Supplementary Note 1 - Mineralogy, textures, and structures of lithologies observed in rock core

The upper two thirds of the rock core consist of lithologies with sedimentary protoliths. In the main text, these lithologies have been divided into turbiditic, pelagic, banded iron formation, and volcaniclastics. Here we will describe the mineralogy, texture, and structure of these lithological groups.

Turbiditic lithologies

The turbiditic group comprises three main lithologies; 1) Plagioclase-chlorite schists, 2) clinzoisite-mica schists, and 3) garnet amphibolites. Plagioclase-chlorite schists occur in the middle section of the core whereas clinzoisite-mica schists occur in the upper section of the core. Intermediate composition occur within a transition zone between 27 m and 25 m. Garnet amphibolites occur sporadically as 1-50 cm thick layers throughout the middle and upper section of the core, but with greater frequency towards the lower portion of the middle section of the core.

Plagioclase-chlorite schists consist of a matrix of chlorite, quartz and minor biotite with abundant sub-rounded to sub-angular plagioclase porphyroclasts. Often, 1-10 cm thick layers can be distinguished with laminated quartz and chlorite in between. These layers are internally graded with larger plagioclase grains (up to 1 mm), less matrix and poorer clast sorting occurring downwards.

Clinzoisite-mica schists consist of a matrix of muscovite and quartz with variable proportions of biotite. Clinozoisite may occur both as small tabular radiating grains in the matrix and as symplectites with quartz, albite, and muscovite. These symplectitic grains form partial or complete pseudomorphs of plagioclase grains. Just as for the plagioclase chlorite schists, layers of clinozoisite-mica schists are often separated by thinner layers consisting of laminar and more finegrained quartz and muscovite grains. Internal gradation is more difficult to constrain due to plagioclase grains largely having been replaced by pseudomorphs, but some layers do show gradual internal variation in the size and frequency of replaced plagioclase grains.

Garnet amphibolites consists of porphyroblastic green amphibole grains and garnets set in a matrix of chlorite, biotite, quartz, and tabular clinozoisite grains. Plagioclase porphyroclasts occur sporadically and some garnet amphibolite sections grade into plagioclase-chlorite schists, which leads us to our interpretation of this lithology as turbiditic.

Pelagic lithologies

Pelagic section occur as finely laminated layers of chlorite or muscovite, with or without quartz and graphite. Whereas chlorite rich pelagic sections only occur in the middle section of the core, muscovite rich sections occur through the meta-sedimentary part of the core. Mineral proportions vary across layers and in some cases do so rhythmically on multiple scales <1 mm. Pyrrhotite and lessor pyrite grains occur abundantly in graphitic sections, whereas titanite is more common in non-graphitic sections. Ovoid plagioclase grains or pseudomorphs thereafter, occur sporadically and contain graphite trails that continue from the layer in which they are contained. In chlorite rich sections, chlorite grains are incorporated into these plagioclase grains.

Banded iron formation and associated iron rich detritals

Banded iron formations are recognized from alternating quartz and magnetite layers. These occur in two ca. 25 cm thick sections and in several sections that are only a few cm thick. Magnetite is also present in adjacent layers where it is interbedded with chamosite (Fe rich chlorite), green amphibole, and garnets. These adjacent sections also contain fine repeating quartz layers similar to what is observed within the banded iron formation sections, suggesting that sequential chemical sedimentation of iron and silica continued during the deposition of surrounding layers, but that it was contaminated by a detrital component. These layered sections are interlayered with thicker layers consisting of green amphibole and chlorite, due to their high Fe concentrations and internal layering, we interpret these layers to be iron rich chemical sediments mixed with a detrital component.

Supplementary Note 2 - Geological background

Regional geology

The Isua Supracrustal Belt (ISB) lies within the northeastern corner of the Itsaq Gneiss Complex (IGC) in Western Greenland (fig. S1a and S1b). The IGC gneisses consist of trondhjemite-tonalite-granodiorites (TTGs, plagioclase rich granitoids), with subordinate granites and enclaves of mafic and ultramafic rocks[1]. Igneous zircons extracted from the TTGs and granites have given ages between 3.9 Ga and 3.6 Ga[1], with population maximums occurring at ca. 30-50 Ma intervals[2]. Granites generally formed between 3.66 Ga and 3.60 Ga concurrent with a regional metamorphic event[3]. Doleritic dykes known as Ameralik dykes intruded the IGC between 3.5 Ga and 3.2 Ga[4,5] and were therefore unaffected by deformation related to the metamorphic event between 3.7 Ga and 3.6 Ga. They therefore record the effects of later metamorphism and deformation, notably at 2.8 Ga to 2.7 Ga where granulite facies metamorphism and deformation affected the southern IGC and lower amphibolite facies metamorphism accompanied by lesser deformation affected the northern IGC[1,6] (where the ISB is located).

Geology of the Isua Supracrustal Belt

The ISB consists of a 35 km long arcuate body of, mainly, supracrustal rocks (fig. S1c). The geological model of Nutman and Friend[7] divides the ISB into an inner and outer arc, delimited by a high strains zone termed the *dividing sedimentary unit*. The outer arc, which faces west, south and south-east, is the older, with an age of ca. 3.81 Ga, whereas the inner arc has ages between 3.72 Ga and 3.69 Ga. Exposed rocks mostly consist of mafic rocks that display structures and textures that are consistent with an origin as pillow lavas, sheeted dykes and gabbroic intrusives[8,9]. Mafic rocks in the outer arc have trace elemental compositions consistent with that of island arc basalts[10], whereas mafic rocks in the inner arc belong to two distinct populations; one with compositions equivalent to boninites formed in fore-arc environments[11] and the other with compositions similar to those found in the outer arc[10,12]. The mafic rocks are variably deformed and are intercalated with chemical sediments (i.e., cherts and banded iron formation; BIF) and ultramafic units, which originated as cumulates[13]. Intrusive tonalites with zircon ages of ca. 3.81 Ga[14] occur as dykes within the outer arc (western, southern and south-eastern facing sides of the ISB). A more extensive tonalite unit, with the same age and trace elemental composition, also occurs within the outer arc, but this unit been subjected to a high degree of deformation, and potassium and carbonate metasomatism[8,15]. Some authors have characterized this unit as a meta-pyroclastic unit[16], whereas others interpret it as intrusive[8].

Chemical sediments within the Isua Supracrustal Belt

The most common meta-sedimentary lithologies occurring within the ISB are cherts and BIF[17]. These lithologies typically occur as few meters thick units in the inner arc and appear repeatedly along strike for several kilometers. Most BIF units consist of magnetite and quartz bands, but many also consist of layers rich in Fe, Mg and Al silicates, such as amphiboles, chlorite and garnet[18]. Many BIF units contain abundant carbonate, which, in many cases, is derived from replacement of the preexisting minerals during infiltration of CO_2 rich fluids[8]. Nd isotopes and trace elements show that the material deposited in Fe-rich layers ultimately was derived from hydrothermal fluids sourced from seafloor volcanism[19].

Proposed detrital meta-sediments within the Isua Supracrustal Belt

Within the eastern arm of the ISB, several rock exposures have been interpreted as having a detrital sedimentary protolith, but for many of these occurrence no sedimentary structures or textures are reported[20,21]. These interpretations often rely on compositional features such as high potassium and aluminum content to argue for sedimentary protoliths, but such features may be replicated by metasomatic processes, which are well documented within some parts of the ISB[8,15,22]. Others have reported the occurrence of conglomerates and sandstones based on structural and textural evidence[23–26], but others again have argued that the proposed structures and textures were formed through deformation of metasomatically altered rocks[8,22,27,28]. As such, the occurrence of detrital sediments in the eastern arm is not clear and the turbiditic and pelagic meta-sediments in the western arm are therefore the only unequivocal detrital sediments observed within the ISB.

Age relations of the turbiditic and pelagic meta-sediments

Following the geological model of Nutman and Friend[7], the pelagic and turbiditic meta-sediments described in our rock core occur within the inner arc of the ISB, placing them within the age range 3.72 Ga to 3.69 Ga. Nutman et al.[29] sampled meta-sedimentary rocks occurring at the surface and extracted a total of four zircons from three samples. These grains gave $^{207}Pb/^{206}Pb$ ages of 3723 ± 6 Ma, 3724 ± 6 Ma, 3712 ± 2 Ma, and 3689 ± 5 Ma, which taken at face value indicates a depositional maximum age of 3689 ± 5 Ma, in agreement with the model age. Rosing[30] reported a ^{147}Sm-^{143}Nd isochron age of 3779 ± 81 Ma calculated from both meta-sedimentary samples and samples retrieved from the enclosing boninitic amphibolites. This is within error of the model age and the three oldest zircons, but not the depositional age, suggesting that the age of the youngest zircon could have been disturbed or that there is a resolvable range in sample ages. In conclusion, reported age constraints are consistent with an age of ca. 3.7 Ga.

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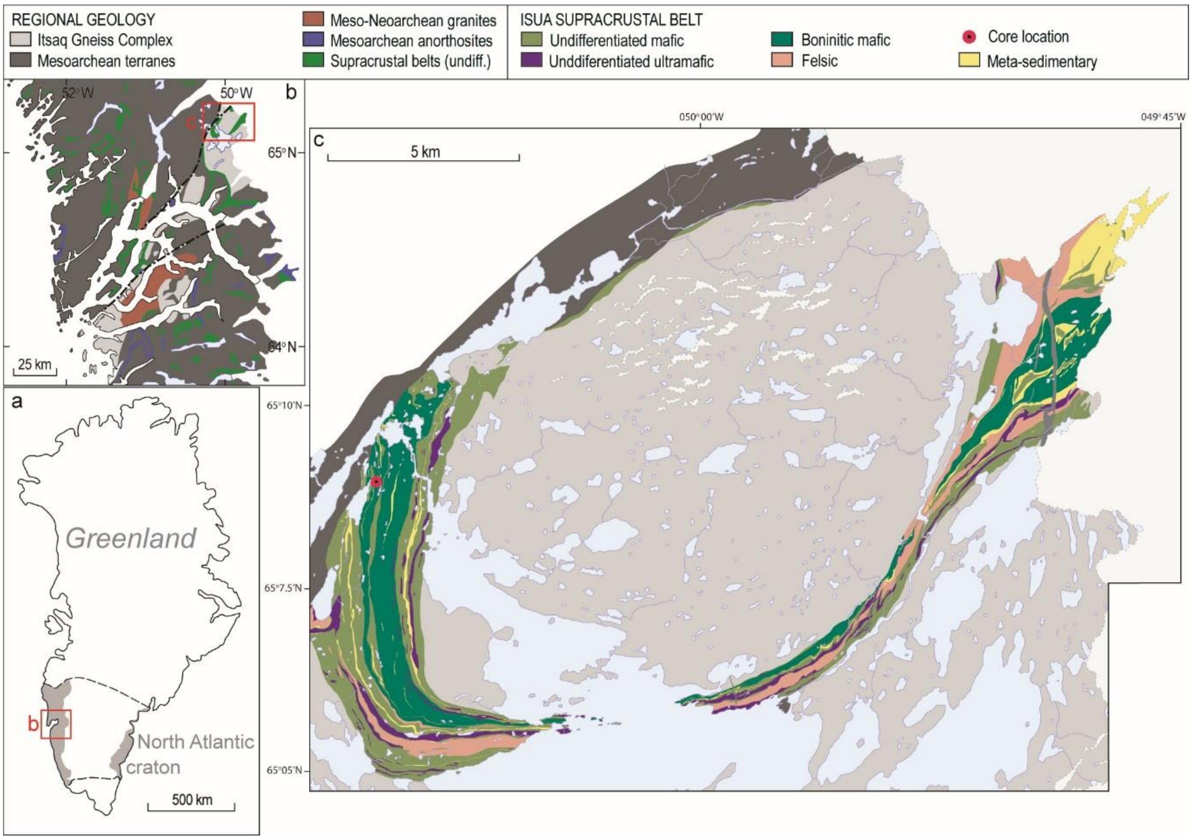

Fig. S1| Maps that illustrate the Isua Supracrustal Belts geological setting. a, Greenland with areas where the Archean North Atlantic Craton is exposed shown in gray. Red box indicates the area depicted in figure b. **b**, regional geology surrounding the Isua Supracrustal Belt adapted from Szilas[31] and based on mapping carried out by the Geological survey of Denmark and Greenland (GEUS). The Isua Supracrustal Belt is located within the red square. **c**, interpreted map of lithology distribution within the Isua Supracrustal Belt adapted from Nutman and Friend[7]. Location of rock cores indicated by red circle.

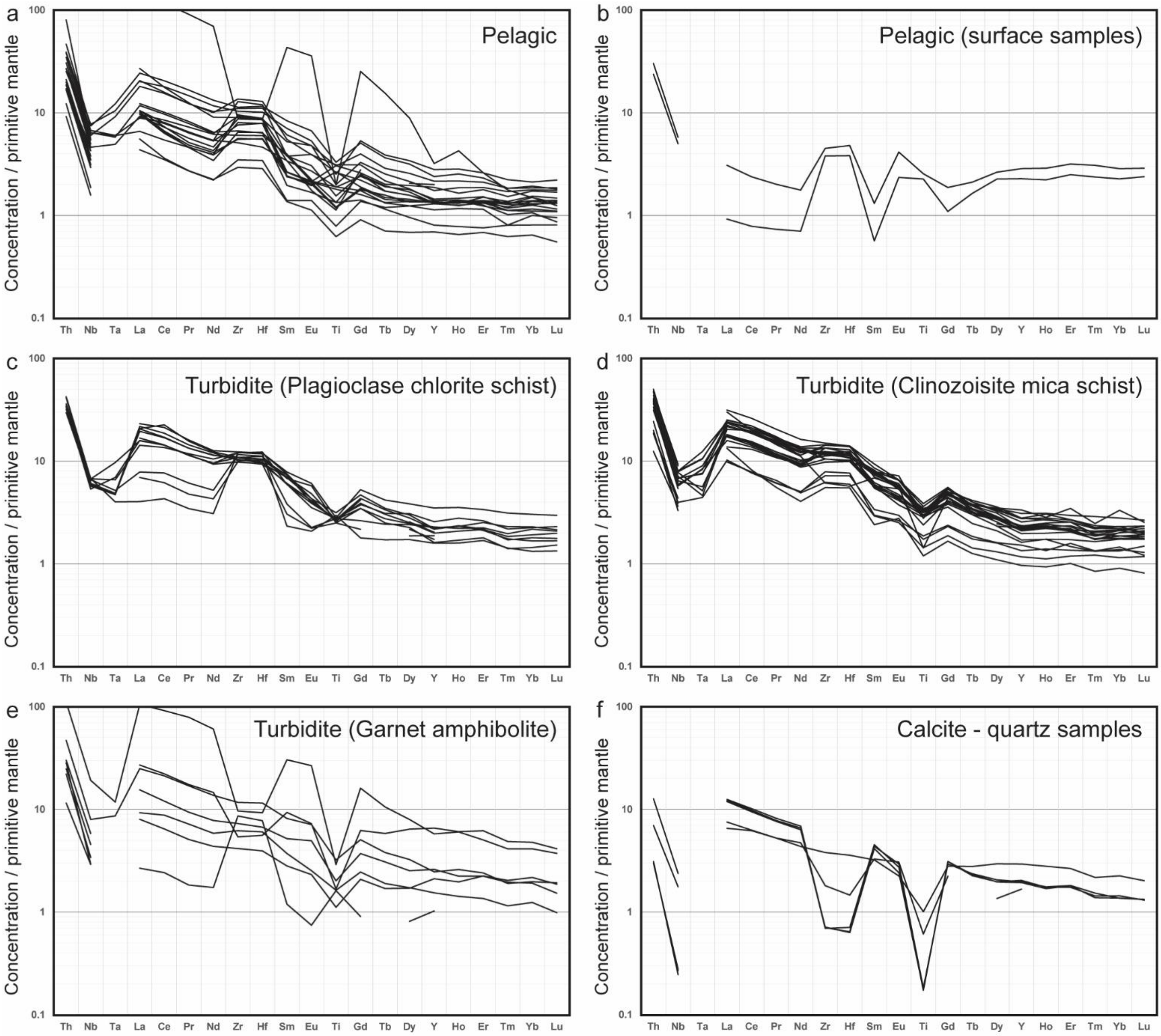

Fig. S2| Immobile trace element plots. The concentration of trace elements that are least mobile during metamorphic fluid flow relative to concentrations in primitive mantle (Palme and O'Neill[32]). **a**, concentrations of samples that are characterized as pelagic. La and Ce values of 161 and 117 respectively of an outlier are not depicted. **b**, concentrations of samples that are characterized as pelagic, but which were not collected from the core. These two samples are taken from two separate layers within a single piece and have trace elemental concentrations that are similar to those of boninites. **c**, concentrations of samples that are characterized as turbidites and that preserve detrital plagioclase grains. **d**, concentrations of samples that are characterized as turbidites where plagioclase grains are largely replace by clinozoisite and Na has been replace by K. **e**, concentrations of samples that are characterized as turbidites and contain abundant garnet and green amphibole. **e**, samples that largely consist of calcite and quartz, diluting trace elements associated with detrital phases.

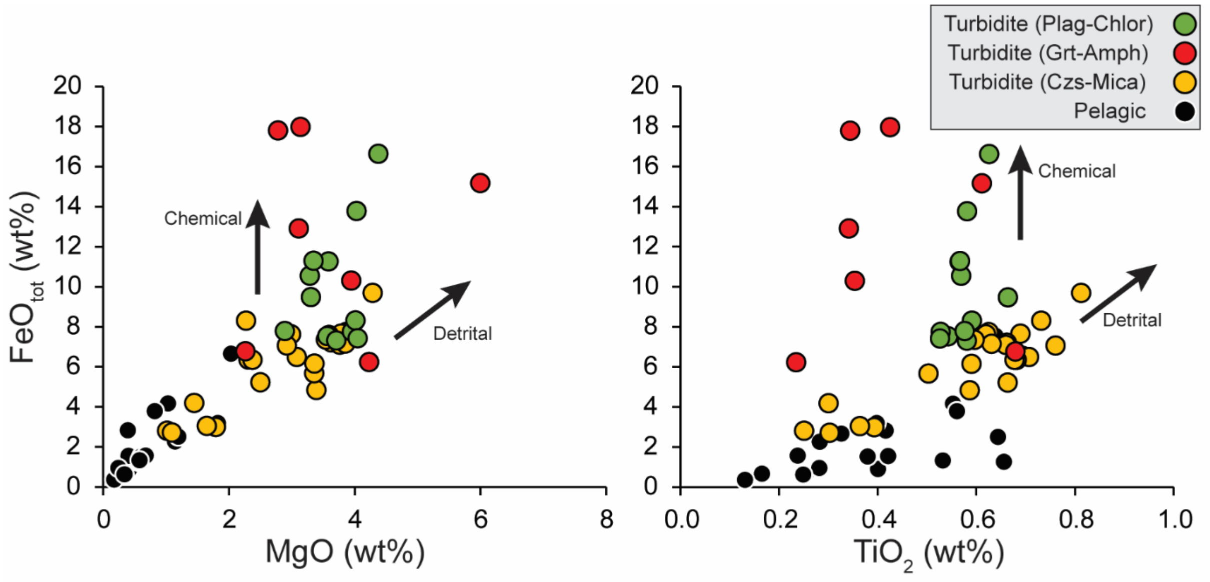

Fig. S3| Detrital and chemical sources of iron. High FeO concentration compared to MgO and TiO_2 in some samples indicate a chemical source for some of the iron as FeO commonly will be transported in MgO- or TiO_2-bearing phases in detrital systems. Turbidite samples are divided into different lithological groups (see supplementary material for discussion of these groups). Plag-Chlor is plagioclase chlorite schist, Grt-Amph is garnet amphibolite, and Czs-Mica is clinozoisite mica schist.