Arc-generated blocks with crustal sections in the North Atlantic craton of West Greenland: Crustal growth in the Archean with modern analogues

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ABSTRACT

The ca. 700 km long, Archean craton of West Greenland consists of six Meso–Neoarchean (ca. 3000–2720 Ma) shear zone — bounded crustal blocks that display similar cross-sections; from south to north Ivittuut, Kvanefjord, Bjørnesund, Sermilik, Fiskefjord, Maniitsoq. Each block has a southerly upper and a northerly lower zone, thus each faces upwards to the south. Upper zones have prograde amphibolite facies mineralogy and have never been in the granulite facies, whereas lower zones reached granulite facies and were partly retrogressed to amphibolite facies. Upper and lower zones consist predominantly of tonalite–trondhjemite–granodiorite (TTG) orthogneisses; geochemistry suggests generation by slab melting in subduction settings of island arcs and active continental margins. The gneisses contain km-thick metavolcanic amphibolite layers typically bordered by km-thick layers containing anorthosite and leucogabbro. Most upper zones contain upper greenschist to amphibolite facies metavolcanic belts including volcaniclastic, andesitic rocks. The two most-prominent metavolcanic belts in the Fiskefjord block at Qussuk (andesitic–volcaniclastic rocks; Garde, A.A., 2007. A mid-Archaean island arc complex in the eastern Akia terrane, Godthåbsfjord, southern West Greenland. Journal of the Geological Society (London) 164, 565–579.) and Ivisaartoq (mafic–ultramafic rocks and anorthosite–leucogabbr from upper and lower parts of a supra-subduction zone system; Polat, A., Frei, R., Appel, P.W.U., Dilek, Y., Fryer, B., Ordóñez-Calderón, J.C., Yang, Z., 2008. The origin and compositions of Mesoarchean oceanic crust: evidence from the 3075 Ma Ivisaartoq greenstone belt, SW Greenland. Lithos 100, 293–321.) have island arc geochemical signatures. The 2 km-thick Fiskenæsset complex (Bjørnesund block) comprises chromite-layered anorthosites, leucogabbros and gabbros, and local pillow-bearing roof pendants from overlying metavolcanic amphibolite. The style of deformation changes downwards within crustal blocks; upper zones are characterised by linear metavolcanic belts deformed by mostly one major phase of isoclinal folding, and lower zones by kilometre-scale double-triple fold interference patterns. Everywhere TTG protoliths have intruded anorthositic and volcanic rocks typically along ductile shear zones, often so extensively that only anorthositic or amphibolitic lenses are preserved. The Meso–Neoarchean crust was thickened by a combination of thrusting, isoclinal folding and continued TTG injection. Dissimilarities in the proportions of anorthositic and metavolcanic rocks in the six blocks suggest that they evolved in several different microcontinents but by similar processes. These crustal blocks provide an exceptional example of how continents evolved in the Meso–Neoarchean. Comparable Archean examples in Kapuskasing and Pikwitonei (Canada) and modern analogues in Fiordland (New Zealand), Kohistan (Himalayas), Southern California batholith, Peruvian Andes, and Hidaka (Japan) demonstrate that processes of continental growth from island arc to continental arc magmatism (and by implication to continental collision) were broadly similar throughout most of Earth history.

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1. Introduction

The mode of growth of the continental crust is a key constraint for understanding the secular evolution of the continents with time (Dewey and Windley, 1981; Taylor and McLennan, 1985; Hawkesworth and Kemp, 2006). The vertical dimension within a 4D perspective provides us with key information on the structure, petrology and evolution of the continental crust that developed in orogenic belts at any period of time (Hatcher et al., 2007). Most information on the lower crust, its relations with the upper crust, and on crustal sections has come from studies of xenoliths (e.g. Rudnick, 1992; Weber et al., 2002) and gravity and seismic data (e.g. Allmendinger et al., 1987; Meissner et al., 2006). Similar studies have demonstrated that the lower continental crust may have a mafic composition (Condie, 1999), that major tectonic boundaries are commonly marked by geophysical reflectors (Snyder et al., 1996), and that structural, metamorphic (e.g. the amphibolite–granulite transition), lithological or isotopic boundaries, observed in geological sections, coincide with gravity or seismic anomalies at depth (Fountain and Salisbury, 1981).

Vertical geological cross-sections through the upper to lower crust enable a continuous, in situ examination of rocks and relationships, but well-exposed continuous cross-sections of any age are rare (Fountain and Salisbury, 1981; Salisbury and Fountain, 1990; Percival et al., 1992, 1997; Percival, 2007). Notable Neoarchean examples, relevant to the present study, are the Kapuskasing Uplift (Percival and West, 1994) and the Pikwitonei section (Ermanovics and Davison, 1976; Fountain and Salisbury, 1981) in the Canadian Shield, and the lesser known Meso–Neoarchean section across the Fermor Line in southern India (Condie et al., 1982; Newton and Hansen, 1986).

There is a long-standing problem with regard to the application of the plate tectonic model in Earth Sciences. Whilst it has been possible to demonstrate that the Meso–Cenozoic plate tectonic model can be successfully applied to a myriad of Paleozoic and Proterozoic orogenic belts, and to a less extent many Archean upper crustal greenstone belts especially from geochemical and isotopic data (Kerrich and Polat, 2006), it has proved very difficult to understand Archean high-grade orogenic belts, partly because of the paucity of deep crustal sections of young orogenic belts to use as modern analogues, partly because the predominance of granulite and high-amphibolite facies gneisses has left a lack of relationships that may be compared with most modern plate tectonic environments and geometry, and because of the overall tectonic conformity of most rock units, as presciently observed by Ramberg (1952) from his study of the craton of West Greenland. However, there are relationships in this craton that permit comparison with modern equivalents, as we demonstrate here. Accordingly it is useful to consider some Mesozoic–Cenozoic analogues, the structural and geochemical development of which can be related to specific plate tectonic settings, and accordingly may provide useful constraints to help understand Archean crustal evolution.

Although vertical crustal sections, of any age, provide a unique opportunity to unravel the development of the crust in space and time, there is still a poor understanding in any detail of the evolutionary relationships between the upper and lower continental crust. To improve this imbalance, we aim in this paper to present a new analysis and understanding of the Meso–Neoarchean crust in West Greenland in terms of a subdivision into several, remarkably well-exposed, crustal blocks, each of which provides a section from the upper to deep crust. We use the time scale of Gradstein et al. (2005): Eoarchean >3600 Ma; Paleoarchean 3600–3200 Ma; Mesoproterozoic 3200–2800 Ma; Neoarchean 2800–2500 Ma.

2. The Archean craton of West Greenland

The North Atlantic craton of southern West Greenland (Fig. 1) largely consists of Mesoarchean (ca. 3075–2820 Ma) orthogneisses with tonalite–trondhjemite–granodiorite (TTG) compositions (Steenfelt et al., 2005) that contain many conformable layers of metavolcanic amphibolite and anorhotosite (Windley, 1969), both of which are up to ca. 2 km thick, and rare metasedimentary rocks; these rocks occur in alternating granulite and amphibolite facies belts (Bridgewater et al., 1976; Owens and Dynek, 1997). The Godthåbsfjord region in the northern part of the craton is noted for Eoarchean rocks (ca. 3900–3600 Ma) of the Itsaq gneiss complex that occur in four terranes, namely Færingehavn, Isukasia, Qarliit Tasersuat and Aasivik (Nutman et al., 2004). In the same region Friend and Nutman (1991) and Nutman et al. (1989, 1993) also defined three Mesoarchean gneiss terranes (Fig. 1) (Tre Brødre, Tasiarsuaq and Akia) on the basis of their different metamorphic, structural and isotopic histories, and Friend and Nutman (2005) added the Kapilsik terrane. Kalsbeek and Garde (1990) provided a description of the 1:500,000 geological map.
from Frederikshåb Isblink to Søndre Strømfjord (Fig. 1), which covers four of the crustal blocks described in this paper.

However, besides local terranes, the craton is constructed of six major crustal blocks, five of which display remarkably similar cross-sections through the Meso–Neoarchean crust, because each one consists of an upper prograde amphibolite facies zone in the south and a lower prograde granulite facies zone in the north; thus each block faces upwards to the south. Metamorphic facies transitions are generally smooth within each block, but abrupt and of tectonic nature (transposed thrusts or shear zones) where block boundaries are exposed. The blocks will be described in turn northwards from the south (Fig. 1). Fig. 2 shows the location of subsequent main maps and figures.

2.1. The Ivittuut block

This block extends from the basal Ketilidian unconformity northwards to the southern side of the metavolcanic Târtoq Group, see Figs. 1 and 3 (Berthelsen and Henriksen, 1975). Only the lower part of this block is exposed, because its upper part extends southeastwards under the Ketilidian unconformity. At the northern end of the Ivittuut block we consider that the boundary against the Kvanefjord block is situated on the southern side of the steeply N-dipping imbricated thrust slices of Târtoq Group rocks that form the top of the Kvanefjord block (following the detailed structural work of Berthelsen and Henriksen, 1975; see Fig. 3).

The Lower Zone consists of granodioritic orthogneisses from which Nutman et al. (2004) obtained SHRIMP U–Pb zircon ages up to 3000 Ma, layers of amphibolite up to ca. 0.5 km thick, and orthogneisses layers up to ca. 2 km thick that contain abundant inclusions of anorthosite, leucogabbro and gabbro. Experience in the Fiskenæsset region farther north (Myers, 1985) leads to the conclusion that the layered anorthositic complex in the Ivittuut block was intruded by sheets of granodioritic protoliths of the gneisses, the present distribution of the inclusions preserving the minimum thickness of the original layered complex. The gneisses, amphibolites and the anorthositic complex were deformed by an early phase of isoclinal folds that were refolded by isoclinal recumbent nappes (Berthelsen and Henriksen, 1975). All the rocks in the Ivittuut block were metamorphosed under high amphibolite facies conditions, and there is no evidence of any earlier granulite grade of metamorphism, even in basic or ultramafic rocks that commonly retain evidence of such mineralogy, when all the quartzo-feldspathic gneisses have been thoroughly retrogressed. This is corroborated by high but northward-decreasing K2O contents in stream sediments (Fig. 4; Steenfelt, 1994). Thus the exposed section of the Ivittuut block is not a retrogressed equivalent of a granulite-grade crustal zone like the lower zone of the other tectonic blocks to the north. In terms of the two zones being described in the tectonic blocks of this paper, the Ivittuut rocks belong to a prograde amphibolite facies lower zone.

2.2. The Kvanefjord block

The 130 km-long Kvanefjord block (Fig. 1) extending from the southern side of the Târtoq Group at Sermiligaarsuk fjord northwards to Frederikshåb Isblink was first mapped in the 1960s, and it was assumed at that time that it had a more or less uniform geological history (Berthelsen and Henriksen, 1975; Higgins, 1990; Kalsbeek et al., 1989). During a later boat reconnaissance McGregor and Friend (1997) distinguished three individual areas with different tectonothermal terranes, separated by folded tectonic boundaries (the Neria, Paamiut and Sioraq blocks in the terminology of the authors). The first of these, around the mouth of Neria fjord in the southwest, exposes orthogneisses that have been retrogressed from granulite to amphibolite facies. The second Paamiut area in the central and southeastern parts of the region consists of prograde amphibolite facies rocks. Thirdly, the northernmost Sioraq belt comprises granulite facies and variably retrogressed rocks. Friend and Nutman (2001) obtained protolith ages of ca. 2940–2900 Ma from the retrogressed rocks around outer Neria, and mutually overlapping, younger ages of ca. 2870–2850 and 2870–2830 Ma from the Paamiut and Sioraq areas, respectively.

Friend and Nutman (2001) interpreted the three components as different tectonothermal terranes, separated by folded tectonic boundaries that post-date the granulite facies metamorphism. However, we view the region as one contiguous crustal segment, the Kvanefjord block, of prograde amphibolite to granulite facies rocks overlain by a major nappe of thoroughly retrogressed rocks. The bulk of the Kvanefjord block thus comprises the ca. 2830 Ma, amphibolite facies Paamiut area in the south and the granulite facies Sioraq area of...
the same age in the north; McGregor and Friend (1997) showed that the boundary between the two areas is the locus of a steep mylonite zone, however, we suggest that the two ca. 2850 Ma areas have a common protolith. The central component of older, 2940–2900 Ma retrogressed rocks around Neria fjord (the Neria block of McGregor and Friend, 1997) structurally overlies the younger rocks in the Paamiut area and does not extend all the way to the icecap. Aided by detailed structural information from the original 1:100,000 scale Survey maps we re-interpret this unit as a major, refolded nappe, the Neria nappe (Fig. 5). The maps show kilometre-scale recumbent, isoclinal folds in its proposed south-facing hinge zone, and the outcrops of its folded sole coincide with mapped high-strain and

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**Fig. 3.** A. Geological map of the Ivittuut block emphasising the main structural relations (simplified from Berthelsen and Henriksen, 1975). B. The 3D interpretation of the triple fold interference pattern by Berthelsen and Henriksen (1975). For location see Figs. 1 and 2.
thrust zones. We tentatively include the synclinal area of retrogressed rocks east of Nerutusoq in the nappe structure as shown in Fig. 5, although no age data exist from this area. Our structural interpretation with the lateral exposure of a major nappe over 50 or 75 km implies that the Kvanefjord block is not steeply tilted. Accordingly the amphibolite facies Paamiut area is very wide and comprises both upper and lower zone rocks (see below).

The Upper Zone, extending from the southern side of the imbricated, steeply N-dipping Târtoq Group to ca. 25 km northeast of Sermiligaarsuk consists of prograde, amphibolite facies hornblende–biotite–epidote gneisses with accessory titanite (McGregor and Friend, 1997) and greenschist–to amphibolite–grade supracrustal rocks, and is further characterised by high K$_2$O in stream sediments (Fig. 4; Steenfelt, 1994), and absence of anorthosite. The orthogneisses contain...

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**Fig. 4.** Two maps of the entire craton of West Greenland showing the distribution of K$_2$O and CaO in stream sediments in relation to the Upper and Lower Zones of the six crustal blocks, which are marked. Note that K$_2$O is highest in the Upper Zones and upper parts of the Lower Zones where granitic rocks are most common, and is lowest in the deep granulite facies parts of the Lower Zones. CaO is highest in the granulite facies Lower Zones and where anorthositic rocks are prominent, and tends to be lower in the Upper Zones. Exceptions are explicable. Modified after Steenfelt (1994).
circular granitic bodies up to ca. 2.5 km across, and a semi-conformable foliated biotite granite 8 by 25 km in outcrop size.

In a 75 km-long belt at the top of this crustal zone the gneisses are intercalated with tectonic layers of the metavolcanic Târtoq Group (Allaart, 1975). The largest 5 km-wide layer is deformed into a major isoclinal, NE-trending synform and contains an original 2 km-thick stratigraphy with a lower mixed sedimentary-volcanic unit and an upper mainly volcanic unit (Berthelsen and Henriksen, 1975). Prominent are: pillow-bearing greenschist up to 100 m thick, actinolitic schist, chlorite–carbonate schist, ferroan dolomite, quartzite up to 60 m thick, quartzitic schist, fuchsite–ankerite schist, quartz–feldspar schist, 30 m-thick serpentinite sills, talc–chlorite schist (Higgins, 1968; Berthelsen and Henriksen, 1975), quartz–magnetite and sulphide facies banded iron formation with an original thickness of ca. 5–10 m, a conglomerate and minor pyroclastic rocks (Appel, 1984). The Târtoq Group also contains shear-hosted gold mineralization (Evans and King, 1993). The best-preserved rocks, in the core of the fold, were metamorphosed in the mid- to upper greenschist facies (Higgins, 1968; Evans and King, 1993), and the rocks increase in grade to amphibolite facies towards the contacts of the synform with adjacent gneisses (Berthelsen and Henriksen, 1975). The boundary between the Kvanefjord block to the north and the Ivittuut block to the south is on the southern side of the main Târtoq layer where four or five volcanic and sedimentary layers are intercalated with an equal number of layers of biotite–chlorite gneiss in a 3 km-wide north-dipping, imbricated thrust zone that is seen in Fig. 3 (Berthelsen and Henriksen, 1975). A U–Pb zircon probe date of 2842±6 Ma from the youngest detrital grains from a Târtoq sediment was interpreted by Nutman et al. (2004) as the time of volcanic-sedimentary deposition. However, a tonalite sheet that crosses a metavolcanic amphibolite has a U/Pb zircon age of 2944±7 Ma (Nutman and Kalsbeek, 1994), implying that part of the Târtoq Group may be still older.

Fig. 5. A. Map of the Kvanefjord crustal block (see Fig. 1) showing how the Neria belt of older (2940–2900 Ma), amphibolite facies rocks thoroughly retrogressed from the granulite facies is interpreted as a sub-horizontal nappe. In the central and southern parts of the block the nappe is resting on (and refolded with) younger, 2870–2850 Ma prograde amphibolite facies rocks in the Lower and Upper Zone. It is uncertain if the nappe was rooted beyond the likewise younger (2870–2830 Ma), un-retrogressed granulite facies rocks in the northern part of the Lower Zone as tentatively shown on the section (B).
The Lower Zone, extending from northeast of Sermiligaaĸ to Frederikshåb Isblink, consists of three different crustal components, namely the northern, relatively low-K₂O part of the prograde, amphibolite facies rocks in the Paamitut area, the granulite facies and partly retrogressed rocks in the Sioraq area, and the Neria nappe of thoroughly retrogressed rocks. K₂O in stream sediments decreases northwards (see Fig. 4, Steenfelt, 1994) in agreement with this division. The northern boundary of the Lower Zone has a steep southeasterly dip but is largely obscured by the glacier of Frederikshåb Isblink (see under Bjørnesund block).

A. The southern part of the Paamitut area consists of amphibolite facies rocks that have never been in the granulite facies; most prominent are tonalitic and granodioritic gneisses within which are layers of amphibolite and aluminous mica schist both up to 4 km wide, but there are no anorthositic rocks (Escher and Pulvertaft, 1995). Although deep in this lower zone a 1 km-wide amphibolite layer contains relict pillows and agglomerates (Walton, 1966). Gneisses have zircon \( ^{207}\text{Pb} / ^{206}\text{Pb} \) protolith ages of 2874±10 Ma, 2872±2 and 2841±40 Ma (Friend and Nutman, 2001).

B. The Sioraq area from Nerutsoq fjord northwards to the Isblink comprises rocks that were metamorphosed under granulite facies conditions and only partly retrogressed to amphibolite facies (Escher and Pulvertaft, 1995, based on observations by A. Steenfelt). They consist of hypersthene- and garnet-bearing tonalitic, granodioritic and dioritic orthogneisses low in K₂O that contain layers of amphibolite up to 3 km wide and one anorthosite layer that were all deformed by at least two phases of isoclinal folds which form major interference patterns. The orthogneisses have zircon \( ^{207}\text{Pb} / ^{206}\text{Pb} \) protolith ages of ca. 2860–2835 Ma, and a metamorphic age of ca. 2820 Ma (Friend and Nutman, 2001).

C. The Neria nappe consists of hornblende-bearing tonalitic to dioritic orthogneisses with layers of amphibolite up to 2–3 km wide, and at least five layers of anorthosite up to 1 km wide and 10 km long. Although no orthopyroxene is preserved, diagnostic textural evidence suggests that the rocks have been thoroughly retrogressed from the granulite facies (McGregor and Friend, 1997). Orthogneisses have zircon \( ^{207}\text{Pb} / ^{206}\text{Pb} \) protolith ages of 2939±11 Ma, 2927±9 Ma, and 2900±8 Ma (Friend and Nutman, 2001).

2.3. The Bjørnesund block

This tectonic block (ca. 78 km N–S) extends from Frederikshåb Isblink northwards to Grædefjord (Fig. 1 and 1:500,000 scale map of Allaaar, 1982). As mentioned above the glacier of Frederikshåb Isblink obscures the tectonic relations between the low amphibolite facies upper zone of this block and the granulite facies gneisses of the lower zone of the Kvanefjord block to the south, however, the tectonic structure of nunataks northeast of the glacier suggest a steeply SE-dipping boundary. From the major difference in the K₂O % and CaO % of stream sediments between the southerly granulites and the northerly gneisses and their granites shown in Fig. 4, Steenfelt (1994) suggested that the Isblink marks a major ‘terrain boundary’ (see also, Nutman et al., 2004).

The Upper Zone of this block, about 50 km wide, extends from Frederikshåb Isblink to Bjørnesund fjord. It comprises prograde, amphibolite facies, tonalitic and granodioritic biotite–hornblende orthogneisses, which have never been in the granulite facies, but reached garnet–kyanite–sillimanite grade (Thomas, 1973; Hopgood, 1973, Williams, 1984). The gneisses contain layers of metavolcanic amphibolite up to 2 km thick in places bordered by layers of anorthosite–leucogabbro up to 1 km thick folded with the amphibolites into refolded isoclines. Kalsbeek and Myers (1973) found that sand samples from this Bjørnesund block reflect the metamorphic grade of the basement rocks. Within this upper zone they mostly contain no hypersthene. Kalsbeek (1976a) confirmed that the hornblende–biotite gneisses in the upper zone are also characterised by primary porphyroblasts of epidote, muscovite, and accessory titanite, and concluded that the rocks had not been in the granulite facies, but were formed by progressive, low amphibolite facies metamorphism. Kalsbeek (1974) reported that the U and U/K values of the gneisses from this upper zone are rather low, but excluded the possibility that granulite facies metamorphism had taken place, based on field observations.

Towards the top of the upper zone the gneisses contain a conformable, linear NE-trending, synclinal metavolcanic belt up to 5 km wide, and ca. 30 km-long in Ikkattup fjord (Escher, 1976); these rocks have only undergone one major isoclinal phase of folding. The belt, referred to here as the Ikkattup Nunaa belt from an island in the middle of the belt, contains a well-preserved stratigraphy with basaltic pillow lavas (Fig. 6F), pillow breccias, tuffs and ash beds with gabbronorite texture (radiating hornblende crystals on foliation surfaces in mafic beds), and an intrusive sill, 1.5 km thick, of rhythmically layered gabbrors and leucogabbros. The rocks were metamorphosed between the greenschist–amphibolite facies transition (Andersen and Friend, 1973; ‘Ravn Store amphibolite belt’) and high amphibolite facies, and are marked by metavolcanic leucocratic schists containing anthophylite, cummingtonite, staurolite, cordierite and garnet (Friend, 1976a). A swarm of still-discordant, amphibolite dykes cuts foliation and folds of the metavolcanic amphibolites, and predates emplacement of the precursors of the regional orthogneisses (Friend, 1976b). SHRIMP zircon dates indicate the Ikkattup Nunaa volcanic and sedimentary rocks have a deposition age of 2908±13 Ma (Nutman et al., 2004). The northern boundary of the Ikkattup Nunaa metavolcanic belt is marked by a 35 m-wide intercalation of volcanic and gneissic sheets, in which the intrusive gneiss sheets decrease progressively in width towards the volcanic belt (Fig. 3 of Windley et al., 1966).

At about 28 km northwest of the top of the Bjørnesund block and along strike of the Ikkattup Nunaa metavolcanic belt biotite gneisses contain layers and lenses of marble, calc-silicate gneiss and anorthosite (Hopgood, 1973; Bollingberg et al., 1976); the trace element composition of the marbles is similar to that of calcareous sediments.

The gneisses in this zone also contain many conformable to discordant layers up to 3–4 km thick of partly foliated, granitic rocks (tonalites, granodiorites, adamellites and muscovite-garnet granites). A discordant tonalitic gneiss sheet has a SHRIMP U–Pb age of 2878±10 Ma (Friend and Nutman, 2001). Zircons from a muscovite granite sheet that intrudes the Ikkattup Nunaa volcanic belt indicate an age of intrusion at 2660±20 Ma. Because the granite exhibits the low amphibolite facies mineralogy of the surrounding area, Pidgeon and Kalsbeek (1978) concluded that the 2660±20 Ma age records the end of the amphibolite facies metamorphism. On nunataks to the east undated, undeformed two-mica granodiorite–adamellite plutons (up to 10 by 15 km across) are locally discordant to the gneiss foliation and some have rapakivi textures (Dawes, 1970). The last deformation to have affected these upper zone gneisses is represented by major thrusts (associated with pseudotachylyte veins) that trend NE–SW subparallel to the general foliation and dip shallowly to the SE, indicating overthrusting to the northwest (Hopgood, 1973). It seems to us that this thrusting was most likely associated with the final northwards accretion of the Kvanefjord block to the Bjørnesund block farther north.

On the eastern side of Bjørnesund fjord a 1.5 km-wide layer of amphibolite is intercalated and isoclinal folded with a ca. 200 m-wide layer of chrome-layered anorthosite to form a ca. 36 km-long synformal, NE–SW, linear belt (Fig. 1). The fact that the anorthosite is chromite-layered and that saphirine-bearing rocks occur along the top amphibolite–anorthosite contact (Williams, 1984) demonstrates that this is a layer within the Upper Zone of this crustal block of the Fiskeaasset complex, which otherwise dominates the Lower Zone (see below); this is an important relationship because it demonstrates that the same distinctive and diagnostic rocks and relationships occur
in both the Upper and Lower zones of a crustal block. In the lowest part of this Upper Zone a 2 km-wide amphibolite layer (near the ice cap) contains well-preserved pillows and pyroclastic structures, and almost along strike a 1.5 km wide amphibolite layer (west side of the mouth of Bjørnesund fjord, Fig. 1) contains recognisable volcanic pillows; this is near an anorthositic layer containing a leucogabbro that has a distinctive cumulate texture (Fig. 6D), which is characteristic of leucogabbros throughout the Fiskenæsset complex (Myers, 1985), and of many other leucogabbros in this craton of West Greenland (e.g. Chadwick et al., 1982, in the Sermilik block).
The boundary between the Upper and Lower zones is marked by a prograde (northwards) amphibolite–granulite facies transition in which orthopyroxene grew at the expense of hornblende (Friend et al., 1988a; Friend, 1989).

The Lower Zone of this block consists of relict areas of granulite facies gneiss up to several tens of kilometres across within retrogressed amphibolite facies gneisses (Escher and Pulvertaft, 1995). The largest area of un-retrogressed granulite facies rocks is near the base of this block (Fig. 1), where the gneisses contain diagnostic hypersthene, diopside, and brownish green to green hornblende (Kalsbeek, 1976a). This Lower Zone has a distinctive tectono-stratigraphy of predominant tonalitic orthogneisses that contain layers up to 2 km thick of metavolcanic amphibolite (LREE-depleted, Weaver et al., 1982) commonly bordered by layers of the Fiskenæsset complex (Figs. 6A–C and 7A; Windley et al., 1973) that reaches 2 km thick and retains a primary igneous layered stratigraphy (upwards) of lower LREE-depleted.

Fig. 7. A. Simplified geological map of the western Lower Zone of the Bjørnesund block showing the stratigraphy of orthogneisses, metavolcanic amphibolites and the Fiskenæsset complex. Deformation of these rocks in the deep crust has produced a triple fold interference pattern, the first synclinal phase of which is indicated only by repetition in reverse order of stratigraphy of the Fiskenæsset complex, shown in the inset (modified from Windley et al., 1973; Myers, 1985). The position of Fig. 6A shown by rectangle. B. Photograph looking northeast in the eastern Bjørnesund block showing a recumbent isocline outlined by an inner dark amphibolite layer and an outer envelope of amphibolite facies gneiss. The hinge of the fold is about 3 km long. For location see Fig. 2.
gabbro, ultramafic rocks, lower leucogabbro, middle gabbro, upper leucogabbro, anorthosite containing chromitite seams — Fig. 6B (up to 20 m-thick, Chisler, 1970), and upper gabbro (Myers, 1985). Anorthosites are calcic — Ar89–96 (Windley and Smith, 1974), and some cumulate plagioclases retain cores of igneous plagioclase (Myers, 1985), like those shown in Fig. 6D from the Upper Zone of this block. Some gabbro and leucogabbro layers are well graded — Fig. 6E (Myers, 1985), and strong currents in the magma chamber gave rise to channel deposits in peridotites, gabbros and chromitites (Myers, 1976a). The chemistry of main silicate phases (including hornblende) through the complex varies with respect to stratigraphy and fractionation (Windley and Smith, 1974; Myers and Platt, 1977; Steele et al., 1977; Bishop et al., 1980). Platinum, palladium and rhodium are relatively high in anorthosites and leucogabbros, but concentrated in ultramafics rocks and chromitites, and show trends with stratigraphic height related to igneous differentiation (Page et al., 1980). Some layers of the Fiskenæsset complex are thinned along strike to less than 1 m-thick, and yet still retain the original igneous stratigraphy with 1-cm thick chromitites near the top (Windley et al., 1973). Where such layers are further thinned, they may continue along strike as layers of gneiss that contain numerous inclusions of the complex; such relations demonstrate how the protoliths of the TTG gneisses were emplaced as sheet-like intrusions into the Fiskenæsset complex (Myers, 1985). The Fiskenæsset complex has a five-point Sm–Nd isochron age of 2860±50 Ma, which Ashwal et al. (1989) regarded as the time of crystallization of the complex. The initial epsilonNd of +2.9±0.4 suggests generation from a depleted mantle source, and the age indicates a prevention of plagioclase crystallisation by lowering of the temperature of formation of plagioclase more than that of olivine and pyroxene enabling it to concentrate as late anorthosite, concomitant increase in the anorthite content of plagioclase allowing crystallization of very calcic plagioclase, increase in the stability of amphibole at the expense of pyroxene, and holding of chromium in the melt to enable the precipitation of very late chromitite.

Abundant sapphireine-bearing rocks with numerous metamorphic phases including regional-scale boron-bearing kormerpune (Ackermand et al., 1984; Grew et al., 1987), clinotone (Ackermand et al., 1986), högbomite (Ackermand et al., 1983), ruby corundum, paragisite, phlogopite and spinel occur along the top metasomatised contact of the Fiskenæsset complex with overlying amphibolite (Herd et al., 1969), and regional geikielite dunite lenses occur along the lower contact of the complex with underlying amphibolite (Windley et al., 1989). Oxygen and hydrogen isotopes indicate that the protoliths of the sapphireine-bearing rocks underwent hydrothermal alteration by seawater when the underlying anorthosite was emplaced in shallow crust and before recrystallization by granulite and amphibolite facies metamorphism (Peck and Valley, 1996); this is consistent with the current idea that the volcanic rocks and anorthosites formed in an island arc.

At about 80 km below the top of this Børneasund block an extensive metasedimentary belt (up to 0.5 km wide and 1–5 km long) is dominated by impure and feldspathic quartzites that contain sillimanite, biotite, garnet and magnetite and pass along strike and with complete gradation into biotite paragneisses in an area of at least 6×17 km (Myers, 1973); the quartzites are locally accompanied by calc-silicate rocks enriched in diopside and hornblende. Other minor quartzitic layers and lenses only a few metres wide occur throughout the lower zone of this block.

The lower zone rocks in the Fiskenæsset region were folded into large-scale double or triple fold interference patterns (Fig. 7A); when unravelled, the earliest isofoldal folds are recurrent nappes (Windley et al., 1973; Myers, 1985); Fig. 7B shows a remarkable, recurrent isoclinal amphibolite and gneiss. These earliest isofoldal folds are locally associated with thrusts along which sheets of tonalitic gneiss were intruded (Myers, 1976b). The last major fold phase to refold the two earlier phases of isoclines in the Lower Zone has NNW–SSW trending axial planes, as seen on Fig. 7A.

Kalsbeek (1974, 1976b) reported that U depletion has taken place over the whole of the area that we designate as the lower zone, and that also Rb/K ratios are depleted. The granulite facies metamorphism in this block has a Pb–Pb whole rock isochron age of 2810±70 Ma (Black et al., 1973), a U–Pb zircon age of 2880–2950 Ma, and a Rb–Sr whole-rock isochron age of 2800–2600 Ma (Kalsbeek and Pidgeon, 1980). Amphibolite facies gneisses have a whole-rock Pb–Pb age of 2820±70 Ma (Taylor et al., 1980), a granitic gneiss sheet that crosscuts anorthosite has a Rb–Sr isochron age of 2819±50 Ma (Moorbath and Pankhurst, 1976), and an enstatite–gedrite rock from a sapphireine-bearing layer along the top of the Fiskenæsset complex has a U–Pb zircon age of 2795±11/7 Ma, which probably dates a late amphibolite facies metasomatic event (Pidgeon and Kalsbeek, 1978).

2.4. The Sermilik block

This block extends from Graedefjord to Biskefjorden and Ameralik fjord (Fig. 1); it ranges from 50 to 100 km wide N–S. Its southern boundary is subvertical, trends E–W, and swings round the northernmost part of the Fiskenæsset complex. The northern boundary is tentatively placed along the southeastern margin of a complex tectonic zone which comprises several different Eo- to Neoarchean terranes (Godthåbsfjord–Ameralik belt, see Section 2.5). Contrary to
other block boundaries this tectonic boundary does not comprise a metamorphic facies change along all of its length. It mostly dips southeast and is folded by late upright folds. Its western part follows a prominent, subvertical high-strain zone several kilometres wide east of Færingehavn and Buksefjorden (Fig. 1). Most of the block was described by Chadwick and Coe (1983).

The Upper Zone on the southern side of this block is unusual, compared with the upper zones of other blocks, in being very thin; mostly 3–10 km wide, increasing to 15–20 km at the western and eastern ends (Fig. 1). Kalsbeek (1976a) showed that the predominant biotite–hornblende gneisses are characterised by similar porphyroblasts of muscovite and epidote and accessory titanite as the upper zone gneisses of the Bjørnesund block, and he likewise concluded that these are prograde, low amphibolite facies rocks that have never been in the granulite facies.

The gneisses contain layers of metavolcanic amphibolite up to 1 km thick (Myers, 1982). One E–W-trending, prominent amphibolite layer that has been deformed by a single phase of isoclinal folding contains primary associations of pillow lavas, volcanic breccias, agglomerates, and leucogabbros, together with discordant amphibolite dykes; geochemical data indicate calc–alkaline protoliths of basalt, andesite, dacite and rhyolite (Wilf, 1982).

Towards the ice cap the upper zone bends northwards to include Nunatak 1390 (Fig. 1) in which low amphibolite facies gneisses include bimodal metavolcanic rocks with amphibolite layers that contain vesicular pillow lavas in a tuffaceous matrix, pillow breccias, ultramafic sills that pass into ultramafic pillows, felsic lava flows and pyroclastic rocks (Escher and Myers, 1975) that probably formed in an island arc setting according to Stendal and Scherstén (2007).

The Lower Zone extends from the northern side of Grødefjord to Buksefjorden and northeastwards towards Ameralik fjord (Fig. 1), where lower-grade rocks reappear (see below). Most of this zone was metamorphosed in the granulate facies, but extensive retrogression has left many areas of hypersthene gneisses up to 30 km across well preserved within pervasive high amphibolite facies hornblende-biotite gneisses.

At the (southern) top of this zone there are many, mostly conformable, layers of porphyritic granite up to 50 km long and 2 km wide, collectively termed the Ilivertalik granite (Fig. 1). These are mезozonal, foliated to homogeneous, in places isoclinally folded, augen gneisses of granitic to granodioritic composition that have gradational contacts with host gneisses. The late structural age of the Ilivertalik granite is post-dates major recumbent isoclinal with host gneisses. The late structural age of the Ilivertalik granite is of granitic to granodioritic composition that have gradational contacts and 50 km long and 2 km wide, preserved within pervasive high amphibolite facies hornblende-biotite gneisses.

Friend and Nutman (1991) concluded that the northern boundary of the Tasiassuq terrane (that is, Sermilik block) was thrust northwards over the Færingehavn terrane to the north after 2750 Ma. Nutman and Friend (2007) reported relics of so-called high-pressure granulite facies metamorphism on this boundary. Amphibolites contain garnet–plagioclase–clinopyroxene–quartz and a metasedimentary rock contains garnets, the cores of which contain kyanite–plagioclase–rutilite–quartz, a paragenesis that indicates crystallisation at 10–12 kbar and 700–750 °C; relevant zircon has a U/Pb age of 2715±8 Ma. In a test-case geochronological study Crowley (2002) concluded that accretion of the crustal blocks was tightly bracketed between 2725±2 Ma and ca. 2720 Ma; 2720–2710 Ma according to Nutman and Friend (2007).

On the island of Qilangaausrut, west of Buksefjorden (Fig. 1), a 200 m to 1 km-wide amphibolite layer in the gneisses (Windley, 1972; Chadwick and Coe, 1984) is partly associated with sedimentary garnet–sillimanite gneiss, marble, fuchsite gneiss and quartz–plagioclase gneiss (Beech and Chadwick, 1980). The amphibolite contains relics of high-pressure granulites (clinopyroxene–garnet–plagioclase–quartz) and paragneisses contain garnet–kyanite–rutilite; the granulate facies metamorphism, which took place at 8–12 kbar, 700–750 °C, has a zircon age of ca. 2715 Ma (Nutman and Friend, 2007).

At the northern end of this zone on the southern side of the Eoarchean Itsaq gneiss complex (see below) is a ca. 8 km wide tectonic strip termed the Tre Brødre terrane by Nutman et al. (1996). This comprises a distinctive, upper amphibolite facies, granodioritic Ilkaattoq gneiss, situated largely between the limbs of the isoclinally folded Færingehavn–Buksefjorden anorthosite (Fig. 1). The Ilkkaattoq gneiss is remarkably uniform, with a fine- to medium-grained homogèneous texture displaying finely dispersed biotite flakes characteristic of prograde amphibolite facies rocks. Friend and Nutman (1991) argued that the Ilkkaattoq gneiss had never been in the granulate facies; it has SHRIMP U–Pb zircon metamorphic ages in the range 2835–2825 Ma (Nutman and Friend, 2005), and metamorphic conditions were calculated to be c. 600 °C and 7 kbar (Wells, 1976; Friend et al., 1987). Accordingly, we conclude that the Ilkkaattoq gneisses belong to a prograde amphibolite facies zone along the northern boundary of the Sermilik block. The reappearence of un-retrogressed amphibolite facies rocks at the northern end of the Sermilik block suggests that this block has been arched into a gentle antiform (see Section 3) in addition to having been tilted southwards. However, the position of the northern boundary itself in the nunatak area at the head of Godthåbsfjord remains uncertain. New protolith ages of orthogneisses from this area of around 2.88–2.87 Ga by Næraa and Scherstén (2008) are similar to those found elsewhere in the Sermilik block and suggest that the boundary is more northerly than originally envisaged by Nutman et al. (1989), see Fig. 1.

2.5. The Godthåbsfjord–Ameralik belt

The NE-trending tract between the Sermilik and Fiskefjord crustal blocks (Fig. 1) is structurally complex, and in spite of numerous studies not yet fully understood. Accordingly we have excluded this tract from both of the neighbouring blocks and labelled it the Godthåbsfjord–Ameralik belt. It comprises several terranes of different ages, besides tectonic slices of the adjacent Sermilik and Fiskefjord blocks (see below), which have all been tectonically accreted and folded together. It also includes tectonic units of the Eoarchean Itsaq gneiss complex (the Færingehavn and Isukasia terranes of Nutman et al., 2004). The western part of the belt is covered by McGregor (1993).

The Itsaq gneiss complex comprises all rocks in the Archean craton that formed before 3600 Ma; 90% of the complex consists of orthogneisses with dominant precursors of tonalite and less abundant
trondhjemite, quartz diorite, diorite and granodiorite (Nutman et al., 1996). Within the Godthåbsfjord–Ameralik belt there are two terranes of Eoarchean rocks (Fig. 1): the Færingehavn terrane (including Akilia island), which is dominated by 3850–3660 Ma migmatics that reached granulite facies in the Eoarchean and contains 3660–3600 Ma partial melts, and the Isukasia terrane (including the Isua supracrustal rock unit with ca. 3800 and 3710 Ma volcanic–sedimentary rocks), which reached upper amphibolite facies in the Eoarchean, but lacks pre-3600 Ma partial melt material (Nutman et al., 2004). The Færingehavn and Isukasia terranes are not strike equivalents, but occupy different levels in the tectono-stratigraphic pile (Friend and Nutman, 2005).

Stø on the north-western flank of the belt (Fig. 1) contains a variety of supracrustal and related rocks of different ages, including Eoarchean components, and pillow-bearing mafic amphibolites (Knudsen et al., 2007), andesitic volcanic rocks, widespread hydrothermally altered (now garnet-rich) metavolcanic rocks hosting gold mineralization, two major units of anorthosite–gabbro, and various metasedimentary rocks. Amphibolites, mica schists and garnetites host gold mineralization from which bulk Pb analyses yield impact ages of 2863±24 Ma for arsenopyrite and 2748±62 Ma for garnet (Juul-Pedersen et al., 2007). Some metavolcanic rocks appear to have been intruded by orthogneisses with U–Pb zircon ages of ca. 3055 Ma, whereas adjacent garnet–mica schists have a depositional age of ca. 2830–2800 Ma. The Støe gneisses underwent upper amphibolite facies metamorphism at ca. 620–520 °C and ca. 4.5–6 kbar at ca. 2700, 2630 and 2550 Ma (U–Pb zircon data, Hollis et al., 2005). The supra- and infracrustal rocks on Stø have thus had a complex history, and there is currently no consensus about their age range, detailed structure and development.

Quartzitic clastic sedimentary rocks on small islands south of Nuuk, on strike with the complex tectonic assemblage of supracrustal rocks with different ages on Stø, contain zircons with U–Pb ages that suggest derivation from orthogneisses as young as 2800 Ma, and sedimentation between 2800 and 2650 Ma (Schiette et al., 1988).

In innermost Godthåbsfjord Friend and Nutman (2005) defined a new 3075–2960 Ma Kapisilik terrane that contains amphibolite facies, tonalitic to granitic orthogneiss and amphibolite-dominated supracrustal layers. However, the age range and metamorphic history of these rocks are very similar to those found in the adjacent eastern Fiskefjord block, and we interpret them as originally belonging to the latter block. Thus, in our interpretation, the Mesoarchean Ivisaartaq metavolcanic belt and the adjacent anorthosite and gneisses were tectonically separated from the Fiskefjord block and folded with the Isukasia terrane. Accordingly, for descriptive purposes the Ivisaartaq belt will be described within the Fiskefjord block.

The Eoarchean Isukasia and Mesoarchean Kapisilik (at Kapisillit, Fig. 1) terranes were juxtaposed and metamorphosed together by 2970 Ma. Tourmaline–garnet–quartz–rich metasedimentary rocks along strike east of Nuuk were interpreted by Appel and Garde (1987) as volcanic–exhalative in origin. The belt is embedded in ca. 3060–3000 Ma TTG gneisses and granites, and has an amphibolite facies, metamorphic age of 2990–2970 Ma.

The prograde orthogneisses in the Upper Zone are part of the rocks that were originally defined by McGregor (1973) as Nûk gneisses, the predominantly earlier tonalitic and later granodioritic precursors of which were emplaced along thrust planes in earlier rocks (McGregor, 1979). These TTG gneisses display characteristic prograde amphibolite facies mineral textures with euhedral hornblende and finely dispersed biotite, and have Pb–Pb and U–Pb zircon ages of ca. 3065–3000 Ma.

The Y-shaped Ivisaartaq metavolcanic belt (Fig. 1), described by Polat et al. (2008) as a greenstone belt, is mainly composed of mafic and ultramafic volcanic rocks. High-Mg basaltic lavas contain pillows with concentric cooling cracks and ocelli, volcanic breccias and hyaloclastites, gabbros display magmatic layering, komatiitic ultramafic lavas have spinifex textures, and ultramafic sills have cumulate structures, and the volcanic rocks contain xenoliths of anorthosite and leucogabbro. The belt has undergone seafloor hydrothermal alteration and amphibolite facies regional metamorphism; least altered mafic rocks are characterized by LREE-enriched, near-flat HREE, and HFSE (especially Nb)-depleted trace elements patterns, indicating a supra-subduction zone, forearc signature (Polat et al., 2007, 2008). High MgO, Ni, and Cr concentrations and Mg-numbers in clinopyroxene cumulates and high-Mg pillow lavas are consistent with island arc picritic compositions, and the picritic lavas and anorthosites are isotopically comparable, permitting a mutual petrogenetic link (Dilek and Polat, 2008).

The Precambrian activity in the Ivisaartaq belt is represented by the metavolcanic rocks that were intruded by 2963±8 Ma old tonalites and granodiorites — now gneisses. Polat et al. (2008) concluded that the Ivisaartaq forearc crust was composed of an upper layer of pillow lavas, picritic flows, gabbroic to dioritic dykes and sills, and dunitic to wehrlitic sills, and a lower layer of anorthosites and leucogabbros that were intrusive into the overlying volcanic rocks. The above Ivisaartaq rocks were engulfed by voluminous tonalities, trondhjemites and granodiorites in a horizontal tectonic regime that led to crustal thickening and consequent high-grade metamorphism (Robertson, 1986).

The linear Qussuk–Bjørneøen metavolcanic belt — Fig. 8 (Garde, 2007; Garde et al., 2007) contains volcaniclastic andesites (Fig. 8B, D) with major and trace element island arc signatures, intercalated with volcano-sedimentary schists, tholeiitic amphibolite and orthopyroxene-rich ultramafic rocks interpreted as mafic tuffs, pillow lavas, sills, stocks, and cumulate rocks. The belt has a U–Pb zircon depositional age of 3071±1 Ma from a volcaniclastic rock in central Bjørneøen (the age of small, stubby, oscillatory-zoned, high-Th/U zircon crystals interpreted as volcanic; Garde, 2007; Garde et al., 2007), and the second author has recently obtained the same age along strike north of Qussuk (AAC, unpublished data). At Qussuk, a ca. 20 km long and up to about 200 m wide zone of quartz–, garnet-, biotite-, and locally sillimanite-rich rocks (Fig. 8F) with gold (–copper) mineralization that have previously been mapped as metasediment, was shown by Garde (2008) to have formed by syngenetic-epithermal leaching of tuffs and volcaniclastic rocks prior to deformation and metamorphism. Such hydrothermal alteration systems with strong acid leaching are characteristic of modern high-level oceanic and continental volcanic arcs (e.g. Sillitoe and Hedenquist, 2003). On Bjørneøen the belt is dominated by both intermediate volcaniclastic rocks and mafic amphibolites containing rare pillows, and was affected by low amphibolite facies metamorphism at ca. 2980–2970 Ma. Tourmaline– and garnet–quartz–rich metasedimentary rocks along strike east of Nuuk were interpreted by Appel and Garde (1987) as volcanic–exhalative in origin. The belt is embedded in ca. 3060–3000 Ma TTG gneisses and granites, and has an amphibolite facies, metamorphic age of 2990–2970 Ma.
The late tectonic, domal, homogeneous Taserssuaq tonalite–granodiorite that covers >1500 km² has a conventional U–Pb zircon age of 2982±7 Ma interpreted as the age of intrusion (Garde, 1997; Garde et al., 2000). The Qugssuk granite, which was mainly emplaced as late kinematic sheets (Fig. 9), has yielded identical Rb–Sr whole
rock and U–Pb zircon ages of 2969±32 and 2975±6 Ma, respectively (Garde et al., 1984; Garde et al., 2000).

The Lower Zone of the Fiskefjord block (Fig. 10) comprises granulite facies rocks, which are well preserved particularly in the western Fiskefjord and Nordlandet regions, and extensively retrogressed to amphibolite facies farther east, see Fig. 11 (Garde, 1990, 1997). Peak metamorphic, granulite facies conditions were estimated at 800±50 °C and 7.9±1.0 kbar (Riciputi et al., 1990). On Nordlandet a core of granulite facies dioritic, quartz–dioritic, and mafic tonalitic gneisses has a SHRIMP U–Pb zircon protolith age of 3221 ±13 Ma. The remainder of the area described by Garde (1997) is dominated by ca. 3050–3000 Ma, grey, K-poor amphibolite facies tonalite–trondhjemite–granodiorite (TTG) orthogneisses mostly retrogressed from brown granulite facies equivalents. Granulite facies gneisses and their retrogressed equivalents contain layers of amphibolite of likely volcanic origin with a maximum thickness of 2 km. The gneisses and amphibolites contain lenses and layers of ultrabasic rocks up to 2–3 km across of dunite and olivine-rich and orthopyroxene-rich peridotite that have migmatic layering and orthocumulus textures, and layers of stratiform layered igneous complexes, up to 150 m thick, of olivine-rich ultrabasic rocks and norites often situated along the base of major amphibolite layers with thicknesses of more than 1 km. On western Nordlandet several 1 km-wide layers of granulite facies dioritic and tonalitic gneiss are packed with enclaves of anorthosite, leucogabbro and gabbro (in a manner similar to many layers in the Ivittuut, Bjørnesund and Maniitsoq blocks). The main anorthosite on the western coast of Nordlandet contains a wide range of plagioclase compositions (−An28–97), has polygonal equilibrated textures, was generated from a protolith mixture of plagioclase and hornblende, and its parental magma had a relatively unfractonated REE pattern (Dymek and Owens, 2001). Anorthosites and their orthogneiss hosts have a Pb–Pb whole rock metamorphic age of 2890±60 Ma (Black et al., 1973), similar to more precise zircon U–Pb metamorphic ages obtained from elsewhere in the block (Garde et al., 2000 and see below). Along the margins of amphibolites and ultrabasic rocks are thin layers of pelitic biotite-garnet schist containing in places sillimanite or cordierite; a granulite-facies metasedimentary rock has zircons with metamorphic zircon overgrowths recording a SHRIMP 207Pb/206Pb age of 2999±4 Ma (Friend and Nutman, 1994), Garde (1997) and Garde et al. (2000) concluded that the precursors of the TTG gneisses were generated at 3050–3008 Ma in a convergent arc setting, most likely by partial melting of subducted basaltic rocks, around a ca. 3220 continental core. Their emplacement was accompanied by thrusting and recumbent isoclinal folding, and followed by a thermal maximum at granulite facies conditions at ca. 2980 Ma. Retrogression to amphibolite facies took place shortly after 2980 Ma (Garde, 1990, 1997), long before the tectonic and magmatic stitching with the complex zone to the east at 2720–2710 Ma (Friend et al., 1996; Nutman and Friend, 2007). The Lower Zone of this block is characterized by a triple fold interference pattern (Fig. 10) formed by refolding of early isoclines, and well illustrated in 3D (Fig. 12) by Berthelsen (1960) at Toqqusap Nuna (Fig. 1).

The Finnefjeld gneiss farther north is a complex intrusion of predominantly tonalitic and trondhjemitic rocks emplaced after the granulite facies metamorphic event, and consequently it has deformed amphibolite facies parageneses. Its contacts with host gneisses vary from, intrusive to hybrid mixed zones; it has the character of a mesozonal pluton emplaced into the deep crust of the lower zone of the Fiskefjord block. It has yielded a well-defined U–Pb zircon intrusion age of 2975±7 Ma (Garde et al., 2000).

East of the Finnefjeld gneiss complex granulite and amphibolite facies gneisses commonly contain trains of inclusions of leucogabbroic...
and anorthositic rocks that occasionally have relict igneous layering (Allaart et al., 1978; Allaart and Jensen, 1979). West of the Finnefjeld gneiss near the bottom of the Fiskefjord block granulite facies gneisses and retrogressed amphibolite facies gneisses contain many lenses up to 0.5 km across of norite, bronzitite, dunite and peridotite; one of these norites contains hornblendites bearing sapphirine, corundum and spinel (Herd et al., 1969). A group of larger, late- to post-kinematic noritic intrusions with well-preserved primary textures such as igneous layering forms a N–S-trending belt east of the Finnefjeld gneiss (Secher, 1983; Fig. 1). Small plugs and sheets of related, post-kinematic plutons of dioritic bulk composition but with up to 4000 ppm Cr are widespread in the Fiskefjord area and have been interpreted as cogenetic with the norites; one of them yielded a U–Pb zircon age of 2976±13 Ma (Garde et al., 2000). These diorites, and probably also the norites east of the Finnefjeld gneiss, were derived from ultramafic magmas that were partially contaminated with continental crust during their ascent and final emplacement (Garde 1991, 1997).

2.7. The Maniitsoq block

This block extends from Søndre Isortoq fjord to Itilleq fjord (Fig. 1), although the northernmost boundary of the Archean craton may lie well within the Nagssuqtoqidian orogenic belt at about latitude 68° N. Its southern boundary is placed along the abrupt but poorly known amphibolite–granulite facies transition along Søndre Isortoq and eastwards, which according to reconnaissance mapping around 1975 is a high-strain zone that dips steeply S and SE. Gneisses to the south of the fjord have protolith zircon dates of 3250–2970 Ma and ca. 2970 Ma granulite facies metamorphism suggesting affinity with the Fiskefjord block (Friend and Nutman, 1994; Nutman et al., 2004), whereas on the northern side of the fjord Garde et al. (2000) reported detrital zircons in a metasedimentary rock with mixed SHRIMP U–Pb ages as young as 2546±6 Ma.

The Upper Zone is only about 10–15 km wide and extends across Søndre Isortoq fjord (for a more detailed map, see Fig. 9 of Garde et al., 2000). It consists of amphibolite facies hornblende–biotite gneisses
that contain only a few layers of amphibolite and mica schist about 1–2 km wide.

The Lower Zone, extending from Søndre Isortoq fjord to at least Ililleq fjord, consists almost entirely of unretrogressed granulite facies gneisses (Bridgwater et al., 1976; Allaart and Jensen, 1979) that enclose numerous layers and inclusion trains up to ca. 1 km wide of garnet–sillimanite biotite schist and amphibolite (Allaart and Jensen, 1979). Parts of this zone contain numerous granitic sheets, which may explain the relatively high K2O contents in stream sediments (Fig. 4; Steenfelt, 1994 and personal communication, 2007). A large body of anorthosite–leucogabbro, 3 km-wide, 20 km-long, is situated on the northern side of Sandre Stromfjord near the lower margin of the block (Allaart, 1982); Allaart and Jensen (1979) reported many inclusion trains of leucogabbro–anorthositic rocks throughout the gneisses of this block. A 0.5 km-wide lens of layered norite near Maniitsoq town contains hornblendites with sapphirine, phlogopite, spinel and plagioclase (Herd et al., 1969). Detrital zircons from a garnet–sillimanite mica schist yield a definite maximum SHRIMP U–Pb age of 2866±8 Ma (Nutman et al., 2004), hypersthene gneisses at Maniitsoq town have a \( 207^{\text{Pb}}/206^{\text{Pb}} \) whole-rock age of 2890±60 Ma (Black et al., 2007).

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**Fig. 11.** Map showing the metamorphic facies of the Upper and Lower Zones of the Fiskefjord block in most of the same area as shown in Fig. 10. The Lower Zone displays granulite facies metamorphism in the west and southwest and variable retrogression to amphibolite facies towards east. In contrast, the Upper Zone in the southeast is characterized by prograde amphibolite facies metamorphism. The Archean metamorphic pattern is disturbed south of Qussuk by a Paleoproterozoic rotational fault along Godthåbsfjord, which juxtaposes granulite facies and amphibolite rocks on either side of the fjord. For location see Fig. 2. Modified from Garde (1989, 1990).

**Fig. 12.** The structure of the Toqqussap Nunaq peninsula in the western Lower Zone of the Fiskefjord block (Fig. 1) showing the geological map and 3D interpretation of the triple fold interference pattern of Berthelsen (1960). Position marked on Figs. 2 and 10.
1973), and a nearby orthopyroxene-bearing granite contains meta-
morphic zircons with a SHRIMP U–Pb age of 2738±6 Ma (Friend and
Nutman, 1994).

In the northeast of the Maniitsoq block within Meso- to Neoarchean
granulite facies gneisses is the ca. 1500 km² Eoarchean Aasivik terrane
that consists of garnet-bearing granulite facies orthogneisses partly
retrogressed under amphibolite facies conditions (Rosing et al., 2001).
Zircons from orthogneisses have SHRIMP U–Pb ages ca. 3600 Ma
with ca. 2700–2550 Ma overgrowths equated with a granulite facies
metamorphic overprint (Nutman et al., 2004).

About 100 km along strike to the east from the entrance of Sandre
Isortoq fjord at Qarliit Tasersuatuk lake orthogneisses, enclosing lenses
of magnetite–quartz ironstone and garnet quartzite (Hall, 1978),
contain zircons with SHRIMP U–Pb ages in the range 3700–3600 Ma
(Nutman et al., 2004).

3. Discussion on crustal evolution of West Greenland in the
Meso–Neoarchean

The development of the continental crust of any age was no doubt
a complicated process, and therefore we should expect any individual
crustal section to have a complicated make-up, structure, and geo-
chemical evolution. In the well-exposed Archean bedrock of West
Greenland great strides have been made by many individuals
to unravel the complexities of the structure (e.g. Berthelsen and
Henriksen, 1975; Myers, 1985), geology (e.g. McGregor, 1973; Myers,
1985; Garde, 1997, 2007), mineralogy (e.g. Herd et al., 1969),
isotopically-and geologically-defined terranes (e.g. Nutman et al.,
1989, 1993, 1996; Friend and Nutman, 2001; Nutman et al., 2004;
Friend and Nutman, 2005), and geochemistry (e.g. Polat et al., 2007,
2008). However, apart from a local attempt by McGregor et al. (1991),
there is still no viable, overall tectonic framework that can explain
the mutual relevance and inter-relationships of the detailed observations.
In this paper we have demonstrated that the Meso–Neoarchean craton
of West Greenland is made up of six major crustal blocks exposing a
complete transition from upper to lower crust, and within or between
which the previously established Eoarchean and younger terranes can
be better understood. Terranes are, by definition, unique, local and
atypical. Each of the blocks, on the contrary, is large enough to display
a representative section of upper to lower Archean crustal compo-
nents and structures. We now compare and contrast the different
crustal blocks to help understand the gross structure of the
continental crust, and to draw implications for mechanisms of crustal
growth and for the possible plate tectonic evolution of the Archean
crust of West Greenland.

1. The four best known blocks of crust, Kvanefjord, Bjørnesund,
Sermilik and Fiskefjord, all have a similar upper zone containing
one or more linear green schist to amphibolite-grade volcanic belts
embedded in amphibolite facies, TTG-type, prograde orthog-
neisses. The metavolcanic belts contain distinctive and diagnostic
pillow-bearing basalts, tufts, pyroclastics, and anidesitic rocks
suggesting an arc origin that is well supported by geochemical
data. The upper zone of the northernmost Maniitsoq block is not
sufficiently well known to be included at present. Also the four
blocks have similar lower zones made up largely of TTG-type orthogneisses (including diorites) that are at granulite grade or
have been partly retrogressed to high-amphibolite facies equiva-
lents. Particularly significant are two observations:

a. All the lower and upper crust zones contain km-thick amphibolite layers that can be demonstrated (by volcanic
structures or geochemical signatures) or can be reasonably
assumed to be volcanic in origin, and many are associated with
minor sedimentary components that are mostly (meta) grey-
vackes or pelites; marbles are sparse and quartzitic rocks very
rare, or of hydrothermal origin (Garde, 2008). There can be no
doubt that these supracrustal rocks were laid down under water
at the surface of the Archean earth and were then buried to a
variety of levels of the continental crust where they reached
different metamorphic grades.

b. All the crustal blocks contain km-thick anorthositic layers in
their granulite facies lower zones and several in their prograde
upper zones (Owens and Dynek, 1997). In the Børnesund block
diagnostic chrome-layered anorthosites have distinctive sapphire-
bearing rocks along their top contact with metavolcanic
amphibolites in both the Upper and Lower zones, demonstrating
beyond doubt the comparable mode of early evolution of these
zones. In most blocks the anorthosites are commonly accom-
panied by leucogabbros and gabbros, and associated spatially
with metavolcanic amphibolite layers. In the Børnesund block
anorthosites of the Fiskeneøset complex have root pendants of
pillow lavas belonging to an overlying amphibolite layer of
volcanic origin, and in the Fiskefjord block anorthosites have
intruded overlying volcanic rocks. From detailed geochemistry
at Ivisaartoq Polat et al. (2008) demonstrated the consanguinity
of overlying picritic lavas and underlying anorthosites.

2. How were the supracrustal volcanic, sedimentary and anorthositic
rocks buried to different levels of the crust? Two observations are
relevant to this question:

a. The bulk of the Archean crust of West Greenland is made-up
of TTG-type orthogneisses, the bulk composition of which was
estimated to be ca. 30% tonalite, 15% granodiorite, 14% granite,
and 22% mafic rocks by Wedepohl et al. (1991), who found no
systematic change in their chemical composition with age. Steenfelt et al. (2005) established that TTGs in the Upper and
Lower zones of the Fiskefjord block have uniform compositions,
close to the average TTGs of Martin (1994), and concluded that
their chemical signatures are compatible with slab melting in a
subduction zone setting, combined with some degree of re-
action of the TTG melts with the overlying mantle wedge. They
showed that some Fiskefjord TTG plutons of both intermediate
and felsic composition contain elevated Ba, Sr and light REE,
which they interpreted as direct evidence of interaction with
carbonatite-metasomatized mantle in addition to slab melting
(see also Martin et al., 2005). From the mineral compositions of
sands Kalsbeek (1971) found little variation in the bulk
mineralogical composition of the rocks in the Upper and Lower
zones of the Børnesund block, except for hypersthenes that
decrees southwards. However, it is well known from many
other studies that certain large-ion lithophile elements will be
redistributed by metamorphic fluid transport. It has long been
recognized that the precursors of the gneisses were emplaced
into the supracrustal rocks as well as into deeper levels of the crust
(e.g. Wells, 1979, 1981; Nutman and Garde, 1989). The
emplacement of the tonalitic precursors into the volcanic rocks
and anorthositic complexes was commonly on such a scale that
only lenses of amphibolite, and anorthositic–leucogabbroic
rocks are left as trails of inclusions in the resultant gneisses;
e.g. in the Ivvittuq block (see map of Berthelsen and Henriksen,
1975), and in the Fiskeneøset region of the Børnesund block
(Windley et al., 1973; Myers, 1985). These TTG orthogneisses
were responsible for appreciable thickening of the crust and, if
emplaced by over-accretion, can mechanically suppress parts of
their supracrustal hosts to deeper crustal levels (Wells, 1979,
1981). We consider that the most likely environment for
formation of such voluminous TTGs was in mature island arcs
and active continental margins, given the fact that today the
North and South American Cordillera is the only site of
voluminous production of TTG rocks (Windley and Smith,
1976; Windley et al., 1981; Lee et al., 2007).

b. Throughout the Archean craton there is considerable evidence
for thrusting of major layers very early in the deformation
history, often in association with isoclinal folding. In some regions double and triple fold interference patterns are prominent on small and km-scales, and unravelling of their fold geometry always shows that the earliest folds were flat-lying isoclines, like that seen in Fig. 7B. These relations led Bridgewater et al. (1974) to propose a horizontal tectonic regime and thrust-controlled crustal thickening. Nutman and Friend (2007) pointed out that so-called high-pressure metamorphic assemblages of 8–12 kbar are present in mafic rocks and paragneissess along some tectonic boundaries, implying possible subduction to mantle depths.

3. In West Greenland the emplacement of many tonalite sheets in the Mesoarchean and perhaps even earlier took place in association with thrusting (Myers, 1985; Hanmer et al., 2002, and our observations). Therefore current data suggest that the combined effects of thrust imbrication, isoclinal nappe folding and syntectonic emplacement of voluminous TTG sheets were responsible for the thickening of the crust in Meso–Neoarchean times, as concluded by Myers (1976b).

4. As one passes from the upper zone downwards in the crustal sections, the structural style of deformation changes concomitantly with the changes in metamorphic grade. In the upper greenschist to amphibolite facies prograde zones the metavolcanic belts are linear and have been deformed mostly by only one large-scale tight to isoclinal phase of folding. In contrast, in the lower zones kilometre-scale, double and triple fold interference patterns are common, having formed from usually two major phases of folding after the first isoclinal phase. It is worth pointing out here that in the Bjørnesund block the stratigraphy of the Fiskenæsset complex is so well-defined and regular that its repetition in reverse order enables definition of the first isoclinal synclinal fold phase — see inset of Fig. 7A (Windley et al., 1973); without such stratigraphy of the layered anorthositic complex it would not be possible to recognise the first deformation phase in the homogeneous gneisses and amphibolites. Figs. 3 and 12 show the refolded folds in 3D reconstructions of the deep crust in the Lower Zones in the Ivittuut block and at Toqqusap Nunaq in the Fiskenesfjord block by Berthelsen and Henriksen (1975) and Berthelsen (1960), respectively. The last generation of major folds in most blocks, which have approximately N–S-trending axial planes, affected particularly the lower zone (see folds in Fig. 6A). This may point to the presence of flat-lying attachment–detachment zones between brittle and ductile parts of the upper and middle crust, as described from younger accretionary orogens such as the Paleoproterozoic Ketilidian orogen (Garde et al., 2002).

5. The similarity in make-up of each of the crustal blocks in West Greenland has been emphasized above, but there are also significant differences, which are:

a. The amount of preserved granulite facies areas decreases southwards (Fig. 13). The northernmost Maniitsoq block consists almost entirely of granulites in a cross-strike width of more than 120 km. The Fiskefjord block has two granulite facies areas that have a total width of about 50 km across strike. In the Sermilik, Bjørnesund and Kvanefjord blocks granulite facies rocks have been extensively retrogressed to high-amphibolite facies and the preserved granulite areas rarely exceed 35 km across. The southernmost former granulite facies rocks in the Neria area of the Kvanefjord block has been so heavily retrogressed to amphibolite facies that evidence of its earlier granulite facies is only preserved in characteristic and diagnostic bleby textures ( McGregor and Friend, 1997 ). The exposed part of the lower zone of the southernmost Ivittuut block has no evidence of having been at granulite facies. It probably only reached at amphibolite facies, rather than being completely retrogressed from granulite to amphibolite facies; we favour the first possibility, because of complete lack of

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**Fig. 13.** Schematic profiles of the six crustal blocks from north to south, which show their Upper Zones of prograde amphibolite facies orthogneisses and granitic rocks, and Lower Zones that largely consist of granulite facies orthogneisses and retrogressed amphibolite facies equivalents, as well as late granitic rocks. Anorthosites and amphibolites occur in most zones of most blocks. To correlate with geology see Fig. 1.
important: the northernmost and southernmost blocks. Two directions may be demonstrating the urgent need for more isotopic data especially from the distribution of ages amongst the blocks varies considerably, 3000 Ma and the thermal maximum and stabilisation of the lower main growth stage of formation of orthogneiss protoliths (ca. 3065 Ma). For instance, in the Fiskefjord block the time difference between the protolith ages of the main rock units in both upper and lower zones. have shown representative samples of the best available zircon growth. In the summary pro
distinguish between early growth of protoliths and late metamorphic directions of crustal growth within or between the blocks? We here ask the question: are there suf
cient data to indicate directions of crustal growth within or between the blocks? We distinguish between early growth of protoliths and late metamorphic growth. In the summary profiles of the six crustal blocks of Fig. 13 we have shown representative samples of the best available zircon protolith ages of the main rock units in both upper and lower zones. For instance, in the Fiskefjord block the time difference between the main growth stage of formation of orthogneiss protoliths (ca. 3065–3000 Ma) and the thermal maximum and stabilisation of the lower crustal zones (2980 Ma) was about 100 Ma (Garde et al., 2000). Clearly the distribution of ages among the blocks varies considerably, demonstrating the urgent need for more isotopic data especially from the northernmost and southernmost blocks. Two directions may be important:
a. The Maniitsoq and Fiskefjord blocks have some protolith ages older than 3000 Ma, in contrast to the four blocks to the south. Does this mean that the southerly blocks formed later than those in the north? This question concerns both the volcanic and sedimentary ages of the main upper zone volcanic belts, and the protolith ages of the granulite facies orthogneisses in the lower zones. At present, the age data are not yet sufficiently widespread to make unreserved conclusions. However, the current information makes these Greenlandic crustal blocks an attractive focus to address this question.
b. Did accretionary crustal growth take place downwards or upwards within any of the individual blocks, or was the growth random? The north-south continuity, varied rock types, and wonderful exposure should enable this important question to be resolved in future, but at present the data are insufficient to detect indubitable trends.

Fig. 14 provides an explanation for many of the differences between the crustal blocks and in particular for some variations from north to south in terms of a model of differential uplift of different blocks. This would explain the general decrease in metamorphic grade of the lower zones from north to south, as well the repetition in the Sermilik block of the upper zone on both northern and southern sides of the lower zone caused by a broad arching of that block.

4. Tectonic model

From all the multi-disciplinary data summarised above, we conclude that the six crustal blocks in the North Atlantic craton are sufficiently different that it is doubtful that they all evolved within the same continent, and more likely that they developed by similar processes but in several separate blocks that were later amalgamated by collision into a single continent. We have yet to constrain whether the amalgamations took place by stepwise or out-of-order collisions. We emphasise that while the metamorphic jumps across the blocks are factual, the block boundaries themselves are not all well exposed, and their tectonic nature is not fully understood in all cases. Thus, the south-eastern boundary of the Fiskefjord block and the north-western boundary of the Sermilik block are variably reworked fault and thrust boundaries, with stitching granite sheets at around 2720 Ma (Friend et al., 1996). Other block boundaries such as the northern boundary of the BjarneSund block are occupied by shear zones (see later). The boundary between the Maniitsoq and Fiskefjord blocks occurs in a poorly known region, and the BjarneSund–Kvanefjord boundary is largely unexposed.

We now consider the history of the blocks in terms of a general model for the evolution of the Meso-Neoarchean continental crust of West Greenland (Fig. 15). Five stages of magmatic-structural growth are considered: formation of island arcs, thrusting and isoclinal folding of the arcs synchronous with emplacement of TTG magmas during formation of Andean-type magmatic arcs, followed by collision of the main blocks and formation of late third phase folds with steep
axial planes, possible detachment between the upper and lower zones of some blocks, and emplacement of late post-tectonic crustal melt granites.

1. The somewhat limited lithological and geochemical data currently available, and mainly from the Fiskefjord block, indicate that the relatively low-grade metavolcanic belts in the upper zones of the crustal blocks formed in different types of island arcs, or at different stages in arc development (Fig. 15A).

According to Polat et al. (2008) the Ivisaartoq belt formed in an incipient stage of arc development in intra-oceanic supra-subduction forearc oceanic crust composed of upper high-Mg pillow basalt with gabbro and dunite-wehrlite sills, and underlying consanguineous anorthosite-gabbro. High Mg-basalts and boninites are conceived as shallow melts of depleted mantle (Van der Laan et al., 1989), as in the extensional regime of the Izu-Bonin forearc (Taylor and Nesbitt, 1995). Polat et al. (2008) then went on to make the following prescient suggestion, which we wish to follow up. If the mantle potential temperature was higher in the Archean and thus gave rise to a thicker oceanic crust than today (Sleep and Windley, 1982; McKenzie and Bickle, 1988), the fact that the anorthosites and leucogabbros at Ivisaartoq are volumetrically minor was maybe because the upper basaltic section of oceanic crust was peeled off and accreted, while the bulk of the lower anorthosite-leucogabbro section was subducted. We suggest this idea is correct for the following supportive reasons.

It is well established that at the subduction zones in the western Pacific the ultramafic rocks, gabbros, and sheeted dykes of the oceanic crust are usually subducted, and therefore only the basalts and overlying sediments are peeled off and accreted (Kimura and Ludden, 1995). This is why in Japan, for example, there are thousands of shreds of basalts, but hardly any ophiolites. Furthermore, the crust of island arcs has commonly been delaminated, the lower part subducting and the upper part accreting (Kay and Kay, 1993). This mechanism is confirmed by seismic reflectors at the arc-arc collision zone at Hokkaido in N. Japan, which indicate that the lower part of the lower crust of the Kuril arc is being subducted, and that the upper part is being peeling off and obducted (Tsumura et al., 1999). In other words, we know today that this is what happens at a typical subducting accretionary plate boundary. In the Bjornesund block where the stratigraphy of the Fiskenæsset complex is best preserved there are seven major units (Myers, 1985) or only five units (Windley et al., 1973). However, throughout the bulk of the region many layers of the complex contain only the top two or three units commonly with their overlying volcanic rocks; the lower half of the complex is consistently missing. We favour the idea that the missing lower section of the complex was removed not by later tectonic deformation (it is unlikely that the extensive thrust deformation would have removed such stratigraphy), but during subduction-accretion. Throughout all the other crustal blocks in West Greenland there are anorthosite-leucogabbro layers, but they contain only one or two units and are unlikely to represent complete sections of a magma chamber in either an oceanic or arc setting. Moreover, whereas many volcanic layers in

Fig. 15. A structured cartoon illustrating the progressive development of the Mesoproterozoic Neoarchean crustal blocks of West Greenland. The five stages demonstrate the strong interrelationships between the arc-generated emplacement of sheets of TTGs, and the subhorizontal thrust-nappe tectonics, which together led to massive thickening of the crust and consequent thermal granulite facies metamorphism in the deeper crust. A. Magmatic accretion of initial ultramafic, mafic and intermediate intrusive and extrusive arc components, including anorthosite, in pre-existing mafic crust during subduction. B. Thrusting. Emplacement of proto-continental TTG magmas in upper crust, possibly prompted by convergence of two oceanic arcs. C. Continued thrusting, now with (mainly recumbent) isoclinal folding. Synkinematic TTG magma emplacement in upper crust. D. Continued isoclinal folding and abundant magmatic TTG accretion into upper and middle crust. Thermal granulite facies metamorphism begins in lower crust. E. Final thickening, maturation and stabilisation of new continental crust about 100 Ma after initial oceanic arc formation. Tectonic attachment–detachment zone separates simple archs and cusps in upper crust from complexly refolded folds in lower crust. Thermal maximum with granulite facies at depth, and crustally derived granites emplaced into upper crust in ‘granite–greenstone’-style. Maximum crustal thickness attained. In the Fiskefjord block local delamination of heavy components at base of crust (and ensuing emplacement of post-kinematic, mantle-derived ultramafic magmas, not shown).
all the crustal blocks are bordered by anorthositic–gabbroic layers, many others are not; absence of the latter may be due to tectonic removal. The Qussuk–Bjørneøen metavolcanic belt in the Upper Zone of the Fiskefjord block contains predominant andesites, lithic tuffs and volcaniclastic rocks, and has the geochemical signature of a mature island arc (Garde, 2007). The Ikktapp Nunaa belt comprising basaltic pillow lavas, pillow breccias, tuffs and a silt of layered gabbros and leucogabbros has a distinctive arc-type stratigraphy (Andersen and Friend, 1973; Friend, 1976a,b), but absence of geochemical data prevents a detailed petrogenetic analysis. Although the Ikktapp Nunaa metavolcanic belt is not itself associated with anorthosite, lower down in the upper zone of the Bjørnesund block metavolcanic rocks are underlain by anorthosite layers, and not far below that there are roof pendants of pillow lavas within an underlying anorthosite, confirming the contemporaneity of the two rocks, and lending support to the idea that the anorthositic complex formed in the underlying magma chamber or in the lower crustal layer of the volcanic arc. The Građefjord belt contains pillow lavas, volcanic breccias, agglomerates, and leucogabbros, and geochemical data indicate protoliths of basalt, andesite, dacite and rhyolite (Wilf, 1982), suggesting a calc-alkaline framework was apparently different from that of the other low-grade metavolcanic rocks to the north. Significantly, the Qussuk–Bjørneøen belt of island arc rocks contains intrusive plutons of tonalite that could be associated with a mature island arc (Fig. 8E). The variable compositions and rock types of the West Greenland arcs would be consistent with growth in basins with highly variable geometry such as the modern Fiji basin, as suggested by Lagabrielle et al. (1997).

2. In the modern Earth, proto-archs with supra-subduction signatures develop in forearc settings that are involved in subduction rollback (Dilek and Flower, 2003), giving rise to extension in the forearc that results in a high degree of partial melting of hydrated upper mantle at shallow depths, as in Ivisaartoq (Polat et al., 2008). Subsequent rollback gives rise to a succession of arcs with potentially different stratigraphy and geochemical fingerprints depending on many factors such as the composition of mantle being melted for ridge growth, the nature of oceanic plate being subducted with or without, for example, ocean islands or seamounts, and subduction angle and the age of ocean floor being subducted. For this oceanic arc environment to change to an active continental margin, a major change is required in tectonic boundary conditions, such as forearc–arc collision or arc–continent collision, and this may involve back-arc basin closure by subduction. Such a collision enables an island arc to become the leading edge of a continental block or of a block of accreted island arcs. In consequence, further subduction gives rise to intrusion of voluminous tonalites or TTGs in plutons and batholiths into the attached island arc (Fig. 15B) and into the new active continental, Andean-type margin (Fig. 15C, D). For example, the Kohistan island arc accreted or collided with the Asian continental margin when its back-arc basin collapsed in the Cretaceous, causing the arc to become the leading active margin of Asia, so enabling an Andean-type tonalitic–granodioritic–dioritic batholith to be intruded into the accreted arc (Coward et al., 1982; Khan et al., 1993).

In most of the crustal blocks of West Greenland there was a major change from formation of an island arc, now preserved in the upper zones, to the emplacement of voluminous TTGs, later to become orthogneisses, that took place in what are now preserved as the upper zones and in particular the lower zones in association with major isoclinal folding and thrusting. We suggest two tectonic scenarios that could account for such a change in tectonic environment.

First, the centre of the Fiskefjord block is occupied by a dioritic gneiss that has a U–Pb SHRIMP zircon protolith age of 3221 ± 13 Ma (Garde, 1997). Garde et al. (2000) suggested that these rocks acted as a continental core or block around which arcs accreted, giving rise to the emplacement in the period ca. 3065–3000 Ma of voluminous dioritic and tonalitic sheets (that later became the common grey gneisses) in association with extensive thrusting and isoclinal recumbent folding. Geochemical and Sm–Nd data suggest that the diorites and tonalites were largely derived by partial melting of hydrated oceanic crust that most likely had been subducted in a convergent setting (Steenfelt et al., 2005). The combination of diorite–tonalite magmatism and thrusting-folding thickened the crust until ca. 2980 Ma when it reached its maximum thickness at granulite facies conditions (Fig. 15E). At present insufficient age data are available to indicate whether other crustal blocks have similar old rocks that could have acted as nuclei for subsequent accretion and collision.

Second, if several island arcs mutually accreted to form larger crustal blocks, then Andean-type magmatic arcs could have developed on their margins. This change in boundary conditions would have triggered the emplacement of large volumes of TTG melts causing burial of many volcanic-anorthosite belts and massive crustal thickening. We favour this model to explain the coalescence and growth of most of the crustal blocks. We do not know the age and composition of most of the volcanic layers in the lower zone high-grade gneisses in all the crustal blocks, but we do know that several contain tuff and pyroclastic rocks, and that very many are bordered by anorthosites, like the island arc-type volcanic belts in the upper zones of the Bjørnesund and Fiskefjord blocks. We envisage that these volcanic rocks with their underlying anorthositic magma chambers were generated in arcs, and that they were progressively buried in the growing crust by thrust-nappe deformation and by associated over-accretion of voluminous TTGs (Fig. 15C, D), as predicted with early insight by Wells (1979). The massive crustal thickening led to granulite facies metamorphism deep in the crustal piles (Fig. 15E). Not only volcanic and anorthositic rocks, but also sedimentary rocks, were buried in the deep crust, such as an 6×17 km area of quartzites at present situated in high-grade gneisses 80 km below the present top of the Bjørnesund block (Myers, 1973). Based on metamorphic conditions, the maximum difference in crustal level between the lowest grade rocks in the upper zones and the highest grade rocks in the lower zones is ca. 15–20 km, and the maximum amount of crustal thickening from surface to granulite facies in the present exposed sections is ca. 30 km.

A consequence of the massive crustal thickening in each block would likely have been the formation at depth of crustal melt granites that arose to be emplaced at different tectonic levels (Fig. 15E). Some came to reside in the lower zones as mesozonal foliated to homogeneous conformable granites commonly with gradational borders, such as the Ilivertalik granite in the Sermilik block that in places has granulite facies assemblages, and in the Fiskefjord block the Finnefjeld granitic gneiss emplaced during amphibolite facies metamorphism (Fig. 1). Post-tectonic norites and diorites dated at 2975±13 Ma (Garde et al., 2000) in the same block stem from ultramafic magmas (shown by their very high Ni and Cr contents, Garde, 1991), and their emplacement may have been prompted by local delamination of the thick new crust. Some granitic melts rose
into the upper zones to form post-tectonic discordant plutons (Fig. 15E) as in the Kvanefjord block, the very extensive Taserssuak tonalite–granodiorite and the Qussuk granite in the Fiskefjord block, muscovite–garnet gneisses and two mica granites with rapakivi textures in the Bjørnesund block, and finally the crustal melt Qørqut granite that was emplaced much later into the Godthåbsfjord–Ameralik belt.

If the above accretionary growth model were applicable, it might be expected that the protolith ages of the tonalitic gneisses increased with depth. However, as Wells (1979) pointed out, the crust can grow by both over-accretion and under-accretion of tonalitic increments, giving rise ideally to isothermal or isobaric PT cooling paths respectively, and to a mixed growth crust. Only a combination of more data from analysis of the structural evolution, geochemistry, PT paths and isotopic ages will constrain these relations. An interesting Neoarchean example of this mode of accretion is provided by the Minto block in the Superior Province in Canada (Percival and Skulski, 2000; Percival et al., 2001). An amalgamation of ca. 2.7 Ga island arcs, oceanic crust and back-arc assemblages gave rise to a proto-craton on the margin of which a granodioritic batholith was emplaced in an Andean-type magmatic arc. The docking of oceanic island arcs and a continental terrane created a composite continent. Collision-related thrusting gave rise to 7–8 kbar granulite facies metamorphism and a ~40 km-thick crust. Minto provides an instructive, near-coeval example of the fundamental processes involved in West Greenland.

3. As pointed out above, current data lead to the most likely conclusion that most of the six Meso–Neoarchean crustal blocks developed as separate micro-continents by similar processes of arc growth and crustal thickening, but with some mutual differences such as their proportions of arc volcanics and anorthosites. Although some of the block boundaries may have been reactivated by Paleoproterozoic movements in response to the Paleoproterozoic Nagssugtoqidian and Ketilidian orogenies on either side of the North Atlantic craton, the logical outcome of such an Archean tectonic scenario is that the micro-continent collided and amalgamated, together with the Eoarchean Godthåbsfjord–Ameralik belt (Nutman and Friend, 2007), to form the present Archean craton. On a more restricted scale, Friend and Nutman (1991) proposed that their terranes between the Fiskefjord and Sermilik fjords were amalgamated by collision tectonics controlled by SE-dipping subduction zones.

In 2008 BFW discovered on the western coast a 400 metre-wide, sub-vertical shear zone precisely on the boundary between the Bjørnesund and Sermilik blocks. This shear zone that separates partly retrogressed granulite facies gneisses to the south from prograde epidote gneisses to the north consists solely of augen gneisses, mylonitic banded gneisses, cataclastic and porphyroclastic gneisses, which have a strong mineral and roding lineation that plunges moderately to the west. The fact that it was possible to predict exactly where this major shear zone would occur between two crustal blocks adds credence to the idea that it does indeed mark a suture zone.

The boundary between the upper and lower zones of the Bjørnesund block is marked by an unthrusted prograde metamorphic transition (Friend et al., 1987, 1988b) and this is where the upper zone is about 40–50 km wide. One might expect that such a boundary could become the locus for later thrusting. The upper zone of the Sermilik block is commonly only 3–10 km wide, and thus we expected it to have been shortened by thrusting. On the northern side of Graefefjord the boundary is marked by a 250 m wide, north-dipping shear zone that contains mylonitic, cataclastic, porphyroclastic and augen gneisses that have a strong, north-plunging, mineral and roding lineation associated with top-to-the-south kinematic indicators. We consider that this provides evidence that the upper zone of this block is narrow because it was shortened by thrusting that carried the high-grade lower zone southwards over the prograde upper crustal zone.

Friend et al. (1987, 1988b) suggested that shear zones in the Nuuk region represent the original thrusts responsible for coalescence of different terranes, but Nutman and Friend (2007) pointed out that some of them may have been reactivated as extensional shear zones, perhaps connected with exhumation. Elsewhere, for example, near the southern border of the upper zone of the Bjørnesund block there are south-dipping late thrusts containing pseudotachylites that we consider most likely resulted from post-metamorphic stacking and imbrication with the Kvanefjord block to the south. However, we are cognisant of the fact that little work has been published on the shear zones marking possible sutures between the different crust blocks in West Greenland. Such zones are particularly prone to glacial erosion, and some are no doubt partly or wholly hidden by fjords, glaciers and/or outwash plains.

5. Other Archean crustal sections

Several Archean crustal sections in the world have lithological and structural features that are remarkably similar to those in West Greenland, and thus provide additional information on how the crust has grown with time in broadly similar tectonic settings. The most comparable are:

5.1. Kapuskasing uplift, Canada

The most comparable section is in the Superior Province in Canada, where the Kapuskasing uplift at 1.9 Ga has exposed a cross-section of shallow-dipping Neoarchean crust through a paleodepth of 25–30 km. A c. 10 km-thick upper crust of greenschist-grade greenstone belts dominated by arc-type metavolcanic rocks and granitic plutons passes downwards to a c. 10 km-thick mid-crust of amphibolite facies tonalitic orthogneiss with largely concordant granites, to a lower crust of granulite facies orthogneiss, paragneiss and anorthosite (Percival and West, 1994). Isotopic data suggest that the lower crust experienced younger metamorphism (2660–2600 Ma) than the upper crust (ca. 2670 Ma; Moser et al., 2008). The upper crust experienced magmatism and metamorphism long before these processes took place at deeper crustal levels, i.e. from ca. 2730 to 2625 Ma (Corfu, 1987). Early arc-type volcanic rocks (komatiites, basalts, dacites) were intruded prior to 2765 Ma by the plutonic compliment (Percival and Cole, 1981) of the Shawmere calcic anorthosite (An95–85) with its minor gabbric, ultramafic and sapphire-bearing rocks (Simmons et al., 1980), whereas emplacement at depth of pre- and syntectonic tonalite sheets that gave rise to tonalitic granulite facies gneisses and intrusion of late plutons was at 2680 to 2668 Ma.

5.2. Pikwitonei, Canada

An oblique cross-section through c. 20 km of Neoarchean crust is exposed in the Pikwitonei region of the northwestern Superior Province (Weber and Mezger, 1990). This is a rare example of an Archean terrane where greenstone belts can be trace continuously from greenschist to granulite facies (Mezger, 1992). In the east, 2830–2700 Ma greenstone belts with arc-type komatitic-tholeiitic to calc-alkaline successions were intruded by granitic protoliths and metamorphosed in the greenschist faces at 3 kbar pressures. They increase in grade and decrease in width and continuity westwards via amphibolite facies through a prograde orthopyroxene isograd at 7–9 kbar dated at c. 2640 Ma. Tonalitic–granodioritic, granulite facies orthogneisses contain layers and trails of metamorphosed volcanic and sedimentary rocks regarded as high-grade equivalents of the greenstone belts. In the granulite terrane are layered igneous complexes (Ermanovics and Davison, 1976; Bell, 1978) up to ca. 6–8 km thick that consist of anorthosite (c. An94–84), gabbro anorthosite and
gabbro (extremely similar in appearance, mineralogy and texture to those in the Fiskenneset complex), and which are bordered by metavolcanic amphibolites and metasedimentary rocks. According to Weber and Mezger (1990) these layered complexes were co-genetic with the greenstone belt basaltic volcanism. Granulite facies metamorphism took place at 2660–2637 Ma, and amphibolite facies retrogression at 2605–2591 Ma; the metamorphism in the deep level crust post-dated that in the high-level greenstone belts (Percival et al., 1992). Mezger (1992) suggested that the Pikwitonei section developed in an Andean-type magmatic arc.

5.3. South India

Meso–Neoarchean prograde amphibolite facies gneisses that contain greenschist to amphibolite facies prograde metavolcanic belts (the Dharwar greenstone belts) pass southwards via a prograde orthopyroxene isograd of the Fermon Line to granulite facies gneisses that have been partly retrogressed in southern India; this is a possible Archean crustal section (Newton and Hansen, 1986) with an approximate difference in crustal level from low- to high-grade of 15–20 km (Condie et al., 1982). In the granulite facies gneisses there are anorthositic-gabbroic layered complexes bordered in places by amphibolite layers that may of volcanic origin; the best example is the Sittampundi complex (Subramaniam, 1956), which is made up of calcic anorthosite (An80–100), gabbroic and noritic anorthosite, gabbro, pyroxenite, troctolite and chromite layers in anorthosite up to 3 m thick and more than 7 km long. Remarkable similarity to the Fiskenneset complex is indicated by its rock components, mineral assemblages and mineralogy, and by the fact that it has been folded into an isoclinal antiform, which can only be recognized by the repetition in reverse order of the chromite-layered stratigraphy (Ramadurai et al., 1975). The whole rock Sm–Nd isochron age of the Sittampundi complex is 2950±60 Ma, and the nearby comparable Bhavani complex 2899±2 Ma (Bhaskar Rao et al., 1996). Leelanandam et al. (2006) speculated, correctly in our opinion, that such mafic-ultramafic complexes in the Sittampundi–Bhavani zone may be “the roots of arc volcanoes”.

In summary, the above crustal sections especially at Kapuskasing and Pikwitonei are broadly similar to those in West Greenland. Particularly striking are the similarities in overall changes with depth from low-grade arc volcanic rocks to deep arc-generated granulites with distinctive layered calcic anorthosite complexes. Nevertheless, although chemotratigraphy and geochemistry go a long way in defining particular tectonic settings in young and old orogenetic belts, Archean belts, particularly those in the deep crust, are still problematic because their lack of geometrical relationships with specific plate tectonic settings prevents any more diagnostic conclusions. Therefore, it would be useful to find some Phanerozoic belts with comparable rocks, lithotratigraphy, structure and age relations, which have known relationships with specific plate tectonic settings.

6. Modern analogues

There are several arc-derived crustal sections especially of Mesozoic–Cenozoic age that are remarkably similar to those in West Greenland, either in their actual mode of occurrence and/or in their principle mechanisms of evolution, and which therefore give useful comparative information, because their structural and petro-geochemical data can be related to specific plate tectonic boundaries and evolution.

6.1. Fiordland, South Island, New Zealand

In Fiordland a 25–35 km-thick tilted section of Mesozoic arc-generated crust is exposed in a metamorphic core complex that formed at an earlier active continental margin (Gibson, 1990; Oliver, 1990). The lower crust includes the 139–129 Ma Arthur River island arc of plutons of calc-alkaline gabbro and diorite that collided with the continental margin of SE Australia and within c. 20 Ma was buried to a deep crustal level with formation of high-pressure granulites (Hollis et al., 2003). The accreted arc was intruded by the >10 km-thick Western Fiordland 119–130 Ma orthogneiss that belong to a batholith that has a distinctive adakitic composition and formed in an Andean-type magmatic arc in the leading edge of the SE Australian plate. The protoliths were buried and metamorphosed to granulate facies at 12–16 kbar in the Early Cretaceous and partly to completely retrogressed to amphibolite facies. These lower crustal basement rocks are overlain by the low-angle 100–300 m-wide Doubtful Sound extensional décollement shear zone that was responsible for loss of c.15 km of crust by attenuation.

The cover rocks of the metamorphic core complex above the décollement comprise a mid–crustal section in Fiordland that is made of Paleozoic 5–9 kbar, amphibolite facies metasedimentary rocks, ortho- and para-amphibolites (including tuffs and arc-type lavas), calc-alkaline granitic rocks, the Mount George layered gabbro, and the 349±5 Ma Black Giants anorthosite (600–800 m thick and up to 40 km long), which is a layered intrusion consisting predominantly of anorthosite (An80–90), gabbro anorthosite and amphibolite. The Black Giants anorthosite complex was intruded into and is now conformably overlain by a 1 km-thick metavolcanic amphibolite; these cover rocks formed in a magmatic arc on the margin of Gondwana. Gibson and Ireland (1999) pointed out that the Black Giants anorthosite is compositionally and lithologically “indistinguishable from the Fiskenneset anorthositic complex”, and suggested that it forms an excellent modern analogue for development of the Fiskenneset anorthosite in an Andean-type arc that was subsequently buried in the deep crust thickened by addition of subduction-generated tonalites and granodiorites.

The upper crust of Fiordland includes the 143–137 Ma Darran complex that consists of calc-alkaline I-type, unmetamorphosed, hornblende-bearing gabbronorite, gabbro and diorite (Muir et al., 1998) that formed in the magma chamber of an island arc by melting of a mantle wedge above a subducting slab. As a result of incipient continental rifting and creation of the back-arc Tasman Sea in the Early Cretaceous, formation of the Fiordland metamorphic core complex uplifted the deep level crust by c. 30 km to within 12 km of the surface by 77 Ma.

The geology and evolution of Fiordland is instructive, because it demonstrates how it evolved in successive stages from island arc to active continental margin to metamorphic core complex, and how extensional tectonics juxtaposed different crustal levels during exhumation. Not only is the Black Giants anorthosite complex very similar to the Fiskenneset anorthosite complex, but the unmetamorphosed island arc Darran gabbronorite complex is very similar to the granulate facies Chilas noritic complex that forms a key component of the Cretaceous Kohistan island arc in the Himalayas of Pakistan.

6.2. Kohistan, N. Pakistan

Kohistan provides a 100–150 km section from greenschist to granulite facies through an accreted Cretaceous island arc in Pakistan (Coward et al., 1982), which continues as the Dras arc in Ladakh of India (Clift et al., 2002). The crustal growth by magmatic accretion is represented by three stages of emplacement of granitic rocks. A mature island arc that formed offshore contains c. 102 Ma plutons of tonalite and diorite. Closure of the back-arc basin to the north accreted the arc to the Asian continental margin, after which the arc occupied the leading edge of the Asian plate, and further northward subduction of the Tethyan oceanic plate gave rise to the intrusion of a c. 54–40 Ma, Andean-type tonalitic batholith. After collision by India and further structural thickening crustal melt granites were emplaced from 34 Ma to 30 Ma (Petterson and Windley, 1985). The Chilas stratiform
complex (10–15 km-thick and >350 km-long) of norites, noritic gabbros and minor chromite-layered dunites formed in the sub-island arc magma chamber during initial rifting of the island arc, and it contains roof pendants of metavolcanic amphibolites derived from the overlying arc (Khan et al., 1993; Jagoutz et al., 2007); being at the sheared base of the arc it was metamorphosed in the granulite facies during arc–continental collision.

The geological history and crustal growth of the Kohistan arc is fundamentally comparable to that of the Mesozoaic crust of West Greenland, in so far as both went though three stages of compositionally similar magmatic evolution: early island arc volcanism with a sub-arc layered anorthositic-noritic complex, Andean-type tonalitic plutonism, and late crustal melt granite emplacement.

6.3. The North and South American Cordillera

Whereas the upper crustal levels of the batholiths of the Andes and Cordillera of western America typically with diapirc granitic plutons are not comparable to the deeper crustal levels exposed in West Greenland, the deep sections of these mountain ranges do present remarkably analogous profiles. In Patagonia and British Columbia (the Coast plutonic complex) sheets of calc-alkaline hornblende-bearing tonalite, trondhjemite, granodiorite or diorite intruded along thrusts and shear zones, became foliated to gneisses, were metamorphosed in the high amphibolite or granulite facies at 9–10 kbar, and underwent partial melting to produce migmatites (Hutchinson, 1967; Bartholomew and Tarney, 1984). These high-grade gneissic rocks and structures developed during crustal thickening in active continental margins in association with sub-horizontal thrust-nappe tectonics in the Mesozoic or Tertiary. Confirmation of the gneissic character of the deep Andes comes from crustal xenoliths in Columbia (Weber et al., 2002) that consist of hornblende tonalitic gneisses, granulite facies gneisses, pyroxenites, pyribolites (two pyroxene amphibolites) and pyriclasites (two pyroxene–biotite–plagioclase schists), the last two similar to those described with the same terminology by Berthelsen (1960) from the Toqqupas Nuna in the lower zone of the Fiskefjord block (Figs. 1 and 12).

The Cretaceous western Peninsular Ranges batholith in southern California is made up of hundreds of mid-crustal plutons of hornblende–biotite tonalite and granodiorite, intruded into island arc-type volcanic and volcanioclastic rocks, some emplaced as foliated and gneissose sheets along ductile shear zones during synkinematic amphibolite facies metamorphism (Cowan and Bruhn, 1992). Tonalitic gneisses and tonalites contain bodies of locally layered and graded norite, gabbro, gabbro anorthosite and anorthosite that consist of very calcic cumulate plagioclase An20–96 (anorthite–bytownite cores have labradorite rims) and inter-cumulus magmatic hornblende; they were derived from a high–alumina basaltic magma (Miller, 1937; Nishimori, 1974; Walawender and Smith, 1980). According to Silver and Chappell (1988) this western batholith formed as a root of a primitive island arc on oceanic lithosphere at a convergent plate margin. Lee et al. (2007) constructed a model for generation of the Peninsular Ranges batholith as follows: in the Triassic a fringing island arc was created off the Paleozoic continental margin of North America, in late Jurassic to early Cretaceous times the back-arc basin closed and this fringing arc was accreted onto the edge of the North American continent, and in the early Cretaceous farther eastwards subduction gave rise to new basaltic arc magmas, which gave rise to the main tonalitic–granodioritic batholiths that were emplaced into the accreted island arc on the active continental margin. They went on to suggest that this environment was applicable along the entire Cordilleran margin from Alaska to Chile, and as a general mechanism even to the Andean. One example is the Jurassic Border Ranges complex in Alaska, which is composed of ultramafic cumulates, massive and cumulate gabbro-norites that represent the plutonic core of an intra-oceanic island arc, and overlying andesitic volcanic rocks (Burns, 1985). The gabbroic rocks contain calcic plagioclase (An75–100), iron-rich pyroxene and magnetite. These plutonic and volcanic island arc rocks were intruded by calc-alkaline plutons of tonalite, granodiorite and quartz diorite in batholithic proportions (Burns, 1985).

In the Cordilleran Occidental of the Peruvian Andes late Cretaceous plutons of the tonalitic–granodioritic Coastal Batholith were emplaced into volcanic rocks of the Albican Casma Basin and their broadly coeval basic plutonic complexes. The 6 km-thick Casma volcanic pile consists of pillow-bearing basalts, hyaloclastites, basaltic andesites, dacites and rhyolites; the volcanic rocks have low Zr/Y vs. low Zr values characteristic of oceanic arcs (Petford and Atherton, 1995). The basic complexes comprise anorthosite (with cumulate plagioclase up to An94) and gabbro with inter-cumulus hornblende that is secondary after clinopyroxene, but derived from a late volatile-rich residual melt (Mullan and Bussell, 1977). The complexes occur in layers and lenses up to 5 km wide, and 40 km long, but grouped in “clusters indicating the former presence of substantial bodies that prior to fragmentation may have approached 1000 km2 in area” (Regan, 1985). The tonalitic–granodioritic plutons were emplaced into the volcanic–plutonic rocks with the result that many of the latter now occur as lenses and inclusions within the tonalites and granodiorites. Similarities in mineralogy, and major and trace element patterns suggest that the volcanic rocks represent the liquid fraction after cumulate crystallization of the basic complexes, both generated from tholeiitic magmas in the mantle. The tonalitic–granodioritic plutons were not cogenetic with the oceanic arc volcanic and basic complexes, they were just intruded into them along the same axis of the overall batholith. Mullan and Bussell (1977) compared these Peruvian basic complexes with similar Cretaceous complexes in the West Mexican batholith that include primary magmatic hornblendites, and layered gabros with cumulate plagioclase and inter-cumulus magmatic hornblende that indicate that these complexes crystallized under a very high fluid pressure; extrusive equivalents are pyroclasitic and basaltic–andesitic volcanic rocks. The Cordilleran batholiths in North and South America provide a viable modern analogue for the arc-generated crustal growth in West Greenland (Windley and Smith, 1976).

6.4. Japan

The Japanese islands that have formed by arc accretion since the late Paleozoic provide important information on arc–arc collision and crustal growth in the Cenozoic. Kimura (1996) pointed out that such arc–arc junctions promote the growth of continental crust at a rate that is higher than that within normal subduction zones. Collision tectonics in Japan is operating at three arc–arc junctions: Kyushu, Central Japan and Hokkaido. In Kyushu the Kyushu–Palau Ridge is at an incipient stage in its collision into and below SW Japan. Biotite–hornblende and hornblende tonalite, trondhjemite, biotite granodiorite and quartz diorite dredged from a seamount on the northern tip of the Ridge have K–Ar ages of 38–37 Ma and were produced by fractional crystallization from basaltic magma during the early stage of oceanic island arc formation (Haraguchi et al., 2003).

In Central Japan the Izu–Bonin arc began its collision into and under the Honshu arc in the late Miocene and is ongoing (Soh et al., 1998). The collision has produced a major syntaxis, a major imbricated thrust stack, and has upthrust onland the middle crust of the Izu–Bonin arc which consists of 10–5 Ma Tanzawa tonalite that ranges from tonalite to quartz diorite and diorite in which the mafic minerals are two pyroxenes, biotite and hornblende, and cumulate textures are distinctive (Haraguchi et al., 2003). Kawata and Arima (1998) concluded that the parental magma of the Tanzawa tonalite was andesitic and was produced in a late stage of arc formation by partial melting of basaltic (amphibolitic) lower crust. The mid–crustal tonalite is overlain by basaltic and andesitic tuff, volcanic breccia and lavas that make up the upper crust of the arc, and is underlain by a lower crust of gabbro according to geophysical data (Haraguchi et al., 2003).
In Hokkaido the Kuril arc began its collision into the paleo-Eurasian continental margin of Japan in the late Eocene before the Japan Sea and Kuril Basin opened in the Miocene. Semi-continuous collision that is ongoing has upthrust a crustal section of the Kuril arc that displays an upward sequence from granulate facies to zeolite facies (Kimura, 1996). According to Komatsu et al. (1983, 1989) and Shiba (1988) the island arc-type crust at Hidaka has the following components from bottom to top: Iherzolite and cumulus peridotite derived from the upper mantle; a lower mafic crust (now mainly amphibolite) comprising granulate facies, tholeiitic olivine gabbro–norite–diorite and olivine amphibolite facies calc-alkaline gabbro–diorite; a middle crust of biotite–hornblende gneisses and gneismites derived from lower garnet–orthopyroxene tonalite, middle tonalite–granodiorite with cordierite–muscovite, and upper granite–granodiorite; and an upper crust of weakly to non-metamorphosed pelites and psammites that may be of forearc origin. Sub-horizontal shearing took place throughout the different crustal levels at granulite to greenschist grades. The structure of the Hidaka island arc section was confirmed by seismic reflection and gravity data by Arita et al. (1998) who compared it to the Kohistan section, described above.

In summary, the arc–arc junctions in Japan demonstrate how tonalitic gneisses are generated in the deep crust of an arc and how a modern island arc crustal section is built up.

7. Conclusions

We have shown that the Archean craton of West Greenland consists of six crustal blocks that were largely generated by growth of island arcs, followed by growth of active continental margin arcs and tectonic accretion. Each of these blocks constitutes a complete and representative section of the Archean crust within its exposed limits of crustal depth. The existence of similar crustal blocks with comparable key rock components and complexes at Kapuskasing, Pikwitonei and South India suggests that this was possibly the most important means of generating the continents in the Meso- to Neoarchean. The crustal sections of Paleozoic, Mesozoic and Cenozoic age demonstrate that the principal mechanisms of growth of the continents from arc formation through accretion to collision have not changed fundamentally in the last 3 Ga.

There is a common thread through several of the tectonic models addressed in this paper. The Kohistan island arc in Pakistan was accreted to the southern margin of the Asian continent, following closure of the back-arc basin to the north, and in consequence magmatism changed from oceanic to continental leading to emplacement of the tonalitic–granodioritic Trans-Himalayan or Kangdese batholith into the island arc in an Andean-type continental margin arc (Petterson and Windley, 1985; Clift et al., 2002). The Peninsular Ranges batholith of California started with an off-shore island arc, and following closure of the back-arc basin to the east, the arc was accreted to the continental margin of North America, enabling a change in magmatism from oceanic to continental and the emplacement of the tonalitic–granodioritic batholith into the active continental margin (Lee et al., 2007). In a comparable manner the Casma oceanic arc in Peru was intruded by the main tonalite–granodiorite batholith of the Andes (Petford and Atherton, 1995).

Moreover, a common feature of most of the quoted arc-root layered complexes of all ages is the presence of relatively late (in the fractional crystallization) calcic anorthosites with bytownite–anorthite plagioclase, of magmatic intercumulus hornblende in leucogabbros, and a common interpretation of their origin was crystallization from a hydrous magma. Such causation was inter-related by Yoder (1969) and Yoder and Tilley (1962), confirmed and foreseen by Müntener et al. (2001) in experimental studies related specifically to high water concentrations in arc magmas, and demonstrated by Clæson and Meurer (2004), Müntener and colleagues demonstrated that, in crystallization experiments at 1.2 GPa corresponding to conditions in the lower crust of a magmatic arc, high water concentrations (>3%) enable ultramafic cumulates such as at first dunites and then olivine-free, high Mg # pyroxenites and wehrlites to form early in the crystallization sequence before plagioclase saturation occurs, and also enables igneous amphibole to form in early ultramafic rocks, as the intercumulus phase in leucogabbros, and in late anorthosites. It is the combined effect of high total pressure and high H2O content in magmas at the base of an arc crust that retains calcium in the melt and suppresses plagioclase until it crystallizes as relatively late calcic anorthosite. These relations provide experimental confirmation for the presence of olivine-free websterites and orthopyroxenites with separate dunite layers in layered complexes in Fiskefjord (Garde, 1997), for separate dunite–wehrlite sills in lower arc crust of the Ivisaartog belt (Polat et al., 2008), for analogous interlayered rocks in the lower levels of the Chilas complex (Burg et al., 1998) that formed in the lower crust of the Kohistan island arc as proposed by Müntener et al. (2001), and for intercumulus hornblende in leucogabbros and post-gabbro/leucogabbro calcic anorthosites in the Fiskefjord complex (Windley et al., 1973; Myers, 1985).

In the modern examples quoted above, island arcs were able to accrete to margins of continents, because continents existed in the Mesozoic and Cenozoic, but in West Greenland, or anywhere else worldwide, there were no major continents in existence in the Mesozoic and Cenozoic. However, e.g. in West Greenland and in the Pikwitonei region of Canada there were many island arcs available to mutually accrete and amalgamate into microcontinents, around which other fringing island arcs could begin the conversion to continental arc magmatism. These speculations imply that the processes of continental growth from island arc magmatism to continental arc/continental magmatism were broadly similar throughout much of Earth history (Windley and Smith, 1976; Tarney and Windley, 1977; Lee et al., 2007), and for this reason tectonic blocks of different ages can be found with similar crustal sections and with broadly comparable components and age relations. Both experimental work and geochemical studies have indicated that there was a change from generation of tonalite–trondhjemite–granodiorite-dominated continental crust at the end of the Archean, as direct partial melting of the subducting slab became less feasible as the Earth became cooler (e.g. Martin, 1994). The Archean gneisses of the North Atlantic craton and the Paleoproterozoic Julianehåb batholith in South Greenland actually provide a good example of this geochemical change. However, it would seem that this transition did not have other fundamental influences on the magmatic and tectonic accretion of new continental crust over time as discussed in this paper.

The results of this study, based on West Greenland and a variety of Archean and modern analogues, suggest that continental growth has commonly taken place by processes dominated by arc generation, as envisaged by Rudnick (1995). From this it is evident that, in order to form a continent, crustal growth potentially goes through a minimum of four stages of development, two magmatic and two structural ones (Fig. 15):

1. An offshore oceanic island arc comprising an upper volcanic crust, and a lower crust or magma chamber consisting of a layered complex that may include anorthosites, gabbros, norites, and ultramafic rocks.

2. Accretion of the island arc either in the Archean to an already accreted collage or microcontinent of island arcs, or in Proterozoic and later time to a continent.

3. In the new active continental margin setting, emplacement of voluminous TTGs in various combinations and compositions, which leads to massive thickening of the crust.

4. Collision of the active continental margin to other crustal blocks to form a final stable continent.

Our model lends support to the viability of a plate tectonic interpretation of the evolution of the Greenlandic crustal sections.
Identification of the six different blocks exhibiting sections through arc-generated crust opens up a new perspective to understand the evolution of the Archean craton of West Greenland. Future research will be able to focus on key problems ultimately and hopefully concerned with how the crust grew in terms of plate tectonic mechanisms.

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