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Research Article

Spatial distribution of ultrahigh-temperature granulites of the Highland Complex of Sri Lanka: Lowermost continental crust above an ultrahot palaeo-Moho

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A B S T R A C T

The spatial distribution of isolated ultra-high-temperature (UHT) granulites in Sri Lanka corresponds to the exposure level of the deepest granulites of the Highland Complex (HC). Their spatial distribution, and that of ultramafic bodies in the HC, defines a relatively thin (several km) lowermost crustal layer that has been folded into a N-plunging asymmetric fold (F_3 in our scheme), *Z*-shaped looking N. We postulate that the UHT granulites represent suitable lithologies within this lowermost crustal layer above an ultrahot palaeo-Moho. Prior to D_3, this layer had already been imbricated and intensely folded with the HT granulites and with ultramafic units after peak metamorphism, leading to a thick package of HT granulites with isolated UHT assemblages. This concept has some important implications for granulite terrains in general: (i) UHT granulites elsewhere may also represent a thin zone above an ultrahot palaeo-Moho and overlain by HT granulite terrains, and their spatial relationships may have been complicated by ductile deformation during early cooling; (ii) many other HT granulite terrains may be underlain by a similar zone of UHT granulites, and, if so, new discoveries of UHT terrains may be expected.

1. Introduction

Ultrahigh-temperature (UHT) granulites are defined as low-pressure to high-pressure (*LP* to *HP)* continental crustal rocks that record peak metamorphic temperatures exceeding 900 °C (e.g., Brown, 2007; Harley, 2004, 2008). They occur in a variety of tectonic settings including, but not restricted to, deep portions of collision zones, continental rifts and regions above subducted oceanic ridges (Brown, 1998; Harley, 2004, 2008; Kelsey and Hand, 2015).

Metamorphic belts in which UHT granulites have been observed are commonly referred to as "UHT terrains", although the majority of rock types encountered are just (HT) granulites, i.e. rocks that belong to the granulite facies. In such belts, metabasites show orthopyroxene-bearing assemblages, charnockitic gneisses are common, and metasediments show pervasive evidence of partial melting with peritectic garnet and/or cordierite in HT granulites (e.g. Kriegsman, 2001). The diagnostic UHT granulites record assemblages such as sapphirine + quartz, orthopyroxene + sillimanite + quartz, osumilite + garnet in aluminous granulites, and grossular + diopside + quartz + wollastonite + scapolite in calcsilicates, that are commonly restricted to a few localities in typical granulite terrains worldwide (e.g. Kelsey and Hand, 2015). This common pattern has raised considerable debate as to whether all rocks encountered in "UHT terrains" have experienced the same UHT conditions or not; whether some domains were subjected to additional heating by localized heat sources; or whether the UHT granulites have been pervasively retrogressed, with the UHT mineral assemblages representing rare relics; or alternatively, whether the UHT granulites in a specific orogen are the rare relics of an earlier metamorphic event which is masked by later HT metamorphism associated with intense deformation (e.g. Sajeev and Osanai, 2004a).

The continental crust can reach UHT metamorphism when the average thermal gradient exceeds 75 °C $kbar^{-1}$ (or 20 °C km^{-1}, e.g. Brown, 2007; Stüwe, 2007). Hence, in view of normally modest to low crustal heat production, it is impossible to generate such extreme metamorphic conditions (e.g., Kelsey and Hand, 2015) implying that the crust cannot be much thicker than the depths recorded by many UHT

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rocks (e.g. Harley, 2004).

Therefore, it is essential to consider the heat sources and mechanisms by which the crust can attain such extreme geothermal gradients. Some workers have suggested that advective mantle heat sources must be involved to generate very high thermal gradients apparent for UHT metamorphism (e.g., Stüwe, 2007). However, syn-metamorphic mafic and/or ultramafic rocks are volumetrically subordinate in most UHT granulite terrains (Collins, 2002; Harley, 2004; Kelsey and Hand, 2015; Kriegsman and Schumacher, 1999). Hence, alternative heat sources have been proposed, such as radiogenic crustal heat production (e.g. England and Thompson, 1984; Sandiford and Hand, 1998), elevated mantle heat flow (Collins, 2002; England and Thompson, 1984; Hyndman et al., 2005), magmatic over-accretion (Galli et al., 2010), ultrahot collisional orogeny (Jamieson and Beaumont, 2013; Sajeev et al., 2010) and local or pervasive strain (viscous) heating (e.g. Duretz et al., 2014).

The mechanisms and tectonic setting(s) responsible for the formation of UHT metamorphism have long been a topic of debate. However recent studies favour accretionary and collisional orogens (Brown, 2007; Kelsey and Hand, 2015). Previous studies showed that UHT metamorphism can be attained in accretionary orogenic systems (e.g. Brown, 2007; Collins, 2002; Hyndman et al., 2005), ridge subduction and subduction slab windows (e.g. Brown, 1998; Santosh et al., 2012), plume bombarded carbonated tectosphere (e.g., Santosh et al., 2012) and large hot orogens with significantly overthickened crust (e.g. Collins, 2002).

There is a wealth of petrological and structural data on Sri Lanka, including detailed mapping by the Geological Survey and Mines Bureau of Sri Lanka, structural maps produced by several workers (e.g. Berger and Jayasinghe, 1976; Kehelpannala, 1997; Kleinschrodt, 1994; Kriegsman, 1994; Tani and Yoshida, 1996), and localities of reported UHT localities in previous studies (Dharmapriya et al., 2015a, 2015b; Mathavan and Fernando, 2001; Osanai et al., 2006; Prame and Prema, 2015; Sajeev and Osanai, 2004a). Building on these data we describe and interpret the spatial distribution of Neoproterozoic UHT assemblages in Sri Lanka, with the aim to establish a feasible tectonic model to explain their occurrences.

2. Geological overview of the Highland Complex of the Sri Lankan basement

This section mainly provides an overview of the geology, petrology and geochronology of the Highland Complex. The central and northeastern part of the HC (Fig. 1) contain mainly granulite facies metasedimentary rocks including metaquartzites, marbles, calc-silicate rocks, pelitic gneisses, and metaigneous rocks such as felsic granitoids, some of which are pyroxene-bearing (charnockite suite), diorites and gabbros that are commonly deformed into orthogneisses (Cooray, 1994; Mathavan and Fernando, 2001; Santosh et al., 2014). All rock types have distinctly older Nd model ages, from 2000 to 3400 Ma, than the rocks in the adjacent complexes (Milisenda et al., 1988, Milisenda et al., 1994).

Metaigneous rocks in the HC have yielded Paleoproterozoic (1950–1850 Ma) and Neoproterozoic (670 Ma) upper intercept ages, interpreted as magmatic ages; and lower intercept ages of 610–530 Ma interpreted as metamorphic ages (e.g. He et al., 2018; Hölzl et al., 1994; Sajeev et al., 2007; Santosh et al., 2014). Recently Kitano et al. (2018) reported the emplacement of some igneous protoliths in the HC at ca. 850 Ma. U-Pb ages of detrital zircons are in the range ~ 3200 Ma to ~700 Ma (Dharmapriya et al., 2015b, 2016; Hölzl et al., 1994; Kitano et al., 2018; Sajeev et al., 2010). Therefore, the 1950–1850 Ma magmatic ages could provide evidence for the existence of a Palaeoproterozoic basement that acted as a platform for sedimentation (Dharmapriya et al., 2016, 2017b). Kitano et al. (2018) have summarized the spatial distribution of detrital zircon U-Pb ages of paragneisses (their Fig. 8) and igneous crystallization ages of orthogneisses (their Fig. 9) Santosh et al. (2014) reported Hf crustal model ages of zircon of mafic and intermediate granulites, and charnockites in the range of

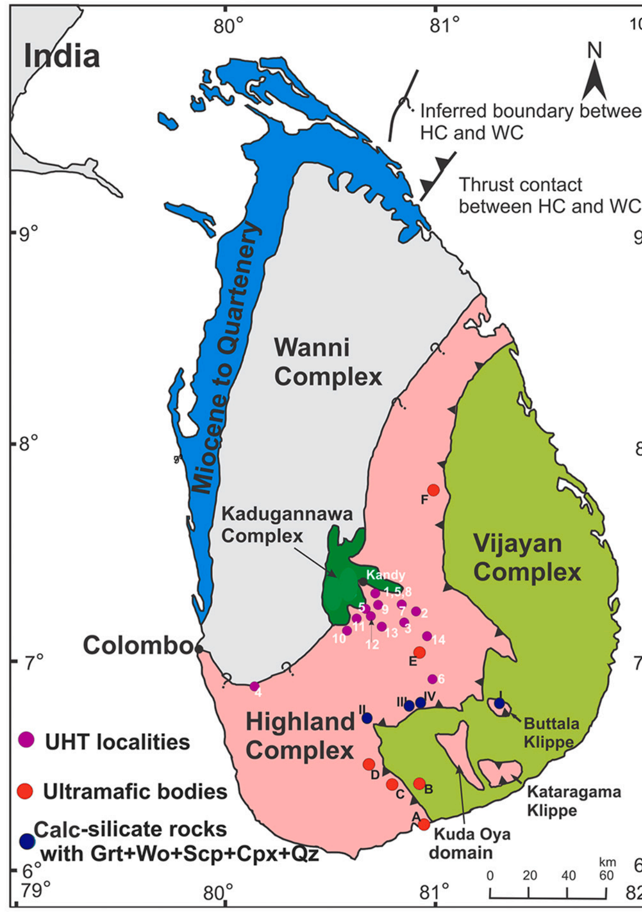

Fig. 1. Geological map of Sri Lanka with lithotectonic subdivision (after Cooray, 1994) and UHT localities and ultramafic bodies reported in the literature. *UHT aluminous assemblages*: 1. Osanai (1989): T = 900 °C; 2 and 3. Kriegsman and Schumacher (1999): T = 830 °C (here corrected to 950 °C); 4. Sajeev and Osanai (2004a): T = 950 °C: 5. Sajeev and Osanai (2004b): T = 1150 °C; 6. Osanai et al. (2006): T = 1000 °C and 7. Sajeev et al. (2007): T = 925 °C; 8. Sajeev et al. (2009): T = 950 °C; 9. Takamura et al. (2015): T = 900–950 °C; 10. Dharmapriya et al. (2015a): T = 950–975 °C; 11. Dharmapriya et al. (2017b): T = 900 °C; 12, Dharmapriya et al. (2015b): T = 920–970 °C; 13, Dharmapriya et al. (2015b): T = 950 °C; 14. Dharmapriya et al. (2015b): T = 900–920 °C. *UHT calcsilicate localities*: I- Mathavan and Fernando (2001); II- Wickramasinghe and Perera (2014); III and IV Prame and Prema (2015). *Ultramafic bodies* (after Tennakoon et al., 2007): A- Ussangoda; B- Indikolapelessa; C- Ginigalpelessa; D- Katupotha; *E*- Rupaha; F- Yodhaganawa.

1500–2800 Ma, that are fairly consistent with the Nd model ages of Liew et al. (1994).

Santosh et al. (2014) interpreted the abundant mafic granulites in the HC in terms of subduction-related mafic magmatism and magmatic underplating. According to these authors, some mafic blocks incorporated into meta-sedimentary rocks of the HC show N-MORB signature, suggesting accretion of oceanic fragments during a subduction-collisional event.

Previous workers have identified a pressure (P) -temperature (T) gradient across the HC. The P-T conditions reported roughly vary from 4.5–6.0 kbar and 700–750 °C in the southwest to 8–9 kbar and 800–900 °C in the east and southeast (Faulhaber and Raith, 1991; Schumacher and Faulhaber, 1994; Raase and Schenk, 1994; Kriegsman, 1996; Braun and Kriegsman, 2003; Prame and Prema, 2015; Dharmapriya et al., 2017a, 2017b). This has been interpreted as a tilted crustal section in which the southwestern part represents the deep middle crust, whereas the southeastern part represents the lowermost crust.

A clockwise P-T trajectory has been proposed for the HT/UHT metasediments in the HC (Dharmapriya et al., 2015a, 2015b; Dharmapriya et al., 2017b; Dharmapriya et al., 2021; Hiroi et al., 1994; Kriegsman, 1996; Osanai et al., 2006; Raase and Schenk, 1994; Sajeev and Osanai, 2004b). During the prograde evolution, the HC metasediments were subjected to a pressure increase from amphibolite facies conditions (Dharmapriya et al., 2017b; Kriegsman, 1996; Raase and Schenk, 1994). Then the rocks were heated further along a prograde decompression path until peak metamorphism (Dharmapriya et al., 2015a, 2015b; Hiroi et al., 1994; Osanai et al., 2006) which was followed by a well-defined near isobaric cooling stage (Dharmapriya et al., 2015a, 2015b; Prame, 1991; Schumacher et al., 1990). It is important to note that the metamorphic field gradient from southwest to the east and southeastern part of the HC mentioned above probably reflects the end of the isobaric cooling stage (see section 3). Subsequently, the rocks have undergone a near-isothermal decompression stage (Dharmapriya et al., 2015b, 2017b, 2021; Prame, 1991; Sandiford et al., 1988; Schumacher et al., 1990). Kriegsman (1996) calculated a retrograde slope of ~15 bar/K for this stage.

Berger and Jayasinghe (1976) first suggested that the Sri Lankan basement has been subjected to at least three deformation phases. The deformation events named D_1 and D_2 were responsible for the formation of the major lineation and foliation (L-S fabric), including the main compositional layering, whereas D_3 mainly resulted in the formation of large-scale upright folds. Later workers (e.g. Kehelpannala, 1997; Kleinschrodt, 1994; Kriegsman, 1994, 1995; Tani and Yoshida, 1996) added evidence for additional deformation events or stages (Table 1). The *syn*-tectonic D_2 event was responsible for the formation of the major lineation and foliation (L-S fabric), resulting in the main compositional layering.

Kehelpannala (1997 and earlier work) and Voll and Kleinschrodt (1991) reported that D_1 produced the main flatting foliation (S_1) and stretching lineation (L_1) and that the D_2 event produced small-scale as well as large-scale isoclinal to tight folds (F_2) with fold axes parallel to L_1. Voll and Kleinschrodt (1991) recognized a second stage of isoclinal folding between F_2 and F_3, with axes perpendicular to L_1. Kriegsman (1994) noticed that steep, high-T shear zones developed as high-strain zones on steep flanks during the late stages of D_3. Kriegsman (1995) defined refolding of D_2 and D_3 structures in the southern part of Sri Lanka as an additional D_4 deformation event that may have developed in the same tectonic setting as D_3. These folds may be correlated with Kehelpannala's (1997) F_4 folds mainly identified from satellite images. Kehelpannala (1997) suggested that the HC and WC have undergone six phases of ductile deformation. D_1 to D_2 are very similar to those described by earlier workers (e.g. Berger and Jayasinghe, 1976). In contrast, minor and large scale recumbent isoclinal folds (F_3) were produced during D_3. The D_4 produced very large gentle, nearly *E*-W trending upright folds whereas D_5 was responsible for the large-scale upright folds (F_5) and D_6, deformation caused local refolding of the F_5 folds.

Large-scale upright folds, with wavelengths of about 7–10 km and an exposed length parallel to their axes up to 50 km). occur throughout Sri Lanka (F_3 according to Berger and Jayasinghe, 1976), and their axial planes curve from NNE-SSW trend in the northeastern part of the HC and WC to N-S and NNW-SSE in the central part of the HC (e.g., Berger and Jayasinghe, 1976; Kriegsman, 1995; Fig. 2a). In the southernmost part, the fold trend becomes NW-SE and W-E or locally even WSW-ENE. Kriegsman (1995) related this arc shape to wrench tectonics during the major convergence between HC and VC. Kleinschrodt and Voll (1994) proposed that the large upright folds (F_3 of Berger and Jayasinghe, 1976) were formed after the peak of metamorphism (800–850 °C, 8–9 kbar) at slightly lower temperatures (700–750 °C), but still under granulite facies conditions.

The KC, WC and VC rocks (see Santosh et al., 2014; Dharmapriya et al., 2015a, 2015b; He et al., 2016; Malaviarachchi et al., 2021 for more details) have yielded Nd-model ages of 1800–1000 Ma, 2000–1000 Ma, and 3300–1000 Ma (Milisenda et al., 1988, Milisenda et al., 1994; Malaviarachchi et al., 2021) respectively. These three complexes were metamorphosed under upper amphibolite to granulite facies conditions during the assembly of Gondwana (e.g. Cooray, 1994; He et al., 2016; Kröner et al., 2013). More details on the WC, KC and VC are summarized in the supplementary document.

Table 1
Correspondence of schemes for ductile deformation events (D) in the Highland Complex (modified after Kehelpannala, 1997).

Author/s	Major compositional Layering	Large-scale upright folds
Berger and Jayasinghe (1976)	Combination of D_1-D_2	D_3
Yoshida et al. (1990)	Second phase of D_1-D_2	D_3
Voll and Kleinschrodt (1991)	D_1	D_4
Kehelpannala (1991)	Combination of D_1-D_2	D_3
Kriegsman (1991, 1995)	D_2	D_3
Kehelpannala (1997)	Mainly D_2	D_5

3. Distribution of UHT granulites and ultramafic complexes

During the last two and a half decades, evidence for UHT metamorphism has been reported from several localities in the central HC. Figs. 1 and 3a show all published UHT localities. They include the classical sapphirine-bearing assemblages in aluminous granulites discussed in Harley (2008) and Kelsey and Hand (2015), as well as occurrences of calc-silicate rocks with the assemblage grossular + diopside + quartz + wollastonite + scapolite. Most UHT localities are located within an elliptical domain with a NNW- SSE trending long axis (Fig. 3a). It is ~50 km wide and ~ 100 km long and the long axis runs from near Kandy to Buttala.

A number of serpentinised ultramafic bodies have been reported from the same domain, close to the HC-VC boundary (e.g. Munasinghe and Dissanayake, 1982; Tennakoon et al., 2007; He et al., 2016; Fig. 1). In addition, there is one serpentinised ultramafic body reported in the central domain of the HC at Rupaha (e.g. Fernando et al., 2017; locality E in Figs. 1 and 3a). These ultramafic units have been interpreted in two contrasting ways: (i) upper mantle enclaves emplaced within granulites during thrusting of the HC on top of the VC (Dissanayake and Munasinghe, 1984; Kriegsman, 1995); and (ii) ultramafic magmatism in a post-tectonic setting (e.g. He et al., 2016; Kriegsman, 1995; Munasinghe and Dissanayake, 1982). We favour the first interpretation, because they form isolated bodies that mainly consist of altered dunites and harzburgites and lack the typical features of layered mafic intrusions. For the serpentinite at Ussangoda, He et al. (2016) reported U-Pb zircons ages in the range of 470–524 Ma (weighted mean 485 Ma), which they interpreted in terms of Ordovician ultramafic magmatism with CO_2 influx and mantle metasomatism. Dharmapriya et al. (2017b) also argued that the near isobaric cooling stage of the HC extended up to 500–510 Ma. Therefore, the incorporation of the ultramafic slivers into granulite gneisses may have taken place during the middle to late Cambrian to early Ordovician periods.

Combining the distributions of all UHT localities and ultramafic bodies, leads to a coherent pattern: they are all located structurally within a few kilometres from the HC-VC boundary. The overall distribution can be explained in terms of late-tectonic refolding (during D_3, after Berger and Jayasinghe) of a single deep-crustal unit in which UHT granulites and upper mantle slivers have been interleaved with HT granulites. Fig. 3b shows a slightly N-plunging *Z*-shaped fold, looking N, in which the Kandy-Buttala domain represents the domain with the shallowly dipping flank. The southern and northern segments each represent a steep flank of the (re)folded UHT granulite zone.

Geological mapping by the Geological Survey Department (1982) and the Geological Survey and Mines Bureau (1996, sheets 14 & 17) revealed the presence of intensely folded (upright folds, F_3) and thrusted

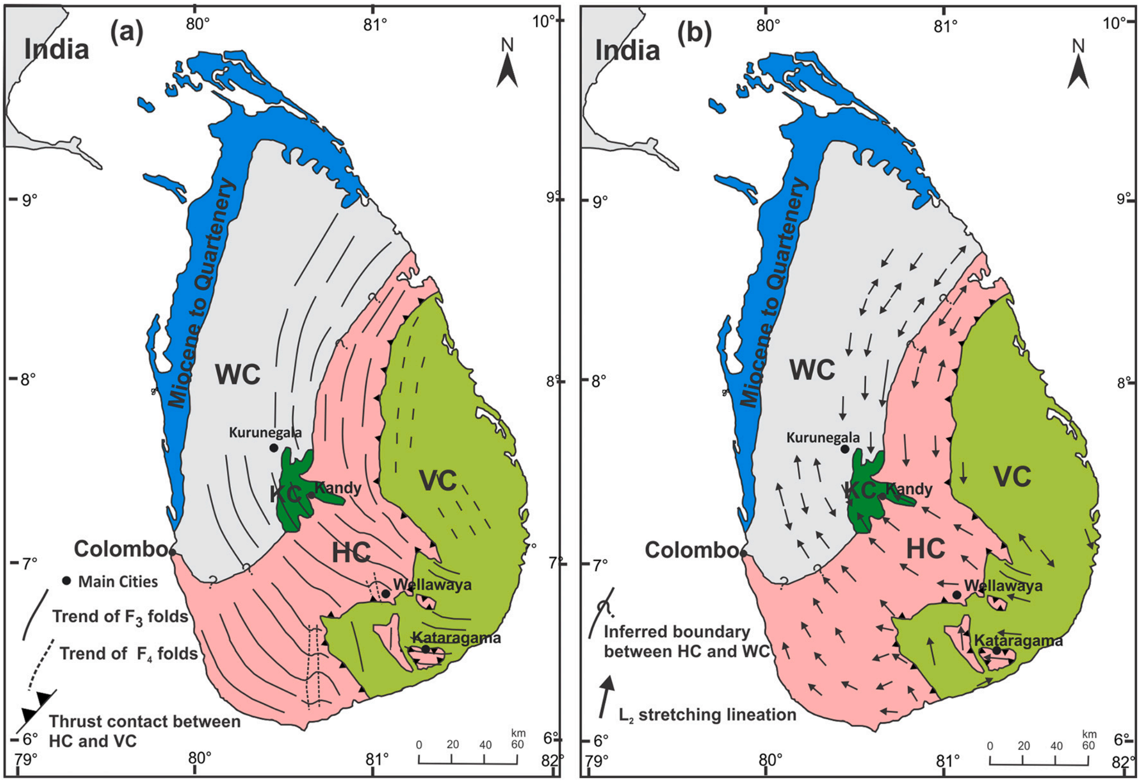

Fig. 2. Structural trend lines of the Sri Lankan basement (after Kriegsman, 1995) on a map with simplified lithological boundaries (after Cooray, 1994): (a) trend of D_3 folds, (b) trend of stretching lineations.

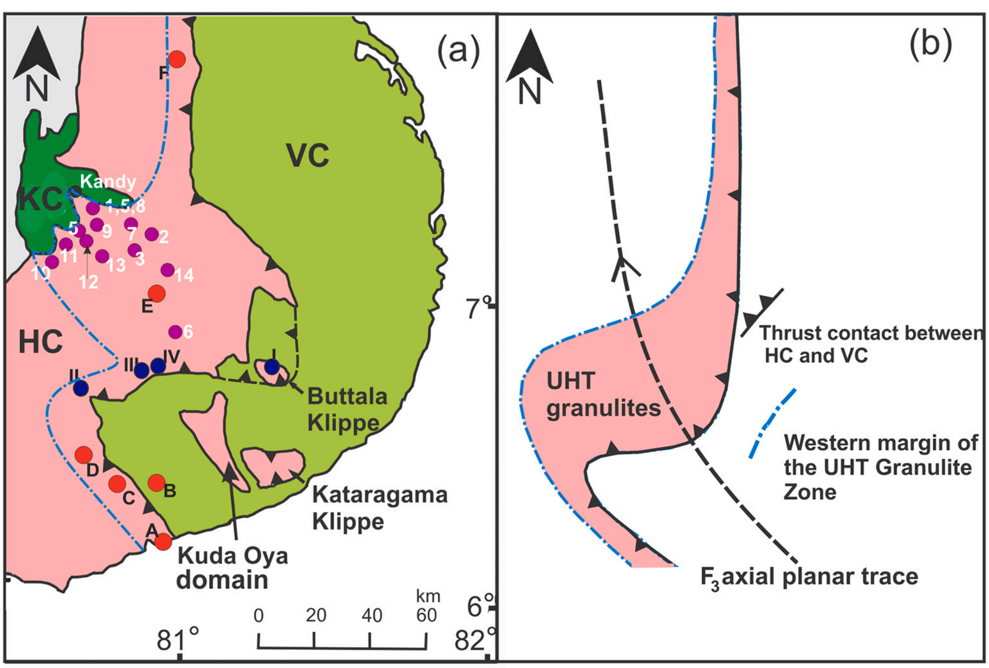

Fig. 3. (a) Enlargement of spatial distribution of UHT localities and ultramafic bodies (UHT aluminous assemblages, UHT calcsilicate localities and ultramafic bodies as Fig. 1). Fig. (b) interpretation in terms of F_3 folding of a single lowermost crustal unit in which UHT granulites and upper mantle slivers have been interleaved with HT granulites.

rock layers in the area of the UHT granulite zone (e.g. Supplementary Fig. S1). The trend of F_3 axial planes varies from NNW-SSE in central part of the country to NW-SE further south and W-E and WSW-ENE close to the HC-VC boundary especially in the Ella area. Kriegsman (1995) reported that in the Ella-Wellawaya area in southeastern HC, the trends of axial planes of F_3 folds and stretching lineations locally swing from NW-SE to N*E*-SW and back to E-W and NW-SE (Fig. 3a,b). He proposed that the final part of the tectonic evolution in southern Sri Lanka can be explained by the 'corner effect' during simultaneous thrusting and wrenching in a continental collision setting. In this model, the lower plate shows only minor syntaxial bending, whereas the upper plate may show a strongly curved arc (see Fig. 3 of Kriegsman, 1995). Eastward thrusting including imbrication, the formation of F_3 upright folds, the formation of a syntaxial bend defined by the axial planes of these folds and by the trajectories of stretching lineations, and refolding of the syntaxial bend by later NW-SE trending upright F_4 folds above a blind thrust, may all have occurred at this stage (D_4 according to Kriegsman, 1995, Kriegsman, 1994). An overall gently NW plunging fold axis could explain the observation that rocks NW of Kandy, for example near Kurunegala, show lower *P-T* conditions.

4. *P-T* estimates for the UHT granulites, and relation to HT granulites in the HC

In contrast to the P-T gradient reported in the western part of the HC (Prame and Prema, 2015; Schumacher and Faulhaber, 1994), no clear *P/T* gradient can be observed along a NW-SE transect within the UHT Belt from Kandy towards Wellawaya and the Kataragama outlier, with peak pressure conditions of ~10 kbar reported for the latter (Prame and Prema, 2015). In addition, no obvious *P-T* gradient has been reported within this UHT granulite zone despite more than 2000 m of elevation difference between the central highlands and the HC outliers (e.g. Kataragama klippe; Figs. 1, 3a) on top of the VC in the southeastern lowlands. Most *P-T* estimates for the UHT granulites range from 900 to 1150 °C at pressures of 9.0–12.5 kbar (e.g., Dharmapriya et al., 2015a,b, 2017a,b; Kriegsman and Schumacher, 1999; Osanai, 1989; Osanai et al., 2006; Sajeev et al., 2007; Sajeev and Osanai, 2004a, 2004b; Takamura et al., 2015). It should be noted that the variation in reported pressures may be due to the methodology chosen, rather than being geologically relevant.

The UHT granulite zone overlaps with the core of the region where Schumacher and Faulhaber (1994) reported the highest P-T conditions of ~9 kbar and 850 °C, based on thermobarometry of metabasites and garnet-bearing charnockites. It should be noted here that these authors, including their earlier papers (e.g, Schumacher et al., 1990), used the reaction assemblage in which Grt, Cpx and Qz break down to form Opx + Pl- symplectites. Along the *P-T* path mentioned in section 2, this assemblage postdates prograde decompression and subsequent isobaric cooling. Hence, it records neither peak *P* nor peak *T*. Thus, the overlap of the UHT granulite zone with the highest *P-T* region merely suggests that the decompression event has affected the whole terrain in a more or less homogeneous manner.

The isothermal decompression of HT and UHT granulites in the HC occurred at ~750–830 °C (Hiroi et al., 1994; Kleinschrodt, 1994; Raase and Schenk, 1994) and 850–900 °C (e.g. Dharmapriya et al., 2015a, 2015b, 2021; Sajeev and Osanai, 2004b) respectively. Hence at this stage, the UHT and HT granulites could have become tectonically interleaved. Fig. 4a indicates the distribution of isotherms in the lower and lowermost part of the crust during the peak metamorphic conditions (i.e. during D_2 deformation) that were probably frozen in during the isobaric cooling stage The later D3 event folded and imbricated the 900 °C isotherms (the lowermost crust) with the overlying high temperature (HT) granulites (Fig. 4b). Hence, we postulate that this process may lead to a thicker package of HT granulites with isolated UHT assemblages. An obvious factor that also contributes to the (apparent) local nature of UHT assemblages is the specific whole-rock composition needed to record such assemblages.

5. Tectonic and geodynamic interpretation

A variety of tectonic models have been proposed, but here we will focus on various stages in the crustal evolution that most workers agree on. We will summarize them in a chronological order.

The earliest recognizable stage of the Panafrican event is a metamorphic stage under amphibolite facies conditions (Dharmapriya et al., 2017b; Raase and Schenk, 1994), most likely at fairly low pressure and possibly in the andalusite stability field ($P < 4$ kbar). This stage is likely to represent pre-orogenic crustal thinning, possibly in a continental shelf setting, coeval with sedimentation (see Malaviarachchi and Dharmapriya, 2021).

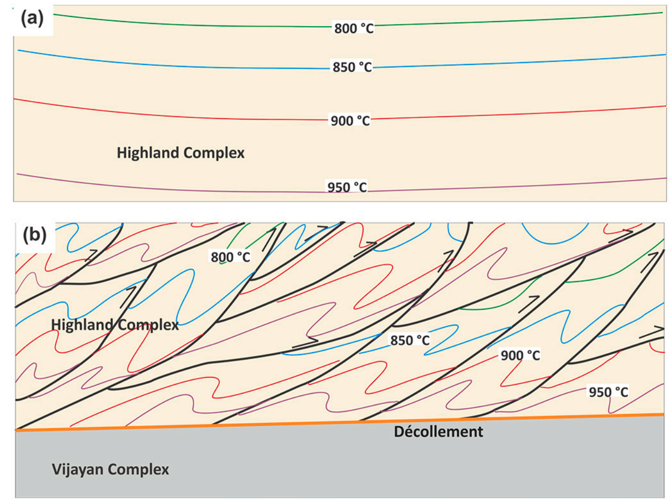

Fig. 4. Sketch diagramme illustrating (a) the distribution of isotherms (isograds) that had been frozen in during the isobaric cooling stage; and (b) their disturbance by subsequent refolding and imbrication (D_3).

A subsequent strong pressure increase, still at amphibolite facies conditions, recorded during prograde metamorphism of HC metapelites (Dharmapriya et al., 2017b; Kriegsman, 1996; Raase and Schenk, 1994) has been attributed by these authors to crustal thickening and fits the suggestion by Hiroi et al. (1994) that the *P-T-t* path resulted from continental collision during the formation of Gondwana. At this stage, the HC probably reached its maximum crustal thickness with maximum recorded *P* around 11–12 kbar (Dharmapriya et al., 2015a, 2015b; Dharmapriya et al., 2017b). The D_1 deformation event of the HC (Berger and Jayasinghe, 1976; Kehelpannala, 1997; Kriegsman, 1994, 1995) may also coincide with this crustal thickening event.

This stage was followed by minor decompression with increased temperature up to UHT conditions (Dharmapriya et al., 2015a, 2015b), which can be explained by lower crustal extension. Extreme flattening occurred at this stage (D_2 of most authors) and led to the formation of an L-S fabric (Kriegsman, 1994, 1995). Decompression may be attributed to isostatic rebound after lithospheric delamination that led to asthenospheric upwelling and UHT conditions (Santosh et al., 2014; Fig. 5).

After HT/UHT peak metamorphism, the HC granulites underwent moderate near isobaric cooling before being uplifted to upper crustal levels along a steep decompression path, simultaneous with folding and thrusting during the Late Proterozoic to Cambrian. Most authors attribute this late event to the thrusting of the HC on top of the VC (Hatherton et al., 1975; Kriegsman, 1995; Voll and Kleinschrodt, 1991). Klippen (outliers) such as Kataragama and Buttala (Fig. 1) in the southern HC suggest displacement >50 km parallel to the transport direction (e.g. Kriegsman, 1995; Voll and Kleinschrodt, 1991).

Dreiling et al. (2020) provide evidence for a major intracrustal layer (18–27 km depth) with high seismic velocity, dipping ~4^0 westward underneath the central HC. Eastwards, the layer continues into the western Vijayan Complex (VC). These data are consistent with the interpretation of the VC or unexposed crustal domains underlying the VC as a colliding promontory (Braun and Kriegsman, 2003), and with the interpretation of complicated structures found along the HC/VC thrust contact in terms of thrusting and wrenching (Kriegsman, 1995). The crustal-scale anticlinorium inferred in this paper, fits well into this scenario.

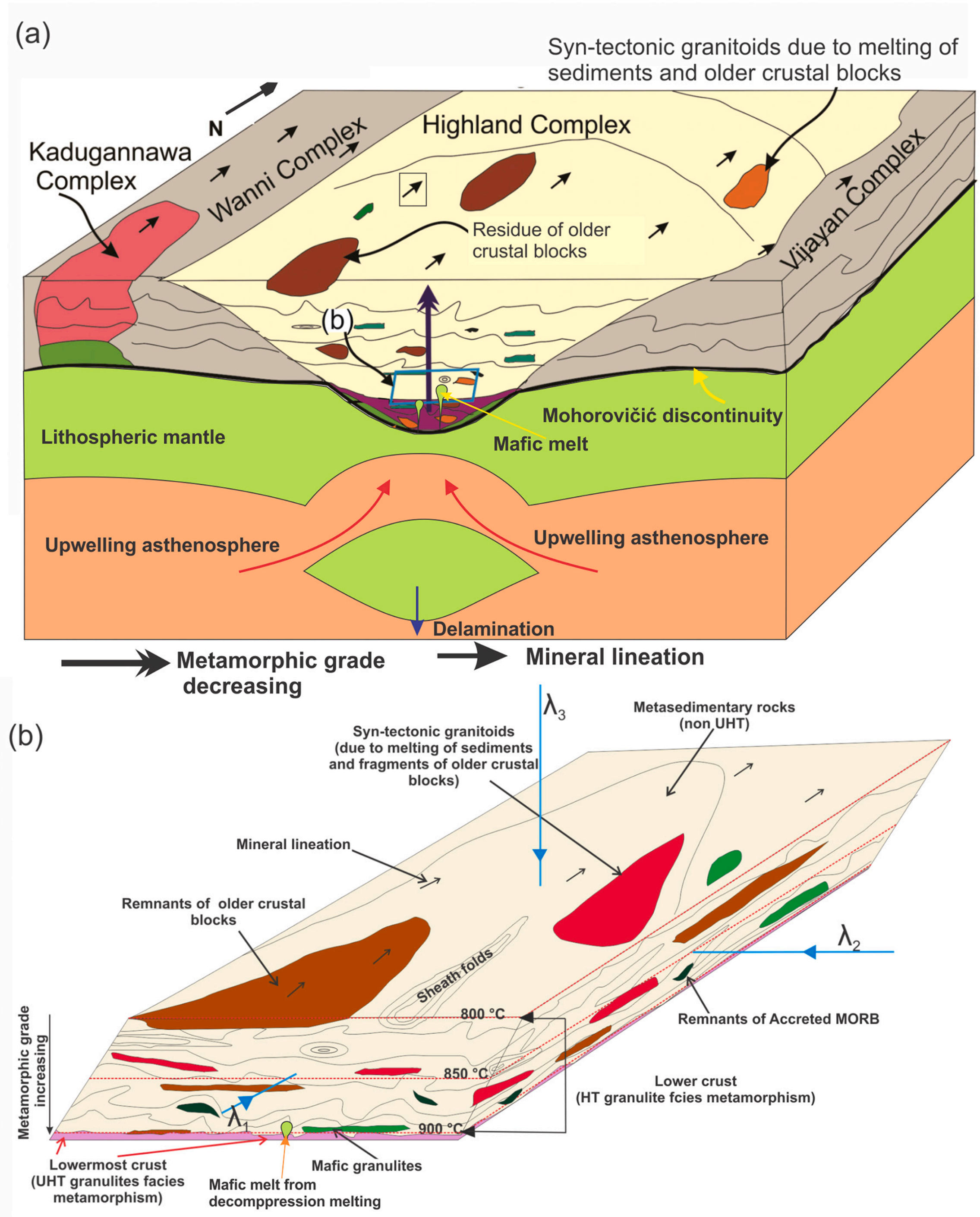

Fig. 5. Sketch diagrammes illustrating (a) delamination of the lithospheric mantle and channel flow of the lower and lowermost crust of the HC (modified after Santosh et al., 2014); (b) 3-D lower crustal block diagram of the HC showing the key petrological and deformational features and the overall asymmetry resulting from non-coaxial flow..

Building on earlier (e.g., Hatherton et al., 1975) and new gravity measurements, Prasanna et al. (2013) modeled a gravity high under central Sri Lanka, with the crust thickening to 40–41 km under the eastern part of the Highland Complex. Dreiling et al. (2020) confirmed Moho depths of 30–40 km, but locate the thickest crust (38–40 km) beneath the central Highland Complex (HC). In either case, HT to UHT granulites that were uplifted from 35 to 45 km maximum burial depth to <14 km (andalusite stability field) during the HC-VC thrusting event and related exhumation, are now underlain by ~40 km continental crust of unknown age and origin.

Using a combination of petrological geochronological, geochemical and structural geological studies, it has been suggested that the WC, HC and VC evolved as separate terranes of distinct origins (e.g. here we refer to it as the arc accretion model: Vitanage, 1972; Hiroi et al., 1994; Kehelpannala, 2004). In these models, the WC would have collided with the HC first, and subsequently, the combined HC/WC unit was thrust over the VC (Kehelpannala, 1997; Kriegsman, 1995; Vitanage, 1972; Voll and Kleinschrodt, 1991), coeval with the amalgamation of Gondwana (Kehelpannala, 1997, 2004; Kriegsman, 1995). Santosh et al. (2014), however, suggested a double-sided subduction model in which the WC and the VC are separate Neoproterozoic continental arcs that collided with the Highland Complex during late Neoproterozoic-Cambrian times. These authors defined the HC as an accretionary belt marking the collisional suture. This model is discussed in some detail in the next section.

6. Discussion

6.1. Tectonic model: Arc accretion or double-sided subduction?

In the arc accretion model, the WC is interpreted as a continental arc and the VC as an intra-oceanic island arc. In the double-sided subduction model, Santosh et al. (2014) proposed that the WC and VC arcs were active on opposite sides of the HC. In addition, it was argued that the HC had been a paleo-ocean where predominantly sediments were deposited. Hence, a continuous collisional event was proposed that would still allow for small differences in relative timing. The incorporation of Paleoproterozoic granitoids in the HC was interpreted as due to micro-continents (as proposed earlier by Kehelpannala, 2004). Santosh et al. (2014) also pointed out that tightly clustering zircon εHf(t) values from −2.2 to 0.1 with Hf crustal model ages (T^{C})in the range of 1501–1651 Ma in some mafic granulites in the HC suggested a mixed source from both juvenile Neoproterozoic and reworked Mesoproterozoic components.

Although the double-sided subduction model is able to explain the geochronological and geochemical framework of the Sri Lankan basement, it gives rise to the following three concerns. Firstly, in view of the ca. 300 Ma difference in timing between the VC and WC arcs, the VC may have formed thousands of kilometres outboard, in a tectonic setting completely unrelated to the HC and it may just be an independent block accreted to the HC more than 400 Ma after its formation. Secondly, having opposing subduction zones on either sides of the relatively small HC would create massive space problems at depth. Thirdly, the UHT granulites would be concentrated in the central part of the HC in the model, which is at odds with the field evidence (Fig. 1).

In this context, the double-sided subduction model of Santosh et al., 2014 and He et al., 2016, was revised by Malaviarachchi (2018) and Satish-Kumar et al. (2021), suggesting a ‘two-staged subduction’ concept, with noncoeval arc segments of WC and VC. In these papers, the HC has been treated as a segment of a suture zone of the Neoproterozoic Mozambique ocean that was dominated by trench-fill sediments. Sedimentation of carbonate rocks started in an open ocean around volcanic OIB type oceanic islands and in passive margins in the Paleo-Mesoproterozoic, before the formation of subduction-generated island arcs. Simultaneously carbonate and pelitic sediments were deposited in a shallow marine passive margin near a continent, which in southern India may have been an Archean craton such as the Dharwar. These sediments may have accreted in a trench during subsequent subduction. Although previous workers envisaged that the VC is an ‘exotic’ block with no petrochemical link to Gondwana (e.g. Kröner et al., 2013), Malaviarachchi et al. (2021) presented Sr—Nd isotopic data suggesting that the VC may have a genetic link to older Proterozoic cratons distant to the Indo-Antarctic peripheral assembly region of Gondwana. The subsequent development of a continental margin arc in the late Neoproterozoic might have led to amalgamation of the East African-Antarctic Orogen during the final stage of formation of East Gondwana.

6.2. Delamination, asthenospheric upwelling, lower crustal flow

In the model by Santosh et al. (2014) it was proposed that post-collisional slab break-off took place after the crust had reached its maximum thickness, allowing asthenospheric upwelling to supply more heat to the thickened lower crustal domains of the HC. However, a number of studies indicate that asthenospheric upwelling under a thickened crust is only possible when the root of that thick crust has partly delaminated, together with the underlying mantle lithosphere (e.g., Kay and Mahlburg-Kay, 1993). Convergent tectonics may lead to lithospheric delamination during or after collision (e.g., Lustrino, 2005). Continent-continent collisions may lead to an orogenic lithospheric keel that becomes gravitationally unstable and detaches, sinking into the asthenospheric mantle (Kay and Mahlburg-Kay, 1993; Lustrino, 2005). That can result in upwelling of the asthenosphere that replaces the delaminated keel (Kay and Mahlburg-Kay, 1993). Decompression melting of this asthenospheric material may explain *syn*-tectonic mantle-derived magmatism in the HC (e.g. Santosh et al., 2014).

As a result of asthenospheric upwelling, the hot and rheologically weakened, melt-bearing lower crustal rocks underwent non-coaxial flow in a N-S direction (e.g. Tani and Yoshida, 1996; Kehelpannala, 2004; Ranaweera and Kehelpannala, 2019; Fig. 6a,b) that led to pervasive L-S fabrics (subhorizontal foliation S_2 and strong stretching lineation L_2), isoclinal, recumbent folds and boudinage of less ductile lithologies. Some garnets show evidence for rotation during the prograde decompression stage of the HC just prior to peak metamorphism (see Fig. 6a, b - Dharmapriya et al., 2021; Fig. 5c, d - Ranaweera and Kehelpannala, 2019). Over-thickened crust can potentially spread laterally while the collisional event is still continuing (e.g. Dewei, 2008;), resulting in a near horizontal L-S fabrics in rocks.

Lower crustal flow has also been suggested for orogens elsewhere. For instance, Holland and Lambert (1969) postulated the formation of regional-scale, pervasive sub-horizontal fabrics with recumbent isoclinal folds can be generated by lower crustal laminar flow during a subduction-accretion phase accompanied by granulite facies metamorphism. A similar reasoning was followed by other authors in collisional settings (e.g., Dewei, 2008).

In the proposed model (Fig. 6b), it is considered that the lowermost crust (just above the Moho) is the segment most influenced by upwelling asthenosphere, leading to extreme heat flow, UHT metamorphism and intense crustal anatexis (at least *P-T* of ~9.5–10.5 kbar and ~ 900–975 °C: Osanai (1989); Kriegsman and Schumacher (1999); Sajeev and Osanai (2004a); Sajeev and Osanai (2004b); Osanai et al. (2006); Sajeev et al. (2007); Sajeev et al. (2009); Takamura et al. (2015); Dharmapriya et al., 2015a, 2015b).

6.3. Final exhumation and cooling

Hiroi et al. (2014) postulated fast exhumation of lower-crustal rocks to the andalusite stability field, i.e. upper crust, via channel flow in a continental collision orogeny. In our preferred model, channel flow occurred earlier, namely during the minor decompression during heating to UHT condition. We advocate the model that uplift to the andalusite stability field is due to thrusting of the HC onto the VC, with concomitant extensional and/or erosional unroofing at shallower, unexposed levels.

Following Kriegsman (1995), we propose an imbricated structure for the HC-VC contact (Fig. 6a), that formed during D_3, after the non-coaxial lateral flow of the lower crustal rocks (D_2). Here we suggest that low-angle underthrusting of the relatively cold VC could have promoted the period of near-isobaric cooling of the HC, as well as local granulite facies overprints on the dominantly amphibolite facies VC gneisses. However, besides this local explanation, several theoretical and modeling studies suggest that given enough time, rocks can undergo a near-isobaric cooling stage during collisional orogeny (e.g. England and Thompson, 1984; Harley, 1991) and, alternatively in the scenario of the HC, stagnation of the asthenospheric upwelling could also cause a period of isobaric cooling of the granulites while still at a lower crustal level. At a later stage of thrusting, ramps may have led to imbrication, folding and rapid uplift of the HC (Fig. 4b; 6b,c), leading to isothermal decompression. The setting of many high-grade complexes can be accounted for by uplift of lower and middle crust to shallower levels by thrusting in collisional orogenies (e.g., Jamieson and Beaumont, 2013).

After the emplacement of lowermost and lower crustal rocks into middle crustal levels by thrusting and duplexing (Figs. 6, 7), exhumation

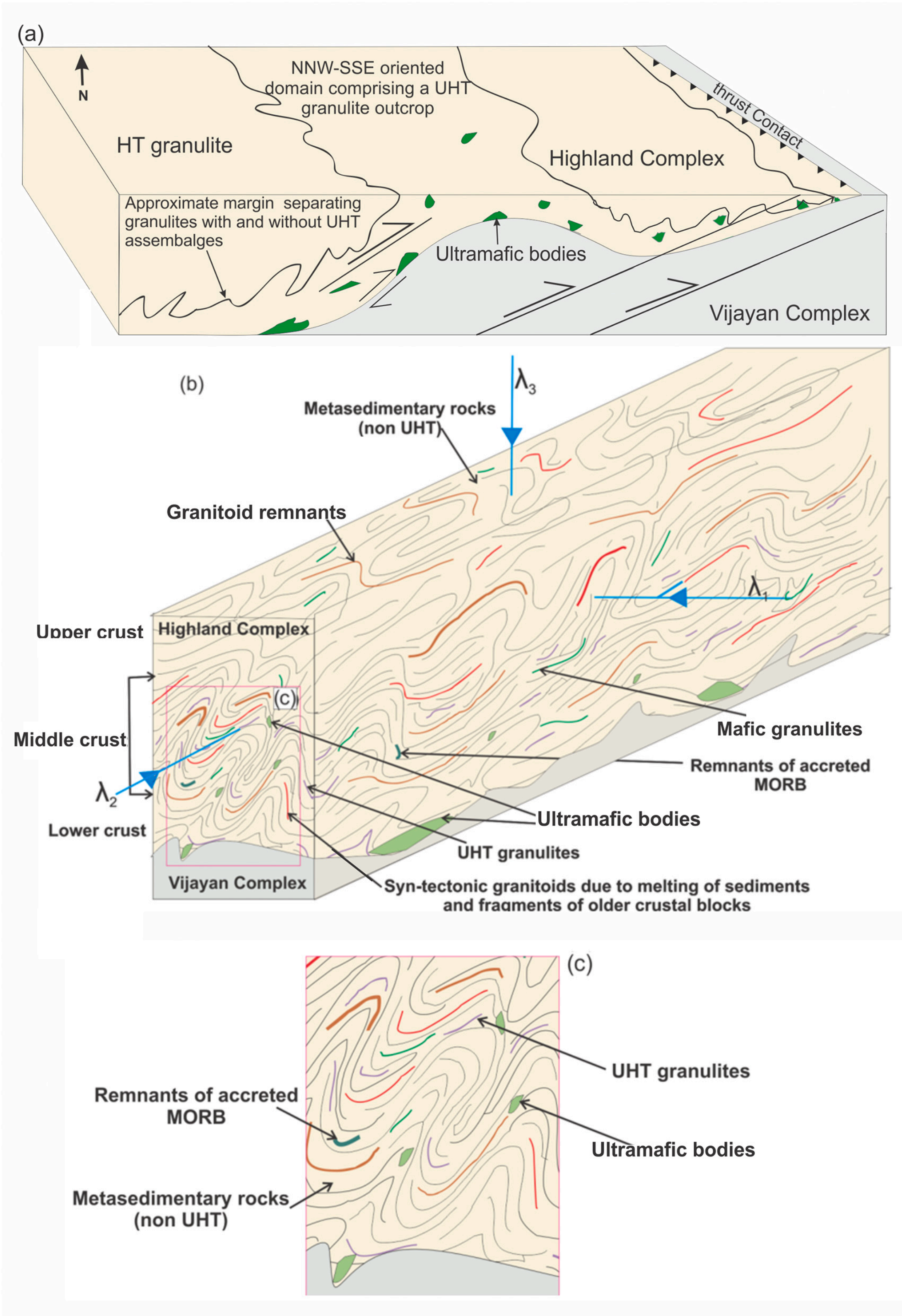

Fig. 6. Sketch diagrammes illustrating (a) continuous thrusting and folding; (b) emplacement of HT and UHT granulites in mid-crustal level due to folding and thrusting; (c) an enlarged view of the lower to middle crustal section.

to near-surface level could be due to extension and erosion of the upper crust during isostatic relaxation. The present erosional surface is due to more recent uplift (Vitanage, 1972). The late-tectonic HC-VC thrusting event facilitated the incorporation of rare upper mantle fragments (Figs. 1, 3a), mainly at the HC-VC boundary in the southern part of the country and, in one case, within the central HC (at Rupaha: e.g. Fernando et al., 2017.

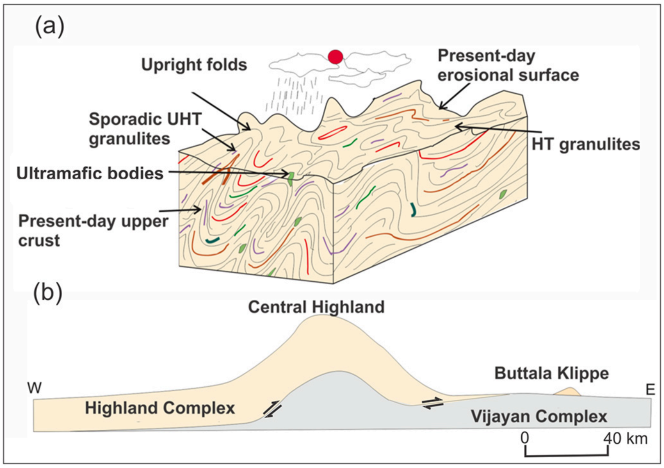

Fig. 7. Sketch diagramme illustrating (a) present-day erosion surface of the HC; see text for details; (b) schematic E-W crustal section across the central part of Sri Lanka after thrusting of the HC on top of the VC, showing only the 20–40 km depth level.

6.4. Field gradients, and spatial distribution of UHT granulites in the HC

Since the large-scale upright folds (wavelengths of about 7–10 km and an exposed length parallel to their axes up to 50 km; Kleinschrodt, 1994) are exposed in the present erosional surface it is possible to encounter lower crustal rocks originally from different levels, with recorded peak metamorphic conditions locally varying 0.5–1.5 kbar and 75–100 °C within a few hundred of meters The best example of an irregular *P* gradient in the field is provided by mineral assemblages of khondalites in the central part of the the HC: some contain ilmenite in the peak mineral assemblages whereas others contain rutile (e.g. Dharmapriya et al., 2015a, 2017b; Hiroi et al., 1994; Raase and Schenk, 1994). The considerable variation in reported peak metamorphic pressures and temperatures, ranging from 790 to over 900 °C and 8 to over 10.5 kbar (e.g. Dharmapriya et al., 2015a, 2017b; Raase and Schenk, 1994; Takamura et al., 2015) may partly reflect these irregular P-T gradients.

Not all rocks in the UHT granulite domain may contain the typical UHT assemblages, which are largely governed by the bulk rock chemical compositions (e.g. Harley, 2008; Kelsey and Hand, 2015). Dharmapriya et al. (2015a, 2015b, 2016, 2021) also demonstrated that the occurrence of rocks having specific minerals and/or mineral assemblages, representing UHT metamorphism as boudinaged blocks, disrupted layers or lenses in the HC, is mainly due to the original compositional variations (inhomogeneity of bulk rock composition) of the particular rock. However, using pseudosection modeling and conventional thermobarometric calculations on specific mineral/s and/or mineral assemblages from UHT rocks as well as their host rock rocks, it could be shown out that all rocks collected from one locality reached the same P-T conditions (Dharmapriya et al., 2015a, 2015b, 2016, 2021).

6.5. Implications for other UHT terrains

Our proposed model has a bearing on granulite terrains in general. UHT granulites may be produced as a thin zone above an ultrahot palaeo-Moho, and overlain by ordinary HT granulite terrains, during collision when an additional heat source is available, such as asthenospheric upwelling after deep lithospheric delamination. Due to intense folding and thrusting of lower to lowermost crustal rocks during post-tectonic uplift still at high temperature conditions, UHT granulites in the lowermost crust can become mixed with and incorporated into the HT ordinary granulites. In addition, the spatial relationships may be complicated by ductile deformation during early cooling. Further, UHT granulites occur only as isolated localities among ordinary HT granulites in many granulite terrains such as Lützow-Holm Bay, Antarctica (Kawakami and Motoyoshi, 2004), Trivandrum Block, India (Morimoto et al., 2004), and Madurai Block, India (Santosh and Sajeev, 2006), with fairly similar field occurrences to those in the HC. Thus, the above suggested tectonic model for the HC could be applicable to UHT granulites in many other HT-UHT terrains as well.

7. Conclusions

UHT metamorphic assemblages have been reported from ~20 localities within the Highland Complex (HC) of Sri Lanka and its outliers. Their spatial distribution, and that of ultramafic bodies in the HC, defines a relatively thin (several km) lower crustal layer that has been refolded into a slightly N-plunging Z shaped fold, looking N. We refer to this domain as the “UHT granulite zone” f the HC. Its core is the shallowly dipping flank, ~50 km wide and ~ 100 km long, trending NNW-SSE. The southern and northern segments each represent a steep flank of the (re)folded UHT granulite zone.

We interpret the UHT granulite zone as the lowermost crustal layer immediately above the hot palaeo-Moho. This layer has been imbricated and intensely refolded with the overlying high temperature (HT) granulites during the post-peak D_3 deformation event. The D_3 deformation caused uplift of HC granulites accompanied with intense folding and thrusting, which disturbed the pre-existing paleo-isotherms, leading to a thicker package of HT granulites with isolated UHT assemblages at the present erosional surface.

This concept has some important implications for granulite terrains in general: (i) UHT granulites elsewhere may also represent a thin zone above an ultrahot palaeo-Moho and overlain by HT granulite terrains, and their spatial relationships may have been complicated by ductile deformation during early cooling; (ii) many other HT granulite terrains may be underlain by a similar zone of UHT granulites, and, if so, new discoveries of UHT terrains may be expected.

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Declaration of Competing Interest

The authors declare that they have no known competing financial interest sor personal relationships that could have appeared to influence the work reported in this paper.

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