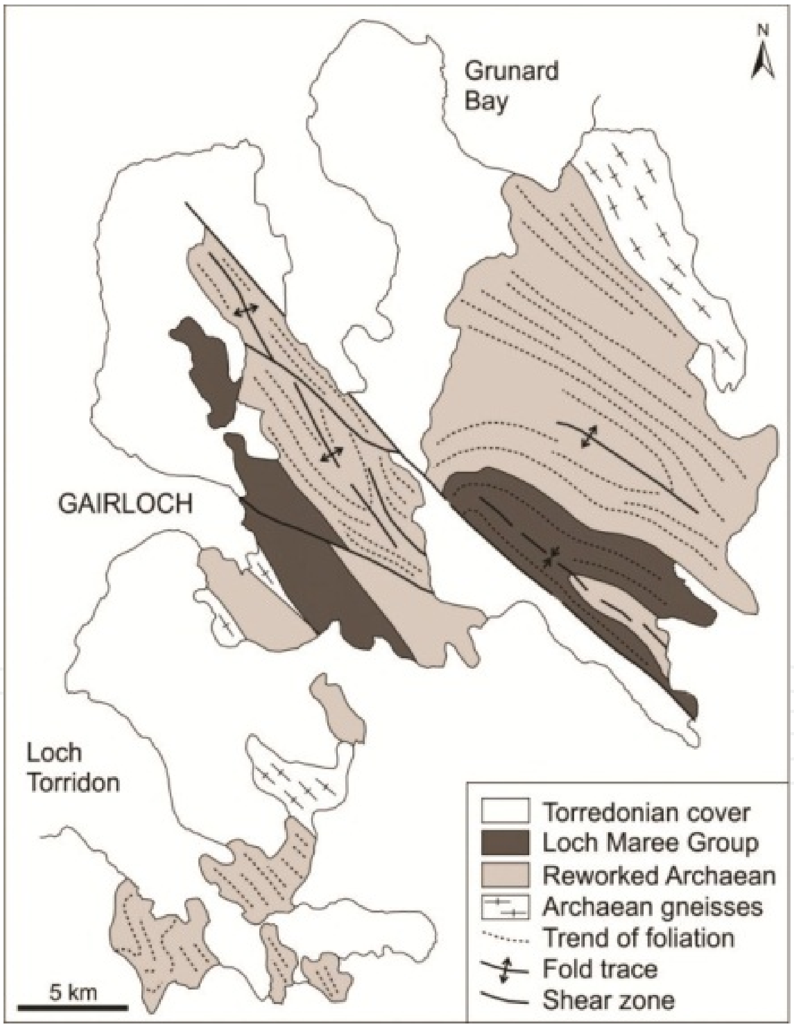

Figure 18. Simplified map of the Palaeoproterozoic Loch Maree Group (shaded) within the Lewisian Complex of the Gairloch area, NW Scotland [110]. This group consists of highly deformed amphibolites and metasedimentary rocks cut by 1.9 Ga granitoid alkaline rocks. Note macro-scale reverse and sinistral folds and related off-sets and lateral shear zone displacement.

The other important tectonic element with similar age and genesis is the South Harris igneous complex in the Outer Hebrides (Figure 4) that comprises a magmatic arc sequence of mafic to felsic intrusive rocks formed at 1.89-1.88 Ga [98, 100]. Clastic sedimentary rocks in this region contain abundant detritus of the same age [7, 98]. Crudely coeval high grade metamorphism strongly overprinted zircon in the 2.5 Ga anorthosite [104] and has also been dated by Sm-Nd mineral systems [113]. This event also caused deformation and metamorphic zircon growth in a shear belt at the northern tip of the Tarbert block at Nis (Figure 6) [98, 99]. As for the Loch Maree assemblage, the South Harris situation has been interpreted 'to represent a magmatic arc, complete with contemporaneously derived clastic sediments, developed in a collisional orogen, which culminated in granulite facies metamorphism' [98].

A third important occurrence of intrusive rocks of this age is within the Laxford shear zone. Granite sheets, with a U-Pb age of 1854 ±13 Ma [72] occur subparallel to the shear zone boundaries but cutting the pre-Scourie dyke fabric (Figure 19a, b). Synchronous shear deformation affected the dykes and aligned them as lenses into shear zones that also affected the surrounding tonalitic and mafic gneisses by mostly contractional deformation (Figure 19b-d). There is some consensus that the Laxford shear zone likely represents a terrane boundary but the timing of juxtaposition is debated, and the role played by the granites unclear. Recent work [80] concludes that the terranes were probably brought together sometime between 2.5 and 2.4 Ga, after the earliest Palaeoproterozoic deformation and retrogression, but before intrusion of the Scourie dykes. Others, however, suggest that the juxtaposition probably occurred around the time of emplacement of the granites, consistent with the timing proposed for the Loch Maree group and South Harris igneous complex [7, 8, 114].

v. Late orogenic events (1.70 – 1.65 Ga)

A metamorphic event at around 1.75 Ga is indicated by a Sm-Nd age of garnet and coexisting metamorphic minerals from a mafic dyke of the Assynt block [109,], and by titanite from rocks near the Laxford shear zone [90, 93]. Because of the localized occurrence of the rocks, however, the tectonic significance of this age is uncertain.

By contrast a later event at 1.70 to 1.65 Ga had a much stronger impact across all of the Lewisian Complex (Figure 6). The main expression of this event is the granitic and pegmatitic migmatite complex associated to, and bordering, the South Harris igneous complex [7, 100, 101, 115, 116]. Felsic dykes and pegmatites of this age have also been found in most other parts of the Lewisian [7, 98, 110] and they also coincide with the ages of rutile and a younger titanite generation [76, 93, 102]. The exact significance of this event is still uncertain. One view is that emplacement of pegmatites of this ages pre-dated, but likely broadly coincided with the late flexuring, steep shearing and greenschist facies retrogression [110], thus attributing these events to contractional processes during late-stage collision of the Lewisian Complex with e.g. a southern block [8].

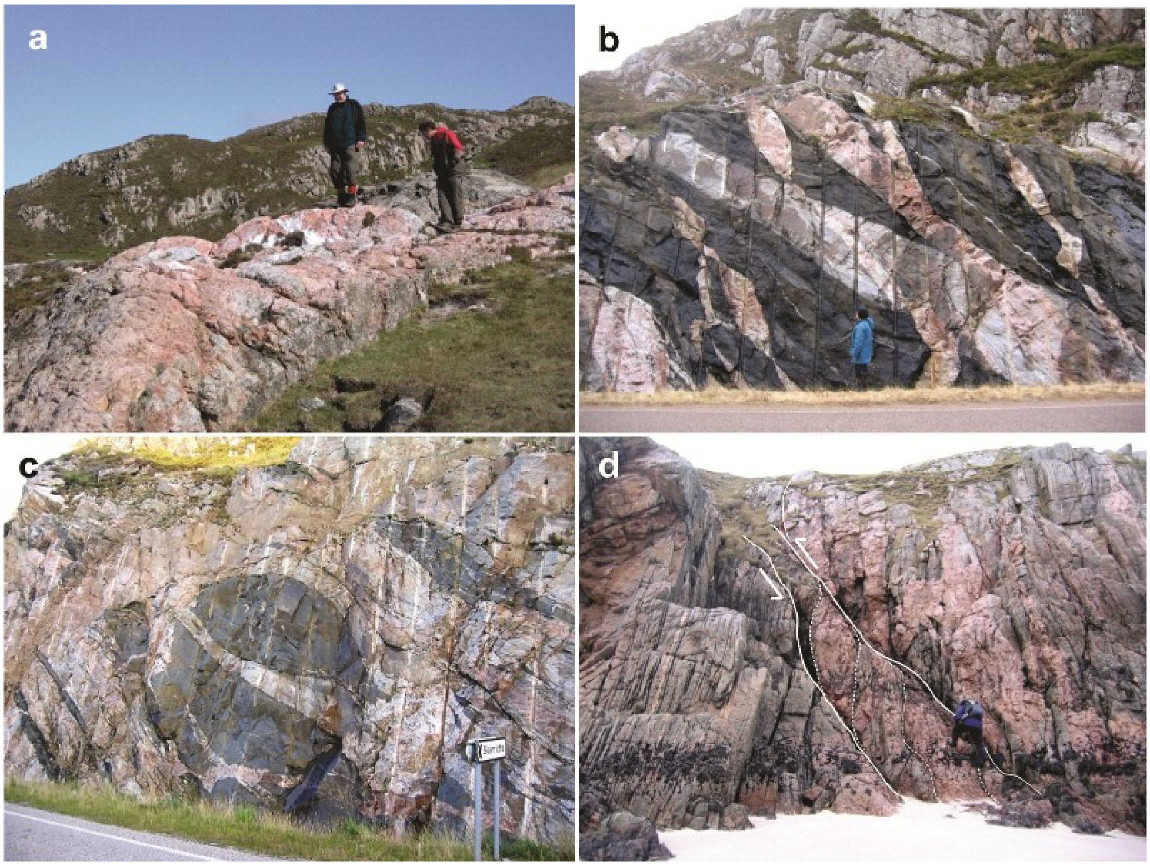

Figure 19. Outcrop features of Palaeoproterozoic intrusive rocks in the Lewisian complex. (a) Granitic pegmatite intrusion into TTG-gneisses of the Rhiconich terrane. (b) Syn-kinematic felsic pegmatite dykes intruded into mafic gneisses and sheared during the main Palaeoproterozoic tectono-thermal events. Note boudinaged and asymmetric lenses indicating top-NE movement sense (right in photo). View is toward SE. (c) Syn- and post-kinematic granite pegmatite dykes injected into mafic Neoarchaean gneisses. Note lensoidal sheaped mafic bodies in the sheared gneisses. (d) Revers ductile shear zone displaying a duplex geometry, cutting obliquely across steeply dipping TTG-gneisses in the Rhiconich terrane. View is to the NW.

4. Discussion

4.1. Similarities of terranes and terrane juxtaposition

In order to discuss terrane aspects and hypothetic assembly history of the West Troms Basement Complex, the Lofoten-Vesterålen province and the Lewisian Complex as a framework for further correlation, we focus on critical similarities such as the presence of crustal segments with contrasting age, tectono-magmatic and/or metamorphic histories, crustal-scale ductile shear zones (sutures), overlapping nature and character of deformation (convergent and strike-slip) and major metamorphic breaks that may have juxtaposed different crustal levels instead of spatial terranes [8, 72]. The first step would be to locate different crustal segments, then to localize the suture(s) formed by the juxtaposition of terranes, and finally, discussing assembly of different crustal levels in order to explain metamorphic and petrologic differences known from all the studied regions [8, 17].

Although metamorphic and structural characteristics of Precambrian continental crust in general is not uniquely diagnostic for correlation [107], the geotransect in western Troms and Lofoten is underlain by age-equivalent TTG-gneisses and granitoid gneisses of Meso- and Neoarchaean age, with only modest differences in metamorphic grades and histories [32]. This indicates that at the end of the Archaean these rocks were likely assembled as a large single terrane, formed during a prolonged cratonization event (2.92-2.67 Ga), though with considerable complexity in detail [32]. In this scenario, the migmatized ductile shear zones bounding the Dåfjord and Kvalsund gneisses on Ringvassøya [17] and, potentially also the major tectonic and metamorphic boundary in the Lofoten-Vesterålen area [10] may reflect Neoarchaean intra-cratonic terrane boundaries. An argument for multiple terranes in the West Troms Basement Complex at the end of the Neoarchaean, however, appears from the contrasting ages, e.g. 2.8 Ga for the Ringvassøya greenstone belt, 2.4-2.2 Ga for the Vanna group [29] and c. 1.9 Ga for the Torsnes belt [31].

By comparison, the Lewisian Complex was earlier considered to have originated from a single Neoarchaean continent that split up into multiple terranes in the early Palaeoproterozoic and later on were juxtaposed during the 1.90-1.67 Ga events [79]. A revised model of Palaeaoproterozoic juxtaposition in the Lewisian, however, included ten terranes and one block [7, 72, 90]. The Laxford shear zone was considered a terrane boundary between the Rhiconich terrain in the northeast and the Assynth terrane in the southwest (see Figure 4). These two terranes were likely accreted after intrusion of the pegmatite sheets in the Rhiconich terrane at c. 1.855 Ga, since they are absent in the Assynth terrane, but prior to the early-Palaeoproterozoic structures common in both regions. One model suggests that the accretion occurred at 1.74 Ga, synchronously with an amphibolite facies metamorphism recorded in the Rhiconich terrane and a metamorphic retrogression associated with the formation of shear zones in the Assynth terrane [7]. Recently, it was argued that juxtaposition of the Assynth and Rhiconich terranes occurred *prior* to the 1.9-1.75 Ga period, e.g. in late Neoarchaean since Scourie dykes are present on both sides of the Laxford shear zone [80].

A second model [8] involved only two continental plates during the Neoarchaean and Palaeoproterozoic history of the Lewisian Complex. In this model the Lewisian Complex was divided into two blocks classified as *upper-plate* and *lower-plate* blocks (Figure 20) that differed considerably with regards to position of the supracrustal rocks relative to the accreted versus the overriding plates, which is critical for the metamorphic conditions. The *upper-plate* block preserved rare and weak Palaeoproterozoic deformation and amphibolite/greenschist facies retrogression of granulite facies gneisses, while the *lower-plate* portion involved prograde and peak amphibolite facies metamorphism and high-strain assemblages. The *upper-plate* blocks displayed weak deformation and could be located on the low-strain areas above the terrane boundary shear zone in the crustal model for the mainland [8, 37], while the *lower-plate* blocks with strong Palaeoproterozoic structures could correspond to the mid-deep level of the shear zone itself, i.e. as during the emplacement of the Loch Maree Group (Figure 20).

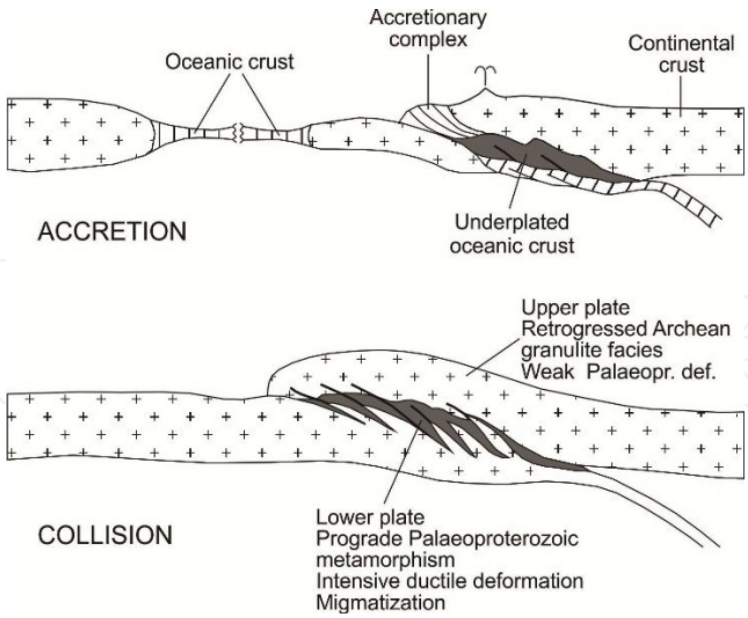

Figure 20. Schematic model of an idealized subduction-accretion collision sequence [8] leading to domains of contrasting deformation and metamorphism, e.g. high-grade/prograde in *lower plate* and low-grade retrogressive metamorphism in *upper plate* settings.

By analogy, in the West Troms Basement Complex, a significant Palaeoproterozoic tectono-metamorphic break is thought to exist southwest of the island of Senja (Figure 2), in a region that separates dominantly amphibolite facies gneisses from granulite facies AMCG-suite rocks of the Lofoten-Vesterålen province [9, 10]. This major boundary is also inferred by contrasting gravity and magnetic characters, and could therefore reflect a Palaeoproterozoic suture [11, 13, 50]. The slightly older ages of the peak Palaeoproterozoic deformation (1.87-1.78 Ga) versus 1.78 Ga in the West Troms Basement Complex (Table 1) suggest progressive southwestward accretion toward an orogenic hinterland near Lofoten (Figure 21), which is also consistent with the observed increase in metamorphic grade. On the other hand, the contrast in U-Pb ages of basement rocks in the southwest (2.7-2.6 Ga) compared to in the north (2.92-2.8 Ga), could indicate a second terrane in the northeast (Figure 21)[10, 17].

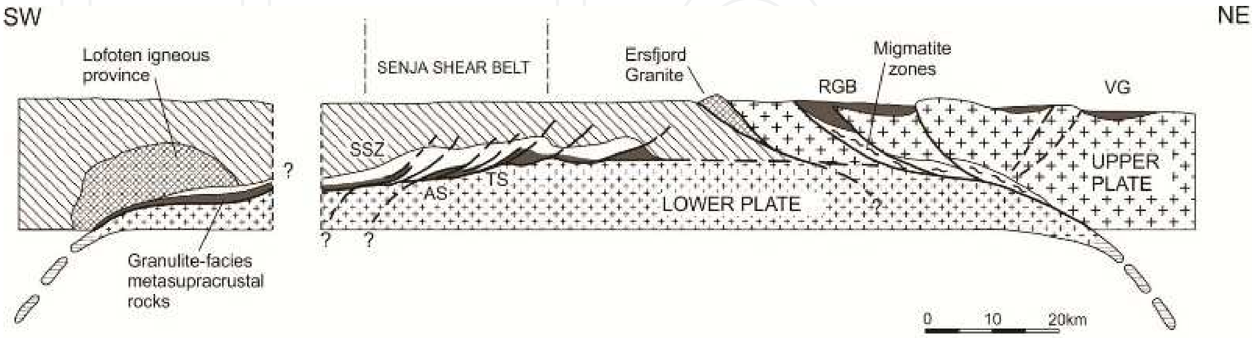

Figure 21. Schematic model of Palaeoproterozoic accretion in the West Troms Basement Complex. Lower – Upper plate model used to explain among others, metamorphic differences [8].

A model involving SW-directed convergence/accretion toward an orogenic front near Lofoten and an oppositely NE-dipping terrane boundary zone in the northeast (e.g. reactivated Neoarchaean migmatite zones in Ringvassøya) may explain both the contrast in deformation styles and metamorphic grades along the geotransect. For example, an accretion/subduction-derived shear zone situated in the Senja shear belt adjacent to eastern Lofoten-Vesterålen would locate them to the *lower plate* of two or more unified colliding terranes (Figure 21, left). This would have formed prograde amphibolite to granulite-facies metamorphic assemblages, granitic melts, migmatite zones and strong ductile deformation, including a detachment in the Senja shear belt that could have become the locus of later-stage partitioned crustal deformation. Conversely, in an upper plate position (Figure 21, right), localized and weakly developed but more likely retrogressed, low-grade metamorphosed cratonic-marginal shear zones and supracrustal platform sequences may appear (as in Vanna). This model would favor assembly of different crustal levels of at least two main crustal segments (terranes) rather than spatially separated smaller terranes [8].

4.2. Comparison between Lewisian, western Troms and Lofoten-Vesterålen

Plate tectonic reconstructions of Precambrian units in the North Atlantic realm with respect to Fennoscandia and Laurentia (Figure 22) have to some extent failed to demonstrate whether these cratons belonged to the same supercraton in the Neoarchean (2.8–2.5 Ga) and Palaeoproterozoic (1.8-1.6 Ga). A tectonic linkage is supported by paleomagnetic reconstructions [8, 119] stating that Fennoscandia was positioned close to the Greenland/Laurentia and Superior supercraton in the late Neoarchean (Figure 22). Geological similarities and differences between domains are important criteria for restoring possible supercontinents, as stated by [3, 4, 119]. In this context, the studied basement outliers west of the Scandinavian Caledonides in North Norway and the Lewisian complex in Scotland both have a pivotal central location within the marginal orogenic belts constituting the presumed Neoarchaen supercontinent (Figure 1) [110, 120]. These units, however, also occupy an interior position of the Caledonian orogen far from the autochthonous shield rocks and are bounded by younger faults. They are, thus, usually not considered part of any shield areas, but instead assigned an uncertain or exotic tectonostratigraphic status [3, 121].

Based on the comparison between the Archaean-Palaeoproterozoic basement suites in North Norway and the Lewisian of Scotland outlined above we can discuss potential correlation of these suites in the context of the North Atlantic realm. Such a correlation can be tested using similarities or dissimilarities in lithology, age, supracrustal units, igneous/ petrogenetic, structural and metamorphic features and evolutionary and tectono-metamorphic history (see Table 1).

4.2.1. Archaean components

Archaean crust forms the backbone of both the western Troms, Lofoten-Vesterålen province and the Lewisian basement complexes. These complexes reveal some broad similarities in terms of lithology and general age patterns but also some differences, which, however, are most pronounced within each of the regions (Figure 6). Thus, we are comparing two heteorgeneous Archaean crustal segments.

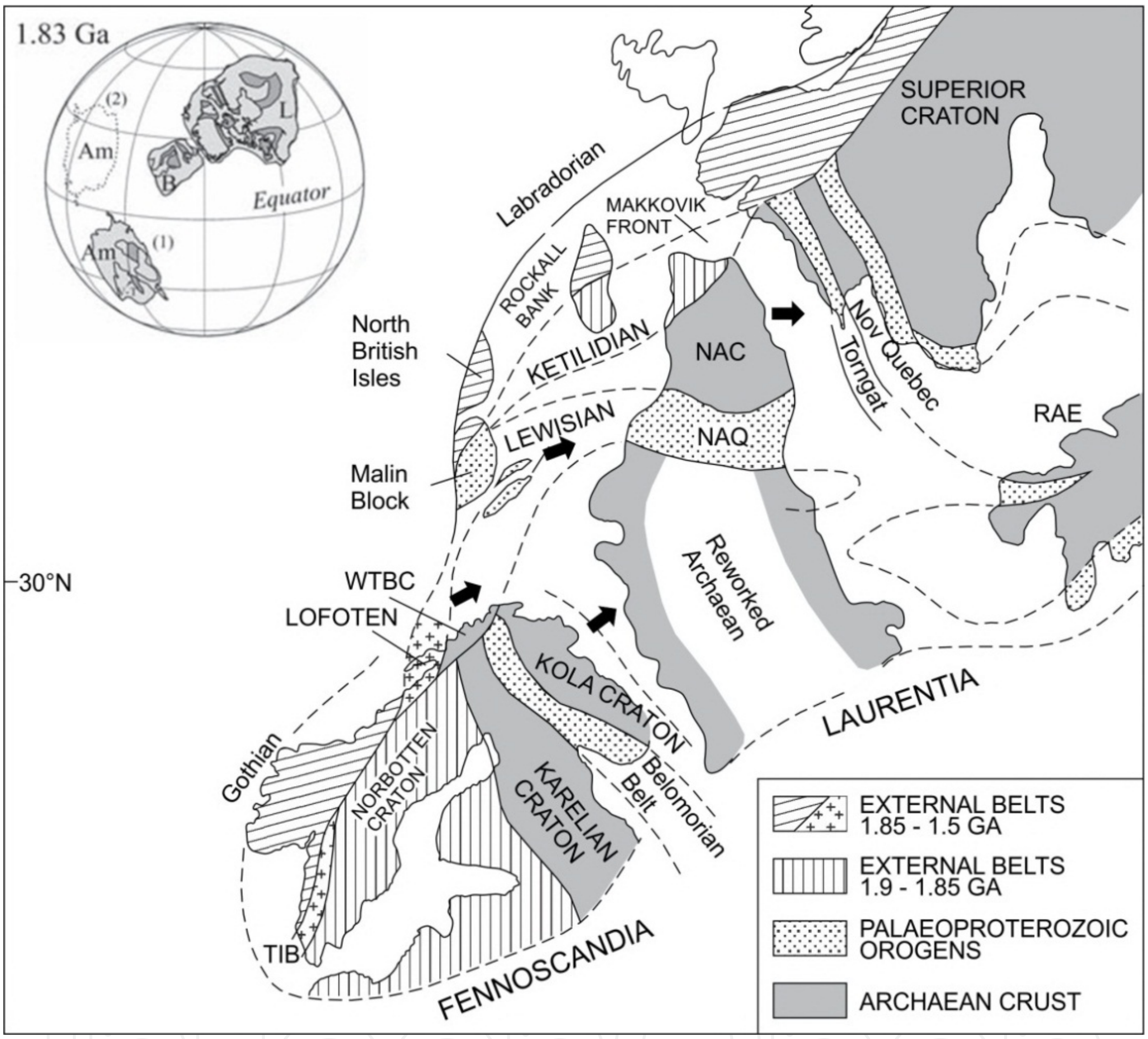

Figure 22. (a) Reconstruction of Laurentia and Fennoscandia during the Palaeoproterozoic based on the palaeomagnetic fit of [118]. Note that the West Troms Basement Complex (including Lofoten) and the Lewisian Complex lies within a continuous Palaeoproterozoic belt extending from the Torngat orogen of Laurentia through the Ketilidian and Nagssugtoqidian orogens to link up with the Kola and Karelian/Belomorian provinces of Fennoscandia. The arrows show inferred movement directions of various crustal segments relative to Laurentia. The map is modified from [4, 110], while the reconstruction of continents at 1.83 Ga is based on [118]. Abbreviations: WTBC =West Troms Basement Complex.

In the West Troms Basement Complex we can distinguish: (i) a Mesoarchaean tonalitic domain formed between 2.9 and 2.8 Ga in Ringvassøya and Vanna, overlain by (ii) the broadly coeval, but tectonically distinct Ringvassøya greenstone belt. These two domains are separated by the (iii) late orogenically active Kvalsund gneiss migmatite zone from (iv) the Neoarchean domain of Kvaløya and Senja farther south, which formed during a short time interval between 2.72 and 2.67 Ga. The Lofoten-Vesterålen province has broadly the same age structure as the southern part of the West Troms Basement Complex, except for the fact that it underwent high-grade metamorphism at 2.63 Ga. The apparent formation of leucosome in southern Senja at around 2.61 Ga [32] may be the expression of the same event, indicating that Senja and Vesterålen may simply represent different crustal levels experiencing metamorphism at different times, a very common pattern of crust construction and maturation [122]. The late Archaean deformation and migmatization that seems to characterize the Kvalsund migmatized zone separating the Dåfjord and Kvalsund gneisses [17, 32] may be another expression of this late orogenic activity.

In the Lewisian there are also some differences in the Archaean history of the contrasting blocks (Figure 6). Several of them appear to have formed mainly in the late Mesoarchaean, between 2.85 and 2.80 Ga. Older ages up to 3150 Ma are recorded in the Rona and Assynt blocks whereas Neoarchean magmatic activity seems to be restricted mainly to the Assynt and Gruinard blocks, although there are local indications of such activity also in the Tarbert-West Uist and Rhiconich blocks.

The single most distint event that is fully missing in the West Troms Basement Complex and Lofoten-Vesterålen is the pervasive high-grade metamorphism and subsequent rehydration and retrogression in the Assynt block at 2.5-2.48 Ga. However, this event is not seen in the other Lewisian blocks except for the anorthosite body in South Harris (Figure 6). One explanation is that it was a specific terrain formed in another orogen, the alternative is that the 2.5 Ga high-grade event reflects a process affecting lower crustal levels but not recorded higher in the crust, in the same way as seen for example in the Superior Province [123].

A comparison of the evolution of these two crustal sectors is difficult because of the inherent internal differences, which may reflect dependencies from crustal level, and probably also the tectonic juxtaposition of different terranes combined with the unequal geochronological coverage in different blocks and the technical difficulties to cleanly date the age of protolith and orogenic tranformations of complex polymetamorphic gneisses. Hence we can conclude that the West Troms Basement Complex, the Lofoten-Vesterålen province and the Lewisian have certain affinities in common (Figure 6), suggesting that they could have been linked to some degree in the Archaean period, but the opposite conclusion is also possible.

2.40 to 1.98 Ga supracrustal rocks and mafic dyke swarms

Intrusion of the huge Ringvassøya mafic dyke swarm in the West Troms Basement Complex (Figure 23b) occurred at c. 2.40 Ga (Figure 6) [34]. This event is part of a major mafic dyke-producing event that affected several Archaean cratons, and as such it does not necessarily represent a unique stitching tool for linking these crustal domains. One argument supporting such a role, however, is the apparent south-westerly shift in age from ca 2.5 Ga events in Kola, to the most widespread phase at 2.45 Ga in Kola and Karelia, and finally to the 2.40 Ga phase in Ringvassøy [34], which is close to the age of the older Scourie dyke generation in the Lewisian. The 2221 Ma dioritic sill intruding meta-sedimentary rocks of the Vanna group [29] is also the expression of a localized but very ubiquitous magmatic phase across northern Fennoscandia [124] and also in Laurentia. No equivalents have so far been described from the Lewisian. However, there is a good temporal correlation, instead, between mafic magmatism at 1.98 Ga in the Mjelde-Skorelvvatn belt of the West Troms Basement Complex [31] and emplacement of the younger generation of the Scourie dykes in the Lewisian, both corresponding to a period of extension and rifting in Laurentia and Fennoscandia.

1.90-1.85 Ga arc magmatism and convergence

The subsequent period of arc magmatism, likely connected to plate convergence and subduction, and final collision was important in the Fennoscandian Shield [4] and also in the Lewisian where it formed the well documented successions at Loch Maree and South Harris in a sequence of events between 1.90 and 1.85 Ga. Granitic sheets of this age are also important along the Laxford shear zone. In Lofoten there was a correlative event at 1.87 Ga, which emplaced granite and local mangerite-charnockite intrusions. Such rocks, however, have so far not been reported from the West Troms Basement Complex, a feature that may reflect a more distal position relative to the orogenic front near Lofoten (Figure 23c; see below). In the Lofoten-Vesterålen province, there is also clear evidence of meta-supracrustal units that post-date the Neoarchaean gneisses [9], but Palaeoproterozoic ages have not yet been documented by radiometric age dating.

1.80 – 1.78 Ga magmatism

The single most important and widespread magmatic event affecting the West Troms Basement Complex and Lofoten-Vesterålen province occurred in a short burst at 1.80 -1.79 Ga. It formed most of the AMCG suite in Lofoten and the major Paleoproterozoic intrusions in Kvaløya and Senja. In Lofoten the event was pre- and post-dated by high grade metamorphism and ductile deformation [9, 10], whereas in western Troms there is no evidence for much activity preceding this magmatic phase. These events can be correlated with a well-defined period of late orogenic magmatism in Fennoscandia [10, 125]. Interestingly, there are no such plutons or strong metamorphic overprint of this age in the northern part of Ringvassøya and Vanna, even though granulite facies metamorphism and partial melting occurred at about 1.78 Ga in Sandøya, just at the edge of this block, supporting an allochthonous origin of the latter [17]. A similar situation is also characteristic of the entire Lewisian which lacks 1.80-1.78 Ga intrusion altogether

1.80-1.75 Ga deformation and metamorphism

Regional deformation and metamorphism are well documented in the West Troms Basement Complex at c. 1.80-1.75 Ga (Table 1). These processes involved high-strain deformation and prograde metamorphism up to granulite facies (1.78-1.768 Ga). The deformation was focused mainly along the boundaries to metasupracrustal belts, e.g. in the Senja Shear Belt (Figure 10, 23c, d), and was probably also superimposed on pre-existing Neoarchaean structures [17, 32, 51]. The deformation started with ENE-directed thrusting and was followed by macroscopic upright folding and combined, late-stage strike-slip shearing and SE-directed thrusting (Figure 24) [17]. The late stages of deformation, not yet documented by age datings (but likely younger than 1.75 Ga), were characterized by partitioned contraction and lateral displacements. In the limbs of the mid-stage macroscopic folds (Figure 24b) the subsequent oblique deformation produced foliation-parallel sinistral strike-slip faults and steeply-plunging folds (Figure 24c), creating a regional lens-shaped structural pattern in the West Troms Basement Complex (Figure 10). The final phase of SE-directed thrusting (Figure 24d) was temporally linked to the strike-slip shearing, thus indicating partitioned transpression as the overall deformation mechanism [17].

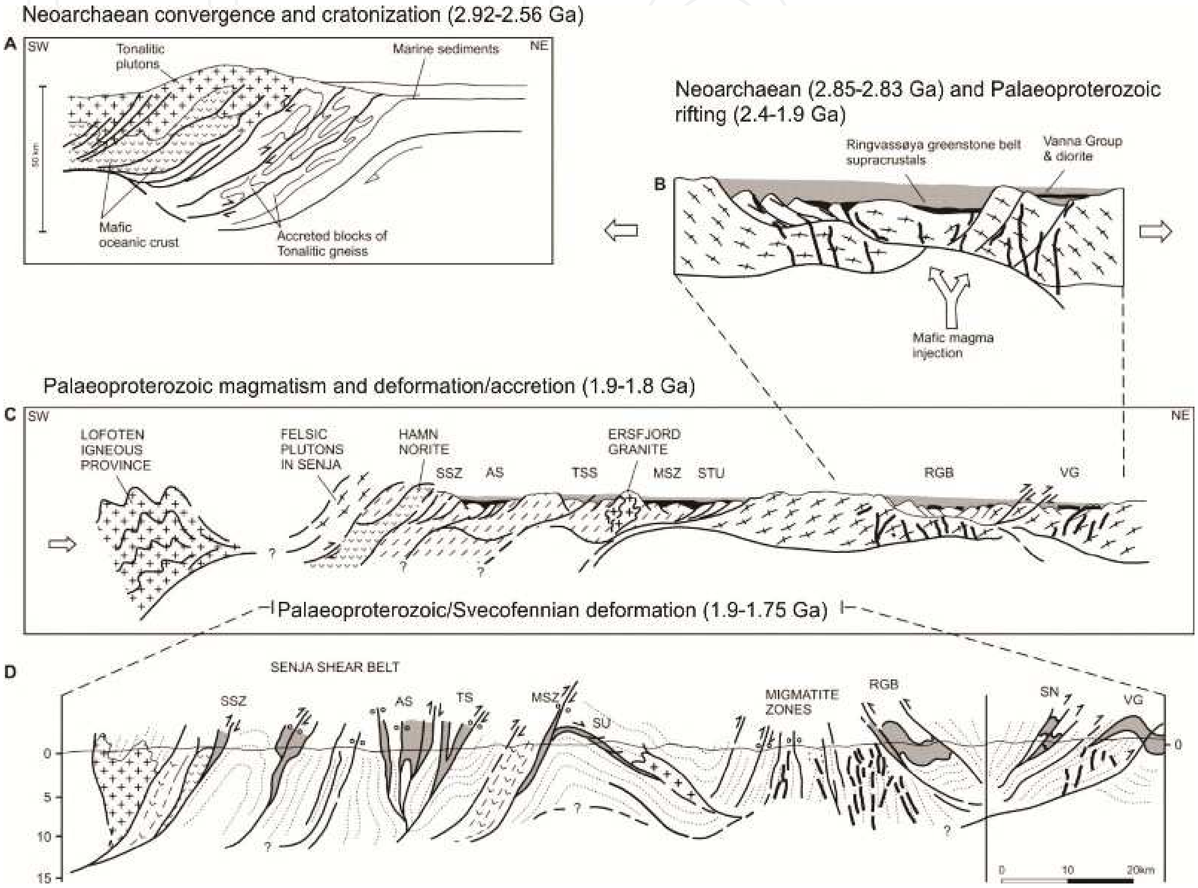

Figure 23. Cartoon sections summarizing the Neoarchaean to Palaeoproterozoic tectonic evolution of the West Troms Basement Complex: (a) Neoarchaean (2.92-2.56 Ga) tonalitic gneiss forming events with crustal accretion and thickening due to underplating. Note position of possible precursory volcanic deposits of the Ringvassøya greenstone belt. (b) Neoarchaean and Early Palaeo proterozoic (2.4-1.9 Ga) crustal extension, basin formation and intrusion of the Ringvassøya mafic dyke swarm. (c) Palaeoproterozoic (1.9-1.8 Ga) continental contraction and probable magmatic arc accretion in the southwest (including the Lofoten AMCG suite). (d) Section illustrating the composite result of Palaeoproterozoic crustal contraction, accretion and continent-continent collision with increasing transpressive deformation through time. For abbreviations, see Figure 2.

In contrast, although the deformation style is similar, there is not full evidence for temporally equivalent deformation events in the Lewisian. The exception is titanite ages of about 1750 Ma near the Laxford shear zone. These ages are considered as the potential expression of a phase of regional metamorphism but the evidence in favor of such an interpretation is dubious. In the Lewisian there are, for example, no datable dykes interspersed with the deformation events like in the West Troms Basement Complex.

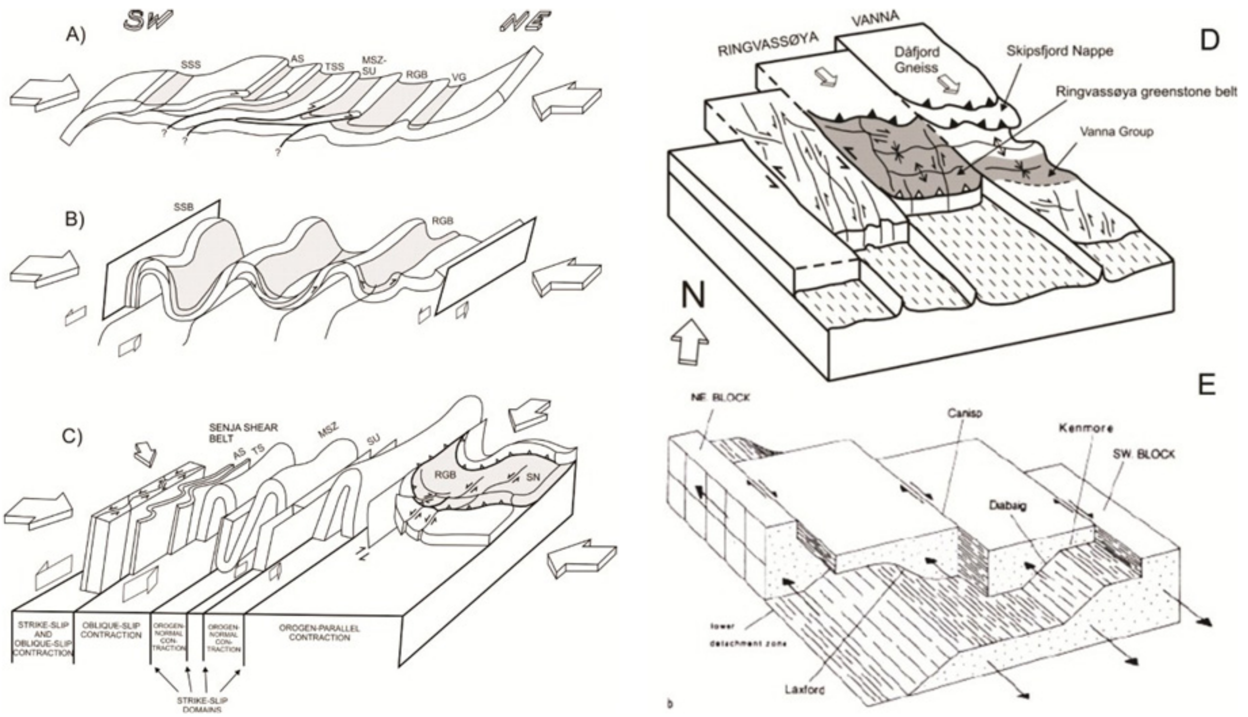

Figure 24. Tectonic model for the Palaeoproterozoic deformation in the West Troms Basement Complex (A-D) [17] compared with the Lewisian (E) [37]. The overall framework is that of NE-SW directed orthogonal shortening and an increasing transpressive component with time. The spatial domains, named in Figure 2, and their kinematic characters are also illustrated. (A) Early-stage formation of NE-directed thrusts and a low-angle main mylonitic foliation in the metasupracrustal belts. (B) Continued orthogonal NE-SW contraction produced upright macro-folds with steep limbs. Note that the main foliation and early thrusts were folded. (C) Late-stage tectonism involved NE-SW orthogonal and/or oblique to orogen-parallel contraction (NW-SE) and mostly sinistral strike-slip reactivation of steep macro-fold limbs, e.g. in the Senja Shear Belt. The eastern, more flat-lying macro-fold hinges (e.g. Ringvassøya greenstone belt) provided the locus for potential low-angle thrust detachments that may have accommodated partitioned NW-SE shortening and SE-directed thrusting. (D) Late-stage Palaeoproterozoic kinematic model for the north-eastern part of the West Troms Basement Complex, where potential low-angle shear zones/detachments accommodated NW-SE directed thrust movements on flats and steep orogen-parallel strike-slip/transfer-type shear zones on ramps [17]. (E) Simplified kinematic model for the Palaeo proterozoic deformation in the Lewisian Complex involving a combination of thrust and strike-slip movements on flats/detachments and ramps/steep transfer-type shear zone [37].

1.70-1.65 Ga deformation and metamorphism

The Lewisian underwent a very distinct set of events at 1.70-1.65 Ga including deformation, the local development of migmatites, the ubiquitous intrusion of pegmatites, and low grade metamorphic overprints reflected in secondary titanite and rutile ages. These events largely post-date similar pegmatite intrusions affecting the West Troms Basment Complex and the Lofoten-Vesterålen province.

The late stages of magmatism and deformation in the Lewisian at 1.70-1.65 Ga involved localized steep ductile reverse (Figure 17) and dextral oblique shear zones developed by strain partitioning likely due to reactivation of steep pre-existing shear zones [77]. This partitioned deformation was interpreted to have been related to major flat-lying detachment zones (Figure 24e) that generated dip-slip thrust movement on flat portions, i.e. the Laxford shear zone, and strike-slip shear zones on steep oblique ramps, i.e. the Canisp shear zone [37]. The 1.70-1.65 Ga stage corresponds to a retrogression of the amphibolite/granulite facies conditions to greenschist facies, indicating exhumation of the rocks from mid to upper crustal levels [79]. This event was associated with the formation of steep-plunging asymmetric folds, retrogressed cleavage, widespread fluid-flow and quartz vein precipitation, and multiple strike-slip shear zones [37].

A similar crustal model is invoked for the late-stage partitioned deformation in the West Troms Basement Complex [17], which, despite lack of critical age dating may correspond to the 1.70-1.67 Ga event in the Lewisian (Figure 24d). In this model, potential flat-lying thrust detachments (shear zones) are present in the northern part of the region.

In summary, the Palaeoproterozoic deformation in the Lewisian Complex and that of the West Troms Basement Complex and Lofoten-Vesterålen, display obvious similarities in terms of tectonic style and partitioning of the deformation, even though they may not be fully temporally correlative (Figure 23). They show: (i) Long-term protracted deformation character, (ii) presence of crustal scale ductile shear zones (iii) partitioned transpressive crustal deformation, e.g. thrusts and orogen-parallel strike-slip shear zones, and (iv) spatial changes in metamorphic grades, i.e. major tectono-metamorphic breaks such as the regional metamorphic boundary in Lofoten and the Laxford shear zone in the Lewisian. These similarities are all consistent with comparable tectonic assembly processes caused by accretion followed by crustal convergence and orogen-oblique/parallel displacement of crustal segments, and a terminal phase of crustal/differential uplift and reactivation [8, 17].

4.3. Linking Palaeoproterozoic terranes and events in the North Atlantic realm

A major problem when trying to restore Precambrian plate tectonics is the nature and processes of assembly of lower crustal blocks or terranes [3, 4, 7, 8]. In terms of the North Atlantic realm Williams et al. (1991)[126] proposed that at the end of the Neoarchean there was a supercontinent, Kenorland, whose breakup led to the formation of several micro-continents which were reassembled together with juvenile terranes in the Palaeoproterozoic (Figure 22). Others, however, have argued for the existence of several micro-continents at the end of the Neoarchaean instead of one single supercontinent [33].

Most workers agree that the Karelia and Superior cratons of Fennoscandia and Laurentia were in close vicinity to each other or connected in the Neoarchaean [24]. The outline of these cratons (Figure 22) is a result of cycles of collision, granitoid intrusion, extension, rifting and basin formation, and if several of these events can be correlated between various cratons, then it is possible to reconstruct former crustal assemblages or supercratons [33]. In particular, timing of large igneous provinces and associated episodes of continental breakup and supracrustal deposits can be used for such analyses, whereby the most detailed record known is that of the Laurentian cratonic fragments [127]. Similar Palaeoproterozoic configurations have also been discussed in the literature, and various models presented [8, 110, 128, 129, 130]. Following breakup of a potential Neoarchaean supercraton, oceanic arcs started to converge from c. 2.0 Ga, with eventually accretion of the cratons along sutures that follow the grain of the Palaeoproterozoic orogens (Figure 22). The model proposed by [110] suggests a rather familiar configuration of the various Fennoscandia/Baltica and Laurentia cratons at the beginning of the Palaeoproterozoic (Figure 22), and despite being a speculative model it addresses the need for more detailed research within these cratons and especially along their margins.

Recent paleomagnetic reconstruction of the Palaeoproterozoic [8, 128] suggests the presence of several large colliding plates, including the North Atlantic and west Greenland plates, the Central Greenland Craton and the Fennoscandian (Baltic-Kola) plate, with the Lewisian somewhere in between (Figure 25). Most workers link the Lewisian to the Palaeoproterozoic Nagssugtoqidian belt in Greenland [5, 129, 131], and consider that this belt may have counterparts both in North America and/or the Fennoscandian Shield [3, 4, 110]. A link between the Lewisian of NW Scotland and the Lappland-Kola and Karelia craton of northern Fennoscandia would then place the West Troms Basement Complex and Lofoten-Vesterålen province exactly along the line of intersection between these major Palaeoproterozoic orogenic belts (Figure 22). A similar reconstruction [129] supports a correlation of Palaeoproterozoic orogens in Greenland and Fennoscandia at the c. 1.8 Ga supercontinent stage.

The scenario proposed by Park (2005) [8] gives a valid plate setting for the end of the Neoarchaean (Figure 25a) and explains the subsequent Palaeoproterozoic tectono-metamorphic events in the Lewisian Complex and tentatively, also the deformation events in the West Troms Basement Complex and Lofoten-Vesterålen province. At ca. 1.9-1.87 Ga, volcanic arcs were created between North American craton and Central Greenland craton/Kola craton due to the subduction/ accretion of the oceanic crust located between them (Figure 25b). The calk-alkaline plutonic intrusions within the Loch Maree Group, the South Harris Igneous complex, and potentially, the earliest phases of magma intrusions in the Lofoten igneous province and West Troms Basement Complex, and accompanied convergent deformation and granulite facies metamorphism manifest this regional accretionary event [8]. At ca. 1.87 Ga, the Central Greenland craton and Kola-Karelian craton collided and was under-thrusted beneath the the North American craton in a NW-SE direction within the Lapland-Kola belt, and resulting in the main phase of deformation (Figure 25c). Granulite facies metamorphism occurred in the down-going slab due to under-thrusting (lower plate). The line of collision between juvenile terranes was likely oriented in the same direction as the Palaeoproterozoic Nagssugtoqidian belt and the orientation of the collision could be given by the orientation of the main NW-SE trend of the Laxford shear zone [8]. At ca. 1.8 Ga, subduction of oceanic crust to the SW of this new continent may have created a volcanic arc trending NW-SE (Figure 25d), and this arc may have been involved with renewed collision at ca. 1.75 Ga (Figure 25d, e), corresponding to the main stages of deformation in the West Troms Basement Complex. There, the intrusion of the calk-alkaline Hamn norite (1.8 Ga) and the Ersfjord granite (1.79 Ga) may have been related to this phase. Similar calk-alkaline intrusive rocks exist further south of the

Lewisian complex [132, 133], known as the 'Malin block' and this block was thought to be part of a belt comprising the Labradorian of NE Canada, the Ketilidian of South Greenland and the Gothian of Scandinavia (see Figure 22). This belt became tectono-magmatic active at *c.* 1.8 Ga and an event at c. 1.7 Ga could be the result of igneous activity and deformation related both to the latest Palaeoproterozoic events in the Lewisian and in the West Troms Basement Complex.

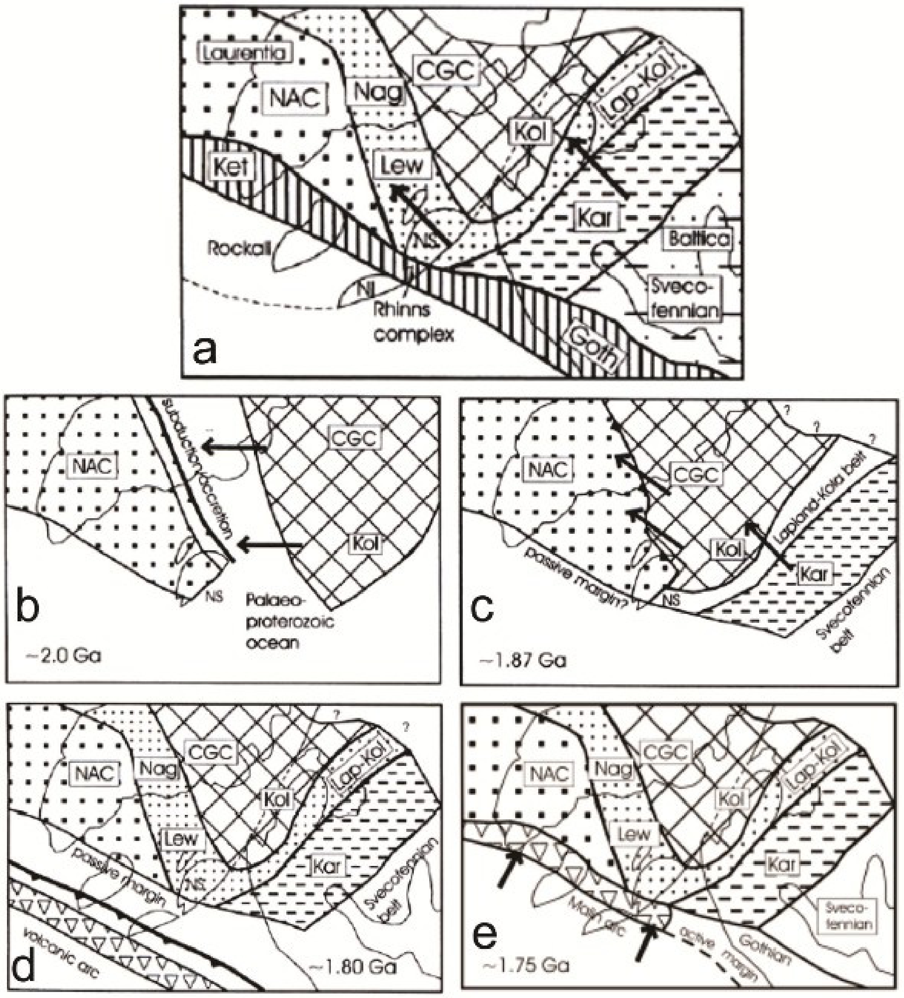

Figure 25. Plate tectonic setting of the Lewisian complex during the Palaeoproterozoic, based on the North Atlantic reconstruction [8, 128]. Abbreviations: CGC=Central Greenland craton, Goth = Gothian belt, Kar = Karelia craton, Ket = Ketilidian belt, Kola= Kola craton, Lap-Kol = Lapland-Kola belt, Lew = Lewisian, NAC = North Atlantic craton, Nag = Nagssugtoqidian belt, NI = north Ireland, NS = north Scotland. (a) Distribution of cratons and orogenic belts during the Mesoproterozoic. (b) 2.0 Ga: Subduction and creation of a volcanic arc in oceanic crust between two continental plates (NAC and CGC/Kol) followed by accretion of oceanic/arc elements along the leading edge of the NAC. (c)1.87 Ga: Collision of the two continents followed by underthrusting of the CGC/Kola craton beneath the NAC, causing the early Palaeoproterozoic deformation and metamorphism. At the same time, collision occurs in the Lapland/Kola belt to the SE caused by collision with the Karelia craton. Note the NW-SE movement direction. (d) 1.80 Ga: Development of a volcanic arc in oceanic crust SW of the amalgamated continent created in b. (e) 1.75 Ga: Collision between the "Malin block" and the continent, causing late-Palaeoproterozoic deformation, metamorphism and granitic melt formation in the Lewisian complex.

5. Conclusions

(1) The West Troms Basement Complex is underlain by Archaean gneisses (2.92-2.56 Ga), metasupracrustal rocks (2.4-1.9 Ga) and mafic dyke swarms (2.4 and 2.2 Ga) that were variably reworked, metamorphosed and intruded by felsic and mafic plutons at c.1.8 Ga. Along strike to the southwest, Neoarchaean high-grade gneisses in the Lofoten-Vesterålen province display magmatic protolith ages of between 2.85 and 2.7 Ga and record a high-grade metamorphic event at c. 2.64 Ga. The Neoarchaean basement rocks have been intruded by a huge 1.8 Ga magmatic suite composed of anorthosites, mangerites, charnockites, gabbros and granites, which corresponds in age with the 1.81–1.77 Ga, NW-trending granitoids in the older part of the Transscandinavian igneous belt of southern Sweden. A similar present structural position of the basement high in the Lofoten-Vesterålen province and the West Troms Basement Complex invokes they are along-strike correlatives.

(2) The Lewisian rocks of NW Scotland comprise a series of Neoarchaean blocks thought to have been amalgamated during a multistage and complex set of Palaeoproterozoic collision events between 1.97 and 1.67 Ga, producing a variety of block-bounding accretional and intrablock shear zones. This province also records Neoarchaean crustal deformation and metamorphism at intervals of 2.7-2.6 Ga and 2.49-2.40 Ga followed by episodes of crustal rifting and mafic dyke intrusion (2.4 and 2.0 Ga), deposition of continental margin-like metasedimentary sequences between 2.0 and 1.9 Ga ago upon the substratum of Neoarchaean gneisses and later on subjected to major orogenic deformation and metamorphism (c. 1.85 and 1.70 Ga).

(3) The West Troms Basement Complex, the Lofoten-Vesterålen province and the Lewisian rocks of Scotland are thus very similar crustal regions in terms of lithology, age, igneous, structural and metamorphic features and tend to share a similar tectono-magmatic and evolutionary history, but there are also sharp differences such as the lack of 1.80 magmatism in the Lewisian and the c.100 m.y. difference in the timing of the latest Palaeoproterozoic deformation overprints in the two regions.

(4) Reconstructing Palaeoproterozoic plate scenarios is a difficult task. Nevertheless, paleomagnetic restorations suggest the presence of several large colliding plates, including the North Atlantic and western Greenland plates, the Central Greenland craton and the Fennoscandian (Baltic-Kola) Shield, with the Lewisian somewhere in between. In this context, the Lewisian has been temporally linked to the Palaeoproterozoic Nagssugtoqidian belt in Greenland and may have its counterpart in North America and/or the Fennoscandian Shield. A link between the Lewisian of NW Scotland and the Lappland-Kola and Karelia craton of northern Fennoscandia would locate the West Troms Basement Complex and Lofoten-Vesterålen province directly along the line of intersection between these major Palaeoproterozoic orogenic belts at the c. 1.8 Ga supercontinent stage.

(5) Tentative similar Palaeoproterozoic terrane models (1.80-1.67 Ga) can be invoked for the basement outliers in northern Norway and the Lewisian Complex. The continental assembly may have involved either multiple small terranes or crustal rejuvenation of one or two large

terranes. The latter model is based on component similarities and metamorphic variations and can be explained by the presence of at least two different crustal blocks and/or depth portions assembled along crustal scale ductile shear zones. The juxtaposition included arc-magmatism and accretion of Neoarchaean continental terranes in the vicinity of the Fennoscandia-Laurentia border, followed by uplift and reworking.

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