



# Geology of the Eoarchean, >3.95 Ga, Nulliak supracrustal rocks in the Saglek Block, northern Labrador, Canada: The oldest geological evidence for plate tectonics



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## ABSTRACT

The Earth is a unique planet, which has been highly evolved, diversified and complicated through geologic time, and underwent many key events, including giant impact, magma ocean, core formation, large-scale mantle differentiation and late heavy bombardment, especially in its dawn. But, our knowledge of early Earth is limited due to the lack of the Hadean supracrustal rocks. The supracrustal rocks with the Eoarchean ages provide key evidence for the Earth's early evolution, but few supracrustal rocks have been comprehensively investigated. Therefore, we mapped in seven areas of the Saglek Block, northern Labrador, where ancient supracrustal sequences are interleaved with a diverse assemblage of orthogneisses. Early studies suggested that some of them have the Mesoarchean ages because of the lack of the Mesoarchean Saglek dyke, but we found the Saglek dykes in the areas to recognize the Eoarchean Nulliak supracrustal rocks and Uivak Gneiss in all the areas. Recent reassessment of U–Pb dating and cathodoluminescence observation of zircons from the oldest suites of the Uivak Gneiss showed that the Uivak Gneiss has the Eoarchean age, >3.95 Ga, and forms the Iqaluk–Uivak Gneiss series. Because our geological survey clearly showed that the Iqaluk–Uivak Gneisses were intruded into the Nulliak supracrustal belts, the Nulliak supracrustal rocks are the oldest supracrustal rock in the world. The supracrustal belts consist of piles of fault-bounded blocks, which are composed of the ultramafic rocks, mafic rocks and sedimentary rocks in ascending order, similar to modern ocean plate stratigraphy (OPS). In addition, small-scale duplex structures are found over the areas. The presence of duplex structure and OPS indicates that the >3.95 Ga Nulliak supracrustal belts originate from an accretionary complex. The presence of the accretionary complex, ophiolite and granitic continental crust provides the oldest evidence for the plate tectonics on the early Earth.

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## 1. Introduction

The Earth underwent many drastic events such as giant impact, magma ocean, core formation, large-scale mantle differentiation, early crust formation, late veneer and late heavy bombardment in her dawn and finally reached the beginning of the plate tectonics, in contrast to the other planets. But, we cannot obtain the evidence for those events in the Earth because so far, the oldest dated rock is 4.03 Ga from the Acasta Gneiss Complex of the Slave Craton, northwestern Canada (Bowring and Williams, 1999), and the oldest supracrustal rock from the 3.83 Ga Akilia island supracrustal rock in the Itsaq Gneiss Complex, southern West Greenland (Mojzsis and Harrison, 2002a). The evidence

for early evolution of the Earth in the Hadean has been obtained from two methods as well as the extraterrestrial materials. One is obtained from inherited and detrital zircons in the Archean orthogneiss and conglomerates: up to 4.4 Ga detrital zircons in the Jack Hills and Mt Narryer conglomerates, Narryer Complex, Western Australia (e.g. Compston and Pidgeon, 1986; Holden et al., 2009; Wilde et al., 2001), a 4.2 Ga inherited zircon, in the Acasta Gneiss Complex (Iizuka et al., 2006), and a 4.08 Ga inherited zircon in a 3.83 Ga orthogneiss in the Akilia island, southern West Greenland (Mojzsis and Harrison, 2002a). The geochemical characteristics such as rare earth element (REE) patterns, Ti-in-zircon thermometry and mineral inclusions support the formation of granitic magma, and their low  $\epsilon_{\text{Hf}}$  values suggest the continental formation in the early Hadean, >4.2 Ga (Amelin et al., 1999, 2000; Harrison et al., 2007; Iizuka et al., 2009; Maas et al., 1992; Watson and Harrison, 2005). Another comprises geological and geochemical evidence from

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the Eoarchean geologic terranes including the Acasta Gneiss Complex (Bowring and Williams, 1999; Iizuka et al., 2007), the Itsaq Gneiss Complex, southern West Greenland (e.g. Nutman et al., 1996), the Mt. Sones, Antarctica (e.g. Harley et al., 2007), the Nuvvuagittuq supracrustal belt in Quebec, Canada (Cates and Mojzsis, 2007; O'Neil et al., 2008), the Anshan area, North China Craton (Liu et al., 1992, 2007), and the Saglek Block, Labrador, Canada (Bridgwater et al., 1975; Collerson et al., 1991; Schiøtte et al., 1989b).

The supracrustal rocks with the Eoarchean ages also provide key witnesses of the early evolution such as tectonic setting (Furnes et al., 2014; Polat et al., 2002), plate tectonics (Friend et al., 1988; Komiya et al., 1999; McGregor et al., 1991; Nutman and Friend, 2009; Nutman et al., 2009), early differentiation (Boyet et al., 2003; Caro, 2011; Caro et al., 2003, 2006; Iizuka et al., 2010; Moynier et al., 2010; O'Neil et al., 2008, 2009, 2012; Rizo et al., 2011, 2012, 2013; Roth et al., 2013; Sharma et al., 1996), mantle evolution (Bennett et al., 1993; Collerson et al., 1991; Komiya et al., 2004; Vervoort et al., 1996), early life (Craddock and Dauphas, 2011; Czaja et al., 2013; Mojzsis et al., 1996; Ohtomo et al., 2014; Rosing, 1999; Ueno et al., 2002; Yoshiya et al., 2015), composition of seawater (Bolhar et al., 2004; Nutman et al., 2010) and the Late Heavy Bombardment Event (Schoenberg et al., 2002). But, few Early Archean supracrustal rocks are preserved only in the Itsaq Gneiss Complex, the Nuvvuagittuq greenstone belt and the Saglek Block. The Akilia association in the Itsaq Gneiss Complex is one of the oldest supracrustal rocks in the world: the 3.81 Ga Isua supracrustal belt (Crowley, 2003; Nutman and Friend, 2009) and the 3.83 Ga banded iron formation in the Akilia Island (Mojzsis and Harrison, 2002a). The Nuvvuagittuq supracrustal belt comprises three main units: a cummingtonite-rich amphibolite, previously referred to as the “faux-amphibolite”, gabbroic and ultramafic rocks, and clastic and chemical sediments including banded iron formation (e.g. O'Neil et al., 2007). They are intruded by 3.66 Ga tonalite (David et al., 2009). Recently, O'Neil et al. (2008) suggested that the mafic and ultramafic rocks may be as old as 4.28 Ga based on a pseudo-isochron of  $^{147}\text{Sm}/^{144}\text{Nd}$ – $^{142}\text{Nd}/^{144}\text{Nd}$ . However, the interpretations of the pseudo-isochron of  $^{147}\text{Sm}/^{144}\text{Nd}$ – $^{142}\text{Nd}/^{144}\text{Nd}$  are highly controversial (Andreasen and Sharma, 2009; O'Neil et al., 2008, 2009, 2012; Roth et al., 2013). On the other hand, zircons recovered from the trondhjemitic gneisses, which cut the supracrustal rocks, provide relatively young ages at  $3751 \pm 10$  Ma (Cates and Mojzsis, 2007) and  $3817 \pm 16$  Ma (David et al., 2009). Therefore, it is important to find old supracrustal rocks, cut by granitoid intrusion with well-determined Eoarchean ages.

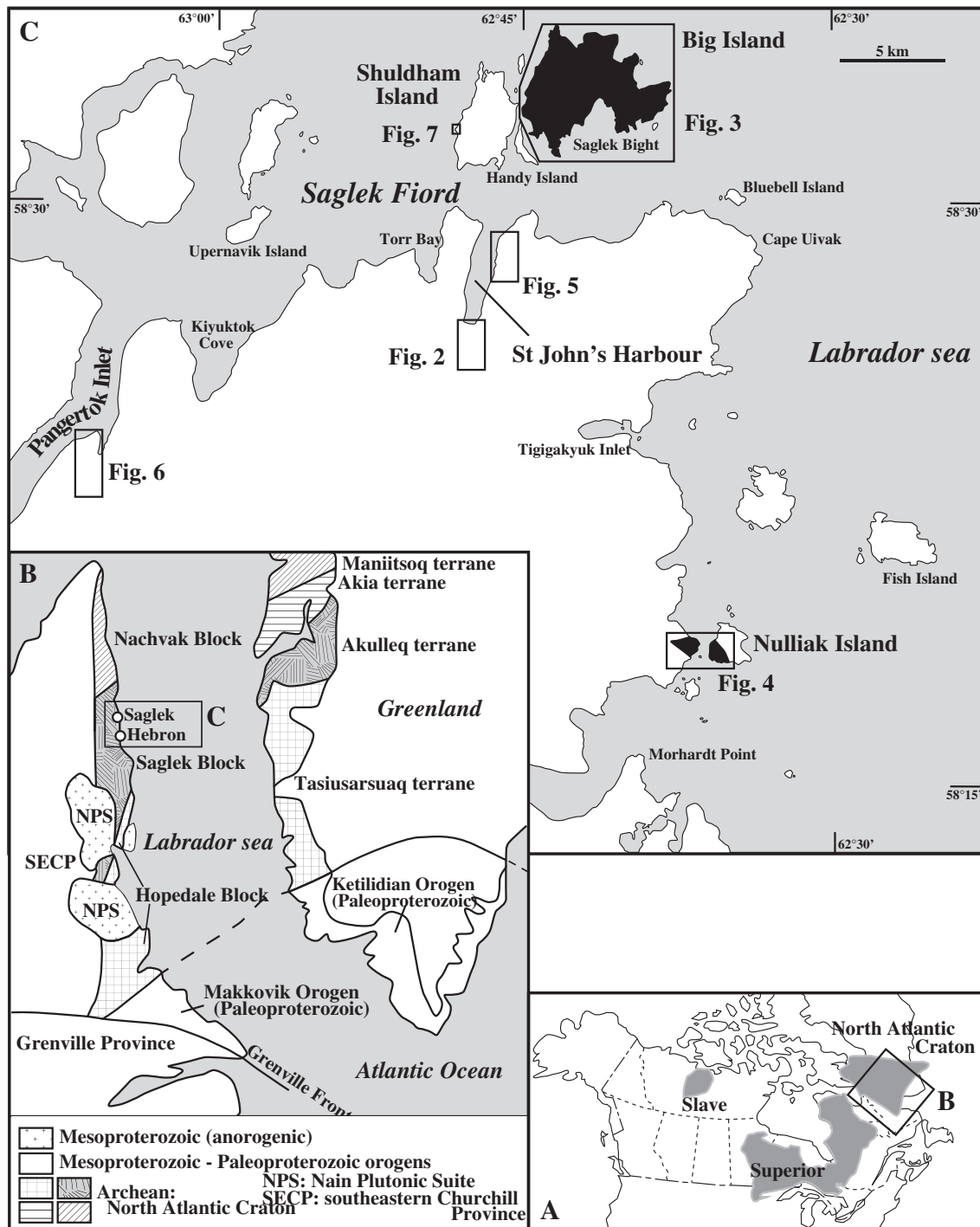
In the Saglek Block, northern Labrador, a plethora of supracrustal belts and blocks are distributed (Bridgwater and Collerson, 1976; Bridgwater et al., 1975, 1978; Collerson et al., 1976a). These were assigned in a relative chronology into two suites based on the cross-cutting relationship with the Mesoarchean Saglek dykes: the Eoarchean Nulliak supracrustal rocks and the Mesoarchean Upernavik supracrustal rocks (Bridgwater and Collerson, 1977; Bridgwater et al., 1990; Collerson and Bridgwater, 1979; Nutman and Collerson, 1991; Nutman et al., 1989). The Nulliak supracrustal rocks comprise chemical and clastic sediments, amphibolite and ultramafic rocks (Bridgwater et al., 1975; Collerson et al., 1991; Komiya et al., 2012), equivalent to the Akilia association in West Greenland (Bridgwater and Schiøtte, 1991; Nutman and Collerson, 1991). The geochronology of the Saglek Block was equivocal, especially for the ages of precursors of the Uivak orthogneisses. Previous works proposed the existence of several generations of granitoid magmatism in the Saglek area based on intrusive relationships: the Nanok Gneiss, the Uivak I and II Gneisses, the Lister Gneiss and the younger granitic intrusions. The Uivak I and II Gneisses are defined by intrusion of the Saglek dykes whereas the Lister Gneiss has no geological evidence for the intrusion (Bridgwater et al., 1975). The U–Pb zircon ages of the Lister Gneiss range from 3200 to 3400 Ma (Schiøtte et al., 1989b). SHRIMP dating of the zircons from the Uivak I Gneisses suggested that they were derived from the

Eoarchean protoliths with a  $^{207}\text{Pb}/^{206}\text{Pb}$  age of 3732 Ma based on U–Pb dating of the U-poor euhedral zircons (Collerson, 1983; Schiøtte et al., 1989b). Krogh and Kamo (2006) reinterpreted the low-U cores with/without zoning as inheritance of igneous type grains, and oscillatory-zoned overgrowths on the cores as magmatic growth, and obtained the U–Pb zircon ID-TIMS age of  $3634 \pm 31$  Ma for the Uivak Gneisses. In addition, the presence of the Nanok Gneiss is quite controversial. The  $>3.9$  Ga Nanok Gneiss was proposed based on the presence of  $>3.9$  Ga zircons (Regelous and Collerson, 1996), but the zircons were also interpreted as inherited (Collerson, 1983; Krogh and Kamo, 2006; Schiøtte et al., 1989a,b). Such controversy is possibly due to poorly understanding the zircons in the highly metamorphosed orthogneisses and geology of the Saglek Block. Recent reassessment of U–Pb dating and cathodoluminescence imaging of zircons from the oldest suites of the Uivak Gneiss showed that the Uivak Gneiss has the Eoarchean age,  $>3.95$  Ga (Shimojo et al., 2012, 2013, 2015). Besides, geological reappraisal of the Saglek Block is necessary because it has been a long time since the detailed geological work in 1970s.

The timing of the beginning of the plate tectonics is one of the most important issues in geological aspect of history of the Earth. The plate tectonics in the Eoarchean were suggested based on geological evidence of accretionary complex (Komiya et al., 1999), ophiolite (Furnes et al., 2007) and collision–amalgamation of terranes (Crowley, 2003; Nutman and Friend, 2009) and petrological study (Polat and Hofmann, 2003; Polat et al., 2002) for the 3.7–3.8 Ga Isua supracrustal belt. In addition, the presence of detrital zircons derived from granitoid magmas, namely formation of continental crust, suggests that plate tectonics go back to even the Hadean (Harrison, 2009; Valley et al., 2002). However, Hamilton (1998a,b) objected to such an Archean plate tectonic model because of the lack of ophiolite, blueschist and Franciscan-type mélange in the Archean greenstone belts. Recently, Condie and Kröner (2008) proposed a more comprehensive set of indicators for ancient plate tectonics including ophiolite, arc assemblages, accretionary complexes/ocean plate stratigraphy (OPS), foreland basins, blueschist, passive continental margins, continental rifts (Wilson cycle), subduction-related mineral deposits, ultrahigh-pressure metamorphism (UHP), paired metamorphic belts, large transcurrent faults/suture zones, orogens, paleomagnetism, igneous rocks with arc-like geochemistry, continental crust and mantle recycling. And, they confirmed that plate tectonics were already operated in the Mesoarchean because most of them were found in the Mesoarchean to Neoarchean terranes. However, continental formation basically occurs only in active plate margins; thus the passive continental margin and large transcurrent faults/suture zones were insignificant before formation of large landmass. Higher mantle temperature (e.g. Herzberg et al., 2010; Komiya, 2004; Komiya et al., 2002b, 2004; Nisbet et al., 1993; Ohta et al., 1996) and hotter subduction (Grambling, 1981; Hayashi et al., 2000; Komiya et al., 2002a; Kusky and Polat, 1999; Maruyama et al., 1996) in the Archean prevent the formation and occurrence of blueschist and ultrahigh-pressure metamorphism as well as the paired metamorphism with distinct metamorphic facies series. In addition, because the Eoarchean greenstone belts underwent severe metamorphism, geochemical and geomagnetic features, sensitive to the secondary modification, are also secondary indicators in practice. As a result, ophiolites, accretionary complexes/OPS and granitic continental crusts are more practical signatures for the ancient plate tectonics, and it is important to study these indicators in the Eoarchean greenstone belts.

## 2. Geological outline of the Saglek–Hebron area

The Saglek–Hebron area is located in the northeastern part of the Labrador Peninsula, northeast Canada (Fig. 1A), and belongs to a coastal, central part of the early Archean terrane, called the Saglek Block. The block is the west end of the North Atlantic Craton, which extends from Scotland through the southern part of Greenland to Labrador (Fig. 1B, Bridgwater and Schiøtte, 1991; Bridgwater et al., 1973;



**Fig. 1.** (A) Distribution of the Archean cratons in northern North America: Slave, Superior and North Atlantic Cratons, respectively. The Saglek Block of the northern Labrador, Canada is a part of the North Atlantic Craton. The box shows the location of (B). (B) Distribution of the Archean terranes in the western part of the North Atlantic Craton. The Archean rocks are present in both sides of the Labrador Sea, and the Saglek Block is equivalent to the Akulleq terrane in the West Greenland (Bridgwater and Schiøtte, 1991). The box shows the location of our study area (C). (C) A map of the Saglek–Hebron area. (B) Modified from Wasteneys et al. (1996).

Gower and Ryan, 1986; Sutton et al., 1972; Wasteneys et al., 1996). The block is underlain by the Early to Late Archean suites including orthogneisses, metasedimentary rocks, metavolcanic rocks, ultramafic rocks, mafic dikes and young granite (Bridgwater and Schiøtte, 1991; Bridgwater et al., 1975; Collerson, 1983; Collerson and Bridgwater, 1979; Komiya et al., 2012; Schiøtte et al., 1989a). The orthogneisses and supracrustal rocks underwent granulite to amphibolite facies metamorphism at 2700 to 2800 Ma (Collerson and Bridgwater, 1979). A major, sub-vertical, NS-trending fault, the Handy fault,

extends from Handy Island through St John's Harbour to the south (Collerson et al., 1982; Schiøtte et al., 1993), and separates two parts: western segment where the Early Archean rocks underwent the granulite facies metamorphism, and eastern segment where sub-granulite facies rocks are sporadically preserved, respectively. It is possibly because of different crustal levels (Bridgwater et al., 1975) or accretion and juxtaposition of two terranes with different tectonothermal histories (Bridgwater and Schiøtte, 1991; Schiøtte et al., 1990).

In this region, the orthogneisses are predominant, and account for about 80% (Collerson et al., 1982; Nutman and Collerson, 1991; Schiøtte et al., 1989a), but supracrustal rocks and mafic to ultramafic mid-Archean dykes, called the Saglek dyke, are also ubiquitously distributed (Bridgwater et al., 1975). Although most of the dykes cannot be described in a geological map because the dykes are thin from some centimeters to meters thick and are highly deformed in most places, the Saglek Block rocks were classified into two groups based on the geological relationship with the Saglek dykes (Bridgwater et al., 1975; Hurst et al., 1975), analogous to the Ameralik dykes in the Itsaq Gneiss Complex (McGregor, 1973). The orthogneisses, older than the Saglek dyke, are called as the Uivak Gneisses, and include the Uivak I and II suites. The Uivak II Gneisses are a suite of deformed feldspar megacrystic granodioritic gneisses with simple fabrics in contrast to the dominantly composite fabrics of the Uivak I Gneisses. The Uivak I Gneisses are widely distributed throughout the Saglek–Hebron block (Bridgwater et al., 1975; Collerson and Bridgwater, 1979; Collerson et al., 1976a). They are grayish in color and relatively fine-grained, with tonalitic–trondhjemitic and granodioritic (TTG) compositions (Bridgwater and Collerson, 1976; Collerson et al., 1976a,b; Schiøtte et al., 1989a). The boundaries between the Uivak I and II gneisses are obscured in most places, but the Uivak II Gneiss intruded into the Uivak I Gneiss in relatively low strain outcrops (Bridgwater and Collerson, 1976; Collerson et al., 1976a). The age of the protolith of the Uivak I orthogneiss is still equivocal (Krogh and Kamo, 2006; Schiøtte et al., 1989a,b; Shimojo et al., submitted for publication). Previous works suggested that the Uivak Gneiss was formed around 3.73 to 3.74 Ga (Schiøtte et al., 1989a; Wasteneys et al., 1996), but the presence of >3.9 Ga zircons suggested pre-Uivak Gneiss rocks, Nanok Gneiss (Regelous and Collerson, 1996). Recent combination work of in-situ U–Pb dating and cathodoluminescence observation of zircons from the Uivak Gneiss indicated that the protoliths of some of Uivak Gneisses had older age, >3.95 Ga (Shimojo et al., 2012, 2013, 2015), much older than the conventional age (~3.7 Ga) given for the Uivak Gneiss. On the other hand, the petrography of the orthogneisses has tonalitic mineral assemblages, different from the Nanok Gneiss. Therefore, the Uivak Gneisses in the Saglek Block is named as Iqaluk ('char' in the Inuit language)–Uivak Gneiss series hereafter. On the other hand, the Lister Gneiss is defined as a post-Saglek dyke gneiss, and has the Mid-Archean age in the Saglek–Hebron area. The SHRIMP U–Pb dating of zircons from the Lister Gneiss in Lister Island obtained  $3235 \pm 8$  Ma (Schiøtte et al., 1989a). The field relation with the Uivak Gneisses is enigmatic: tectonically juxtaposed against the Early Archean "Uivak continent" (Schiøtte et al., 1990) or intrusive into the Uivak Gneisses (Collerson et al., 1976a; Schiøtte et al., 1993).

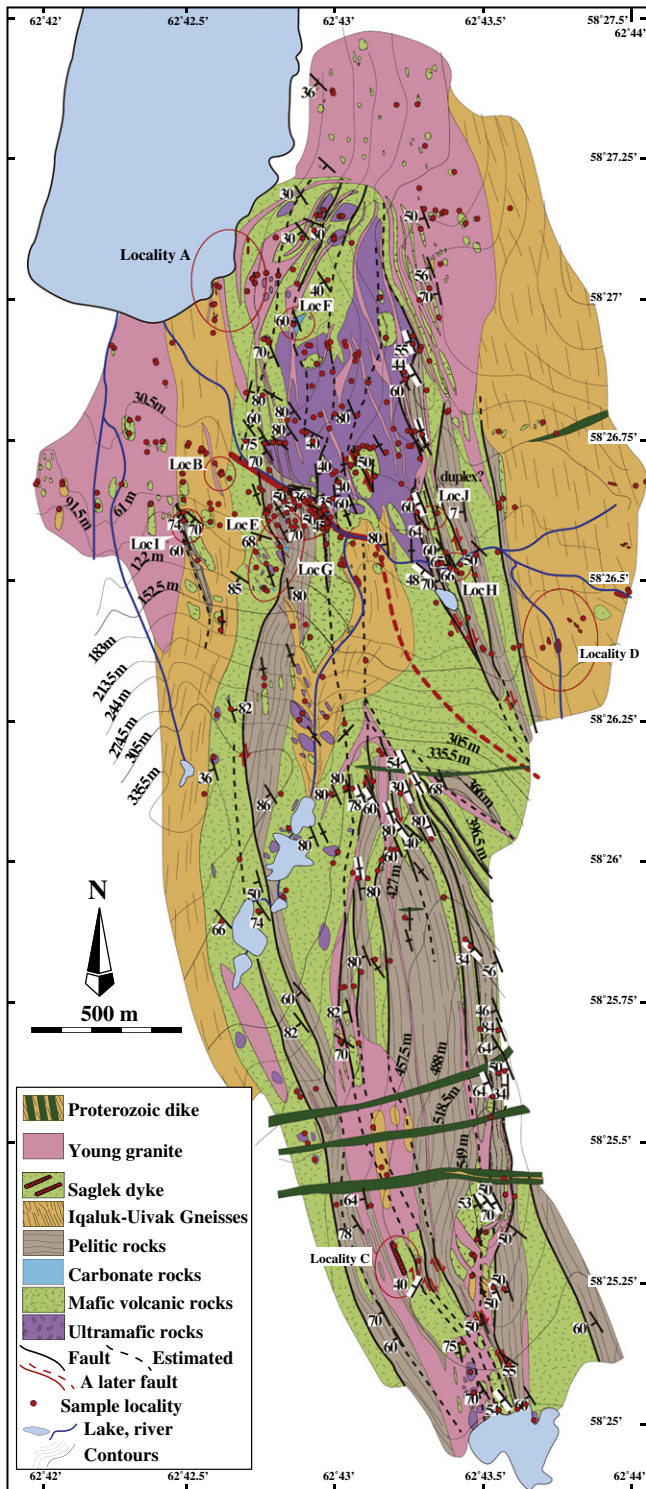
Supracrustal rocks in the Saglek Block are classified into two groups based on the relationship with the Saglek dyke: the pre-Saglek Nulliak and the post-Saglek Upernavik supracrustal rocks, respectively (e.g. Bridgwater et al., 1975; Collerson et al., 1976a). The Nulliak supracrustal rocks contain mafic and ultramafic rocks, gabbroic rocks, anorthosite, chemical sedimentary rocks including carbonate rocks, chert and banded iron formation (BIF), and clastic sedimentary rocks. On the other hand, the Upernavik supracrustal rocks comprise mafic and ultramafic rocks, chemical sedimentary rocks of carbonate rocks and chert, and clastic sedimentary rocks including quartzite, pelitic and psammitic rocks and conglomerate. Because the Saglek dykes are thin and highly deformed, and the supracrustal belts also contain amphibolites, it is difficult to recognize the Saglek dyke, especially in small supracrustal blocks. In those cases, the lithological assemblages, especially the presence of the BIF, are often used to recognize the Nulliak supracrustal rocks. Previous works showed that the Nulliak supracrustal rocks are present in the Nulliak Island and the opposite coast, Pangertok Inlet, and Fish Island (e.g. Ryan and Martineau, 2012). The boundaries between the Nulliak supracrustal rocks and Uivak Gneisses are commonly equivocal, but it is considered that the Uivak Gneiss intruded into the Nulliak supracrustal rocks in some places. Although the geological

evidence shows that the Nulliak supracrustal rocks are older than the Uivak Gneiss, the exact ages of the Nulliak supracrustal rocks are still ambiguous. Baadsgaard et al. (1979) reported that conventional U–Pb dating of zircons from the Nulliak supracrustal rocks gives dates of ~2500 Ma, and interpreted that the zircons have been completely reset by ~2500 Ma thermal events. Schiøtte et al. (1989b) showed that SHRIMP U–Pb dating of zircons from a garnet–biotite gneiss in Bluebell Island, interpreted as a metavolcanic rock of the Nulliak supracrustal rocks, yielded  $3776 \pm 8$  Ma, and interpreted the age as original magmatic age. Nutman and Collerson (1991) reported that zircons from a quartzite in 2 km south of St John's Harbour have maximum age of 3845 Ma to constrain the minimum age of the deposition of the Nulliak supracrustal rocks. Collerson et al. (1991) reported an Sm–Nd isochron age of  $4017 \pm 194$  Ma for the Nulliak metakomatiites.

### 3. Lithology and deformational structures in the Saglek Block

The Saglek Block comprises supracrustal rocks and several generations of orthogneisses. Figs. 2 to 7 show the detailed geological maps of the seven areas in the Saglek Block. The Eoarchean suite is intruded by mid-Archean mafic dikes (Saglek dyke), white granitic intrusions, and Proterozoic basaltic dikes. The Proterozoic dikes occur as linear bodies, and apparently lack deformation and recrystallization. Figs. 8 to 12 display the photographs of outcrops of the supracrustal rocks, orthogneisses and Saglek dykes, and microscopic observation of some supracrustal rocks. The Saglek dyke is characterized by mafic composition (Fig. 8B, D and F) and plagioclase megacrysts in some places (Fig. 8E). The occurrence of the Saglek dyke is critical because the presence is utilized to traditionally distinguish pre-Saglek suites of the Nulliak supracrustal rocks and Uivak Gneisses from post-Saglek suites of the Upernavik supracrustal rocks and Lister Gneisses, respectively (e.g. Bridgwater et al., 1975; Collerson et al., 1976a), analogous to the Ameralik and Tarsartoq dykes in the southern West Greenland (McGregor, 1973). However, the Saglek dykes are much thinner and more deformed than the Ameralik dykes (Fig. 8B, D, E and F). In addition, some of the Ameralik and Tarsartoq dykes escaped the Late Archean metamorphism and preserve the original, igneous minerals such as olivine, plagioclase, pyroxenes and brown hornblende (Komiya et al., 2004) whereas the Saglek dykes underwent high-grade metamorphism, as well as the amphibolites in the Nulliak supracrustal belt. As a result, it is difficult to distinguish the Saglek dykes from amphibolites in the supracrustal rocks so that the classification between the Nulliak and Upernavik supracrustal rocks based on the occurrence of the Saglek dykes is always ambiguous. Another complementary criteria of the presence of BIFs are often employed to select the Nulliak supracrustal rocks (Ryan and Martineau, 2012). Mafic and ultramafic rocks in another belts of the neighborhood of Nulliak supracrustal belts are also considered as parts of the Nulliak supracrustal rocks. On the other hand, pelitic and carbonate rocks were regarded as representatives of the Upernavik supracrustal rocks.

Field occurrence and U–Pb dating of the orthogneisses show that the Iqaluk–Uivak Gneisses are varied in their lithological, mineralogical and geochronological aspects. Fig. 8A shows the composite occurrence of mafic rocks of the Nulliak supracrustal rocks and Iqaluk–Uivak Gneisses in a small outcrop, and displays at least three generations of the Iqaluk–Uivak Gneisses, intruded into older mafic rocks. The oldest Iqaluk–Uivak Gneiss suite is a gray orthogneiss, and has clear banding, which is cut by white, deformed orthogneiss. The youngest, white and coarse-grained, Iqaluk–Uivak Gneiss suite cuts all of them. Fig. 8C shows the typical occurrence of Iqaluk–Uivak Gneisses, and at least three generations of them are recognizable. The oldest suite is dark gray to brown in color, and intruded by thin white intrusions. They occur as blocks within white banded-gneisses. Fig. 8D also shows the multiple generations of mafic and felsic igneous rocks. The first generation is composed of dark green, mafic rock-dominated supracrustal rocks, intruded by thin, highly deformed



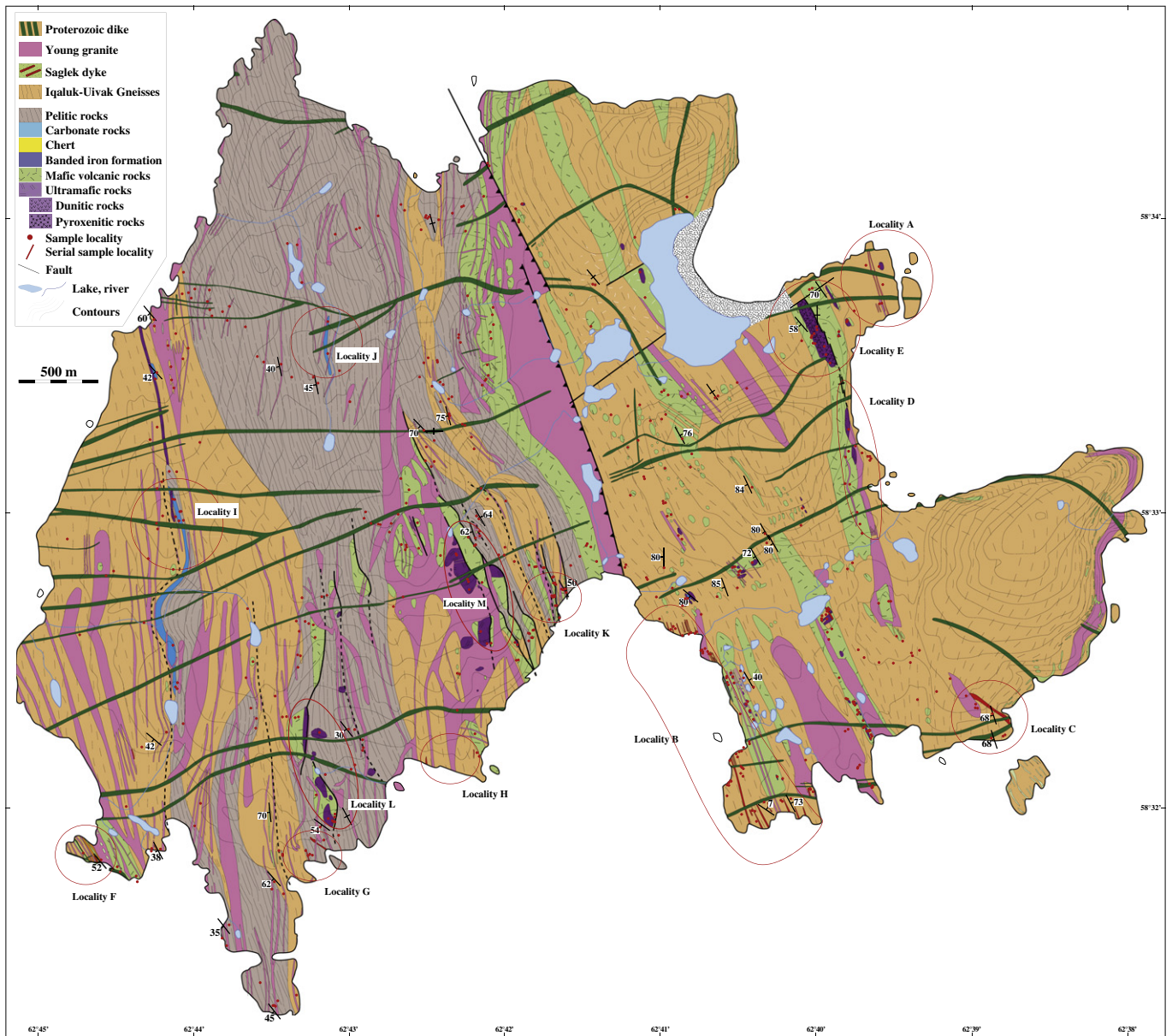
**Fig. 2.** A geological map of St John's Harbour South. The area is composed of the supracrustal rocks, Iqaluk-Uivak Gneisses, Saglek dyke, young granite intrusion and the Proterozoic mafic dikes. The supracrustal rocks form a NS-trending belt, and are intruded by the Iqaluk-Uivak Gneisses and subsequent Saglek dykes. The supracrustal belt is composed of fault-bounded blocks with similar lithostratigraphy from mafic rocks with/without ultramafic blocks to pelitic rocks from each other. The faults merge to the south.

Iqaluk-Uivak Gneiss series. Both are intruded by thin, fragmented, black, Saglek dykes. Some folded coarse-grained pale-yellowish white granitic dikes were intruded into them, and then a straight white, granitic dike was intruded into all of them.

The supracrustal rocks contain ultramafic rocks, mafic rocks, BIFs, carbonate rocks, cherts and clastic sedimentary rocks. The ultramafic rocks occur as blocks and enclaves within granitic rocks in most places because it is difficult for the granitic intrusions to completely rehomogenize ultramafic rocks (Figs. 8H and 9B). Large ultramafic bodies are sporadically present through the Saglek Block (Fig. 8G). In some places, the reaction periphery between the ultramafic blocks and the surrounding granitic intrusions is clearly preserved (Fig. 9B). In addition, most ultramafic rocks underlie mafic rocks or are sandwiched by mafic rocks. Ultramafic rocks are commonly massive and homogeneous, but some bodies display clear banding. In the Pangertok Inlet area, two ultramafic bodies, separated by an Iqaluk-Uivak Gneiss intrusion, are characterized by the presence of large minerals of olivine and orthopyroxene (Fig. 9C and D). In an eastern coastal point of the Shuldham Island, some kinds of ultramafic rocks are present: harzburgitic (Fig. 15D), dunitic (Fig. 15F), wehrlitic (Fig. 15C), and pyroxenitic ultramafic rocks (Fig. 15E). Especially, the harzburgite is characterized by large needle-like textures of olivines (Fig. 15D). Collerson et al. (1976b) considered the texture as a harrisitic texture of cumulate of a sill and thick flow, analogous to modern harrisitic texture in adcumulate (Donaldson, 1974; O'Driscoll et al., 2007; Wager et al., 1960). However, recent high-pressure metamorphic petrology shows that metamorphic olivine growth under high-pressure dehydration also produces the large needle structures of olivine in harzburgite rocks (Padrón-Navarta et al., 2011; Trommsdorff et al., 1998).

Carbonate rocks are also enigmatic in the Saglek Block, similar to the debate on the origin of the Isua carbonates between chemical sedimentary (Bolhar et al., 2004; Friend et al., 2008; Nutman et al., 2010) and metasomatic origins (Rose et al., 1996; Rosing et al., 1996). Metasomatism is not pervasive in the Saglek Block but there are signatures of the metasomatism such as the formation of metasomatism-type pelitic rocks. The carbonate rocks are also classified into at least two types based on the field occurrence and petrography. In the future, the geochemical signatures will help for the selection of primary carbonate rocks, analogous to the carbonate rocks in the Isua supracrustal belt (Bolhar et al., 2004; Nutman et al., 2010). Both types contain carbonate minerals, amphibole, quartz, magnetite, and sulfide, and the modal abundances vary according to the degree of metamorphism, silicification, and secondary alteration. The first type is accompanied with sedimentary rocks of BIF, pelitic rocks and chert (Fig. 10A to D) whereas the latter is accompanied with mafic and ultramafic rocks, intruded by granitic and dioritic intrusions (Fig. 10H). Fig. 10A shows the transition from carbonate rock through argillaceous carbonate layers to pelitic rocks, intruded by Iqaluk-Uivak Gneiss series in the western part of the Big Island. The carbonate rocks at the bottom of the sequence display wavy structures (Fig. 10B). Fig. 10C and D shows the carbonate rocks in the northeastern area of the St John's Harbour area, which are accompanied with chert and pelitic rocks (Figs. 5 and 13). Especially, the layered carbonate rocks have some siliceous nodules, and display a stromatolite-like, domal structure (Fig. 11C). Although most of them are almost completely recrystallized, some preserve relatively early textures. Fig. 10E shows the texture of the first type, and rounded or irregularly-formed globules of fine-grained carbonates are distributed in sparry carbonates. On the other hand, the latter contains polygons of fine-grained carbonates in coarse-grained carbonate spar with amphiboles (Fig. 10G). Especially, the shapes are similar to euhedral to subhedral olivine within anhedral pyroxene matrix in peridotites, suggesting that they are pseudomorphs of the olivine. In addition, some of carbonate rocks suffered from silicification, especially in the Pangertok Inlet area (Figs. 1, 6, 10H and 14). The silicified carbonate rock contains quartz and clinopyroxene with subordinate amounts of magnetite, and displays greenish white to pale-gray color with dark-green nodules and boudins (Fig. 10H).

Two types of chert can be also recognized: pure white/green chert and white to pale-gray chert with dark green nodules and boudins, respectively. The former occurs in the northeastern side of the St John's



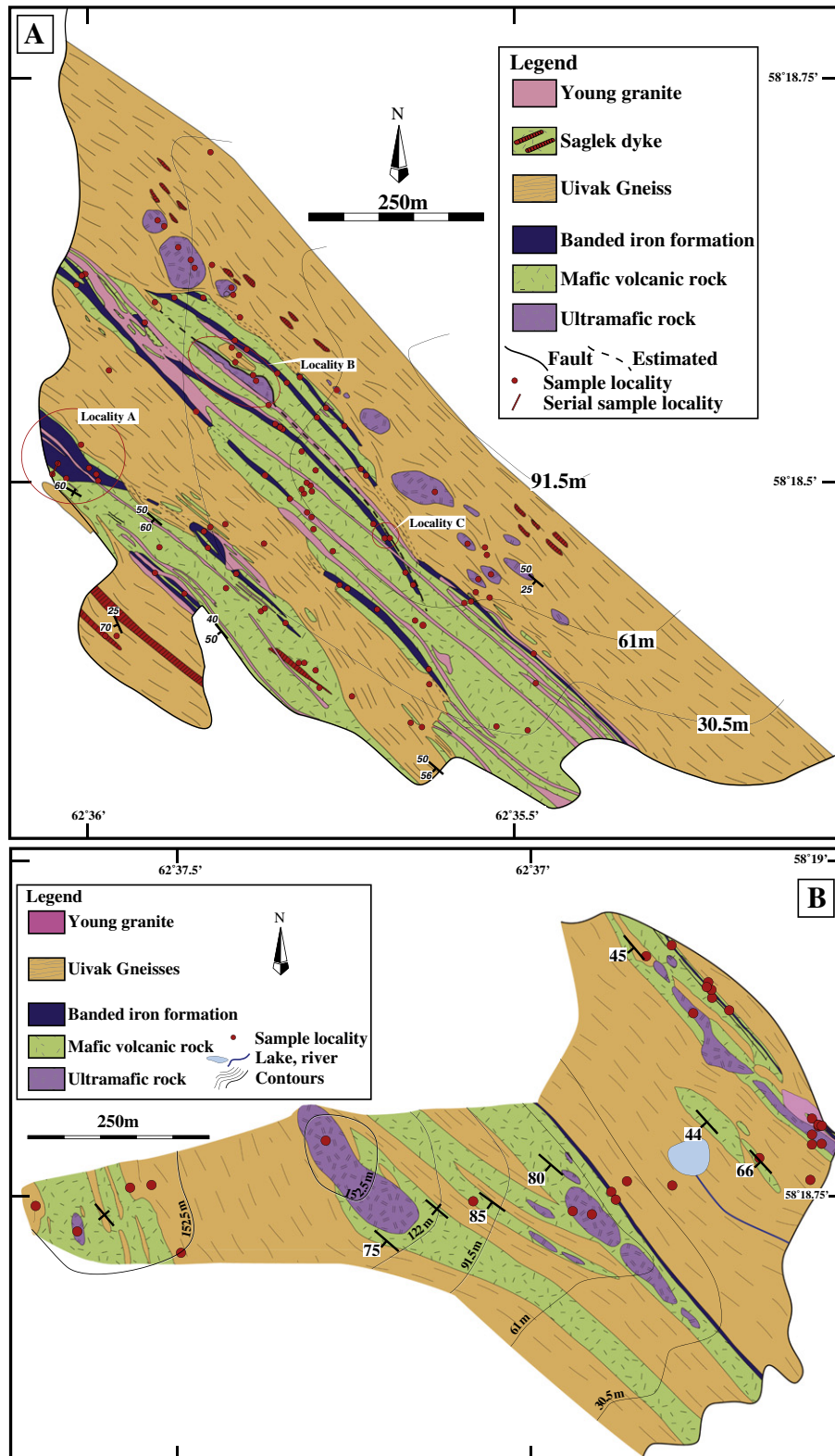
**Fig. 3.** A geological map of Big Island. The area is subdivided into two parts by a NS-trending fault. The eastern side is composed of the supracrustal rocks, Iqaluk-Uivak Gneisses, Saglek dykes, young granite intrusions and the Proterozoic mafic dikes. The supracrustal rocks form NS-trending belts, and are intruded by the Iqaluk-Uivak Gneisses and subsequent Saglek dykes. The supracrustal rocks contain ultramafic and mafic rocks with subordinate amounts of BIFs and cherts. The western side is predominant in pelitic rocks, and contains ultramafic, mafic and carbonate rocks.

Harbour Fjord (Figs. 5, 9H and 13). On the other hand, the latter displays rough or dissolved surface in most places, and are present especially in more severely metamorphosed, western side of the Saglek Block, e.g. Pangertok Inlet area (Figs. 10H and 15). They locally preserve carbonate minerals and are accompanied with carbonate rocks in weakly silicified areas.

Pelitic rocks are ubiquitously present through the Saglek Block (Ryan and Martineau, 2012). Especially, previous works regarded the pelitic rocks as representatives of the Late Archean Upernavik supracrustal rocks (Ryan and Martineau, 2012). But, the pelitic rocks are enigmatic in their origin, and can be differentiated into two types: true and putative pelitic rocks, as fully discussed later. All of the rocks, except for the Proterozoic basaltic dikes, underwent severe metamorphism up to granulite facies condition and strong deformation. In addition, young granite dikes, ranging from tens of cm to hundreds of meter thick, were ubiquitously intruded into all over the area. The intrusion of the granitic dikes resulted in alteration, metasomatism and

migmatization of the host rocks of mafic and ultramafic rocks, and modified the mafic and ultramafic rocks into the putative pelitic rocks (Fig. 11C to F). Because both pelitic rocks are brownish in color and display layering in most places, recognition between true and putative pelitic rocks is confusing. Most pelitic rocks, derived from clastic sediments, contain garnet, biotite, quartz and graphite (Fig. 11A and B). In a few places, possible sedimentary structures such as cross-lamination are still preserved (Fig. 11H). On the other hand, the putative pelitic rocks display transition from mafic rocks to granitic rocks (Fig. 11C and E), and contain plagioclase and amphibole in addition to biotite, garnet and quartz (Fig. 11D).

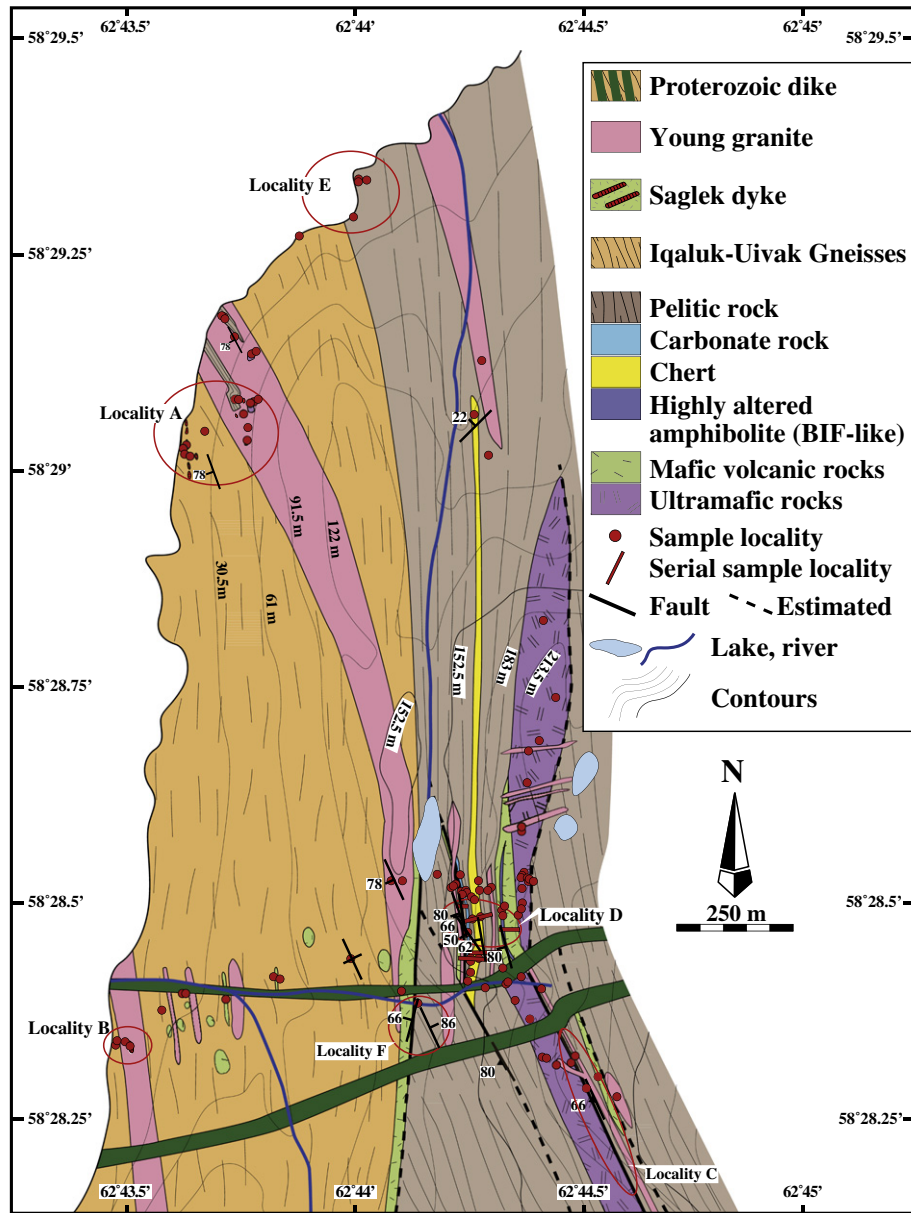
Conglomerates occur, accompanied with the pelitic rocks in some places, especially at the mouth of a small river in the northeastern side of the St John's Harbour Fjord (Fig. 12A to C). The clasts are mostly quartzite, and the largest clast reaches over boulder size. The shapes of the clasts in the conglomerates are utilized to estimate the degree of deformation. Fig. 12A shows a conglomerate observed from the



**Fig. 4.** (A) A geological map of the southwestern part of the Nulliak Island. The area is underlain by supracrustal rocks and the Uivak Gneiss, intruded by Saglek dykes and young granitic intrusions. The supracrustal rocks contain ultramafic and mafic rocks, interlayered with many thin BIF layers. (B) A geological map of the opposite side of the Nulliak Island. The area is also underlain by supracrustal rocks and the Uivak Gneiss, intruded by Saglek dykes and young granitic intrusions. The supracrustal rocks contain ultramafic and mafic rocks, and three thin BIF layers.

south whereas the clasts of a conglomerate in Fig. 12B and C are observed from the east. The aspect ratios of the clasts in conglomerate range from 1.9 to 46.8, indicating more significant extension along NS-direction.

Banded iron formation (BIF) is often regarded as the oldest chemical sedimentary rocks in the Eoarchean greenstone belt (e.g. Mojzsis and Harrison, 2002a,b; Mojzsis et al., 1996; Nutman et al., 1996, 1997; O'Neil et al., 2007). BIF is defined as a chemical sedimentary

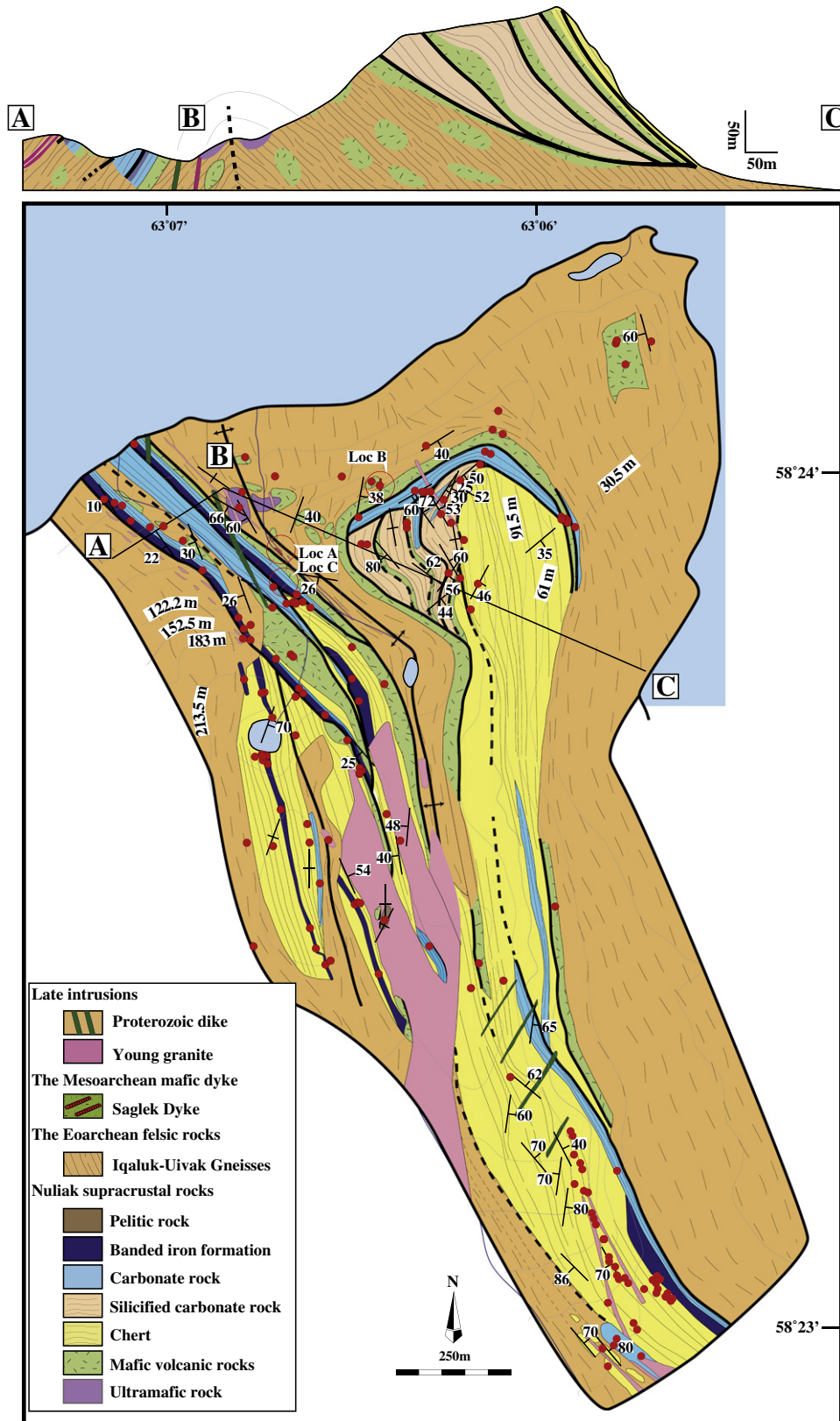


**Fig. 5.** A geological map of St John's Harbour East. NS-trending supracrustal belt and Iqaluk-Uivak Gneisses are present in this area. The supracrustal rocks contain ultramafic and mafic rocks, pelitic rocks, carbonate rocks and cherts.

rock (James, 1954) and was classified into four types: Algoma, Superior, Minette and Clinton-types (Gross, 1965). The classification of the Superior and Algoma-types is often used for the Precambrian BIFs (Klein, 2005). In the Saglek Block, previous works showed that the BIFs occur in the Nulliak Island, its opposite side, and Pangertok Inlet areas, but we found more places where the BIFs are present, e.g. the Big Island and Shuldham Island (Fig. 3). On the other hand, a previous work showed BIF to the southern of the St John's Harbour area (Ryan and Martineau, 2012), but no BIF was found there. The BIFs are also classified into two types: chemical sedimentary and metasomatic types. The BIF in the Akilia Island, southern West Greenland, is controversial whether metasomatized ultramafic rock (Fedo and Whitehouse, 2002) or chemical sediment (Friend et al., 2002; Mojzsis and Harrison, 2002b). In the Saglek Block, mafic rocks metasomatized by thin granitic intrusions look like BIF. The BIFs of chemical sediments occur in Big Island (Figs. 3 and 9G), Nulliak Island (Figs. 4A, 9E and F), its opposite side (Fig. 4B), Pangertok Inlet (Fig. 6) and the southern part of the Shuldham Island (no map) whereas the

metasomatic-type BIFs sporadically occur all over the Saglek Block such as the northeastern part of the St John's Harbour area (Fig. 5) and to the southern of the St John's Harbour area (Ryan and Martineau, 2012). Most of the BIFs occur as thin (<5 m) layers but a BIF layer in the Nulliak Island reaches ca. 32 m (Fig. 4A). The thickness of BIFs changes drastically, and they laterally fade out in tens to hundreds of meter. The layering of white and black bands in the BIFs is relatively clear in most places, and slumping structures are also obvious in some places (Fig. 11E). The BIFs contain magnetite, quartz, orthopyroxene and clinopyroxene with subordinate amounts of cummingtonite, hornblende, pyrite and apatite. The white bands mainly comprise quartz, which forms large, 10 to 1000  $\mu\text{m}$  across, anhedral to euhedral crystals and shows wavy extinction. On the other hand, the black bands consist mainly of magnetite, clinopyroxene, orthopyroxene with subordinate amounts of cummingtonite and hornblende. The clinopyroxenes are present as discrete grains, parallel to the bands, whereas the orthopyroxenes are as discrete grains or exsolution lamellae in clinopyroxenes.

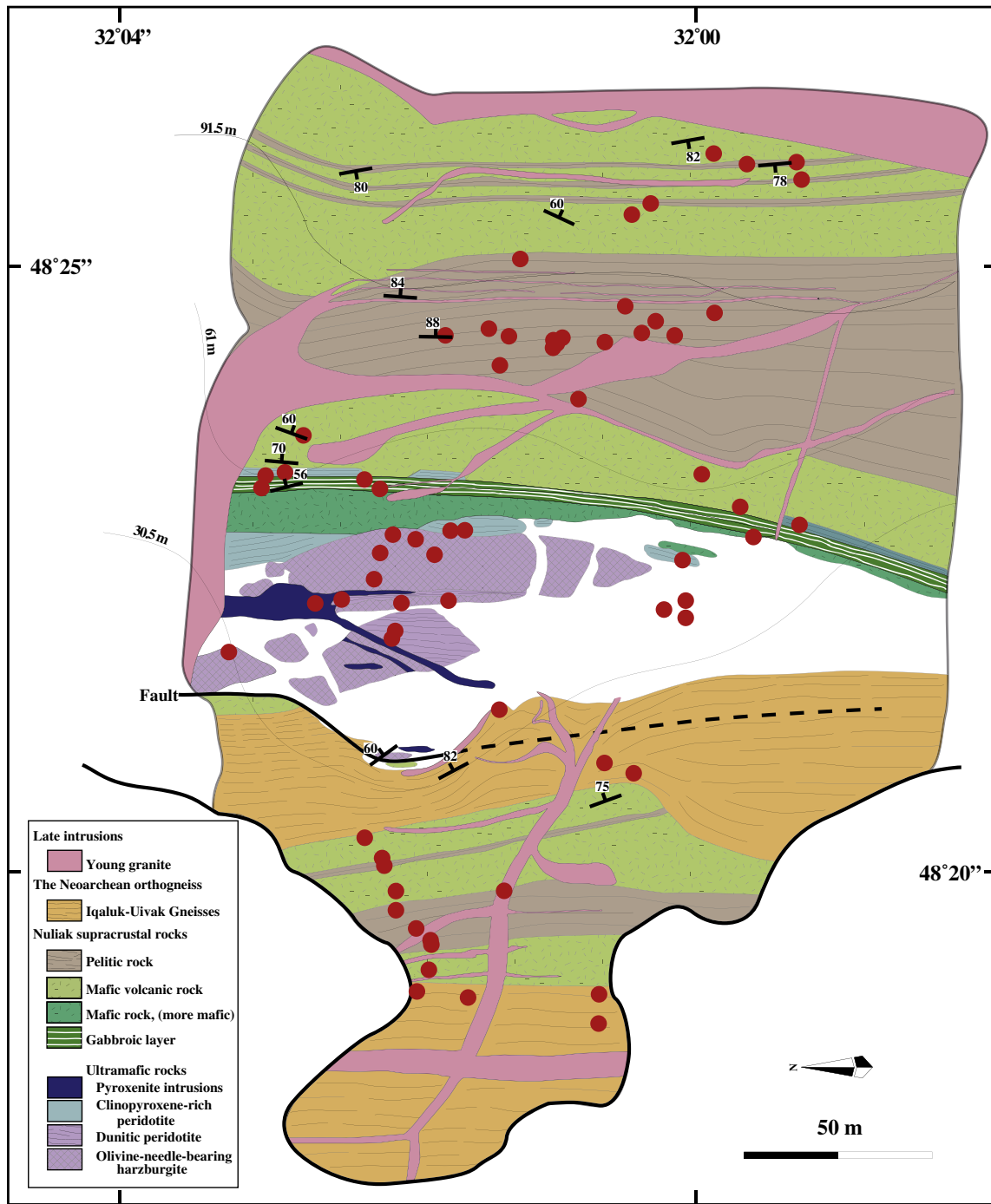




**Fig. 6.** A geological map of the Pangertok Inlet area. Two NS-trending supracrustal belts run within the Iqaluk–Uivak Gneiss body, intruded by the Saglek dykes. The supracrustal rocks have mafic rocks, cherts, carbonate rocks, BIFs and pelitic rocks. Mafic and ultramafic blocks occur as enclaves within the Iqaluk–Uivak Gneiss body. The branching of sedimentary layers in some places indicates faults. The lithostratigraphies of the fault-bounded blocks are commonly from mafic rocks, carbonate rocks, cherts and BIFs in ascending order.

The Nuliak supracrustal rocks and Iqaluk–Uivak Gneiss series repeatedly suffered from tectonothermal events; it is difficult to recognize the first tectonic structures of foliations, faults and folds. Because at

present, the predominant structures of them display NS-trending, parallel to the granitic intrusions (Figs. 2 to 7), the original deformation structures were erased by the intrusions in most places. However, we can



**Fig. 7.** A geological map of a small point of the western coast of the Shuldam Island. The area is characterized by ultramafic rocks with large olivine needle-like structures. The ultramafic rock-bearing body consists of harzburgitic ultramafic rocks, olivine-clinopyroxene rocks, clinopyroxene-hornblende, banded mafic rock of plagioclase-phyric and clinopyroxene-hornblende layers, fine-grained amphibolite and pelitic rocks, in ascending order. The lithostratigraphy is similar to an ophiolite.

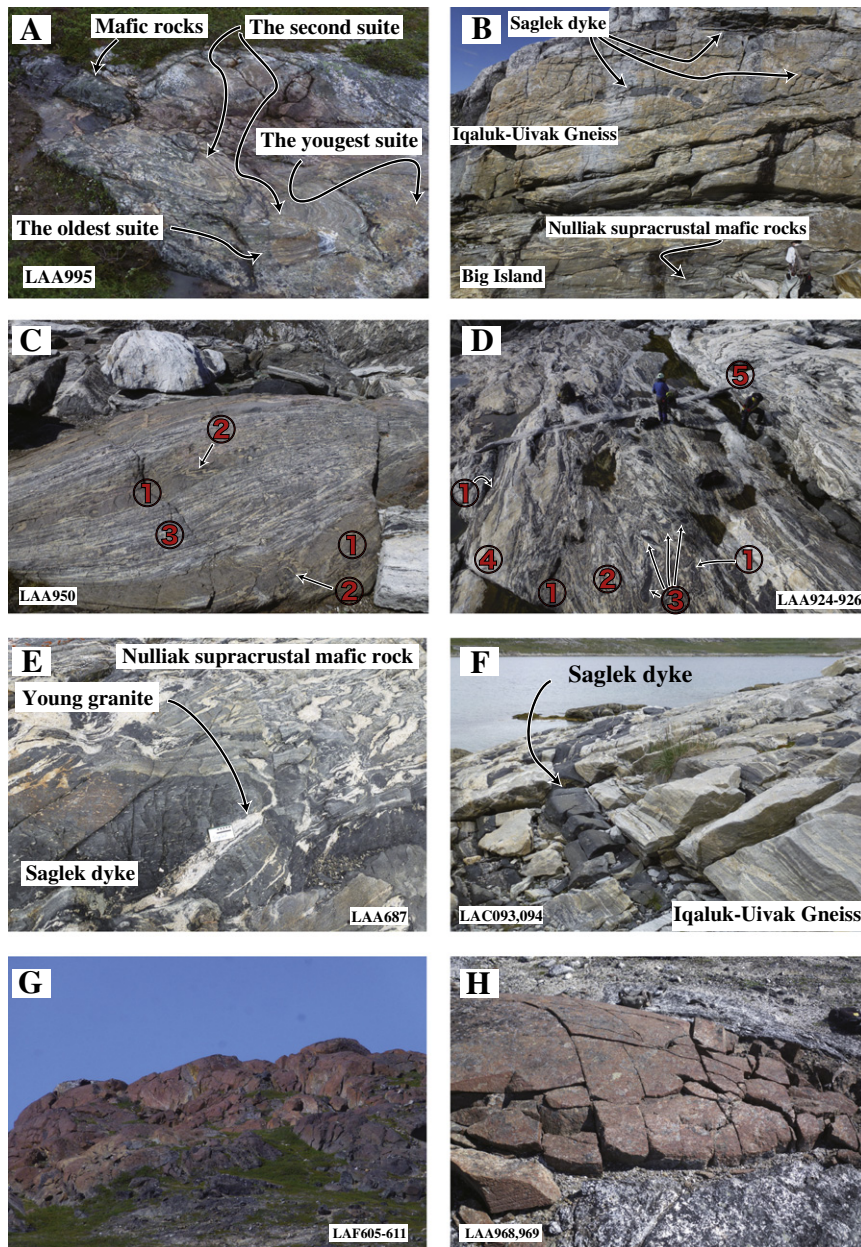
recognize a few original deformation structures but those all over the areas. Fig. 12D shows a boundary between the mafic rocks and pelitic rocks in the southern area of the St John's Harbour. The deformation is concentrated along the boundary and two types of foliations are obvious. Some foliations are parallel to the boundary whereas others are oblique to the parallel foliations and boundaries, showing a duplex structure, defined by roof-, floor- and link-thrusts (McClay, 1992). Such branching structures, displaying that the foliations in mafic rocks are oblique to those in pelitic rocks, can be observed in many places (Fig. 12E, and F), suggesting that faults were originally present along the boundary. In addition, small-sized branching or oblique structures

are ubiquitously observed all over the areas. Fig. 12G shows a small duplex structure around the boundary between the mafic and pelitic rocks. Fig. 12H displays the upper half of a duplex structure within a silicified carbonate rock in the Pangertok Inlet region.

#### 4. Geology of the six areas in the Saglek Block

##### 4.1. Geology of the St John's Harbour South

Fig. 2 shows a detailed geological map of the southern part of the St John's Harbour area (Fig. 1C), which contains a supracrustal belt with

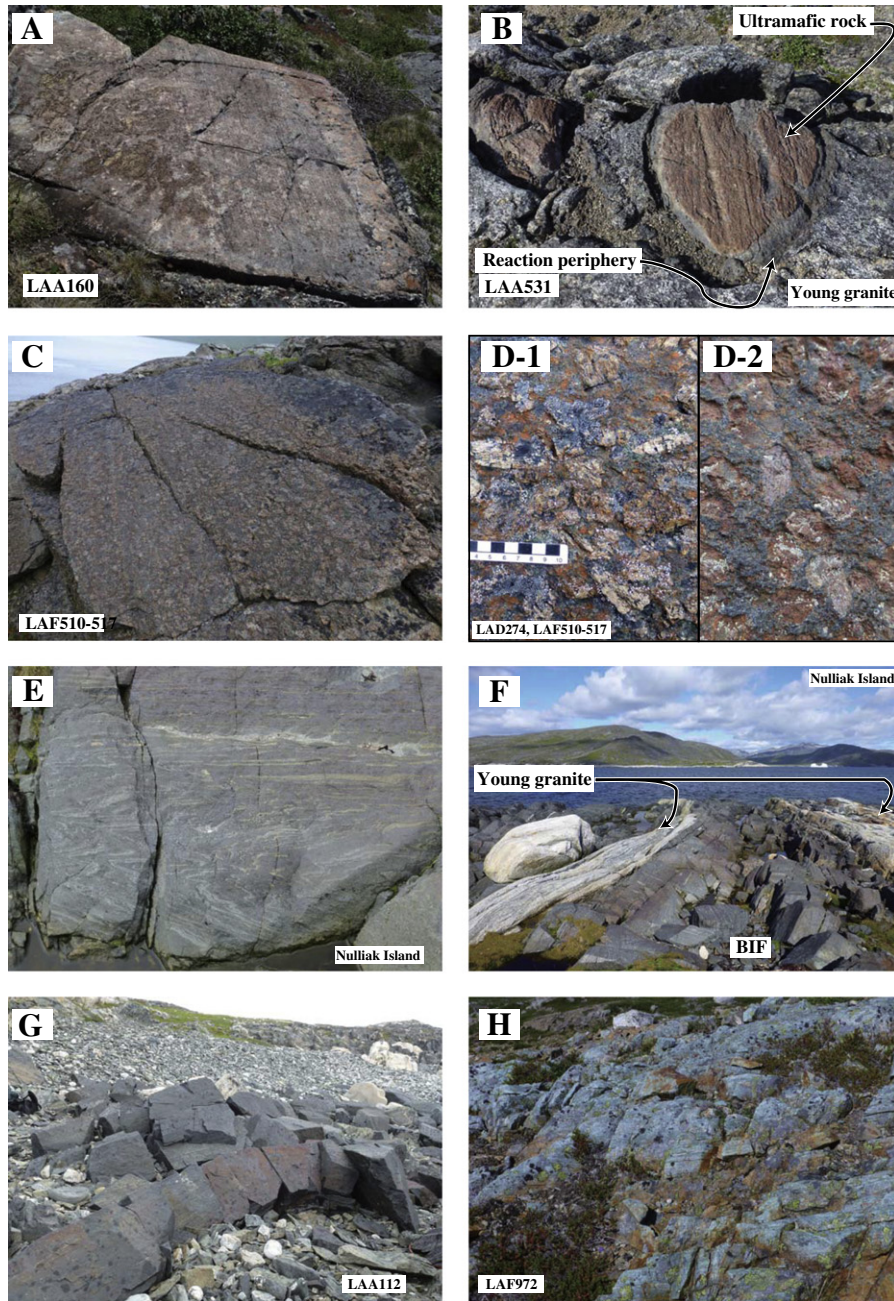


**Fig. 8.** Photographs of outcrops of the Iqaluk–Uivak Gneiss complex, Saglek dykes and ultramafic rocks. (A) An outcrop of Iqaluk–Uivak Gneisses and mafic enclaves. The Iqaluk–Uivak Gneisses are composed of orthogneisses with at least three generations based on cross-cutting relationship. The mafic rocks occur as enclaves within the oldest suites of the Iqaluk–Uivak Gneisses. (B) Iqaluk–Uivak Gneisses, intruded by thin Saglek dykes. (C) Iqaluk–Uivak Gneisses with at least three generations. (D) A composite body of the Nulliak supracrustal rocks (1), Iqaluk–Uivak Gneisses (2), Saglek dykes (3), white coarse-grained gneiss (4) and a young granitic intrusion (5). (E) Nulliak supracrustal rock, intruded by a Saglek dyke with large plagioclases and subsequent granitic intrusions. (F) Iqaluk–Uivak Gneisses, intruded by a Saglek dyke. (G) An ultramafic body. (H) An ultramafic block within a granitic intrusion.

N–S trending, Iqaluk–Uivak Gneiss bodies, and granitic and the Proterozoic basic intrusions. We named the area St John's Harbour South hereafter. The supracrustal belt contains ultramafic bodies, mafic rocks, pelitic rocks, and carbonate rocks. The mafic and pelitic rocks mostly strike NWN and dip steeply to moderately west. On the other hand, the Iqaluk–Uivak orthogneiss bodies are composed of tonalitic to granitic gneisses with some generations. Because no Saglek dyke was reported before our geological survey and pelitic gneisses are widely cropped out (Ryan and Martineau, 2012), the supracrustal rocks had belonged to the Upernavik supracrustal rocks (e.g. Bridgwater and Schiøtte, 1991; Ryan and Martineau, 2012). But, our detailed geological survey found many, thin (<2 m thick) and fragmented (< tens meter long), Saglek dykes with plagioclase megacrysts in Localities A to D (Fig. 2). Besides, the supracrustal belt is intruded by a Iqaluk–Uivak Gneiss body in the

central part (Fig. 2). The presence of the Saglek dyke and geological relationship that the supracrustal belt was intruded by the Iqaluk–Uivak Gneiss indicate that the supracrustal belt belongs to the Nulliak supracrustal rocks. In addition to the Saglek dykes, ultramafic bodies and carbonate rocks were first found in this area. Especially, the ultramafic rocks in the coastal area were described as pelitic gneisses in the previous works (Ryan and Martineau, 2012). We could not find banded iron formation and chert in this area. We collected 650 samples from the supracrustal belt and the surrounding orthogneisses in St John's Harbour South.

In addition to severe metamorphism up to granulite facies condition and strong deformation, the ubiquitous intrusion of the granitic dikes resulted in alteration, metasomatism and migmatization of the host rocks of mafic and ultramafic rocks. Previous studies showed that the

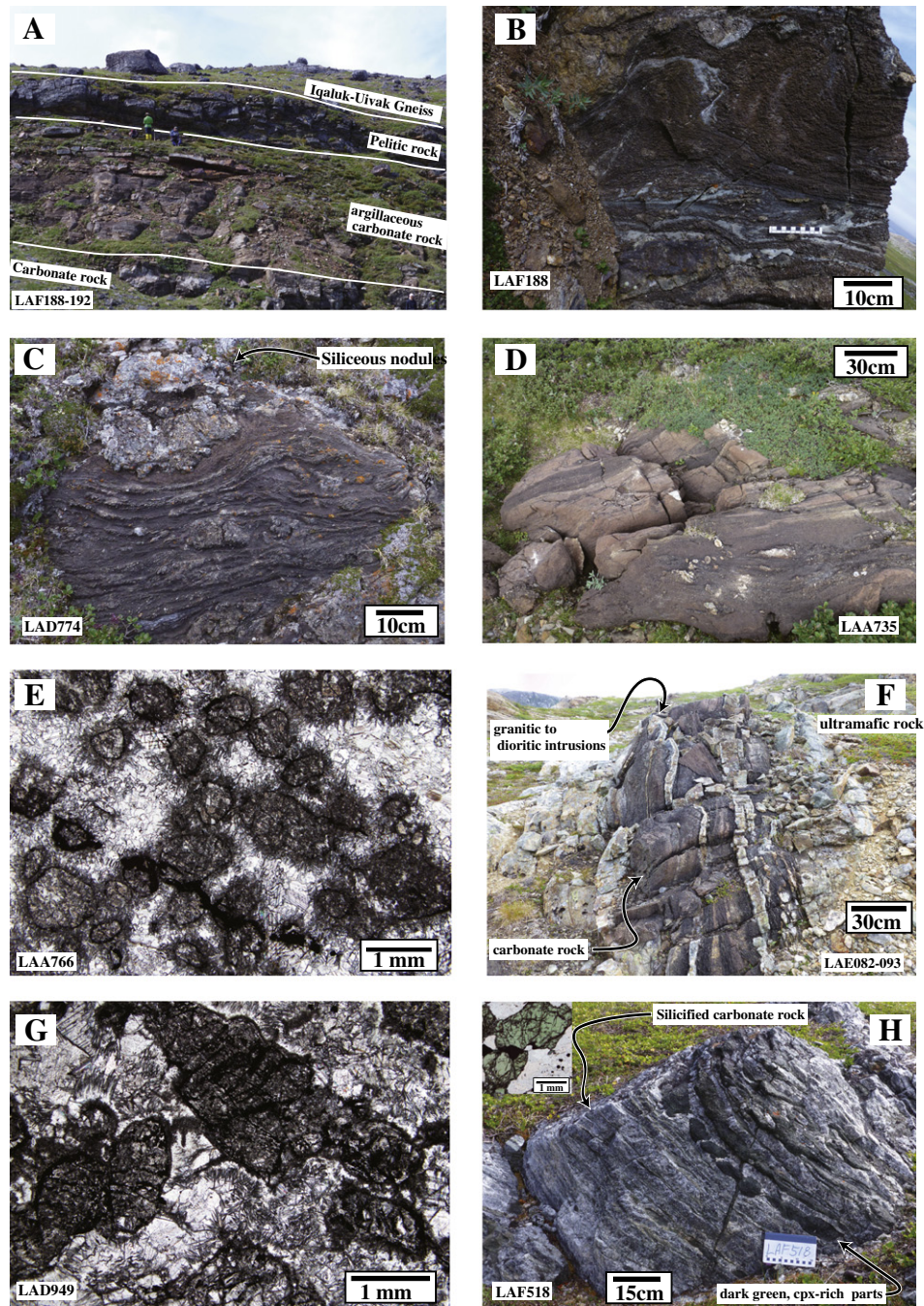


**Fig. 9.** Photographs of outcrops of the Nulliak supracrustal rocks. (A) An outcrop of the ultramafic rocks with clear layering at the St John's Harbour South. (B) Ultramafic rock enclaves within a young granite intrusion. Micaceous reaction periphery is found between the ultramafic rocks and granite intrusion. (C) An ultramafic rock with olivine and pyroxene megacrysts in the Pangertok Inlet area. (D) The pyroxene (1) and olivine megacrysts (2) within the ultramafic rock (C). (E) A banded iron formation (BIF) with slumping structure interlayered with parallel layers at Nulliak Island. (F) A thick banded iron formation with clear layering, intruded by young granite intrusions at Nulliak Island. (G) A thick banded iron formation at Big Island. (H) Green cherts with white and green layers at St John's Harbour East.

pelitic gneisses are widely distributed in this region but the pelitic gneisses should be classified into pelitic gneiss and putative pelitic gneiss based on the occurrence and petrography, as shown above and discussed below. The intrusion of granitic rocks modified the mafic and ultramafic rocks into the putative pelitic rocks (Fig. 11C to F). The true pelitic rocks contain biotite, garnet, quartz, and graphite, and display clear banding (Fig. 11A and B). Especially, fresh pelitic rocks are preserved in the Locality E. The ultramafic rocks are pervasively distributed in the southern part of this area, and sporadically present as blocks in mafic rock bodies and felsic rock bodies of Iqaluk–Uivak Gneisses and young granite intrusions (Fig. 2). Some of them display clear banding (Fig. 11A) but most are homogeneous. In some places, the ultramafic

rocks are near pelitic rocks. The carbonate rocks sporadically occur at least eight points of Localities F and G near the ultramafic rocks and within the ultramafic bodies (Fig. 10F). Most of them were intruded by granitic and dioritic intrusions (Fig. 10F). The microscopic observation displays pseudomorphs of olivine and pyroxenes, replaced by fine-grained carbonates, within coarse-grained carbonate matrix (Fig. 10G), suggesting metasomatic-type carbonate rocks. We cannot find chemical sediment-type carbonate rocks in this region.

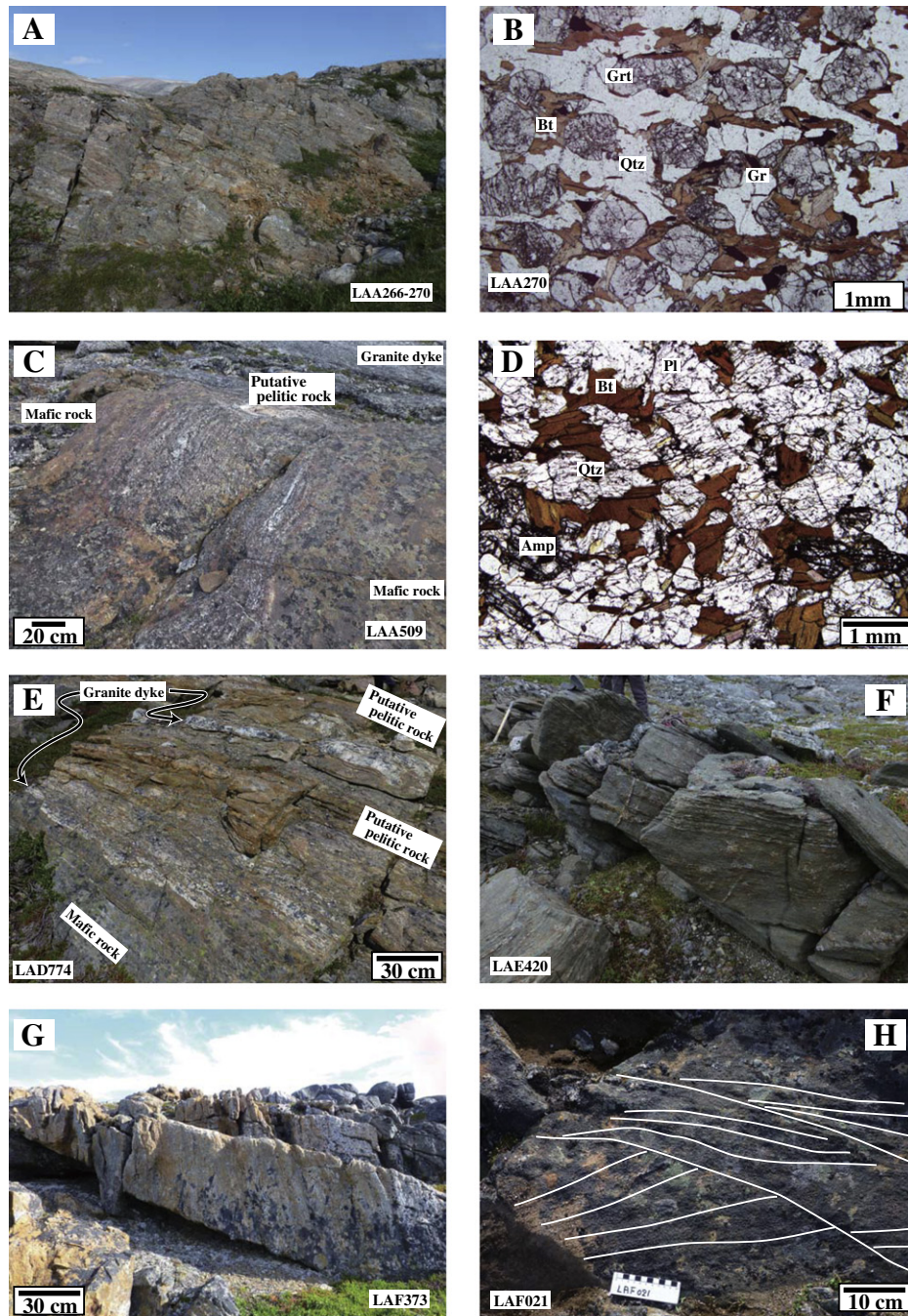
It is well known that the Saglek–Hebron area underwent some thermal tectonic events. The field occurrence of orthogneisses clearly shows that some orthogneisses with different ages are present even in small areas. Shimojo and colleagues showed a detailed 1:10 sketch



**Fig. 10.** Photographs of the Nulliak supracrustal rocks and thin sections of carbonate rocks. (A) A large outcrop of the Nulliak supracrustal rocks from the carbonate rocks through argillaceous carbonate rock to pelitic rock, intruded by the Iqaluk–Uivak Gneisses at Big Land. (B) A close view of the carbonate rock (A), showing a domal structure. (C) A carbonate rock with siliceous nodules at St John's Harbour East, showing domal structure in the center. (D) A representative microscopic image of the carbonate rock (C), showing fine-grained globular structures within clear coarse-grained carbonates. (E) A carbonate rock, associated with ultramafic rocks at St John's Harbour South. (F) A representative microscopic image of the carbonate rock (E), showing fine-grained polygonal structures within clear coarse-grained carbonates. (G) White chert with green boudins, derived from silicification of carbonate rock, at Pangertok Inlet area. (H) A representative microscopic image of a green pyroxene in quartz is also shown at the upper left side.

map and described field relationships among the multiple-formed mafic rocks and orthogneisses in an outcrop of the northern part of this area (Shimojo et al., 2012, 2013, 2015). They showed that the orthogneisses with five generations and mafic rocks with two generations are present even in the small area based on cross-cutting relationship of orthogneisses and mafic rocks. In addition, they reported cathodoluminescence imaging and in-situ U–Pb dating of zircons separated from the oldest and youngest suites of the Iqaluk–Uivak Gneiss series in the outcrop (Fig. 8A). The former is a banded gray

gneiss, which is composed of highly folded white and black thin layers, and is intruded by the white gneiss intrusions because the gneissic structures are clearly truncated by the white gneiss (Fig. 8A). The latter is a coarse-grained, white granitic intrusion, which is mostly massive and faintly gneissic in only some places (Fig. 8A). The banded gray gneiss contains igneous zircons up to 3.95 Ga, indicating that the precursor was formed at 3.95 Ga. On the other hand, the white granitic intrusion was formed at 3.87 Ga (Shimojo et al., 2012, 2013, 2015).

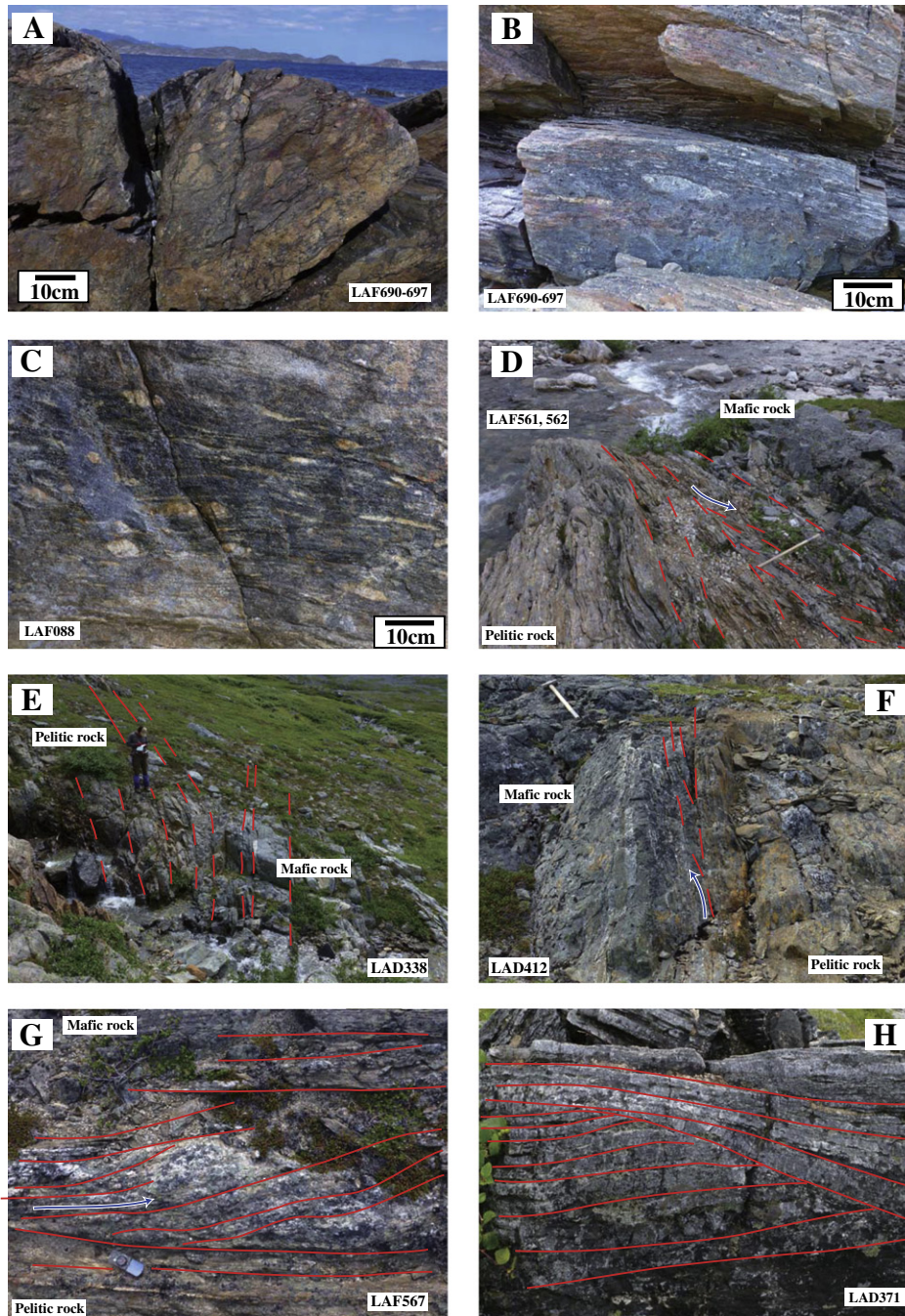


**Fig. 11.** Photographs of the Nulliak supracrustal rocks and thin sections of pelitic rocks. (A) An outcrop of the pelitic rocks at the St John's Harbour South. (B) A representative microscopic image of the pelitic rock (A), containing biotite (Bt), garnet (Grt), quartz (Qtz) and graphite (Gr). (C) An outcrop of putative pelitic rock, showing transition from brownish green mafic rock (both sides) to the brown pelitic rock (center) possibly due to thin granitic intrusions at St John's Harbour South. (D) A representative microscopic image of the putative pelitic rock (C), containing biotite (Bt), amphibole (Amp), and quartz (Qtz). (E) An outcrop of putative pelitic rock, showing transition from brown-grayish mafic rock to the brown pelitic rock, possibly due to thin granitic intrusions at Big Island. (F) A weathered, brownish foliated ultramafic rock at St John's Harbour South. (G) A weathered, brownish granitic intrusion at Shuldham Island area. (H) A pelitic rock with possible cross-lamination at Big Island.

The boundary between the supracrustal rocks and Iqaluk–Uivak Gneisses was ambiguous. Because previous works considered the supracrustal rocks in this region as the Upernavik supracrustal rocks, it was considered that the supracrustal rocks unconformably overlay the Uivak Gneiss. But, any basal conglomerates cannot be found in this area. On the other hand, varied-sized outcrops of mafic rocks, ultramafic rocks, and pelitic rocks are found within the Iqaluk–Uivak Gneisses (Fig. 2). The Iqaluk–Uivak Gneisses are intruded into the mafic and ultramafic rock bodies and pelitic rock belt in some places (Fig. 2). The

geological relationship indicates that the supracrustal belt was intruded by the Iqaluk–Uivak Gneisses.

The Nulliak supracrustal rocks repeatedly suffered from tectonothermal events; it is difficult to recognize the original tectonic structures of foliations, faults and folds at the emplacement. In most cases, it seems that the relationship among ultramafic, mafic and pelitic rocks is conformable, and they occur repeatedly. But, it can be immediately noticed that the sedimentary belts of pelitic rocks branch in many places (Fig. 2), and faults can explain the branching structures. In



**Fig. 12.** Photographs of outcrops of the Nulliak supracrustal rock and deformation structures within the supracrustal belt. (A) An outcrop of the conglomerate at the St John's Harbour East, taken from the south. (B) An outcrop of the conglomerate at the St John's Harbour East, taken from the east. A large siliceous clast in the conglomerate displays north to south extension. (C) An outcrop of the conglomerate at the St John's Harbour East, taken from the east. Many clasts in the conglomerate were highly extended along the NS-direction. (D) The boundary between lower pelitic rock (left side) and upper mafic rock at St John's Harbour South. The foliations are concentrated at the upper part of the pelitic rock and oblique to the fabric of the upper mafic rock, indicating a fault between them. A duplex structure is also found there. (E) A boundary between pelitic rock (left side) and mafic rock (right) at St John's Harbour East. The fabrics of pelitic rocks are oblique to those of the mafic rocks. Although their relation looks conformable, a fault is present between them. (F) A boundary between pelitic rock (right side) and mafic rock (left) at St John's Harbour South. The fabrics of mafic rocks are oblique to those of the pelitic rocks. Small faults, oblique to the fabrics in pelitic rocks, form a half duplex structure. (G) A duplex structure in pelitic rocks, at St John's Harbour South. (H) A half duplex structure in chert, a silicified carbonate rock, in the Pangertok Inlet area.

addition, the oblique structures of internal fabrics and small duplex structures, which are ubiquitously distributed in this region, also support the faults. Fig. 12D shows that bedding planes of a pelitic rock are oblique to the fabric of overlying mafic rocks at Locality H. A small duplex structure with floor, roof and link thrusts can be recognized in a pelitic rock at Locality J (Fig. 12D). Fig. 12F also displays the oblique structure between the internal fabric of mafic and pelitic rocks at Locality I.

Fig. 2 shows the distribution of the faults, reconstructed based on the branching of the pelitic belts, and that the belt is composed of many

fault-bounded tectonic blocks. The ultramafic blocks occur only in two tectonic blocks, and the ultramafic bodies are mostly concentrated in the lower parts of the mafic rock belts. As a result, the lithostratigraphy of each ultramafic rock-bearing tectonic block consists of ultramafic, mafic and pelitic rocks, in ascending order. The tectonic blocks merge with the lower blocks to the south whereas with upper blocks to the north. The former can be observed in many places whereas the latter in some places (Fig. 2). The structure is similar to a large-scale duplex structure.

#### 4.2. Geology of the Big Island

The Big Island is located within the Saglek Fjord, and the northern extension of the St John's Harbour South (Fig. 1C, Bridgwater et al., 1975). A detailed geological map is available, and shows that Big Island is underlain by the Upernavik supracrustals and Uivak and undifferentiated gneisses, intruded by NS-trending granite intrusions and EW-trending Proterozoic dikes (Bridgwater et al., 1975; Collerson and Bridgwater, 1979; Collerson et al., 1976a). The supracrustal rocks contain metasedimentary rocks, amphibolite and ultramafic rocks, and strike NWN to NS and dip steeply to moderately west. No original fabrics of pillow structure and hyaloclastite are preserved but the mafic rocks are considered as lavas and sills (Bridgwater et al., 1975; Collerson et al., 1976a). Because clastic sediments are predominant in the supracrustal rocks and no BIF was found, despite no obvious boundary between the Uivak Gneiss and supracrustal rocks, it was considered that the supracrustal rocks belong to the Upernavik supracrustal rocks, namely younger than the Uivak Gneiss (Bridgwater and Schiøtte, 1991). And, the orthogneiss in the eastern part was considered as the Uivak Gneiss because they are intruded by Saglek dykes (Bridgwater et al., 1975; Collerson and Bridgwater, 1979; Collerson et al., 1976a,b). On the other hand, Bridgwater and Schiøtte (1991) suggested that young orthogneiss, ca. 3200 Ma Lister gneiss, widely occurs as intrusions in the western and central areas.

Fig. 3 shows our geological map of the Big Island, which is underlain by the Nulliak supracrustal rocks, Iqaluk–Uivak Gneiss, Saglek dyke and Proterozoic dikes. We collected 833 samples in the Big Island. The Big Island mostly comprises two parts, bounded by a vertical fault and granite intrusion (Fig. 3). The eastern part consists of the Iqaluk–Uivak Gneisses with subordinate amounts of mafic rocks, ultramafic rocks, BIFs, cherts and carbonate rocks. Many Saglek dykes are sporadically distributed over the area, especially in Localities A to C (Fig. 8B, C and E). Pale-green to silver, micaceous rocks occur as blocks, accompanied with granitic intrusions and ultramafic rocks along a fault in the eastern part (Locality D). The supracrustal rocks occur as blocks or enclaves within the Iqaluk–Uivak Gneisses or as belts with NW–SE trending, and strike NNE and dip vertically to steeply to the west (Fig. 3). The mafic rocks are ubiquitously present, and ultramafic rocks are mostly accompanied with the mafic rocks (Fig. 8G). The ultramafic rocks contain dunite, peridotite, and pyroxenite, and the dunite and pyroxenite occur as intrusions in the Locality E (Fig. 8G). The ultramafic rocks mostly occur within the mafic rock belts, and the original structural relationship between them is ambiguous in most outcrops because they adhere to each other due to later metamorphism (e.g. Collerson et al., 1976a). It was suggested that the peridotite is cumulate part of a basaltic and basaltic komatiite flows and sills because of chemical transition from peridotite through clinopyroxenite to amphibolite (Collerson et al., 1976b). However, a large ultramafic body at Locality E overlies mafic rock belt, bounded by a fault, indicating that the boundary between eastern side of the ultramafic body and the mafic rocks is tectonic. On the other hand, the western boundary between the ultramafic rocks and overlying mafic rocks seems to be conformable. In addition, two pyroxenite parts occur as intrusions, in contrast to previous interpretation of transition from ultramafic to mafic rocks. The occurrence is common in some places, where the primary structure was still preserved, for example in the Shuldham Island.

Banded iron formation (BIF) layers and chert occur at Locality B. This work first describes both BIF and chert in the Big Island. The thin chert layer occurs in the northern part. A thin (1 m thick) BIF layer with clear banding of white and bluish black layers on mafic rocks occurs in the middle part (Figs. 3 and 9G). Another BIF layer occurs as a small fragment within a mafic rock belt in the southern part.

The western part of the Big Island is characterized by dominance of pelitic rocks (Fig. 3). The supracrustal rocks contain mafic and ultramafic rocks, pelitic rocks, and carbonate rocks. Both gneisses at Localities F and H and supracrustal rocks at Locality I are intruded by the Saglek

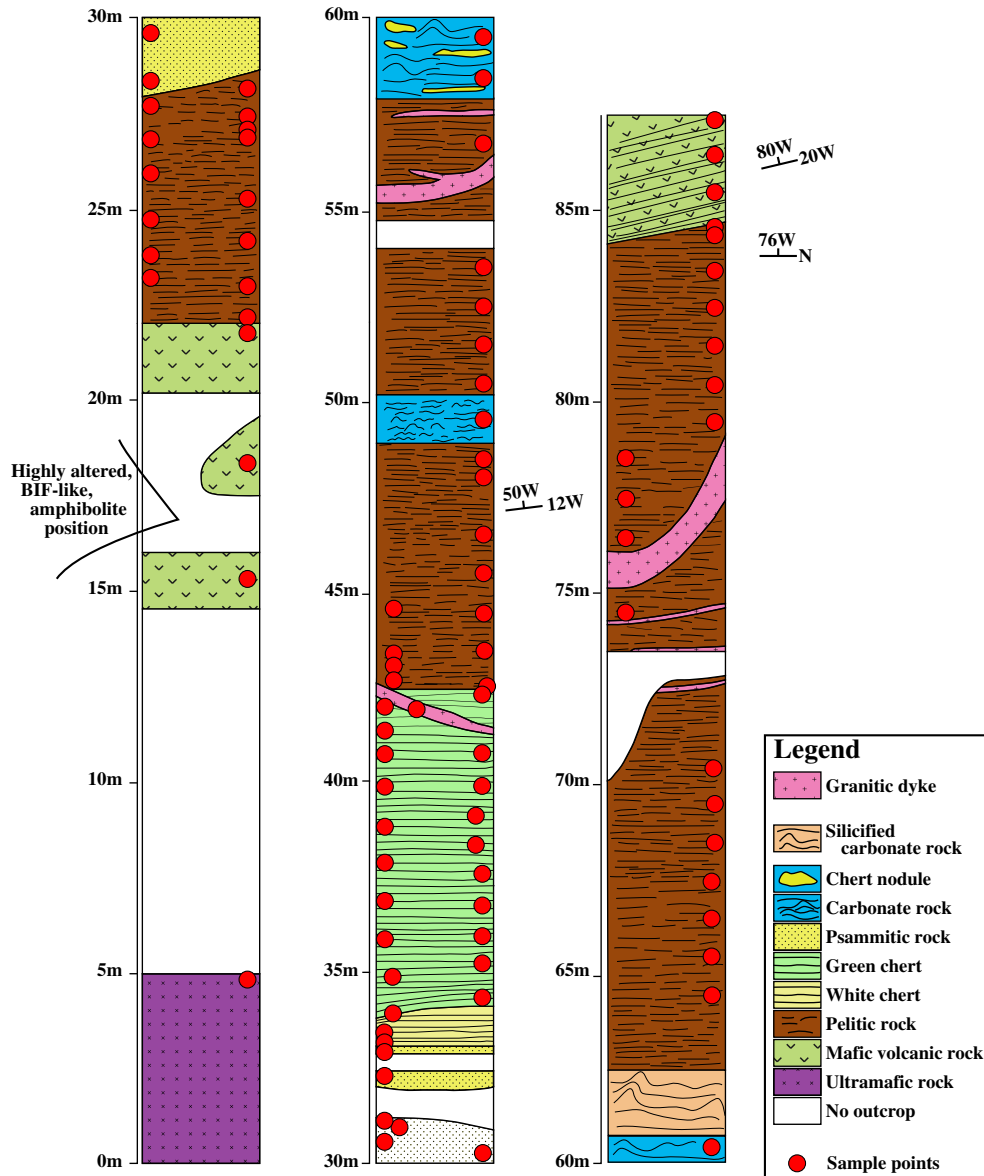
dykes, indicating that they are Iqaluk–Uivak Gneisses and Nulliak supracrustal rocks (Fig. 3). The ultramafic rocks occur as blocks within pelitic rocks and granitic intrusions or lower parts of mafic–ultramafic rock bodies. The carbonate rocks are found in two areas, Localities I and J (Fig. 3). The carbonate layers are changeable in thickness, up to 10 m thick but laterally fade away. In the Locality I, the carbonate layer displays transition from pure carbonate rocks through argillaceous carbonate rocks to pelitic rock (Fig. 10A). Especially, the lower part seems to have a domal structure (Fig. 10B). The pelitic rocks are pervasively distributed in the western part. Especially, the pelitic rocks in the Locality K contain garnet, biotite, quartz, zircon, plagioclase and graphite, and obviously originate from clastic sediments.

Some ultramafic blocks occur within the pelitic rocks, and seem to be *mélange* in modern accretionary complexes (e.g. Cowan, 1985; Matsuda and Ogawa, 1993). However, the geological structure in the western part can be also explained as tectonic piles of fault-bounded blocks similar to that of the St John's Harbour South. The tectonic block consists of the ultramafic rock, mafic rock and pelitic rocks, in ascending order, especially at Localities L and M.

#### 4.3. Geology of the southwestern part of the Nulliak Island and the opposite side

Supracrustal rocks occur in the southwestern part of the Nulliak Island (Figs. 1 and 4A). The supracrustal rocks are composed of ultramafic rocks, garnet–amphibolites and BIFs. The supracrustal belt is bounded by the Uivak Gneiss on both the eastern and western sides, and is intruded by the Uivak Gneiss intrusions in the center (Fig. 4A). The supracrustal rocks and the Uivak Gneisses are intruded by the Saglek dykes, supporting that they belong to the Nulliak supracrustal rocks (e.g. Nutman et al., 1989; Ryan and Martineau, 2012). Although all the granitic intrusions are not shown in the geological map because it is difficult to show the exact distribution of the granitic intrusions in the Uivak Gneiss areas, many N–S trending granitic intrusions are distributed all over the area. The supracrustal rocks mostly strike NW–SE and dip moderately to gently west. The ultramafic rocks occur in two zones: the first occurs as blocks within the Uivak Gneiss and the other consists of two blocks within the mafic rock zone at Localities B and C (Fig. 4A). We found at least eight layers of BIFs, which are fragmented due to intrusion of Uivak Gneiss. The BIF layers are interlayered with mafic rocks, and varied in thickness. The thickness reaches 31.5 m in the northwestern part, Locality A (Fig. 9F), but most layers become thin to the southeast and fade away. The layering of white and dark blue layers is clear in most places, and especially, the very thick BIF in the Locality A displays even slumping structures as well as rhythmical layering (Fig. 9E). The BIFs are composed of magnetite, orthopyroxene, clinopyroxene, and quartz with subordinate amounts of cummingtonite, hornblende, pyrite and apatite. White and black bands are distinct even on thin sections. The white bands mainly comprise relatively large quartz whereas the black bands consist mainly of magnetite, clinopyroxene, and orthopyroxene with subordinate amounts of cummingtonite and hornblende. The magnetite forms bands, and the pyroxenes are present as discrete grains, and almost parallel to the bands. Most clinopyroxene grains have exsolution lamellae of orthopyroxene. The ultramafic rocks occur in two zones as mentioned above. The ultramafic rocks in the northeastern zone occur as block within the Uivak Gneiss but continue laterally. On the other hand, the ultramafic rocks in the southwestern zone separately occur at two areas, Localities B and C, and no ultramafic rocks are present between them. In addition, there is no extension to the northwest and southeast (Fig. 4A). The geological structure suggests a fault under the ultramafic rock bodies. Namely, this area comprises at least two fault-bounded blocks, which consist of lower ultramafic rocks and upper mafic rocks interlayered with BIF layers. We collected 221 samples in the Nulliak Island.





**Fig. 13.** Lithostratigraphy of the Nulliak supracrustal rocks of a fault-bounded block in the St John's Harbour East. The lithostratigraphy comprises ultramafic rocks, mafic rocks and pelitic rocks, interlayered with chert and carbonate rocks, in ascending order. The lithostratigraphy is truncated by upper mafic rocks.

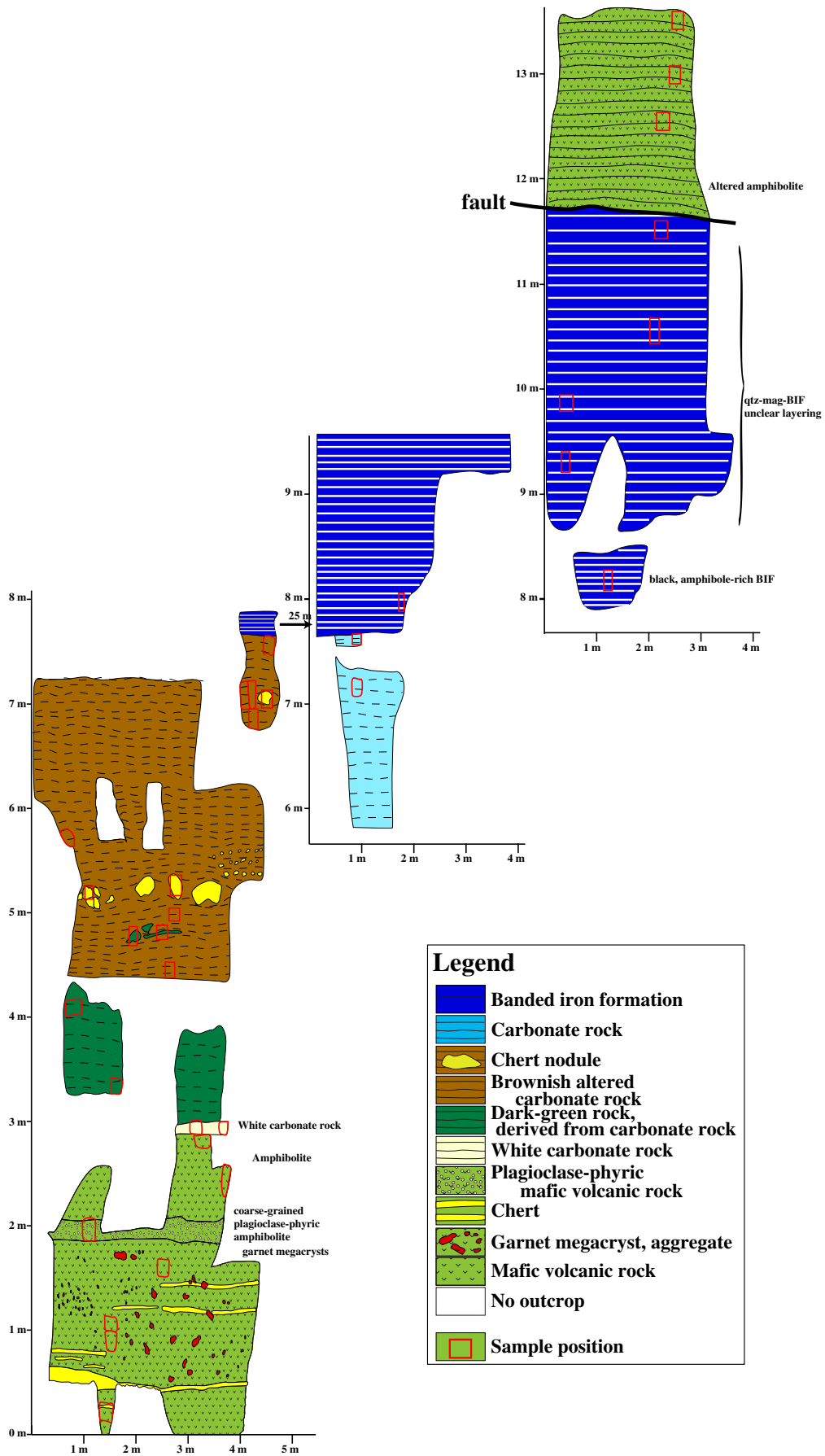
Fig. 4B shows a geological map of the opposite side of the Nulliak Island. The region is underlain by the supracrustal rocks and Uivak Gneiss, which are intruded by granitic intrusion with NW–SE-trending. The supracrustal rock contains BIFs as well as mafic and ultramafic rocks. They are intruded by the Uivak Gneiss so that they occur as blocks or belts within the Uivak Gneiss (Fig. 4B). The BIFs occur in two areas: two layers in the northeastern area and one layer in the central area (Fig. 4B). We collected 61 samples in this area.

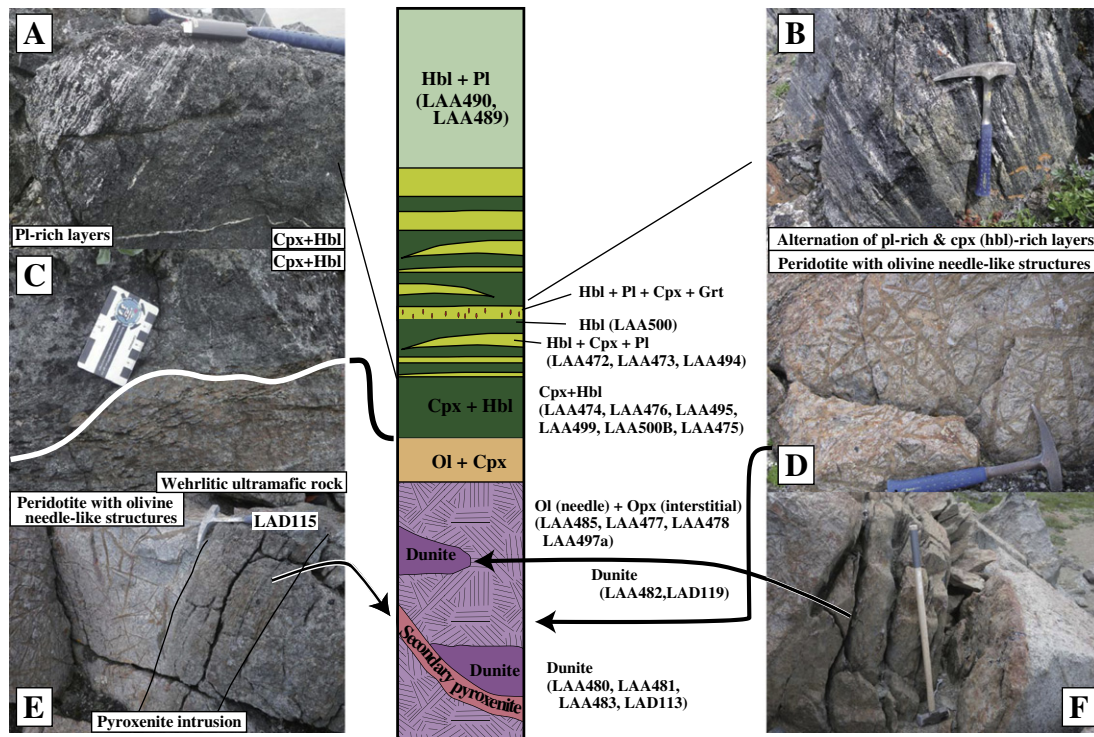
#### 4.4. Geology of the St John's Harbour East

Fig. 5 shows our geological map of the northeastern part of St John's Harbour Fjord, named St John's Harbour East (Fig. 1C). There are the Iqaluk–Uivak Gneisses and supracrustal rocks, together with Saglek

dykes, young granite intrusions and Proterozoic mafic dikes. The Saglek dykes are present in the Localities A and B (Fig. 5). The Iqaluk–Uivak Gneiss is mainly distributed in the western side, and intruded by the Saglek dykes and granitic intrusions. Because both the Iqaluk–Uivak Gneiss and granitic intrusions are white in color and the granitic intrusions are thin and have network-like distribution, it is difficult to distinguish granitic intrusions within the Iqaluk–Uivak Gneiss. The granitic intrusions may be more distributed there, but reproduction of all the young, thin granite intrusions is beyond the scope of this study. The supracrustal rocks contain pelitic rocks, mafic rocks, ultramafic rocks, carbonate rocks, chert and conglomerate. Because pelitic rocks are predominant, the supracrustal rocks were regarded as the Upernavik supracrustal rocks (Bridgwater and Schiøtte, 1991; Ryan and Martineau, 2012; Schiøtte et al., 1992). However, the presence of the

**Fig. 14.** Simplified, sketch maps of a sequence from mafic rocks through carbonate rocks to BIFs in the Pangertok Inlet area. The mafic rocks have garnet megacrysts and are interlayered with thin chert beds. The lower carbonate rock is silicified into whitish carbonate rocks, piles of thin cherts or green rocks. At the upper part, the degree of weathering of the carbonate rocks is highly variable laterally from brown, highly weathered to bluish, relatively fresh dolostone. The carbonate rocks contain siliceous nodules. The BIFs, which suffered from severe weathering, overlie the carbonate rocks.





**Fig. 15.** A lithostratigraphy of an igneous body of ultramafic and mafic rocks at the Shuldham Island area, and photographs of outcrops. The lithostratigraphy comprises harzburgitic ultramafic rocks with olivine (Ol) needle-like structures, intruded by pyroxenitic and dunitic intrusions, wehrlitic ultramafic rock, clinopyroxene (Cpx)–hornblende, interlayering of the hornblende with plagioclase (Pl)–phyric white layers, and fine-grained amphibolite in ascending order. (A) A lower part of the coarse-grained amphibolite, interlayered with thin Pl–phyric layers. (B) Alternation of white Pl-rich and black, Cpx–hornblende layers. (C) Boundary between black, Cpx–hornblende (upper) and brownish wehrlitic ultramafic rock (lower). (D) Harzburgitic ultramafic rock with large needle-like olivine crystals. (E) A pyroxenite intrusion into the harzburgitic ultramafic rock. (F) A dunitic intrusion into the harzburgitic ultramafic rock. Opx: orthopyroxene, Hbl: hornblende, and Grt: garnet.

Saglek dykes in this region indicates that the supracrustal rocks belong to the Nulliak supracrustal rocks. They underwent amphibolite facies metamorphism, but there is no evidence for granulite facies condition in this area (e.g. Schiøtte et al., 1992). The supracrustal rocks of ultramafic, mafic and pelitic rocks are sporadically present within the Iqaluk–Uivak Gneiss body in the western side. On the other hand, the pelitic rocks are abundant in the eastern side. The boundary between the ultramafic rock bodies and pelitic rocks is ambiguous in most places, but the boundary, which can be observed in Locality C, forms a shear zone so that it is tectonic. Fig. 13 shows the lithostratigraphy, reconstructed in Locality D, from the ultramafic rocks through mafic rocks to sedimentary rocks such as pelitic rock, carbonate rock, and chert in ascending order. The bottom boundary of the ultramafic rock is tectonic, as mentioned above. On the other hand, the upper boundary between the mafic rock and pelitic rock is also tectonic because the fabric of the mafic rock is oblique to that of the pelitic rock (Figs. 5 and 13). Namely, the fault-bounded tectonic block in this region has lithostratigraphy from the ultramafic rocks through mafic rocks to sedimentary rocks such as pelitic rock, carbonate rock, and chert in ascending order. A putative BIF layer is interlayered with the mafic rocks (Fig. 5). The putative BIF rock has much magnetite and is quite heavy; similar to the BIF, but the relationship with a granitic intrusion and transitional change from mafic rock indicates metasomatic origin. There are two stratigraphic levels of carbonate layers (Figs. 5, 10C, D and 13), and some carbonate rocks display stromatolite-like, domal structures (Fig. 10C). The microscopic observation shows that most carbonate rocks consist of marble or coarse-grained homogeneous carbonate minerals but some of them contain rounded to irregular globules of fine-grained carbonates within coarse-grained carbonate matrix, possibly still preserving primary textures such as ooids or micrite grains within sparry carbonate (Fig. 10E). In the Locality E, conglomerates are present (Fig. 12A to C). The clasts are mostly quartzite, and the largest clast reaches over

boulder size. The clasts in the conglomerates are highly deformed, utilized to estimate the degree of deformation. Fig. 12A shows a conglomerate observed from the south whereas Fig. 12A and B shows the clasts of a conglomerate observed from the east, indicating more significant extension along NS-direction. If the original shapes were rounded, the aspect ratios of the clasts range from 1.9 to 46.8. But, the measured aspect ratio would be minimum and much larger because there are many thin, fragmented quartz boudins, which cannot be measured. The chert displays interlayering of green and white layers (Figs. 9H and 13), and is predominantly composed of quartz.

In addition to the occurrence of the shear zone and fault at Localities C and D, we can observe some tectonic contact between mafic and pelitic rocks. In the Locality F, the bedding planes of pelitic rocks are oblique to the fabric of the mafic rocks (Fig. 12E). It is difficult to find any boundary in the pelitic rock-dominant region, free of mafic rocks, but at least four tectonic blocks can be recognized in this area.

#### 4.5. Geology of the Pangertok Inlet area

Fig. 6 shows a geological map of the eastern side of the Pangertok Inlet (Fig. 1C), and that the supracrustal rocks occur as two belts within the orthogneiss. The Saglek dykes are found in Localities A and B, indicating that the supracrustal rocks and orthogneiss belong to the Nulliak supracrustal rocks and Iqaluk–Uivak Gneisses, respectively. The supracrustal rocks and Iqaluk–Uivak Gneisses mostly strike from 20° west to 20° east, but dip moderately to gently to the west in the western side and steeply to moderately to the east in the eastern side, which displays antiform (Fig. 6). The supracrustal belts consist of mafic rocks, BIF, chert, and carbonate rocks. Ultramafic rocks and mafic rocks occur as enclaves within the Iqaluk–Uivak Gneiss body (Fig. 6). The ultramafic rocks are highly coarse-grained, contain olivine, orthopyroxene and clinopyroxene together with amphibole and display compositional

banding (Fig. 9C and D). The rocks are locally serpentinized and metasomatized to partially form carbonate rocks. Carbonate rocks and cherts are widely distributed in the supracrustal belts. The chert is composed of silicified carbonate rocks as well as pure white chert (Fig. 10H). The silicified carbonate rocks contain green clinopyroxene and quartz with amphibole, and occur only in the Pangertok Inlet area, consistent with the higher metamorphic grade. The carbonate rocks are present in both the eastern and western sides, and some carbonate rocks directly overlie the mafic rocks. In the western side, the carbonate rock and related partially-silicified carbonate rocks are interlayered with BIFs. The carbonate rocks contain siliceous nodules. The degree of weathering of carbonate rocks varies laterally from fresh bluish dolostone to highly weathered, sheared brownish gray rock with rough surface (Fig. 14). In the eastern side, the carbonate rocks are more severely silicified, and accompanied with chert in many places (Fig. 10H).

Seven BIF layers are present in this region, and they are thin, and fade away to the south, especially in the western side (Fig. 6). In addition, they do not directly overlie mafic rocks but occur on carbonate rocks or cherts in most places (Figs. 6 and 14), in contrast to those in the Nulliak Island and the opposite side (Fig. 4A and B). They contain magnetite, clinopyroxene and quartz with minor amounts of hornblende and pyrite, and comprise white and black bands. The white bands are mainly composed of anhedral quartz whereas the black bands mainly comprise magnetite and clinopyroxene with/without hornblende.

The BIF layers and possibly carbonate rock layers apparently branch in some places (Fig. 6). The branching cannot be considered as a primary structure but can be explained by faults because the BIF and carbonate rocks originate from sediments. In addition, a similar lithostratigraphy from mafic rock through carbonate rock to BIF can be found all over the area, and their duplication can be observed especially in the western side. Fig. 14 shows the detailed lithostratigraphy from the mafic rock through carbonate rocks to BIFs in the western side, described at Locality C. The mafic rocks are interlayered with thin white cherts, and contain large garnet grains and aggregates in the lower part and plagioclase grains in the upper part. Carbonate rocks overlie the mafic rocks, and comprise three parts: white, weathered carbonate rocks at the bottom, dark-green carbonate rocks or piles of thin white chert layers with dark-green nodules and boudins at the middle, and bluish dolostones or highly weathered, sheared brownish gray rocks with rough surface at the upper part, respectively (Fig. 14). The highly weathered rocks contain many white siliceous nodules. The BIFs with white and black bands overlie the carbonate rocks. In most places, the BIF is highly weathered, and reddish brown in color. The sedimentary sequence is overlain by mafic rocks through a fault (Figs. 6 and 14). The branching of sedimentary rocks and duplication suggest that the supracrustal belts in the Pangertok Inlet comprise fault-bounded tectonic blocks with similar lithostratigraphy. The distribution of the BIF layers, carbonate layers and faults suggests that the faults merge to the south, and appears a duplex structure.

#### 4.6. Geology of the western coast of the Shuldham Island

A small point of the western coast of the Shuldham Island is famous for the occurrence of large needle-like olivines in ultramafic rocks (e.g. Collerson et al., 1976b, Fig. 1C). Fig. 7 shows a geological map of the point where ultramafic rocks crop out well. The supracrustal rocks consist of ultramafic rocks of needle-like olivine-bearing harzburgitic peridotite, dunitic peridotite, clinopyroxene-rich peridotite, pyroxenite intrusions, clinopyroxene-rich hornblende and clinopyroxene-poor hornblende, interlayering of plagioclase-rich white layers with clinopyroxene + hornblende-rich dark green layers, fine-grained mafic rock and pelitic rocks, in ascending order (Figs. 7 and 15). The clinopyroxene-rich peridotite occurs along the boundary between the ultramafic and mafic rocks. The dunitic peridotite is accompanied with pyroxenite intrusions, and dunitic peridotite and pyroxenite intrusions

are oblique to the fabric of the harzburgite with needle-olivines (Fig. 7). The pyroxenite intrusions were possibly derived from secondary intrusions of mafic or felsic magma or siliceous fluid. Any compositional or mineralogical transition cannot be observed within the harzburgitic peridotite, as well as no olivine spinifex structures. The mafic rocks mainly comprise three parts. Massive black rocks of clinopyroxene and hornblende occur at the bottom. The lower part of mafic rock unit is characterized by interlayering of black and white layers; upward increasing and thickening of the plagioclase–phyric white layers. The upper part is fine-grained amphibolite. Although granitic intrusions run along the boundary, the pelitic rocks overlie the mafic rocks (Fig. 7). Finally, the Iqaluk–Uivak Gneiss is intruded into the sequence. The lithostratigraphy is similar to an ophiolite stratigraphy but is quite different from a peridotitic komatiite sheet flow because of harzburgitic ultramafic rocks and gabbroic parts (e.g. Arndt et al., 1977; Viljoen et al., 1982). The peridotitic komatiite, generally speaking, was not saturated in orthopyroxene during the fractional crystallization (Parman et al., 1997). But the total thickness from the ultramafic rocks to the pelitic rocks is about 170 m, apparently much thinner than calculated Archean oceanic crust (e.g. Davies, 1992; Komiya, 2004; Komiya et al., 1999, 2002b, 2004; Ohta et al., 1996; Sleep and Windley, 1982). However, there are many modern analogies of thin basaltic oceanic crusts on ultramafic bodies such as south-western Indian ridge, Australia–Antarctica Discordance (AAD), Vema fracture zone and Hess deep, and some sections have no sheeted dike complex (Juteau and Maury, 1999). In addition, the aspect ratios of clasts in conglomerate indicate significant extension along NS direction, as mentioned above. We can calculate the minimum thickness of the oceanic crust to be ca. 8000 m using the estimated extension ratio.

## 5. Discussion

### 5.1. Origins of pelitic and carbonate rocks

Pelitic rocks and accompanied conglomerate are ubiquitously distributed through the Saglek Block (Figs. 2 to 7). The pelitic rocks are interlayered with mafic volcanic rocks, and are accompanied with carbonate rocks in the western part of Big Island and ultramafic rocks in the northern part of the St John's Harbour South. However, the origin of pelitic gneiss is very enigmatic not only in the Saglek Block but also in all the Eoarchean greenstone belts. For example, in the Isua supracrustal belt, an old geological map showed pelitic rocks as “variegated schist formation” (Nutman, 1986), but the pelitic rocks are removed in a new geological map because the pelitic rocks were interpreted as altered amphibolites (Nutman and Friend, 2009).

Fig. 11 shows the representative occurrence of some types of pelitic rocks. Fig. 11A shows the true pelitic rocks based on the occurrence and petrography. The pelitic rocks contain biotite, garnet and quartz with subordinate amounts of graphite and pyrite and nearly lack amphibole and plagioclase (Fig. 11B). The biotites form layers whereas the garnets occur randomly, consistent with modern pelitic rocks. On the other hand, the occurrence and mineralogy of some pelitic rocks were possibly formed from alteration and metasomatism of mafic and ultramafic rocks and granitic intrusions (Fig. 11C to G). They display transitional change from the host rocks of mafic and ultramafic rocks and granitic rocks to the brownish putative pelitic rocks (Fig. 11C, E to G). Their mineralogy is also distinct from the true pelitic rocks, and for example, the putative pelitic rocks, derived from mafic rocks, contain more amphibole and plagioclase (Fig. 11D). The apparent similarity between the true and putative pelitic rocks, which display brownish colors and banding, makes distinguishing the true pelitic rocks from the secondary pelitic rocks difficult but the detailed observation of the field occurrence and petrography is useful to find the true pelitic rocks. The geochemistry will be also useful to classify the pelitic rocks in the future (Bolhar et al., 2005). The presence of pelitic rocks indicates that older sialic continental crust already existed and supplied K–Al-rich mature detritus to the depositional environment even in the Eoarchean, in contrast to a

conventional view of the lack of evolved continental crust. The graphite clots in the pelitic rocks will be utilized to discover evidence for early life.

The origin of carbonate rocks is commonly enigmatic in the Eoarchean greenstone belts. For example, the origin of the carbonates in the Eoarchean Isua supracrustal belt is disputed between chemical sedimentary and metasomatic origins, as mentioned above. The extensive carbonization of greenstones and ultramafic rocks is common in the Archean greenstone belts (e.g. [Duchač and Hanor, 1987](#); [Hofmann et al., 2007](#); [Nakamura and Kato, 2004](#); [Shibuya et al., 2007b](#)) due to high CO<sub>2</sub> contents of atmosphere, seawater and consequent hydrothermal fluid ([Nakamura and Kato, 2004](#); [Shibuya et al., 2007b](#)). In the Nulliak supracrustal rocks, there are two types of carbonate rocks. The first type is accompanied with supracrustal rocks of BIFs and pelitic rocks, and displays rounded to irregular, fine-grained carbonate globules within sparry coarse-grained carbonates for weakly recrystallized samples. The other is accompanied with ultramafic rocks, and contains possible pseudomorphs of olivine, replaced by fine-grained carbonate, within coarse-grained carbonate matrix. The former originates from sedimentary carbonates, deposited in seawater whereas the latter was possibly formed by carbonation of mafic and ultramafic rocks. The occurrence of carbonated ultramafic rocks, intruded by granitic and dioritic intrusions, indicates that the carbonation was due to intrusion of felsic magmas ([Fig. 10F](#)), but some carbonated ultramafic and mafic rocks, free of the felsic intrusions, were possibly derived from regional carbonation by high CO<sub>2</sub> fluid in the Eoarchean during metamorphism, analogous to the Paleoproterozoic and Mesoproterozoic carbonated greenstones ([Nakamura and Kato, 2004](#); [Polat et al., 2007, 2008](#); [Shibuya et al., 2007b](#)).

### 5.2. The relationship between the Nulliak supracrustal rocks and ambient Iqaluk–Uivak Gneisses

The supracrustal rocks and orthogneisses intruded by the Mid-Archean Saglek dykes are defined as the Nulliak supracrustal rocks and Uivak Gneisses, respectively ([Bridgwater et al., 1975](#)). Besides, it is considered that the Nulliak supracrustal rocks are older than the Uivak Gneiss ([Bridgwater and Collerson, 1977](#)). However, the relationship between the supracrustal rocks and Uivak Gneiss is obscure in most places, thus it is ambiguous to distinguish the Nulliak supracrustal rocks from young, if present, supracrustal rocks. As a result, speculation that clastic sediments were scarce in the Early Archean whereas they became relatively abundant since the Middle Archean (e.g. [Ronov, 1972](#)) leads to considering that the supracrustal rocks with abundant pelitic rocks are young Upernavik supracrustal rocks. However, there is geochemical evidence for the clastic sediments even in the Hadean ([Ushikubo et al., 2008](#)). Recycling of continental crusts has been emphasized for a long time ([Armstrong, 1981](#); [Fyfe, 1978](#); [Veizer, 1976](#)), and recent geochemical studies support that there were significant amounts of continental crust in the Hadean and Early Archean ([Harrison, 2009](#); [Komiya, 2011](#)). Another lithological difference, namely BIF, may be applicable to distinguish the Nulliak supracrustal rocks, but the thickness of BIFs is highly variable laterally, indicating that the occurrence of BIFs does not depend on the depositional age but on depositional environment, namely distance from the volcanic centers or hydrothermal vents. Besides, no basal conglomerates between the supracrustal rocks and Uivak Gneisses are inconsistent with the idea that the supracrustal rocks with abundant pelitic rocks overlay the Uivak Gneiss.

This study newly reports the Saglek dykes in the supracrustal belts of the St John's Harbour South and Big Island, which were classified as the Upernavik supracrustal rocks before ([Bridgwater and Schiøtte, 1991](#); [Bridgwater et al., 1975](#); [Ryan and Martineau, 2012](#)). [Glikson \(1977\)](#) claimed that the presence of the Saglek dyke is not appropriate to distinguish the Early Archean rocks because distribution of mafic dikes is to a large extent controlled by differential properties of the

host rocks. In addition, the fact that the Saglek dykes are too thin and metamorphic rocks, apparently similar to amphibolites in the supracrustal rocks, strengthens the objection. On the other hand, our geological survey shows that the supracrustal rocks were intruded by the Iqaluk–Uivak Gneisses in some places, and indicates that the supracrustal rocks are older than the Eoarchean Iqaluk–Uivak Gneisses. Namely, to obtain ages of ambient orthogneisses allows us to constrain the ages of the supracrustal rocks. Recent reappraisal of the U–Pb dating and detailed cathodoluminescence observation of zircons in the oldest suite of the Iqaluk–Uivak Gneiss series, defined by geological occurrence, showed that the age of the precursor of the Iqaluk–Uivak Gneiss was over 3.95 Ga ([Shimojo et al., 2012, 2013; 2015](#)). Because the Iqaluk–Uivak Gneiss was intruded into the supracrustal rocks, the supracrustal rocks are formed at >3.95 Ga. As a result, the Nulliak supracrustal rocks are the oldest supracrustal rocks in the world because the current oldest supracrustal rocks occur in the 3.83 Ga Akilia Island and 3.81 Ga Isua supracrustal belt, southern West Greenland ([Crowley, 2003](#); [Mojzsis and Harrison, 2002a](#); [Nutman and Friend, 2009](#); [Nutman et al., 1996](#)). Because orthogneiss is, of course, lithologically distinct from amphibolites, it is relatively easy to recognize the relationship between the amphibolites and felsic rocks. In addition, we can obtain the ages of precursors of the orthogneisses because they contain zircons. As a result, the orthogneiss is a much better geological marker than the Saglek dykes. In addition, all of the supracrustal rocks are basically composed of similar lithologies of ultramafic and mafic rocks, BIF, carbonate rock, chert and pelitic rock, and have similar strike and dip one another (e.g. [Bridgwater et al., 1975](#); [Ryan and Martineau, 2012](#); [Figs. 2 to 7](#)). The regional geology and their field occurrence support that all of the supracrustal rocks have the same origin, and were formed at >3.95 Ga.

The internal structures of the supracrustal belts did not receive so much attention for a long time. The supracrustal rocks on previous geological maps display many tight to isoclinal folds and branching of mafic rocks and pelitic rocks in many places ([Bridgwater et al., 1975](#); [Ryan and Martineau, 2012](#)). However, most of them are due to tracing sporadically distributed outcrops in orthogneisses, granitic intrusions and vast pelitic rocks. We traced BIF and pelitic rock layers one by one to reconstruct their original distribution, and found some branching of them, as mentioned above. The branching of sedimentary rocks can be explained by faults, which merge to the south in St John's Harbour South and Pangertok Inlet. In addition, the fault-bounded blocks have similar lithostratigraphies from ultramafic rocks through mafic rocks to sedimentary rocks in ascending order, and are piled up. Namely, the supracrustal belts display imbricate structure. The similarity of the structure and lithostratigraphy as modern accretionary complexes implies that the supracrustal belts in the Saglek Block originated from accretionary complex in the Eoarchean, analogous to the Archean greenstone belts in Pilbara, Western Australia ([Ohta et al., 1996](#)), Superior ([Kimura et al., 1993](#)) and southern West Greenland ([Komiya et al., 1999](#)).

### 5.3. Depositional environment of the Nulliak supracrustal rocks and emplacement of the Nulliak supracrustal belts and Iqaluk–Uivak Gneisses

The tectonic setting of the Archean greenstone belts, generally speaking, is controversial. Uniformitarians explain tectonic settings of greenstone belts inside framework of plate tectonics (e.g. [de Wit et al., 1987, 1992](#); [Furnes et al., 2007](#); [Komiya et al., 1999](#); [Kusky et al., 2001, 2013](#); [Polat et al., 2002](#)) whereas non-uniformitarians propose tectonic models peculiar to the early Earth (e.g. [Bédard, 2006](#); [Hickman, 2004](#); [Kröner, 1985](#); [van Hunen et al., 2008](#); [Zegers and van Keken, 2001](#)). In addition, the relationship between greenstone belts and ambient granitoid batholiths is also controversial; some propose that basaltic lava and chemical sediments in the greenstone belts were allochthonous, and were intruded by young granitoid batholiths after their

emplacement (e.g. Kitajima et al., 2001; Terabayashi et al., 2003) whereas others emphasize that they were autochthonously formed on preexisting sialic crust (e.g. Nijman et al., 1998; van Kranendonk et al., 2002) or were in-situ intruded by granitoid during the formation of basalts and sediments in plume-related large igneous provinces (e.g. Pease et al., 2008; Reimink et al., 2014; Rudnick, 1995; van Kranendonk et al., 2002). The basaltic greenstones and chemical sedimentary layers appear to be interlayered in most greenstone belts. The origins of the structure are often under debate whether tectonic or conformable, for example those in the Paleoproterozoic greenstone belts, Pilbara (e.g. Nijman et al., 1998; Terabayashi et al., 2003; van Kranendonk et al., 2002, 2004). On the other hand, it is widely considered that the continental masses were grown and enlarged through collision–amalgamation after formation of continental crusts of granite–greenstone belts based on a concept of terrane tectonics of plate tectonics (Coney et al., 1980) even for the Eoarchean and Paleoproterozoic terranes (e.g. Friend et al., 1988; Hickman, 2004; McGregor et al., 1991; Nutman and Collerson, 1991; Nutman and Friend, 2009; van Kranendonk et al., 2002).

The Isua supracrustal belt is one of the most intensively studied terranes with the Eoarchean ages. The supracrustal belt also contains ultramafic rock, mafic rock, banded iron formation, chert, carbonate rock, pelitic rock and conglomerate, and is intruded by the Itsaq Gneiss (Appel et al., 1998; Komiya et al., 1999; Myers, 2001; Nutman, 1986; Nutman and Friend, 2009). Recently, it is proposed that the supracrustal belt was formed through collision–amalgamation of two terranes with different histories based on U–Pb dating of zircons from sedimentary and volcanic rocks in the belt and ambient orthogneisses (e.g. Crowley, 2003; Nutman and Friend, 2009; Nutman et al., 2009). But, the tectonic setting of mafic magmatism, depositional environment of chemical sediments of BIF, chert and carbonate rock, and mechanism of their emplacement in a continental crust are still enigmatic. Komiya et al. (1999) proposed that the oceanic crust of basaltic rocks, derived from mid-ocean magmatism, and chert, deposited in pelagic and deep-sea environment, was covered with clastic sediments at a trench, and that the package was accreted to a continental crust during subduction based on the geological characteristics. On the other hand, suprasubduction environment for the tectonic setting of basaltic magmatism was more widely considered based on the geochemistry of basaltic greenstones (Furnes et al., 2014; Polat et al., 2002). If the two models are accepted, formation of the Isua supracrustal belt can be explained by accretion of an oceanic crust of abyssal arc basalts, derived from immature subduction magmatism in intra-oceanic environment, and chemical sediments, deposited until the subduction, to a mature continental crust, analogous to modern Izu–Mariana arc or Mineoka ophiolite in Japan (e.g. Hirano et al., 2003; Tani et al., 2010).

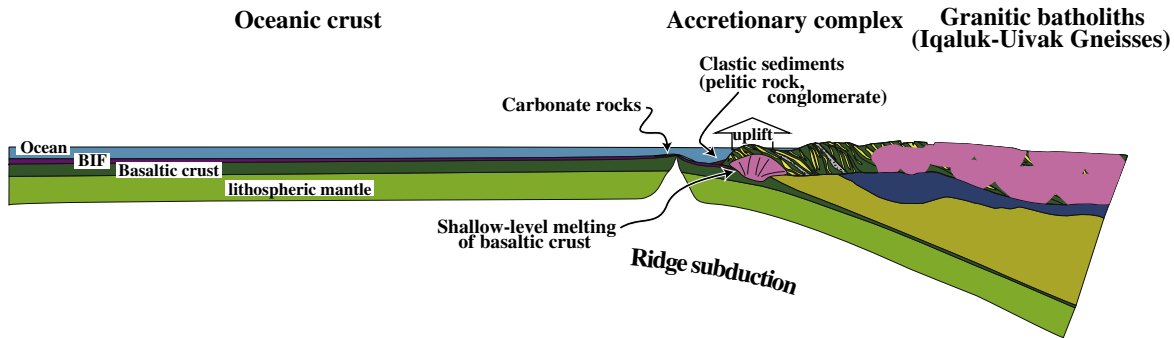
The Nulliak supracrustal rocks have many similarities, such as the presence of mafic and ultramafic rocks, BIF, pelitic rocks and carbonate rocks, and geological structures of imbricate and duplex structures, to the Isua supracrustal belt. But, abundances of pelitic rocks and cherts are distinct between them. The Nulliak supracrustal rocks have much more abundant in the pelitic rocks and almost lack chert. The lack of chert, pelagic and deep-sea sediment, makes it difficult to recognize ocean plate stratigraphy because the relationship between chert and clastic sediments is useful to find thrusts (Komiya et al., 1999; Kusky et al., 2013; Matsuda and Isozaki, 1991). In modern accretionary complexes and ophiolites, the thickness of the pelagic sediments depends on the distance from mid-ocean ridge to subduction zone (Isozaki, 1996; Ueda and Miyashita, 2005). Some modern ophiolites such as Trinity, Josephine and Del Po ophiolites have lithostratigraphy that clastic sediments directly overlie mafic and ultramafic rocks, and they are considered as suprasubduction ophiolite (Gillis and Banerjee, 2000). On the other hand, Taitao ophiolite, one of the youngest ophiolites, also has the similar lithostratigraphy of clastic sediments directly on basaltic volcanic rocks, but it is considered that the ophiolite

was formed at mid-ocean ridge based on the geochemistry of the basaltic lavas and the absence of fore-arc volcanism (Bourgeois et al., 1993; Guivel et al., 1999; Shibuya et al., 2007a). There are also some examples of accretion of oceanic materials during ridge subduction, such as the Late Cretaceous to Paleogene Tomuraushi greenstone complex and Shimanto belt (Kiminami et al., 1994; Miyashita and Katsushima, 1986). In those greenstone belts, basaltic flows with MORB affinity are interlayered with clastic sediments, basaltic dikes are intruded into the clastic sediments, and there are no radiolarian cherts (Kiminami et al., 1994; Miyashita and Katsushima, 1986). The lithostratigraphies in those accretionary complexes and ophiolites provide a modern analogy to the Nulliak supracrustal rock.

Fig. 16 shows a cartoon of the formation of the Nulliak supracrustal rocks and intrusions of Iqaluk–Uivak Gneisses, derived from shallow-level slab melting of the subducted hydrated basaltic crust during the ridge subduction, analogous to the Isua supracrustal belt (Komiya et al., 1999; Polat and Frei, 2005). The carbonate rock was deposited around, relatively shallower, mid-oceanic ridge, and the BIF was deposited off the ridge. The presence of abundant pelitic rocks and lithostratigraphy of pelitic rocks directly on the mafic rocks suggests that the accreted oceanic crust was so young that pelagic and deep-sea sediments of BIF and chert were not deposited because of short distance from mid-oceanic ridge to subduction zone, as mentioned above. In a modern tectonic setting of such young plate subduction, a volcanic front of the Calc-alkaline rock series moves to the trench (e.g., Anma et al., 2006), and the hanging wall is uplifted on a regional scale because of buoyancy (DeLong and Fox, 1977; Kusky et al., 1997). In these cases, thick piles of turbidite cap the subducting oceanic crust before subduction. Namely, large amounts of clastic sediments, including conglomerate, are supplied to the trench, and cover on the basaltic mafic crust. In addition, ridge subduction promoted settlement of ophiolite on continental crust (Dilek and Furnes, 2011; Van den Beukel and Wortel, 1992; Wakabayashi and Dilek, 2003), for example, the Taitao ophiolite (Anma et al., 2006; Bourgeois et al., 1993; Le Moigne et al., 1996; Nelson et al., 1993) and Resurrection Peninsula and Knight Island ophiolite, Alaska (Lytwyn et al., 1997).

The orthogneisses in the Archean granite–greenstone belts commonly have several generations in relatively small areas, and their cross-cutting relationships can be observed in many places such as the Acasta Gneiss Complex (Bowring and Housh, 1995; Bowring and Williams, 1999; Iizuka et al., 2007; Mojzsis et al., 2014), Isua area (Crowley, 2003; Crowley et al., 2002; Nutman and Friend, 2009; Nutman et al., 1996, 2009), and Saglek Block (Collerson, 1983; Krogh and Kamo, 2006; Schiøtte et al., 1989a,b; Shimojo et al., 2012, 2013, 2015). The age distributions are intermittent but continue for more than 100 m.y. For example, in the Acasta Gneiss Complex, the granitoid formation intermittently took place from 4.03 to 3.58 Ga, and even the first period continued from 4.0 to 3.94 Ga (Bowring and Housh, 1995; Bowring and Williams, 1999; Bowring et al., 1989; Iizuka et al., 2007). In the Isua area, the granitoid formation continued from 3810 to 3600 (Crowley, 2003; Crowley et al., 2002; Nutman and Friend, 2009; Nutman et al., 1996, 2009). In the Saglek Block, the Iqaluk–Uivak Gneiss series also have several generations (Fig. 8A, C and D), and over five generations in some points (Shimojo et al., 2012, 2013, 2015). And, their formation intermittently continued from 3.95 Ga to ca. 3.7 (Collerson, 1983; Krogh and Kamo, 2006; Schiøtte et al., 1989a,b; Shimojo et al., 2012, 2013, 2015). If the Lister Gneiss and young granite intrusions are included, the duration exceeded six million years (Schiøtte et al., 1989a). Assuming that plate tectonics were operated, such a long duration is inconsistent with plume-related continental formation because an ocean plateau moves off from a plume center. If the Archean oceanic plates were smaller and shorter-lived (de Wit and Hart, 1993; Hargraves, 1986; Komiya, 2004), the duration was much shorter. On the other hand, there are many examples in the Phanerozoic accretionary complexes, where granitoid formations intermittently continued for over 200 Ma. For example, in Tateyama

## Plate history of the >3.95 Ga oceanic plate, and its emplacement on a continental crust



**Fig. 16.** A cartoon of plate history of the Eoarchean oceanic lithosphere. The oceanic lithosphere was formed at mid-oceanic ridge, and then carbonate rocks and subsequent banded iron formation were deposited on the oceanic crust. The upper part was off-scraped and accreted to a continental crust during the subduction underneath the continental crust. Ridge subduction accounts for lack of thick cherts and BIFs, predominance of pelitic rocks and formation of ophiolite.

area of the middle part of Japanese Island, a 250 to 200 Ma accretionary complex is intruded by four generations of granitoids at 1.7–7, 52–65, 65–100 and 170–200 Ma within <5 km<sup>2</sup> and three of them display cross-cutting relationship one another at the same place (Harayama et al., 2000). In addition, the accretionary complex is bounded by a 170–200 Ma tonalite batholith in the western side and by a 65–100 Ma granite batholith in the eastern side within only 2 km (Harayama et al., 2000). In Gorny–Altai tectonic unit, an Ediacaran–Cambrian accretionary complex is intruded by ca. 400, 350 and 250 Ma granodiorites, and the 350 and 250 Ma granodiorites are in contact with each other (Glorie et al., 2011). Namely, the intermittent granitoid formation for over 200 Ma is quite common in an active plate margin because a trench often moves back and forth depending on the age of a subducting oceanic plate. On the other hand, collision–amalgamation of different terranes, namely a terrane tectonic model, is often proposed based on the geological structure that a supracrustal belt is bounded by granitoids with different ages in the Archean terranes (e.g. Nutman and Friend, 2009; Nutman et al., 2009). However, the modern examples suggest that the geological structure does not necessarily need terrane tectonics but can be formed through accretion of oceanic materials and later intermittent granitoid magmatism. Because granitoid formation takes place repeatedly but formation of accretionary complex is only once in the place, the oldest suites of the granitoids constrain the minimum age of the accretionary complex, namely supracrustal rocks. And, the old, >3.95 Ga, suite of the Iqaluk–Uivak Gneiss series is sporadically distributed and is erased by young granitoid suites in most places, thus some of the Nulliak supracrustal rocks are not directly intruded by the old suite. However, all the Nulliak supracrustal rocks should be formed at >3.95 Ga.

The timing of the beginning of plate tectonics on planet Earth is highly controversial (summarized by Condie and Kröner, 2008). Condie and Kröner (2008) proposed sixteen criteria to recognize the ancient plate tectonics, including ophiolite, accretionary complex and continental crust, and mentioned that the plate tectonic indicators are widely obtained by 2.7 Ga. But, the presence of accretionary complex, ophiolite, mélange-like structures of ultramafic and mafic blocks in pelitic rocks, and granitoid in the Saglek Block favors the plate tectonics even in the Eoarchean, possibly providing the oldest evidence for the plate tectonics on the Earth.

## 6. Conclusions

We carried out geological survey to make geological maps in seven areas of the Saglek Block, northern Labrador. Although previous works suggested that some of the supracrustal belts and the orthogneisses have the Mesoarchean ages because of the lack of the Mesoarchean basaltic intrusions, Saglek dyke, we found the Saglek dykes in the

areas to recognize the Eoarchean Nulliak supracrustal rocks and Iqaluk–Uivak Gneiss in all the areas. Recent reassessment of U–Pb dating and cathodoluminescence observation of zircons from the oldest suites of the Iqaluk–Uivak Gneiss showed that the Iqaluk–Uivak Gneiss has the Eoarchean age, >3.95 Ga. Because our geological survey clearly showed that the Nulliak supracrustal belts are intruded by the Iqaluk–Uivak Gneiss in the areas, the Nulliak supracrustal rocks are the oldest supracrustal rock in the world. The Nulliak supracrustal belts are composed of ultramafic and mafic rocks, chemical sedimentary rocks of banded iron formation (BIF), chert and carbonate rock, and clastic sedimentary rocks of pelitic rocks and conglomerate. The supracrustal belts consist of piles of fault-bounded blocks, which merge to the south in the St John's Harbour South and Pangertok Inlet. The fault-bounded blocks comprise the ultramafic rocks, mafic rocks and sedimentary rocks such as carbonate rocks, cherts, BIFs, pelitic rocks and conglomerates in ascending order, similar to modern ocean plate stratigraphy. In addition, small-scale duplex structures are found in all the areas. The combination of duplex structure and ocean plate stratigraphy indicates that the >3.95 Ga Nulliak supracrustal belts originate from the Eoarchean accretionary complex, analogous to the Isua supracrustal belt. The presence of accretionary complex, ophiolite and granitic continental crust provides the oldest evidence for the plate tectonics on the early Earth.

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