



GR Focus Review

Metacraton: Nature, genesis and behavior

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ABSTRACT

In this paper, we show with examples that cratons involved in intercontinental collisions in a lower plate position are often affected by orogenic events, leading to the transformation of their margins. In some cases, craton interiors can also be shaped by intense collisional processes, leading to the generation of intra-cratonic orogenic belts. We propose to call these events “metacratonization” and the resulting lithospheric tract “metacraton”. Metacratons can appear similar to typical orogenic belts (i.e. active margin transformed by collisional processes) but are actually sharply different. Their main distinctive characteristics (not all are present in each metacraton) are: (1) absence of pre-collisional events; (2) absence of lithospheric thickening, high-pressure metamorphism being generated by subduction, leading to high gradient in strain and metamorphic intensity; (3) preservation of allochthonous pre-collisional oceanic terranes; (4) abundant post-collisional magmatism associated with shear zones but not with lithospheric thickening; (5) presence of high-temperature–low-pressure metamorphism associated with post-collisional magmatism; (6) intracontinental orogenic belts unrelated to subduction and oceanic basin closures. Reactivation of the rigid but fractured metacratonic lithosphere will cause doming, asthenospheric volcanism emplacement, and mineralizations due to repetitive mineral enrichments. This paper provides several geological cases exemplifying these different metacratonic features in Scandinavia, Sahara, Central Africa and elsewhere. A special focus is given to the Saharan Metacraton because it is where the term “metacraton” originated and it is a vastly expanded tract of continental crust (5,000,000 km²). Metacratonization is a common process in the Earth's history. Considering the metacraton concept in geological studies is crucial for understanding the behavior of cratons and their partial destruction.

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1. Introduction

Cratons are defined as “part of the crust which has attained stability and which has not been deformed for a long time” (Bates and Jackson, 1980), thus they are Precambrian in age (Kusky et al., 2007). Such stability is attributed to the presence of a thick lithospheric mantle giving a high rigidity to cratons (Black and Liégeois, 1993 and references therein). However, cratons can be involved in continental collisions and be partly reactivated to generate continental tracts that are no longer cratons but that are not typical orogenic belts either. Such continental regions have been initially called “ghost craton” (Black and Liégeois, 1993) and were subsequently referred to as “metacraton” (Abdelsalam et al., 2002). A metacraton has been defined as “a craton that has been remobilized during an orogenic event but is still recognizable dominantly through its rheological, geochronological and isotopic characteristics” (Abdelsalam et al., 2002). “Meta” is a Greek prefix meaning “after” (in time or in space), but this prefix does not imply a direct temporal subsequence to the main event as the prefix “post” implies. For example, the term “post-collisional” refers to events that occurred shortly after collision. In contrast, metacratonic events can occur long after the craton was formed. “Meta” can also mean succession, change, or transformation. All these meanings are well-suited for describing the processes affecting cratons during continental collisions.

Abdelsalam et al. (2002) original definition of the term “metacraton” was strictly descriptive, and did not present explanations for the metacraton’s genesis. Since the introduction of the term, several regional studies have applied the metacraton concept and brought important constraints for the understanding of metacratonic processes in NE Africa (Abdelsalam et al., 2003; Bailo et al., 2003; El-Sayed et al., 2007; Finger et al., 2008; Küster et al., 2008), in Hoggar (Acef et al., 2003; Liégeois et al., 2003; Bendaoud et al., 2008; Henry et al., 2009; Fezaa et al., 2010), in the Zambian Irumide (De Waele et al., 2006), in the Moroccan Anti-Atlas (Ennih and Liégeois, 2008), in Cameroon (Kwekam et al., 2010; Shang et al., 2010a), in Brazil (da Silva et al., 2005) and in China (Zhang et al., 2011a,b), in addition to numerous publications that have accepted the use of the term “Saharan Metacraton” to define this part of the continental crust in northern Africa.

It is thus of timely importance to properly define the concept of metacraton, discuss metacratonization processes, and present several examples to highlight the diversity existing within the common structure of metacraton and metacratonization. This is of paramount importance because the evolution of many metacratonic regions were often not understood because they are misinterpreted as only orogenic belts. Defining the nature, genesis and behavior of metacratons will enable the geoscientific community to recognize these important continental blocks and understand their tectonic characteristics; this is the aim of this paper.

2. Main characteristics of a metacraton

A craton is underlain by a thick, rigid and cold lithosphere. Events such as the initiation of subduction zones transform the craton’s edges into active margins. Such events will lead to the loss of the craton’s lithospheric rigidity and coldness, ultimately transforming the cratonic margins into orogenic belts in which old lithologies are

mixed with juvenile ones. In this case, it is not easy to demonstrate that these old lithologies were part of a cratonic area and all cratonic characteristics will be lost. Metacratons, however even when they were formed through severely modifying tectonic processes, can still preserve major cratonic characteristics, especially rheological properties. Metacratonization can occur either at the margins of cratons or within their interiors (including the hinter parts of subduction-related margins), depending on the intensity of the metacratonization processes.

Due to the high level of force needed to destabilize rigid and thick cratonic lithosphere, it is most likely that metacratonization occurs dominantly during collisional or post-collisional events. Thus, this sequence of tectonic events can be used to establish several constraints on the metacratonic features abbreviated here as (mCf). These mCfs are outlined below and their significance in the evolution of metacratons will be demonstrated through several examples. It should be noted that it is not necessary for a given metacraton to display all mCfs to be qualified as a metacraton but that all metacratons must not bear features that contradict any of the mCfs.

2.1. mCf-1

A collision resulting from an oceanic basin closure involves by definition an active margin and a passive margin. In contrast to the active margin, the passive margin is not affected by major orogenic events before the collision. This description qualifies metacratons to be characterized by the absence of pre-collisional orogenic events.

2.2. mCf-2

During collision, the former passive margin, being located in the lower plate, will be subducted. Hence, in the case of a cratonic passive margin, a thick lithosphere is subducted. This results in a sharp increase of the pressure unrelated to lithospheric thickening of the cratonic plate but due to lithospheric plunge (continental subduction). Also, due to the thick nature of the subducted lithosphere, increase in temperature will be limited. The cratonic rigidity imposes a relatively static environment; hence high-pressure–low temperature (HP–LT) metamorphic paragenesis will develop only in regions where sufficient movements take place and are accompanied by fluids percolation, i.e. along shear zones of all scales. This makes metacratons to be characterized by syn-collisional HP–LT metamorphic conditions not linked to lithospheric thickening but locally developed along zones of high strain in selected lithologies preferentially accommodating this strain. Such strain and metamorphic localization allows for the preservation of original and undisturbed cratonic lithologies within regions of low-strain.

2.3. mCf-3

In a position of passive continental margin, cratonic margins can be accreted and overthrust by oceanic terranes (island arcs and ophiolites) that are preserved in this structural position because they are protected by the rigid cratonic lithosphere. Metacratonization can be associated with such early oceanic accretionary event, but it will generally be of limited extent. When this cratonic margin, which is covered by oceanic material, is involved in a later major

continental collision, it could be metacratonized, but the cratonic rigidity will allow for at least a partial preservation of the oceanic terranes. This defines metacratons to be characterized by the presence of well-preserved remnants of allochthonous pre-collisional oceanic units, whose ages are significantly younger than the cratonic basement.

2.4. mCf-4

The subducted cratonic margin is in a metastable state. Any rupture in the cratonic lithosphere in response to high stress will induce asthenospheric upwelling that can lead to melting and magmatism. This would especially occur in post-collisional events when the subducted cratonic margin was subjected to uplift and affected by transcurrent movements. Failures in the craton's rigid parts will favor magmatism while more ductile terranes will accommodate the stress through folding and plasticity. This produces a scenario in which metacratons are characterized by post-collisional magmatism associated with transcurrent movements but not related to lithospheric thickening. This magmatism is much younger than the cratonic basement and its source is either the old lithosphere or the asthenosphere, depending on the intensity of the metacratonization (these source-contrasted magmatisms can occur successively or simultaneously).

2.5. mCf-5

In the case of igneous events that resulted in emplacement of large magmatic bodies within the metacratonic lithosphere, high-temperature–low-pressure (HT–LP) metamorphism could develop in association with the intrusion of the magmatic bodies. Hence, metacratons can be characterized by batholith-related, post-collisional HT–LP metamorphism, superimposed on the much older metamorphic parageneses of the cratonic basement.

2.6. mCf-6

During continental collisions, the interior of cratons can be affected by far-field stresses leading to reactivation of pre-existing zones of weakness. The latter are inherited from the pre-cratonic geological history reflecting the fact that cratons were initially assemblages of juvenile terranes and/or of continental fragments. As in cratonic margins, such reactivation will form brittle fractures in the cratonic lithosphere, leading to asthenospheric upwelling but also, when more intense, to ductile hot belts and lithospheric magmatism. Being located within continents, no subcontemporaneous oceanic and subduction-related lithologies are present. This makes the interior of metacratons to be characterized by asthenospheric volcanism and doming when these reactivations are of low intensity. Differently, high intensity reactivation results in the development of intracontinental zones of HT–LP metamorphism and lithospheric magmatism (hot belts), the origin of which is not related to subduction and oceanic basin closure processes as indicated by the absence of oceanic lithologies within these zones.

3. (Nearly) amagmatic metacratonization: the Caledonian evolution of Baltica (“cold orogen”)

The Baltic Shield, or Baltica, is part of Fennoscandia and was mainly formed during the Archean and the Paleoproterozoic (Gorbatshev and Bogdanova, 1993). Its south-western margin was subsequently subjected to the late Mesoproterozoic/early Neoproterozoic Sveconorwegian orogeny that started with accretion of oceanic terranes and ended with the intrusion of massif-type anorthosites (Schärer et al., 1996; Bingen et al., 2005).

The western margin of Baltica is covered by ~2000 km long and ~350 km wide Caledonian nappes (Fig. 1). These nappes were overthrust during the closure of the Iapetus Ocean, resulting in the collision between Laurentia and Baltica during the Silurian (430 Ma). The Caledonian nappes have been divided into the Lower, Middle, Upper and Uppermost Allochthons (Fig. 2; Roberts, 2003). The main vergence of these allochthons is towards the SE on distances in excess of several hundred kilometers. The Archean–Paleoproterozoic metamorphic and magmatic rocks of Baltica crop out as inliers within the nappes (Fig. 2). The Lower and Middle Allochthons are made-up of Baltica crystalline rocks and siliciclastic sedimentary rocks translated from Neoproterozoic margins on the shield as demonstrated by their detrital zircons (Bingen et al., 2011). The Upper and Uppermost Allochthons include Neoproterozoic to Paleozoic sedimentary rocks, volcanic-arc magmatic rocks and ophiolite complexes, initially formed within the Iapetus Ocean or along the margin of Laurentia. By contrast, there are no Caledonian magmatic rocks within the allochthons made-up of Baltica basement (hereafter called “Baltica allochthons”).

The absence of Caledonian magmatic rocks in Baltica allochthons (“cold orogen”) highlights the fact that the evolution of the Norwegian Caledonian Belt does not fit the classical orogenic belt model. Nevertheless, the peculiarity of this geodynamic setting has rarely been addressed. Here, we make the case that the geodynamic setting of the Norwegian Caledonian Belt can be explained through metacratonization processes.

Before the Caledonian orogeny, Baltica was a craton, the latest orogenic events affecting its SW margin occurred around 930 Ma (Bingen et al., 2005 and references therein). This region is underlain by a 200–300 km thick heterogeneous lithosphere as defined by the Lithosphere–Asthenosphere boundary (LAB) with a sharp discontinuity to the south along the Tornquist Trans European Suture Zone (Eken et al., 2007 and references therein).

During the Caledonian orogeny, the western margin of Baltica was strongly reactivated but kept its rigid cratonic behavior as expected for a tectonic evolutionary path leading to metacratonization. This

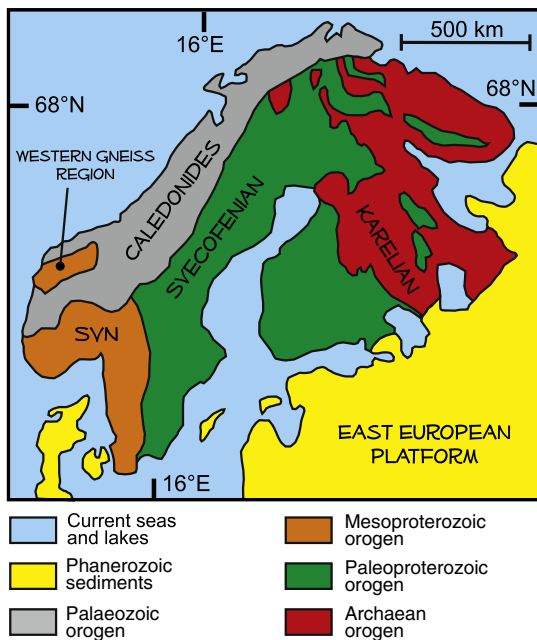


Fig. 1. Main geological subdivisions in Scandinavia showing the importance of the Caledonides nappes thrust over the Precambrian shield. Adapted from Gaál and Gorbatshev, 1987.

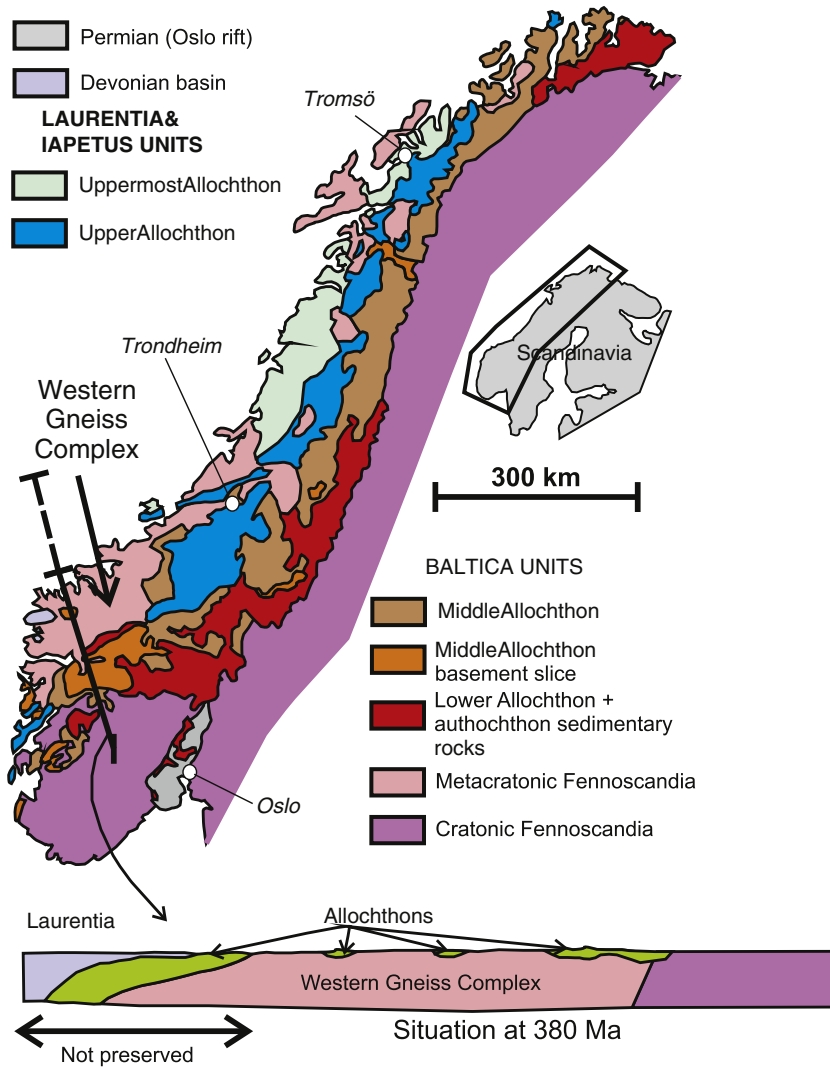


Fig. 2. Geological map of the structural units constituting the Scandinavian Caledonides. Simplified from Be'eri-Shlevin et al., 2011.

metacratonization affected the Baltica lithologies present in the Lower and Middle Allochthons but also the westernmost parautochthonous basement, called the Western Gneiss Complex (WGC, Fig. 2). The WGC is composed of the Proterozoic Fennoscandian gneisses (from c. 1650 Ma and 950 Ma) belonging to the Sveconorwegian orogen (e.g. Skår and Pedersen, 2003). The Caledonian orogeny can be summarized in three successive tectonic events (Hacker et al., 2010): (1) 435–415 Ma: closure of the Iapetus ocean and emplacement of allochthons onto Baltica, which began to be subducted; (2) 415–400 Ma, continental subduction within the Baltica–Laurentia collisional framework. Baltica was the lower plate (former passive margin) and was subducted westward (Fig. 3A). Subduction of the Baltica basement and portions of the allochthons to critical depths led to the generation of ultra-high pressure metamorphism (up to 3.6–2.7 GPa) within the Fennoscandian basement (Lappin and Smith, 1978), without any thickening of the Baltica lithosphere (Fig. 3B). This corresponds to a high-pressure metacratonization; (3) 400–395 Ma: exhumation to shallow crustal levels associated with a dextral transcurrent movement resulted in localized melting (HP–LT leucosomes; Labrousse et al., 2011) but did not produce mobile magmas (Fig. 3B). This corresponds to a low-pressure metacratonization. Later (395–380 Ma), the exhumation was associated with extensional structures resulting from the relative movement between the exhuming WGC and the slipping allochthons.

The development of HP–LT metamorphic parageneses are linked to movements and to fluid percolations indicating activation of shear zones at all scales. North of the Scandinavian Caledonides, in the Lofoten islands, Caledonian eclogites were developed in <4 m wide shear zones in Paleoproterozoic gabbro-anorthosite in association with Cl-rich fluids (Küllerud et al., 2001). These Lofoten eclogites are older (505–450 Ma) than those in the Western Gneiss Region and record a long exhumation history (Steltenpohl et al., 2011). This indicates that the Lofoten eclogites have an old Baltic protolith and occur within the autochthonous Baltic basement in agreement with Steltenpohl et al. (2011) who stated that “continental crust in a collisional setting can be subducted to mantle depths and show only very sparse evidence of this tectonic history”. This tectonic setting is strongly suggestive of a metacratonic evolution.

The globally synchronous evolution (although infrequent diachronism may exist) between the onset of collision and peak metamorphism stretching for hundreds of kilometers along the Caledonian orogen in Scandinavia indicates that a large portion of the margin of Baltica behaved uniformly during collision with Laurentia (Spengler et al., 2009). This is in good agreement with what is expected for the subduction of a rigid, cold and thick cratonic margin. Such a subduction also explains the observed extreme P–T conditions (6.3 GPa and 870 °C) pointing to a 200 km subduction depth and a long period (c. 30 m.y.) of UHP metamorphism (Spengler et al., 2009). The important role

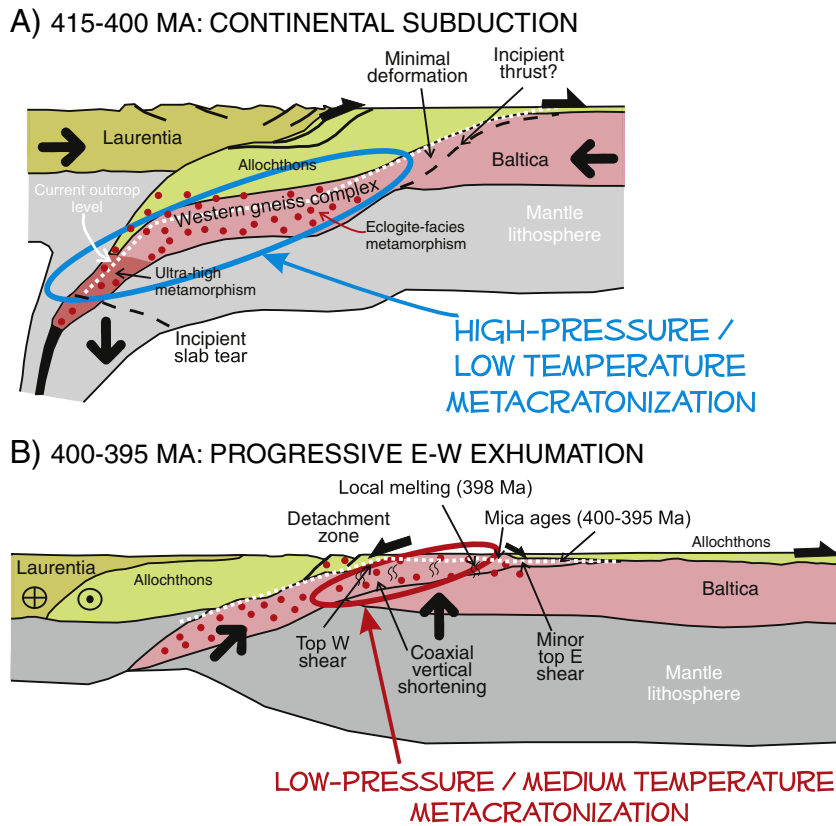


Fig. 3. Continental subduction of the western margin of the Baltica craton during the Caledonian orogeny (NW–SE schematic cross section; slightly modified from Hacker et al., 2010). The present exposure level is indicated by white dotted line. (A) After allochthons emplaced onto Baltica craton, both are subducted to UHP depths coincident with continued thrusting and cooling in the foreland. (B) Baltica margin (Western Gneiss Region) rises to crustal depths. Metacratonization occurs at great depth (A: HT–LP metamorphism) but also at shallower depth (B: higher temperature/lower pressure metamorphism with some migmatitization).

of strike-slip movements with minor oblique-slip component during the exhumation is demonstrated by the recorded maximum metamorphic P–T conditions that vary systematically between the hinterland and the foreland, but discontinuously parallel to the orogen strike (Spengler et al., 2009 and references therein).

Differently, some Fennoscandian lithologies (such as the Hustad complex constituting a c. 1.65 Ga granite batholith and a thick c. 1.25 Ga dolerite dyke) have been preserved in the WGC. These regions of low-strain are well preserved and are nearly devoid of Caledonian deformation and metamorphism. In the Hustad complex, the U–Pb and the Rb–Sr chronometers gave the same (within error

limits) late Paleoproterozoic Statherian ages fundamentally different from the Caledonian event which is only marked by the intrusion of pegmatites (c. 390 Ma) and by Caledonian ages determined from some zircons and baddeleyites U–Pb lower intercepts (Austrheim et al., 2003). The presence of well-preserved low-strain-regions within highly deformed and metamorphosed regions (represented by UHP rocks) is typical in metacratons. Such tectonic setting can be attributed to the presence of rigid, thick and cold, but fractured, metacratonic lithosphere.

The Caledonian Baltica tectonic evolution illustrates several metacratonic features: (1) mCf-1, the absence of Baltica pre-collisional orogenic events (i.e. in the 500–430 Ma age range); (2) mCf-2, the UHP metamorphism and associated structure are syn-collisional not linked to lithospheric thickening. These UHP metamorphic belts are separated by well-preserved regions of low-strain; (3) mCf-3, the preservation of large areas dominated by Paleozoic age pre-collisional oceanic assemblages (Iapetus) and former active margin assemblages (Laurentia) within the Upper and Uppermost Allochthons much younger than the Paleozoic and Mesoproterozoic ages of Baltica basement. Features mCf-4 and mCf-5 do not apply to the proposed metacratonization model of Baltica since this case of metacratonization was not associated with magmatism and thus no HT metamorphism had occurred during the Caledonian orogeny in Baltica (“cold orogen”). Presence of trondhjemitic/tonalitic leucosomes attest for partial melting but the latter was limited in extent and occurred at high pressure (>25 kbar and 750 °C; Labrousse et al., 2011). This indicates that Baltica thick continental lithosphere was thermally shielded and has not been disrupted throughout the metacratonization process, WGC thrusting being intra-crustal (Fig. 3). The metacratonization and high-pressure partial melting could be linked

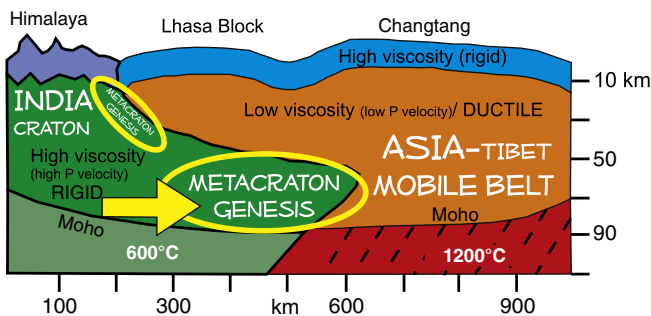


Fig. 4. Section of the lithosphere showing the current relation between the thick, cold and rigid Indian lithosphere and the thin, hot and ductile lithosphere of Tibet. Craton boundaries are submitted to high stress, pressure and temperature that can lead to their partial reactivation without lithospheric thickening, i.e. to their metacratonization, at high and low depth.

Adapted from Bendick and Fleisch, 2007.

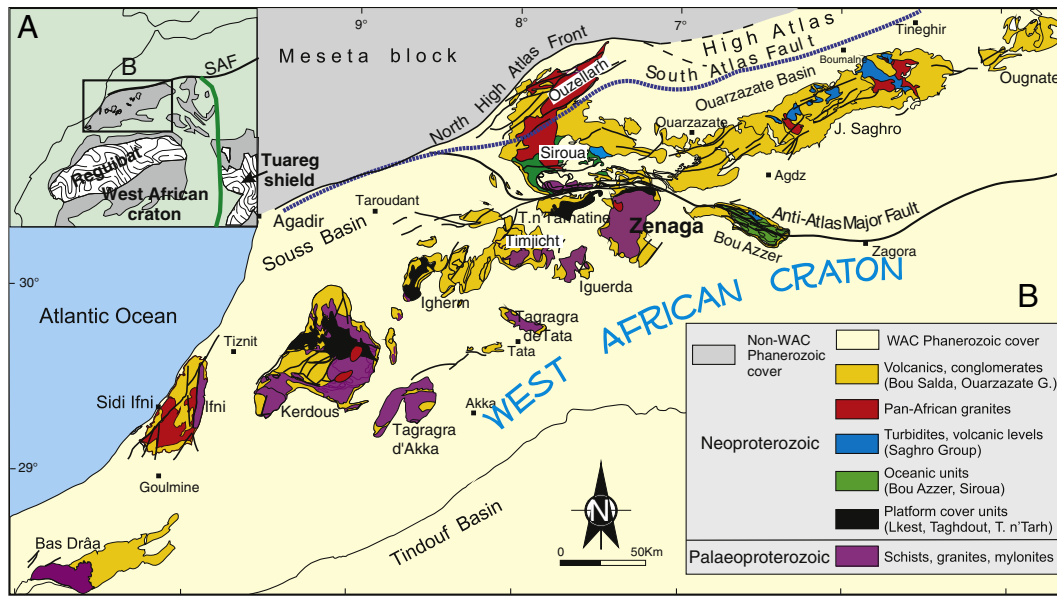


Fig. 5. Geological map of the Anti-Atlas mountain range (Morocco) based on Thomas et al., 2002, 2004; Gasquet et al., 2008.

to the detachment of the subducting oceanic lithosphere, leading to the initiation of exhumation (Labrousse et al., 2011).

A similar present-day analog of metacratonization can be argued for the India–Tibet tectonics. The rigid and cold Indian lithosphere which is currently subducted under the ductile and hot Tibet (e.g. Bendick and Flesch, 2007) is subjected to high pressure and stress at greater depth and to fluid movements and stress at shallower depth (Fig. 4). Preservation of underthrust Greater India, which has advanced sub-horizontally northward by some 600 km beyond the collisional front is attributed to the compositional buoyancy and high strength of its thick lithospheric mantle (Chen et al., 2012). However, it is likely that the subducted Indian lithosphere is currently undergoing metacratonization.

The contrasting lithospheric structure between Baltica craton and its metacratonic western margin could explain some of the Cenozoic geological and morphological features characterizing the Norwegian Caledonian Belt. Indeed, the origin of the anomalous anorogenic uplift of the Norwegian passive margin to produce > 2000 m high mountain ranges, is still highly debated and not fully understood (e.g.: Nielsen et al., 2009 and Chalmers et al., 2010). We propose that such uplift is due to far-field stress (mostly mid-ocean ridge push) accommodated by the rigid but fractured Norwegian metacratonic boundary in the form of tectonic rock uplift. Such far-field stress is not expected to translate into uplift along the margins of a typical craton underlain by a coherently rigid and rheologically homogeneous lithosphere or within orogenic belts that can accommodate the stress through folding and thrusting. This would imply that that metacraton rheology influences the way stress is accommodated long after their formation.

4. Magmatic metacratonization of a craton margin: the Pan-African evolution of the West African Craton northern boundary, the Anti-Atlas (Morocco)

The West African Craton (WAC) was formed during the Archean and the Paleoproterozoic (~2 Ga, Eburnian orogeny). Mesoproterozoic quiescence from ~1.7 to 1.0 Ga allowed this large area to develop into a craton. At the end of the Neoproterozoic, the WAC was subjected to convergence on all its margins along the Anti-Atlas (Morocco) in the north, Trans-Saharan belt (from Algeria to Nigeria) in the east, Rockelides and the Bassarides (from Liberia to Guinea) in the south, and the

Mauritanides (Senegal and Mauritania) in the west. This led to partial remobilization of the craton's margins. In the north in the Anti-Atlas (Fig. 5), this remobilization was marked by transpressional and trans-tensional tectonics, intrusion of granitoids (Assarag and Bardouz suites; Thomas et al., 2004) and extrusion of voluminous volcanic rocks such as the Ouarzazate Supergroup and by alteration of the 2 Ga basement (Ennih and Liégeois, 2001, 2008).

The Zenaga inlier in the central part of Anti-Atlas (Fig. 5) is a depression of about 500 km² containing mainly Paleoproterozoic gneisses and granitoids unconformably overlain by the late Neoproterozoic Ouarzazate volcanic Group or, in places, directly covered by the Cambrian Tata Group (Thomas et al., 2004). Neoproterozoic rocks, also present within the inlier, are represented by passive margin sedimentary rocks (Taghdout Group; Bouougri and Saquaque, 2004; Fig. 5), pre-Pan-African doleritic dykes and sills, and a late Pan-African alkaline ring-complex (red dot in Zenaga; Fig. 5). The northern boundary of the inlier is limited by the Anti-Atlas Major Fault (AAMF) where remnants of the Cryogenian Bou Azzer–Siroua oceanic terranes are present to the east and west (Samson et al., 2004; Fig. 5). The Zenaga inlier is the northernmost outcrop of the West African craton but the actual lithospheric boundary of the inlier is located further north along the South Atlas Fault according to Ennih and Liégeois (2001, 2008, Fig. 5) or, further north along the North High Atlas Front following Gasquet et al. (2008, Fig. 5).

Although often considered as an undisturbed Eburnian (c. 2 Ga) basement, the Zenaga rocks have been affected by the Pan-African orogeny (Ennih and Liégeois, 2008). The Zenaga Paleoproterozoic metamorphic rocks, dominantly schists, represent a high-grade metamorphic supracrustal series. These schists have not been dated but the presence of inherited zircons dated at c. 2170 Ma within the c. 2035 Ma cross-cutting granitoids could be approximated to be the age of the Zenaga schists (Thomas et al., 2002). The Zenaga Paleoproterozoic plutons enclose frequent quartzo-feldspathic layers separated by biotite and garnet layers, locally associated with gneisses and anatectic components. They contain rare metasedimentary xenoliths and no mafic microgranular enclaves (MME). These plutons have been dated in the 2030–2040 Ma age range (U–Pb zircon ages, Thomas et al., 2002). The Paleoproterozoic Zenaga basement is overthrust by the Neoproterozoic Bou Salda Group. These rhyolite rocks represent the lower part of the Ouarzazate Supergroup and give U–Pb zircon date of 605 ± 9 Ma (Thomas et al., 2002). This

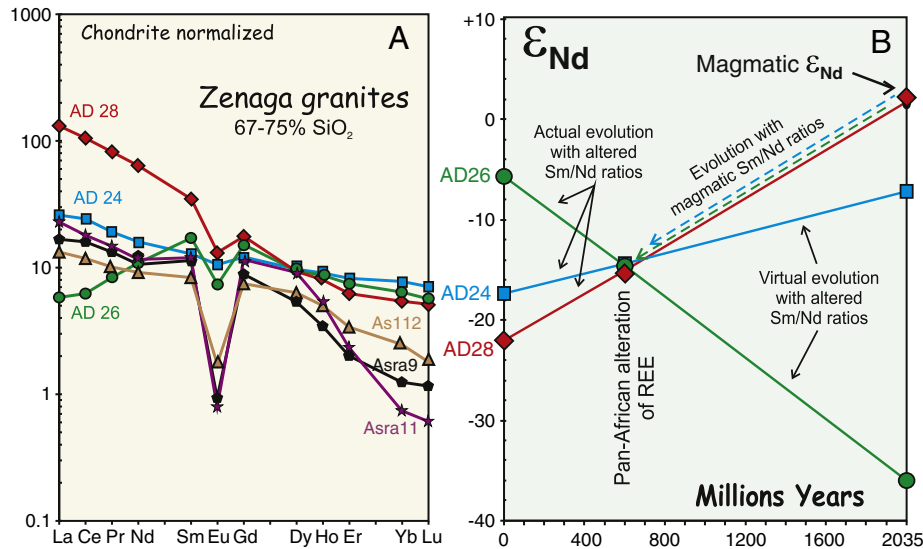


Fig. 6. (A) REE patterns of representative granites from the c. 2.035 Ga Eburnian basement from the Zenaga inlier (Anti-Atlas; Morocco). Both LREE and HREE have been modified by F-rich fluids. (B) This alteration occurred during the Pan-African extrusion of the volcanic Ouarzazate Supergroup, period where the Nd isotopes have been reset. Data and interpretation from Ennih and Liégeois (2008).

group is overlain by the thick pile of volcanic rocks of the Ouarzazate Group (the upper part of the Ouarzazate Supergroup) dated between 581 ± 11 Ma and 543 ± 9 Ma (Gasquet et al., 2005; Fig. 5). Within the Zenaga inlier, the Sidi El Houssein alkaline granitic ring-complex was intruded at 579 ± 7 Ma as suggested by the U–Pb zircon systematics (Thomas et al., 2002), at the same time as the extrusion of the Ouarzazate Group.

The Pan-African deformation occurred locally under greenschist facies metamorphic conditions producing NW–SE oriented fabric along the AAMF corridor (Ennih et al., 2001). The Zenaga Paleoproterozoic granitoids, although showing fresh magmatic appearance in the field, have been strongly chemically altered during the Pan-African period. All these plutons are felsic and strongly peraluminous (ASI from 1.15 to 2.60, decreasing with silica), which is marked by abundant garnet. Considering that all rocks are within the 67–75% SiO_2

limited range, their rare earth elements (REE) display very different patterns (Fig. 6A). They display a large variability in HREE abundance, very low abundance in REE with no Eu negative anomaly, tetrad effect, normalized HREE richer than LREE, strong negative Eu anomaly, and strong depletion in LREE or in HREE. This suggests the influence of F-rich fluids (for the tetrad effect) coupled with the destabilization of HREE-rich minerals (such as garnet) present in these granites. Initial $^{143}\text{Nd}/^{144}\text{Nd}$ values at 2035 Ma relative to bulk Earth (ϵ_{Nd}) are highly variable in these plutons ranging between -36 and $+2$ (Fig. 6B) and even can be as high as $+16$ (not shown; Ennih and Liégeois, 2008). The rather homogeneous ϵ_{Nd} for these rocks at 600–700 Ma (Fig. 6B) coupled with a great variability of the Sm/Nd ratios demonstrate that the destabilization occurred during the Pan-African period. Sr isotopes have been also largely modified during the Pan-African event. The likely cause of these effects was the circulation of F-rich fluids linked to the extrusion of the voluminous Ouarzazate Group and the emplacement of associated plutons (580–545 Ma) along reactivated faults and shear zones (beryl has been mined in the area). It must be stressed that the reactivation event occurred during important vertical movements (the current thickness of the Ouarzazate Group varies from 0 to more than 2200 m depending on the vertical fault-bounded block considered) but without major crustal or lithospheric thickening. This is demonstrated by the low-grade character of the Pan-African metamorphism (greenschist facies) and by the excellent preservation of the c. 800 Ma Taghdout passive margin sedimentary rocks which still show primary structures such as desiccation cracks and ripple marks, and by the preservation of the early c. 750–700 Ma ophiolitic complexes.

The model proposed for the Anti-Atlas Pan-African period advocates for continental subduction of the northern margin of the West African craton first below Cryogenian island arcs (accretion of ophiolitic sequences) and second below Peri-Gondwanan terranes during the Ediacaran (for paleogeographical reconstructions, see Nance et al., 2008 and references therein). As far as we are aware, the Cryogenian accretions did not affect the craton, the gneisses which were previously interpreted as reworked Paleoproterozoic were found to have Neoproterozoic protoliths (D'Lemos et al., 2006). By contrast to the rather orthogonal Caledonian Baltica–Laurentia collision, the Pan-African (c. 600 Ma) West African craton–Peri-Gondwanan terranes collision was highly oblique (Ennih and Liégeois, 2001, 2008; Ennih et al., 2001), limiting the subduction depth of the West African craton margin. However, due to intense transpressional movements (marked by the

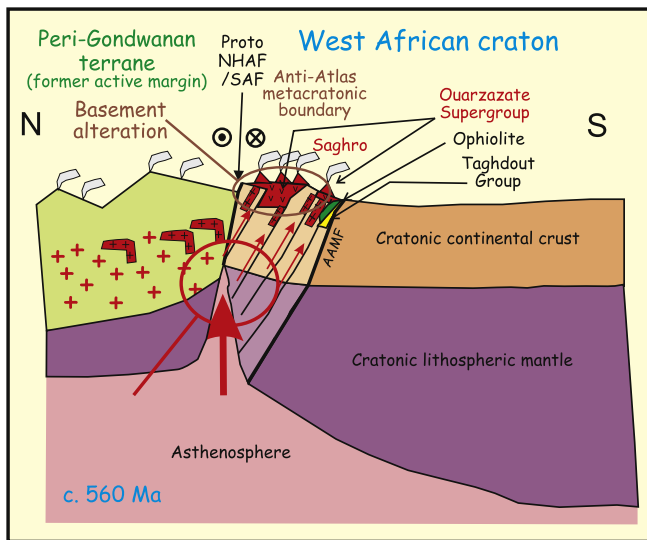


Fig. 7. Schematic cross-section of the Anti-Atlas and Peri-Gondwanan terranes adjacent at the time (c. 560 Ma). Anti-Atlas constitutes the metacratonic boundary of the West African craton. The latter, fractured but not thickened, was invaded by huge amounts of alkali-calcic magmas during Ediacaran but preserved the Eburnian basement, the c. 800 Ma passive margin sediments (Taghdout Group) and the Cryogenian ophiolite and associated rocks. NHAF=North High Atlas Front; SAF=South Atlas Fault. Modified from Ennih and Liégeois (2001).

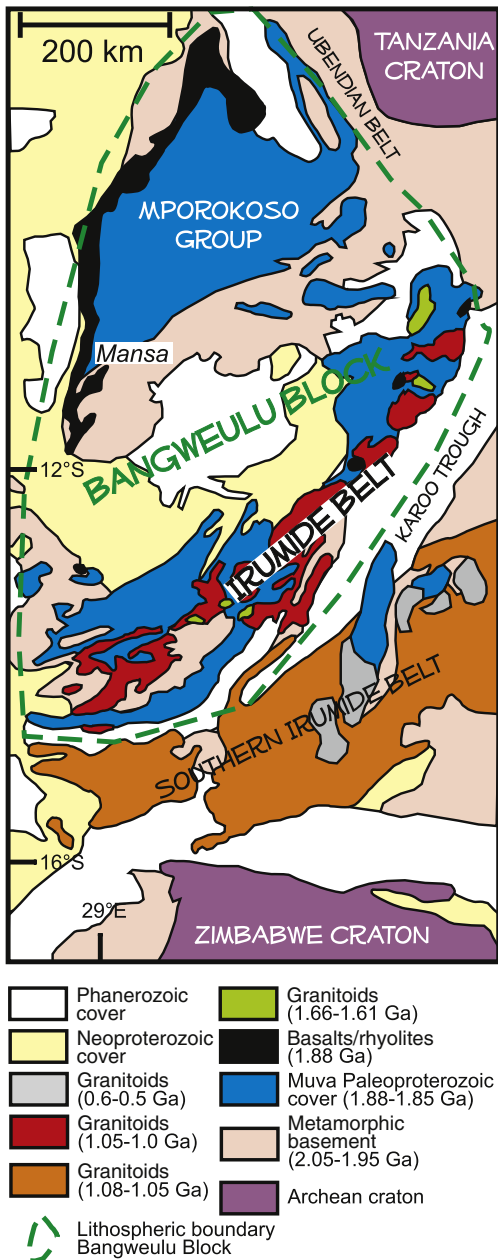


Fig. 8. Geological map of the Bangweulu Block, the Irumide Belt and adjacent areas. Modified from De Waele et al. (2006).

deposition of the Saghro Group; Liégeois et al., 2006; Gasquet et al., 2008), the cratonic margin was subjected to an intense dissection allowing a large magmatic outpouring when the transpressional movements are replaced with transtensional movements (Gasquet et al., 2008; Fig. 7).

The geological and tectonic observations above are typical of what is expected to accompany a metacratonic evolution. The cratonic margin was dissected by faults and invaded by magmas but preserved most of its lithospheric rigidity and primary features. Such a metacratonic reactivation was favorable for fluid circulation, in this case able to mobilize rare-earth elements.

Since the Pan-African, the metacratonic Anti-Atlas has been reactivated several times, especially during the Variscan orogeny, an event that was responsible for the reconcentration of mineral deposits. The Anti-Atlas is characterized by polyphase mineral deposits (e.g. Pelleter et al., 2007). Importantly, all these reactivations were

Phanerozoic in age and they took place well after the Pan-African metacratonization. Metacratonic areas, allowing fluid movements, appear to be excellent loci for element concentration and genesis of mineralizations, for which the Anti-Atlas is internationally renowned.

The Cenozoic Africa–Europe collision induced the High Atlas Range along the West African craton and the northern metacratonic margin of the WAC has been elevated to an altitude of 2200 m (the Anti-Atlas range, running along the High Atlas). This was accompanied by the emplacement of recent alkaline–peralkaline volcanism (Sirwa and Saghro volcanism; Liégeois et al., 2005; Berger et al., 2009b, 2010).

The Pan-African Anti-Atlas metacratonization is supported by the presence of features mCf-1 (absence of pre-collisional orogenic events), mCf-3 (well-preserved allochthonous oceanic remnants), mCf-4 (abundant post-collisional potassic magmatism without lithospheric thickening), and mCf-5 (HT–LP metamorphism linked to post-collisional magmatism affecting the Eburnian basement). The metacratonic reactivation also influenced Cenozoic uplift and asthenospheric volcanism.

5. Repeated magmatic metacratonization of a craton margin: the Proterozoic evolution of the Irumide Belt (Zambia)

The Irumide belt in Zambia corresponds to the southern metacratonic margin of the Bangweulu Block (De Waele et al., 2006; Fig. 8). In addition to rare Archean remnants (c. 2.73 Ga, G₀ granitoids; De Waele, 2005), it comprises large tracts of Paleoproterozoic amphibolite-facies orthogneisses and metasedimentary successions. Paleoproterozoic granitoids intruded the Irumide Belt in two pulses at 2.04–2.03 Ga (G_{1A}) and at 1.95–1.93 Ga (G_{1B}) (Rainaud et al., 2003; De Waele, 2005; Rainaud et al., 2005). The granitic intrusion event was followed by the development of a passive sedimentary margin (represented by the Muva Supergroup including the Mporokoso Group) and the extrusion of basalts and rhyolites (G_{1C}, c. 1.88–1.85 Ga; De Waele, 2005) (Fig. 8). Localized and more alkaline granitoids intruded at 1.66–1.55 Ga (G_{2A} and G_{2B}) (De Waele et al., 2006). The above lithologies were intruded by voluminous high-K granitoids during the late-Mesoproterozoic Irumide orogeny at ca. 1.02 Ga (G₄) (De Waele et al., 2006 and references therein; in red in Fig. 8). The existence of G₃ granitoids (c. 1.4 Ga, Rb–Sr data; Daly, 1986) has not been confirmed by recent SHRIMP ages (De Waele et al., 2003).

The late Mesoproterozoic event at c. 1.02 Ga was associated with the Rodinia assembly and produced the dominant structural features in the Irumide Belt, largely obliterating previous structural grain. However, older lithologies are well-preserved as exemplified by the c. 1.88–1.85 Ga G_{1C} volcanic rocks. It is remarkable that the geochemistry of the various Proterozoic Irumides magmatic generations share the same geochemical signature (Fig. 9) as well as similar Nd T_{DM} model ages, all being Archean. These observations suggest that all successive generations of magma pulses have the same Archean lithosphere source represented by the G₀ granite gneiss, with minor juvenile inputs (De Waele et al., 2006). Additionally, it must be noted that: (1) the c. 1.88–1.86 Ga granites and volcanic rocks present in the Mansa area (NE Bangweulu Block; Fig. 8) display a similar geochemical signature and Nd T_{DM} model ages (Fig. 9D), pointing to a similar source and magma generation on the NE margin of the Bangweulu Block at that time; (2) the more alkaline G_{2A} (1.66–1.61 Ga; Fig. 9E) granite has similar Nd T_{DM}, but it has a slightly different geochemical signature reflecting a lower degree of partial melting of the same Archean source (De Waele et al., 2006). This also shows the total absence of oceanic subduction-related magmatism during the Proterozoic in the Irumide Belt.

In contrast, subduction-related juvenile terranes are present in the Southern Irumide belt (Johnson et al., 2005a,b) south of a major shear zone now marked by a Karoo trough (Fig. 8). The subduction-related magmatism is slightly older (1.08–1.04 Ga) than

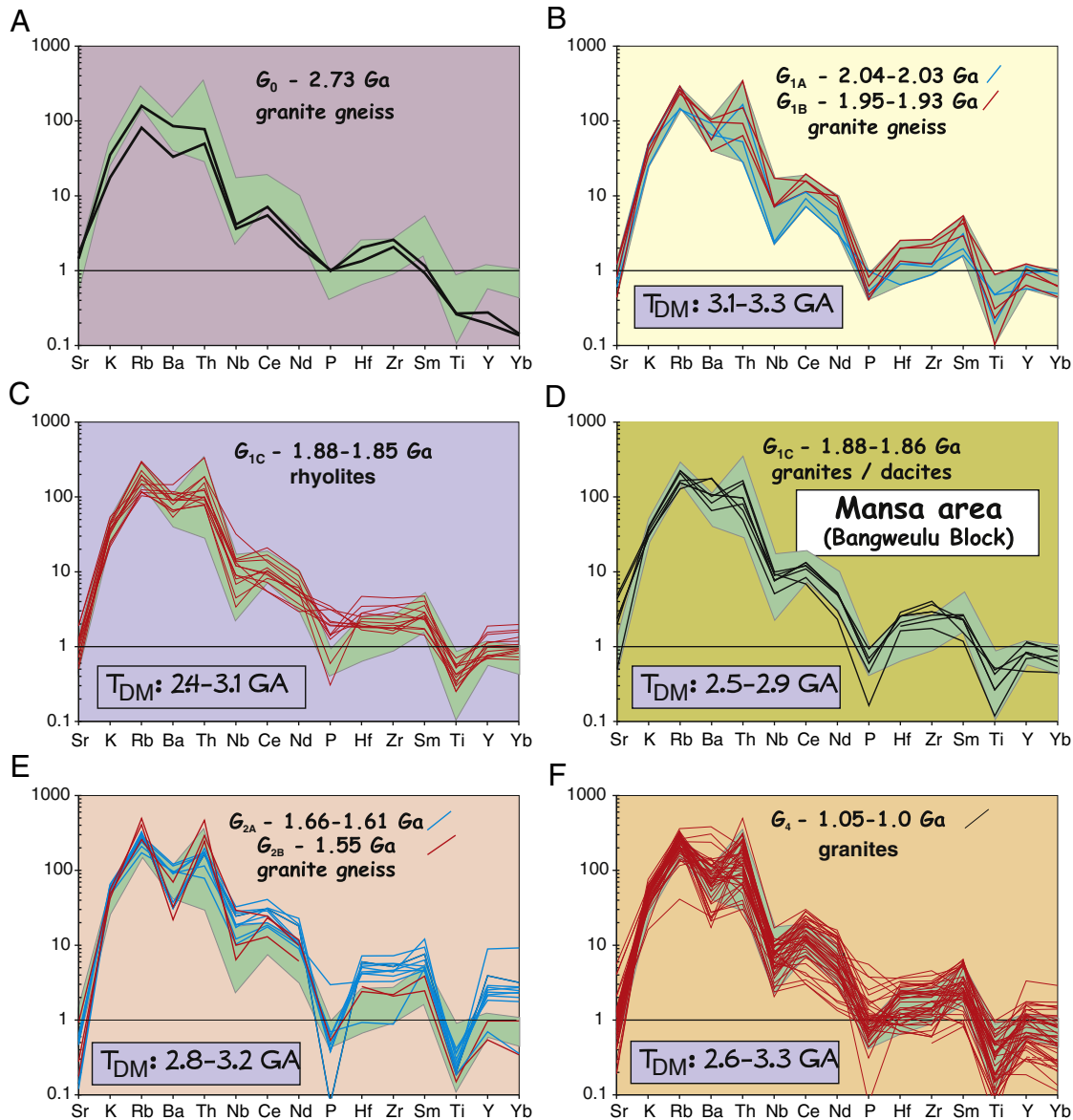


Fig. 9. MORB normalized trace element diagrams for magmatic rocks of northern Zambia (Irumide and Bangweulu Block (Mansa area)). The green underlying pattern represents the envelope of the G1A–B analyses for an easier comparison. Modified from De Waele et al., 2006.

that in the Irumide Belt (G_4 episode at c. 1 Ga). This is consistent with the generation of these juvenile terranes prior to the collision that led to the generation of the Irumide G_4 episode. The latter can thus be explained as due to metacratonization of the south-eastern margin of the Bangweulu Block at the end of the Mesoproterozoic (De Waele et al., 2006).

During the Paleoproterozoic (2.05–1.93 Ga; G_{1A} to G_{1B} generations), magmatism and tectonism appears to have affected the entire Bangweulu Block, suggesting that this block was entirely metacratonized during this period. The Bangweulu Block might have been the metacratonic margin of a larger Archean craton (Tanzania or Congo craton). The G_{1C} (c. 1.85 Ga) and G_2 (c. 1.65 and c. 1.55 Ga) granitic intrusion events represent additional reactivation of the metacratonic margins of the Bangweulu Block, but with less intensity. This means that the Bangweulu Block, and its south-eastern margin, the Irumide Belt, was never an active margin during its entire Proterozoic development.

The Irumide can thus be considered as a belt that had witnessed multiple metacratonization events, but with two notable major

events; one that have affected the entire Bangweulu Block (c. 2 Ga) and another one that was limited to the current trace of the Irumide belt (c. 1 Ga). Hence, observations from Irumide belt satisfy most of the criteria set forth to characterize it as a metacraton including the lack of pre-collisional events (mCf-1), abundant metacratonic magmatism (mCf-4), high-temperature metamorphism but also preservation of older lithologies (mCf-5), and alkaline magmatism with low-degree partial melting at c. 1.65–1.55 Ga (mCf-6).

6. Magmatic metacratonization of a craton margin evolving towards a continental interior: the Pan-African evolution of LATEA (Central Hoggar, Tuareg shield, Algeria)

The Tuareg shield, composed of Hoggar, Air and Iforas regions in Central Sahara (Fig. 10) was assembled during the Pan-African orogeny, at the end of the Neoproterozoic, as a result of collision between the Tuareg terranes and the West African craton (Black et al., 1994). The western part of the Tuareg shield, with the notable exception of the In Ouzal terrane (200 km to the west of LATEA; Ouzegane

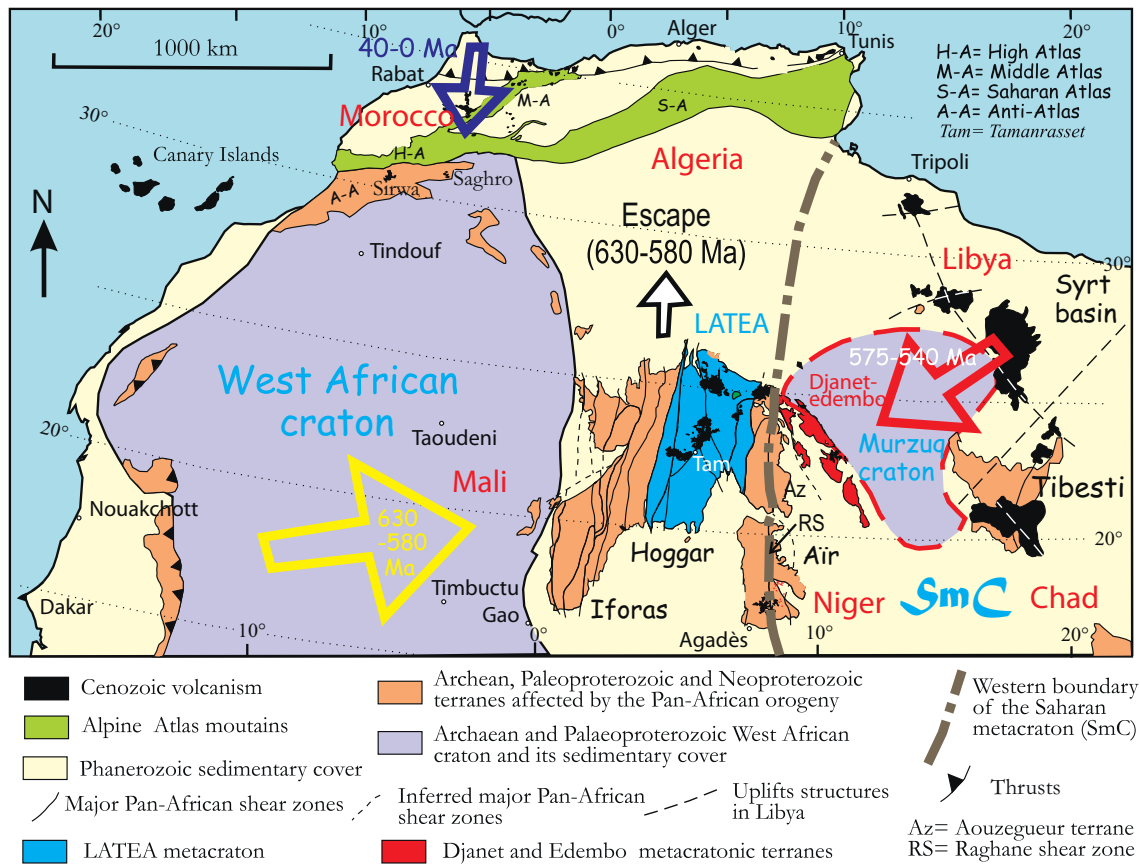


Fig. 10. Main rheological domains from North-West Africa with enhancement of the LATEA metacraton and of the Djanet and Edembo metacratonic terranes (see text). The Cenozoic volcanism is represented showing that it is located within the LATEA metacraton and within the metacratonic area around the Murzuq craton. Adapted from Liégeois et al. (2005) and Fezaa et al. (2010).

et al., 2003) dominantly constitutes terranes that were formed during an island arc and continental arc subduction-related events (730–630 Ma) followed by Ediacaran collisional and post-collisional events (630–580 Ma) (Liégeois et al., 1987; Caby, 2003).

Central Hoggar, historically named “Polycyclic Central Hoggar” (Bertrand and Caby, 1978) had a different evolution, being located 400 km east of the suture between the Tuareg shield and the West African craton (Fig. 10). It has been shown that the four terranes constituting the Central Hoggar (Black et al., 1994) were part of a single pre-Pan-African passive margin (Liégeois et al., 2003). The acronym LATEA (from the first letters of the names of the four terranes, Laoui, Azrou-n-Fad, Tefedest, Egéré-Aleksod; Fig. 11) was given to this region (Liégeois et al., 2003). LATEA, made of Archean and Paleoproterozoic metamorphic and magmatic rocks (Peucat et al., 2003; Bendaoud et al., 2008 and references therein) behaved as a craton during the Mesoproterozoic and the Early and Middle Neoproterozoic where oceanic terranes (such as the juvenile Iskel terrane and the Tin Begane eclogite-bearing nappes) were accreted along its margins during Cryogenian and Ediacaran periods (Fig. 11; Caby et al., 1982; Liégeois et al., 2003; Bechiri-Benmerzoug et al., 2011). No Neoproterozoic events older than 630 Ma have been recorded in the LATEA basement itself, a time that marks the beginning of the Tuareg/West African craton collision. During that collision, the LATEA craton became a metacraton, being dissected into several terranes and intruded by high-K calc-alkaline batholith largely from a preponderant Paleoproterozoic/Archean crustal source (Acef et al., 2003; Liégeois et al., 2003; Abdallah et al., 2007). LATEA lithosphere was subsequently intruded by shallow circular plutons such as the Temagoussine pluton (c. 580 Ma; Abdallah et al., 2007), marking the end of the metacratonization process in LATEA. The dissection of LATEA

lithosphere was accomplished through the activation of mega-shear zones that accommodated several hundred kilometers of horizontal displacement. These shear zones are interpreted as escape roots of the northward expulsion of the Tuareg terranes as LATEA was squeezed between the West African craton to the west and the Saharan metacraton to the east (Fig. 10).

During LATEA’s metacratonization, most of the Archean/Paleoproterozoic basement was preserved. In the Tidjenouine area, the c. 2.06 Ga granulitic parageneses can be documented in greater detail, with nearly no Pan-African metamorphic overprinting (Bendaoud et al., 2008). These rocks show a peak metamorphism at 880 °C and 8 kbar with a loop going to 750 °C and 5 kbar with a final retrogression to amphibolite facies at 580 °C and 3.5 kbar (Fig. 12; Bendaoud et al., 2008). Closer to the mega-shear zones, and thus to Pan-African batholiths, a Pan-African reheating is seen, the orthopyroxene recrystallizing around the amphibole (amphibole + quartz → orthopyroxene + plagioclase; Fig. 12; Bendaoud et al., 2008). This event has been dated at c. 615 Ma (zircon recrystallization age, Fig. 12; Bendaoud et al., 2008), which corresponds to the climax of the batholith intrusions in the area (Bertrand et al., 1986; Liégeois et al., 2003; Talmat-Bouzeguella et al., 2011). Similar P–T–t paths, even if with slightly different absolute values, are observed elsewhere in LATEA (Bouregghda et al., 2010).

Since 580 Ma, LATEA was reactivated along the same shear zones sporadically. Some of these reactivation episodes triggered igneous activities such as the formation of the Taourirt magmatic province (c. 530 Ma; Paquette et al., 1998; Azzouni-Sekkal et al., 2003; Liégeois et al., 2003) and the Cenozoic silica-undersaturated volcanism (Liégeois et al., 2005). Additionally, it is likely that these shear zones had controlled Paleozoic sedimentation suggestive by the

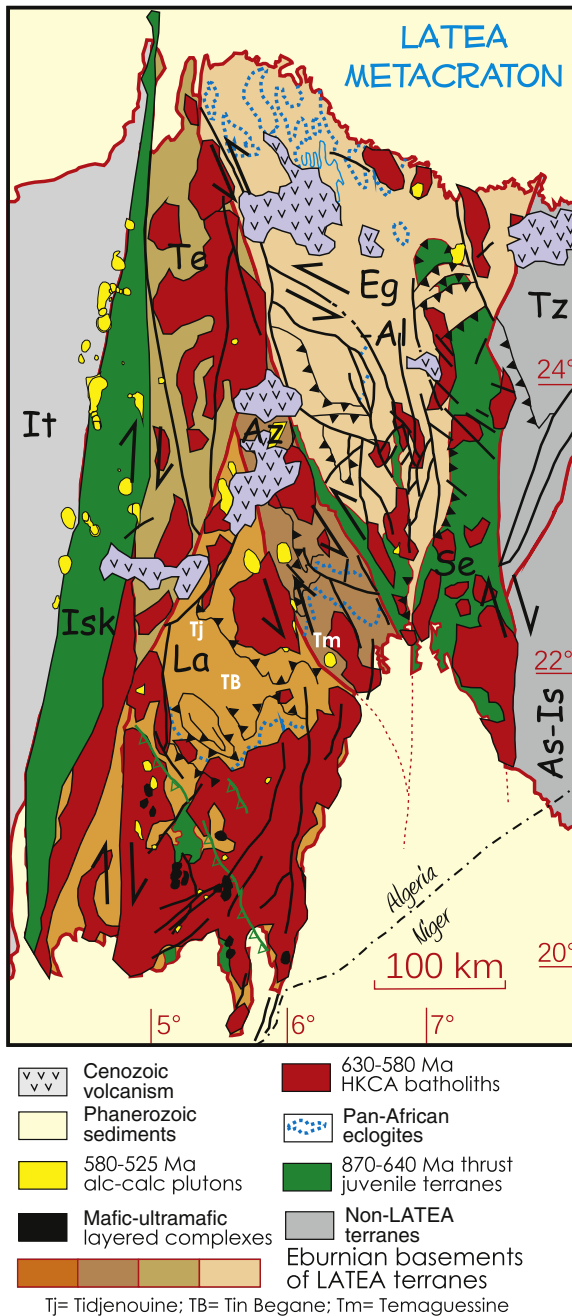


Fig. 11. Geological map of the LATEA metacraton showing the preserved Archean–Paleoproterozoic basement (metacraton) dissected in four terranes (La=Laouni, Az=Azrou-n-Fad, Te=Tefedest, Eg-Al=Egéré-Aleksod), the thrust juvenile Cryogenian terranes (ISK=Iskel, Se=Serouenout with additional more localized material such as the eclogitic bands), the high-K calc-alkaline (HKCA) batholiths (630–580 Ma), the mantle-derived mafic–ultramafic layered complexes, the alkaline/alkali-calcic plutons (principally Taourirts Province) and the Cenozoic Tuareg volcanism mainly located in the LATEA metacraton. Neighbor terranes are In Tedeini (It), Tazat (Tz) and Assodé-Issalane (As-Is). Adapted from Liégeois et al. (2003).

large thickness variations of the Paleozoic sedimentary section across the shear zones (Beuf et al., 1971).

The above analysis indicates that LATEA behaved as a craton during the Tonian and Cryogenian periods when there were no marked collisional orogenic events, except arc accretions not affecting the LATEA basement. Hence, LATEA metacraton displays mCf-1 (absence of pre-collisional orogenic event in the basement), mCf-3 (the presence of well-preserved allochthonous terranes and nappes). Post-collisional

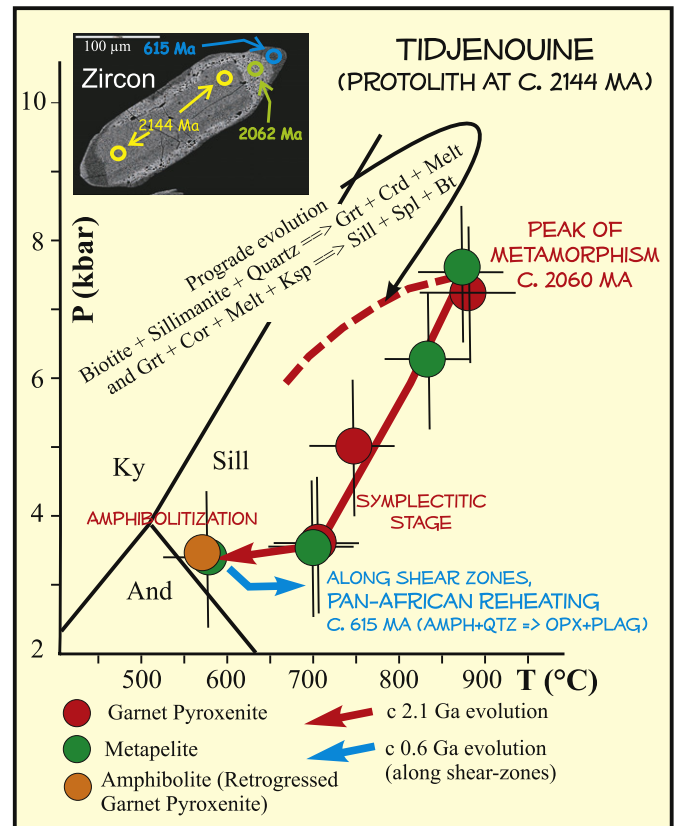


Fig. 12. P–T–t path of the Tidjenouine metapelites and an example of zircon dated showing a magmatic core at c. 2144 Ma, a first metamorphic rim of Eburnian age (c. 2062 Ma) and a second often very thin rim of Pan-African age (c. 615 Ma). Adapted from Bendaoud et al. (2008).

potassic magmatism linked to the activation of mega-shear zones is abundant while no major lithospheric thickening occurred (mCf-4) allowing the preservation of pre-collisional allochthonous units. Well-preserved post-collisional HT–LP metamorphism, spatially associated with the intrusion of granite batholiths is recorded superposed on the 2.06 Ga basement metamorphism (mCf-5). During the Cenozoic, the LATEA metacraton was affected by doming (the current Tuareg Shield) accompanied by asthenospheric volcanism (Fig. 10; mCf-6). Ascent of the asthenospheric volcanism was facilitated by the reactivation (dominantly vertical movement) of the Pan-African metacratonic mega-shear zones and related superficial faults in response to the stress induced by the Africa–Europe collision (Liégeois et al., 2005).

7. Magmatic metacratonization of a cratonic continental interior: the Murzukian evolution of Djanet/Edembo terranes (Eastern Hoggar, Algeria)

Eastern Hoggar is bounded to the west by the Raghane mega-shear zone (Liégeois et al., 1994), which is also the western boundary of the Saharan metacraton (Abdelsalam et al., 2002). Eastern Hoggar belongs to the Saharan metacraton (Fig. 10). It is composed, from west to east, of the Aouzegueur, Edembo and Djanet terranes (Fig. 10). The Aouzegueur terrane represents the westernmost part of the Saharan metacraton together with other oceanic terranes accreted along the Saharan metacraton western margin between 900 Ma to 640 Ma (Henry et al., 2009). Within Eastern Hoggar, there exists a well-preserved 1.9 Ga basement which rarely crops out to the surface (Nouar et al., 2011).

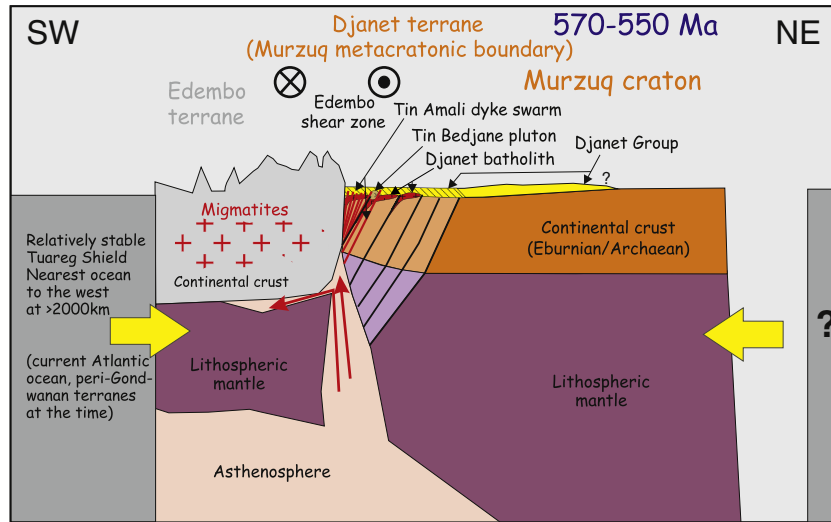


Fig. 13. Schematic model section at 570–550 Ma of the Murzuq craton, the metacratonic low grade Djanet terrane and high-grade Edembo terrane. Metacratonization resulted from the intracontinental convergence of the Murzuq craton, relaying a northern unknown continent push and the relatively stable Tuareg shield, leant against the West African craton. No ocean at the time in a circle of 2000 km. Adapted from Fezaa et al. (2010).

Aouzegueur terrane comprises also the Tiririne sedimentary Group (Bertrand et al., 1978) representing the molasse sedimentary rocks of Central Hoggar. This group is known as the Proche-Ténére Group in Aïr (Liégeois et al., 1994) and as the Djanet Group in the Djanet and Edembo terranes (Fezaa et al., 2010). It was deposited between 600 and 575 Ma (Henry et al., 2009; Fezaa et al., 2010). A point of paramount importance is that the Djanet Group, equivalent to the Tiririne Group, constitutes the oldest rocks in the Djanet and Edembo terranes (Fezaa et al., 2010; Liégeois et al., in prep.), demonstrating that Eastern Hoggar was a stable cratonic lowland at c. 600 Ma. On the other hand, these two terranes were affected during 575–545 Ma by metamorphic events of different metamorphic grades (greenschist facies in Djanet terrane, amphibolite facies in Edembo terrane) and magmatic events of different geochemical characteristics (potassic granitoids in Djanet, peraluminous granites in Edembo) both largely of crustal origin (Fezaa et al., 2010). The Djanet Group was deposited after 590 Ma as indicated by detrital zircon age constraint. This group was subsequently intruded successively by the Djanet Batholith at 571 ± 16 Ma, by high-level subcircular plutons such as the Tin Bedjane Pluton at 568 ± 5 Ma, and by the felsic Tin Amali Dyke Swarm at 558 ± 5 Ma as indicated by SHRIMP U–Pb zircon age determinations (Fezaa et al., 2010). During this period, the Tuareg shield to the west of the Raghane shear zone was largely stable, and only witnessed the intrusion of few and isolated high-level alkali-calcic circular plutons (Azzouni-Sekkal et al., 2003; Abdallah et al., 2007; Fig. 11). This led Fezaa et al. (2010) to propose a separate event termed the Murzukian orogenic episode that affected the entire Eastern Hoggar. This orogenic event did not form in relation to the stress induced by the West African craton in the west, but rather on the western margin of a newly-discovered rigid entity to the east referred to by Fezaa et al. (2010) as the Murzuq craton (Fig. 10). During the same period, the adjacent Edembo terrane was affected by a 568 ± 4 Ma old migmatization as documented by U–Pb zircon SHRIMP age determinations (Fezaa et al., 2010) as well as high-temperature metamorphism that lasted until 550 Ma (Liégeois et al., in prep.), corresponding to the age of the end of the intrusion of the Tin Amali dyke swarm in the Djanet terrane.

It must be stressed here that not a single oceanic-related lithology is known from the Edembo and Djanet terranes and that these terranes were stable low lands before 575 Ma. The Murzukian orogenic event corresponds thus to the destabilization of the interior of

the Saharan metacraton in response to compression stress originating from convergence at plate boundaries beyond the limits of the Murzuq craton. The late Ediacaran Murzukian event is thus an intra-continental and even an intracratonic event. It has been proposed that this event was due to vertical planar lithospheric delamination during transpressive movements along pre-existing weakness zones inherited from the Paleoproterozoic evolution of these terranes as a result of the indentation of the Murzuq craton (Fezaa et al., 2010; Fig. 13). This delamination allowed the uprise of the asthenosphere, generating a high heat flow at the origin of the high-temperature amphibolite facies metamorphism characterizing the Edembo terrane (Fig. 13). The Murzuq craton, protected by its thick lithosphere was not affected except on its margin, fractured and invaded by magmas and affected by a greenschist metamorphism (Djanet terrane; Fig. 13). A similar event probably occurred for the same reasons on the other side of the Murzuq craton, in the Tibesti region of Chad and Libya (Fig. 10) where contemporaneous similar events occurred (Suayah et al., 2006). The Murzuq craton was only passively transmitting compression stress that occurred in the NE, within the current location of the Sirt basin in Libya. This situation could be compared to the current Altai intracontinental intraplate transpressional orogen sandwiched between the more rigid Junggar Basin block and Hangay Precambrian craton, whose convergence is a consequence of the India–Asia collision occurring more to the south (Cunningham, 2005). It is observed that abundant Cenozoic volcanism occurred around the Murzuq craton margins but not within the interior of the craton itself (Fig. 10).

The Djanet–Edembo situation displays a number of the metacratonic feature characteristics including mCf-6, the metacratonization of a craton interior. It also displays mCf-4 (potassic granitoids) and mCf-5 (high-temperature metamorphism induced by the intrusions). The Edembo terrane can be considered as an intracontinental hot belt (Liégeois et al., in prep.).

8. The complex case of the Saharan metacraton (Africa)

The vast region (~5,000,000 km²) located between the Tuareg Shield and the Arabian-Nubian Shield (Fig. 14) is probably the largest geologically poorly known tract of a continental crust in the World, as a consequence of a remote location. Due to the presence of metamorphic and magmatic rocks and the early dating of Archean rocks in the

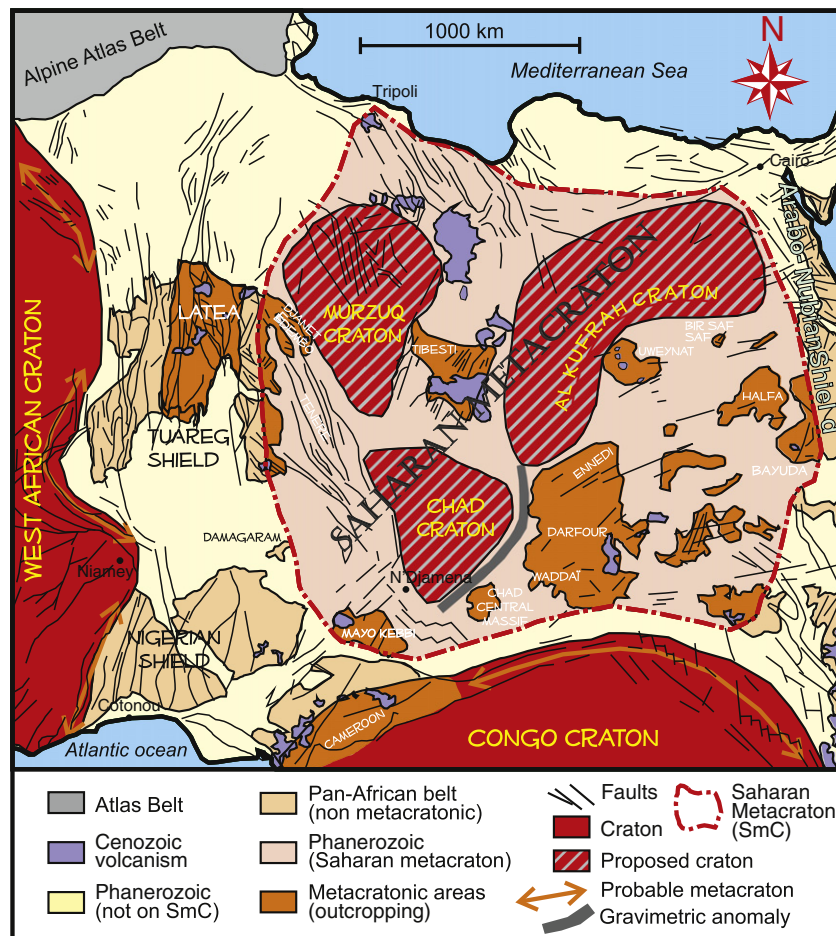


Fig. 14. Main rheological domain of Saharan Africa centered on the Saharan metacraton. Contours and structural features are based on the Tectonic map of Africa (Milesi et al., 2010). Subdivisions within the Saharan metacraton are from Fezaa et al. (2010) for the Murzuq craton, Djanet–Edembo terranes and Tibesti, Küster et al. (2008) for Bayuda and from this paper for the remaining parts, especially the proposed Al Kufrah and Chad cratons. Gravimetric anomaly is the major Haraz anomaly (Cratchley et al., 1984).

Uweynat area (Klerkx and Deutsch, 1977), this continental tract has first been considered as a craton and was named the Nile Craton (Rocci, 1965), the northern part of the large Sahara–Congo Craton (Kröner, 1977), and the Eastern Saharan Craton (Bertrand and Caby, 1978). However, increasing evidence for an abundant presence of Pan-African granitoids renders the craton interpretation untenable. This led Black and Liégeois (1993) to propose that if this large region was indeed a craton during the Mesoproterozoic, it was partly destabilized during the Neoproterozoic. They propose to name it the “Central Saharan Ghost Craton”. Finally Abdelsalam et al. (2002) compiled all the geochronological and isotopic data on the region and demonstrated that if Pan-African events are largely present, protoliths are largely Paleoproterozoic or Archean in age and that the whole area share numerous pre-Neoproterozoic characteristics and behaved as a single block during the Phanerozoic. They named it “Saharan metacraton”, defining a metacraton as a craton that has been remobilized during an orogenic event but is still recognizable dominantly through its rheological, geochronological and isotopic characteristics. This name is now widely used and adopted in the literature.

Although heterogeneous at the surface (Fig. 14), its homogeneity at the lithospheric scale has been recently demonstrated by geophysical data, the Saharan metacraton presenting a peculiar lithospheric structure, nearly unique in the world (Abdelsalam et al., 2011). Looking at its continental size (Fig. 14), it is likely that several events and several processes have been operating to metacratonize the preexisting Saharan craton. To the west, as described above, Fezaa et al. (2010) have established the existence of the Murzuq craton, included

within the Saharan metacraton. It is marked at the surface by the circular Murzuq Phanerozoic basin, flanked to the SW by the metacratonic Djanet and Edembo terrane and also probably to the SE by similar metacratonic terranes in Tibesti (Fig. 14; Suayah et al., 2006). A series of uplifted areas within the Saharan metacraton including Mayo Kebbi, Chad Massif Central, Waddai, Darfour to Bayuda (Fig. 14) display similar characteristics: (1) the presence of metamorphic Paleoproterozoic and Archean protoliths (Adamawa–Yade block in Mayo Kebbi; Penaye et al., 2006; Pouclet et al., 2006), (2) the presence of overthrust island arc lithologies towards the Saharan metacraton (Pouclet et al., 2006) and (3) the presence of Neoproterozoic high-K calc-alkaline batholiths mainly of crustal origin. These regions correspond to metacratonic areas as exemplified by Bayuda (Fig. 14; Küster et al., 2008). The metacratonic character is shown in Bayuda where c. 900 Ma continental granites and gneisses are well preserved despite the main c. 600 Ma Pan-African event that affected the area (Küster et al., 2008). Similarly, in the Halfa region (Fig. 14), just north of Bayuda, c. 720 Ma anorogenic alkaline ring-complexes have been preserved from the main Pan-African orogenic event (Shang et al., 2010b). Differently, in the vicinity of the two above regions, high-K calc-alkaline granites and associated high-temperature migmatitic metamorphism have been generated at c. 600 Ma, all rocks having old Nd T_{DM} model ages (Küster et al., 2008; Shang et al., 2010c). Contrastingly, just to the east, in the Arabian–Nubian Shield, all Nd T_{DM} model ages are Neoproterozoic, even those obtained from high-grade metamorphic rocks (Liégeois and Stern, 2010), reflecting the existence there of a large ocean whose closure at c.

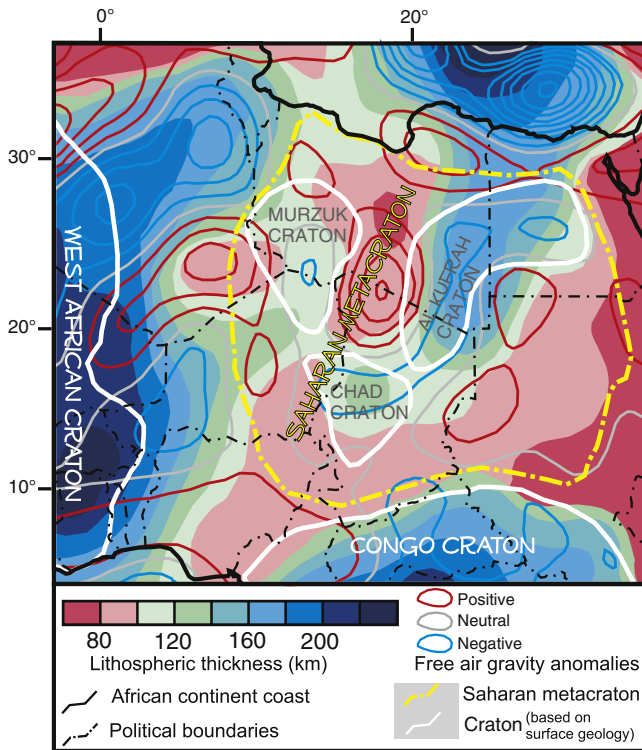


Fig. 15. Comparison between the outlines of the Saharan Metacraton, Murzuq craton, proposed Al Kufrah and Chad cratons based on surface geology and the lithospheric thickness based on tomography. The latter is extracted from the African map estimating the lithospheric thickness based on tomographic data and long-wavelength free-air gravity anomalies of Fishwick and Bastow (2011).

600 Ma led to the formation of Gondwana supercontinent, including the accretion of the oceanic Arabian-Nubian Shield towards the Saharan metacraton (Johnson et al., 2011 and references therein).

The peculiar structure of the Saharan metacraton can be tentatively attributed to the fact that the preexisting Saharan craton was subjected, at the end of the Neoproterozoic, to important collisional events along all its margins against the Tuareg Shield (with the West African craton behind) in the west, against the Congo craton and intervening Pan-African belts to the south, against the Arabian-Nubian Shield to the east, and against an unknown continent to the north. As no tectonic escape was thus possible, the Saharan craton has been metacratonized not only on its margins but also within its interior, as described for the Djanet–Edembo terranes in Eastern Hoggar (at the origin of the Murzukian episode; Fezaa et al., 2010). Such an interpretation can be applied to Tibesti, Darfur and Bir Saf Saf (Fig. 14) where Pan-African granitoids generated from old continental source are known coupled with the absence juvenile oceanic lithologies (Kusnir and Moutaye, 1997; Suayah et al., 2006; Bea et al., 2011a). No Pan-African granitoids are known in the Uweynat inlier (Bea et al., 2011b) but the presence of Cenozoic volcanism (Franz et al., 1983) and especially of silica-oversaturated alkaline ring-complexes (André et al., 1992) suggests that this inlier is not anymore a craton but a metacraton.

The Murzuq craton, an intact remnant of the former Saharan craton, is now covered by the Phanerozoic Murzuq basin. This feature suggests that the Al Kufrah Basin (similar to Murzuq basin; Klitzsch, 2000; Le Heron and Howard, 2012) and, to the SW, the Chad Basin, could be taken as evidence for the potential presence of two more cratonic blocks within the Saharan metacraton that escaped the metacratonization process (Fig. 14). Juxtaposition of metacratonic and intact cratonic blocks at all scales is characteristic of metacratons that evolved from the metacratonization of the interior of a craton

(mCf-6). Additionally, Al Kufrah and Chad basins evolved as intra-continental sag basins (Heine et al., 2008); a basin-forming mechanism that does not require accommodation of tectonic subsidence through faulting and rifting. Rather, subsidence in intra-cratonic sag basins can be derived by negative buoyancy caused by thick cratonic lithosphere (Ritzmann and Faleide, 2009). Cratons are not limited to the basin they can bear and adjacent flat, not reworked areas can be considered also as cratonic areas. Using these surface geology criteria, outline of Al Kufrah and Chad cratons can be proposed (Fig. 14). It is remarkable that thicker lithosphere is indeed present below these proposed cratonic remnants as shown by a recent review of passive-source seismic studies of the African crust and upper mantle (Fig. 15; Fishwick and Bastow, 2011). This is similar to the Congo basin which extends within much of the surface expression of the Congo craton and is underlain by a lithosphere as thick as 250 km (Craig et al., 2011). The proposed Chad craton is bordered to the SW by the Ténéké rifts and to the SE by the Haraz major gravimetric anomaly (Cratchley et al., 1984), still enigmatic (Braitenberg et al., 2011). This anomaly is not recorded to the north (Figs. 14; 15). The proposed Al Kufrah craton is bordered by uplifted metacratonic zones to the W (Tibesti), to the E (Uweynat) and to the SE (Ennedi) (Fig. 14). All these regions are marked by thinner lithosphere (Fig. 15).

We propose that the Murzuq, Al Kufrah and Chad cratons are remnants of the pre-Neoproterozoic Saharan craton (Fig. 14). The current stress applied to the African plate, due to Europe–Africa convergence, mid-ocean ridge push (Mahatsente et al., 2012) and mantle activities may induce Rayleigh–Taylor instabilities due to rheological contrasts existing within blocks inside cratons (Gorczyk et al., 2012) or even more likely inside a structure like the Saharan Metacraton, resulting in uplifting of metacratonic areas sometimes accompanied by intra-plate volcanism (Liégeois et al., 2005).

These cratons relics are bordered by metacratonic outcropping regions, Mesozoic rifts (superposed on depressed metacratonic areas) and by Cenozoic magmatism (Fig. 14). The latter emplaced preferentially within metacratonic regions surrounding cratonic blocks (Liégeois et al., 2005). It is important to point to that the cratonic blocks were not affected by Cenozoic volcanism and do not show penetrative structural features, in agreement with a cratonic structure (Fig. 14). Currently, the stress applied to the African plate, due to Europe–Africa convergence, mid-ocean ridges and mantle activities, induces uplift of metacratonic areas in the vicinity of non uplifted cratons, due to their rigid but fractured structure, as it is also the case for the LATEA metacraton in the Tuareg Shield (Liégeois et al., 2005). The metacraton model predicts that other smaller cratonic areas exist within the Saharan metacraton but more geological data are needed for verifying this point. Distinguishing the brittle and ductile metacratonic terranes (i.e. Djanet and Edembo terrane types; Fezaa et al., 2010) is needed for proper understanding of the structure of the Saharan metacraton. Here also more field work is needed. The Saharan metacraton is thus an exceptional geological feature, probably unique in the world, by the size and the complexity of the metacratonic processes that were at work at the end of the Neoproterozoic.

9. Metacratons elsewhere?

Metacratonization occurs when a craton is involved in a continental collision, and possibly when the margins of the cratons are accreted by island arc terranes. These tectonic events are common processes that have been operated on Earth for a good part of its geological history. Hence, it is quite possible that additional examples of metacratons will be found elsewhere on Earth.

The NW Congo craton in Cameroon (Fig. 14) was involved in a Pan-African continental collision and its northwestern margin has been metacratonized, which has been observed in the Archean

basement (Shang et al., 2010a) and suggested through the presence of Pan-African magmatism (Kwekam et al., 2010). It is possible that the entire northern boundary of the Congo craton (Fig. 14) has been metacratonized but more geological, geochemical, isotopic and geophysical data are needed to test this notion.

The eastern margin of West African craton has been overthrust by the Amalaoulou (Berger et al., 2009a, 2011) and Tilemsi (Caby et al., 1989) island arcs at c. 700 Ma and collided at c. 630 Ma (Jahn et al., 2001) with the Tuareg shield; an event that triggered basement uplift and HP–LT metamorphism in the passive margin sedimentary rocks of the craton (Caby et al., 2008). The successive tectonic events that have shaped the eastern margin of the West African craton can be integrated within a metacratonic evolution. Later, during the Atlantic opening, intraplate stress induced the uplift of this metacratonic margin, allowing the emplacement of Permian–Jurassic nepheline syenites and carbonatites in the Tadhak region (Liégeois et al., 1991). The northeastern boundary of the West African craton might also be considered as a metacraton as it has been shown above.

Other areas in the world have been already proposed to be metacratons. In Brazil, based on detailed U–Pb SHRIMP age studies, da Silva et al. (2005) defined the extent of the Brasiliano (Pan-African) overprinting over large basement polycyclic units, where U–Pb age resetting and tectonic transposition are ubiquitously recorded into the Archean and Paleoproterozoic orthogneisses. This led these authors to conclude that the southeastern margin of the São Francisco Craton underwent metacratonization. Several other areas, corresponding to partly reactivated old basement, sometimes repeatedly, could be easily interpreted as metacratons, e.g. the c. 1.2 Ga shear-related O'okiepian and Klondikean episodes affecting the 1.9 Ga basement in Namaqualand, South Africa (Duchesne et al., 2007).

Metacratonization could also occur on craton margins in intracontinental setting, for example at the hinder regions of major active margin such as the Andes. To the east of the Altiplano, along the lithospheric discontinuity separating Andes from the Amazonia craton, recent lavas are observed aligned along this margin showing Sr–Nd signature indicating that these lavas were derived from the craton, demonstrating the remobilization of the Amazonia craton by the Andean convergence (Carlier et al., 2005).

In China, the North China craton has received much attention in the recent years. The craton was formed during the Archean–Paleoproterozoic (Kusky, 2011; Zhai and Santosh, 2011, and references therein) but has been reactivated during the Mesozoic (Zhang et al., 2011a). This reactivation led to widespread lithosphere remobilization and extension in the eastern margin of the North China craton (e.g. thinned crust and lithosphere in the Bohai Bay Basin; Chen, 2009), and also possibly in the central part of the craton (Chen, 2009). The Early Permian Guyang high-K calc-alkaline batholith intruded in the northern margin of the North China Craton with a spatial–temporal association of mafic and felsic magma suites (Zhang et al., 2011a,b). The emplacement of this batholith has been interpreted to have occurred during the metacratonic evolution of the northern margin of the North China craton, possibly in response to linear lithospheric delamination and hot asthenospheric upwelling along crustal-scale shear zones within a post-collisional transensional regime of a passive continental margin (Zhang et al., 2011b). The intrusion of the c. 130 Ma Longbaoshan alkaline complex in the southeastern part of the North China Craton is interpreted as the result of complex interactions between mantle and crust interpreted as generating the destruction of the craton (Lan et al., 2011) implying its refertilization (He et al., 2011). This Cretaceous “destruction” of the North China craton lithosphere (Tian and Zhao, 2011) is localized, leaving large portions of the craton intact. We thus believe that using “metacraton”, “metacratonic” or “metacratonization” is more suited to describe the transformative processes that have shaped the North China Craton through

geological time. A metacraton is not a completely destroyed craton; it still possesses a series of cratonic features.

10. Conclusion: the concept of metacraton, genesis and behavior

Africa represents 20% of the emerged continents and has been mostly shaped during the Pan-African orogeny at the end of the Neoproterozoic. Since then, Africa has been subjected to various stresses but never of high intensity which allowed for its preservation as a wonderful laboratory for studying intracontinental processes, especially when compared with other outstanding regions such as the Norwegian Caledonides. Initially, the paradigm shift that accompanied the introduction of plate tectonics has mistakenly precipitated overlooking intraplate processes by only considering the lithospheric plate as rigid and not affected by intraplate deformation. This has also caused the lack of sufficient consideration of the events occurring in the subducting lower plate when the continental lithosphere is subducted, especially the thicker cratonic lithosphere. It is largely considered (because of the nature of kinematic models proposed for plates motion within the plate tectonics context) that the cratonic passive margin which collides with an active margin will suffer little or no deformation and metamorphism, all transforming processes being concentrated in the active margin. Geophysical studies carried out on modern orogenic belts have refuted the assertion that, during collision, the cratonic passive margin will not be modified. However, it is not easy to study metacratonic processes in the present-day setting because continental subductions occur at great depth. In the case of Precambrian terranes, geological processes affecting the lower plate during collision were generally interpreted within the framework of a colliding active margin, leading to a variety of contradictory models.

Cratons are rigid and cold due to a thick continental lithospheric mantle (Black and Liégeois, 1993 and references therein). By definition, they cannot be active margins characterized by high heat flow associated with subduction-related magmatism. Nevertheless, cratons can be involved in collisional processes as former passive margins, i.e. in the lower subducting plate configuration. Subducting a rigid cratonic margin will result in a set of distinctive modifications, only some of them are grossly similar to those occurring during collisional processes affecting a former active margin. In addition, cratonic interiors can also be reactivated. This led us to use the expression “metacratonization” to describe these modifications and infer their genesis and to re-introduce the term “metacraton” to identify additional tracts of continental crust resulting from metacratonization processes beyond the initial limited use of the term metacraton which was introduced by Abdelsalam et al. (2002) to restrictedly describe the Saharan metacraton.

Review of geological processes acting on craton's margins and interior allowed us to establish a set of criteria for metacratonization and the emergence of metacratons. These are referred to as metacraton features (mCfs). It is not necessarily that all mCfs are to be found in each metacraton because the dominant metacratonization processes are different from the craton's margin to its interior. In fact, the variation of metacratonic processes is an intrinsic part of the metacratonic characteristics. However, all these processes have common characteristics due to the initial cratonic rigidity and coldness, and the old age of the craton relative to the destabilizing collisional events. Metacratonic features can be summarized as:

(1) mCf-1: metacratons will be characterized by the absence of pre-collisional orogenic events. (2) mCf-2: metacratonic syn-collisional HP–LT metamorphism is linked to subduction and not to lithospheric or crustal thickening and will be only developed in regions of high strain while low-strain areas are preserved; (3) mCf-3: allochthonous pre-collisional oceanic, metamorphic and magmatic units are preserved from collisional effects in metacratons, due to the remaining cratonic rigidity. (4) mCf-4: in metacratons, post-collisional magmatism

associated with transcurrent movements (but not to lithospheric thickening) could be abundant and fracturation of the rigid cratons is favorable to magma ascent. Magmas are much younger than the cratonic country-rocks and the source of these magmas is either the old lithosphere or the asthenosphere, depending on the modalities of asthenospheric uprise, itself depending on the intensity of the metacratonization. (5) mCf-5: in the case of a large magmatic input in the metacratonic region, a high-temperature-low pressure metamorphism could develop at the vicinity of the intrusions and can become regional in extent, although leaving some zones unaffected. (6) mCf-6: intracontinental metacratonization, occurring when the craton is subjected very high stresses resulting in reactivation of pre-existing zones of weakness, generating intracontinental hot belts (HT–LP metamorphism) unrelated to local subduction and oceanic basin closures (absence of oceanic lithologies). Later intracontinental reactivations of metacratons will generate uplift, doming sometimes accompanied by asthenospheric volcanism, especially close to cratonic margins and will favor mineralizations through repeated mineral enrichment because fractured rigid metacraton facilitates fluid movements. Intracontinental metacratonic uplifts in response to stress applied to plate margins will favor the development of cratonic basins.

Considering that a collision occurs most often between an active margin and a passive margin, and that the latter will regularly include cratons, it can be concluded that the presence of metacratons on Earth is more frequent than has been currently identified.

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