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Cratons, mobile belts, alkaline rocks and continental lithospheric mantle: the Pan-African testimony

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Abstract: Several late-collision and intraplate features are not entirely integrated in the classical plate tectonic model. The Pan-African orogeny (730–550 Ma) in Saharan Africa provides some insight into the contrasting behaviour of cratons and mobile belts. Simple geophysical considerations and geological observations indicate that rigidity and persistence of cratons are linked to the presence of a thick mechanical boundary layer, the upper brittle part of the continental lithospheric mantle, well attached to an ancient weakly radioactive crust. The surrounding Pan-African mobile belts, characterized by a much thinner mechanical boundary layer and more radioactive crust, were the locus of A-type granitoids, volcanism, tectonic reactivation and basin development during the Phanerozoic. During oceanic closures leading to the assembly of Gondwana, lithosphere behaviour was controlled by its mechanical boundary layer, the crust being much less rigid. We suggest that the 5000 km wide Pan-African domain of Saharan Africa, a collage of juvenile and old reactivated basement terranes, has suffered regional continental lithospheric mantle delamination during the early stages of this orogeny, as has been postulated for the more recent Himalayan orogeny in Tibet. Delamination of the continental lithospheric mantle and juxtaposition of crust against hot asthenosphere can explain many features of the late Pan-African (around 600 Ma): reactivation of old terrains, abundant late-tectonic high-K calc-alkaline granitoids, high temperature-low pressure metamorphism, important displacements along mega-shear zones and mantle-derived post-tectonic granitoids linked to a rapid change in mantle source. Recycling into the asthenosphere of large amounts of continental lithospheric mantle delaminated during the Pan-African, can provide one of the reservoirs needed to explain the isotopic compositions of ocean island basalts. Lastly, the lithospheric control over the location of the alkaline rocks enjoins us to consider the thermal boundary layer (the lower ductile part of the continental lithospheric mantle) as a major mixing source zone for these rocks.

Plate tectonic theory does not explain directly some intracontinental features, such as rapid uplift, widespread post-collisional crustal deformation or igneous activity far from plate boundaries (Bird 1979). The contrasting behaviour of cratons and mobile belts during and after collisions, repeated intraplate magmatism at the same place, basin development and reactivations of shear zones, also require explanation.

Based on simple geophysical notions already developed in the literature, parallels drawn with Tibet and our experience in Pan-African geology and alkaline rocks, we propose a composite but coherent model which, in the plate tectonic framework, clarifies interpretations in structural geology, mantle geochemistry and global geodynamics.

The Pan-African, comparable in scale to the currently active Alpine–Himalayan–Circum Pacific orogenic belt, is unique in that evidence of the end of a Wilson cycle with creation of a supercontinent, in this case Gondwana, is well preserved even if still largely undeciphered (Black 1978; Hoffman 1991).

Attention will be focused on the Pan-African (750–550 Ma) Trans-Saharan belt (Fig. 1) limited to the west by the West African craton stable since 2 Ga. This belt, extending from Mali to Saudi Arabia, comprises numerous accreted island arcs and related tonalite-trondhjemite-granodiorite associations, principally at both extremities. Elsewhere it is characterized by reactivated Archaean and lower Proterozoic terranes, thrusted ophiolitic sutures, abundant high-K calc-alkaline granitoids, major shear zones parallel to craton margins, and widespread development of Phanerozoic anorogenic alkaline magmatism.

There is apparently only one solution to explain all these features: to give a major role to the continental lithospheric mantle.

Continental lithospheric mantle and cratons

The lithospheric mantle is the lower part of the plates which move on the convecting asthenospheric mantle. In the oceans, its lower limit is classically given by the *c.* 1300 °C isotherm (Kay *et al.* 1970), the peridotite dry solidus. However, the base of the mechanical boundary layer, the layer able to support loads of long duration (Karner *et al.* 1983), is much shallower and corresponds to the brittle-ductile transition in the mantle (Parsons & McKenzie 1978). The thickness of this mechanical boundary layer probably approximates to the elastic thickness, the value of which can be determined from the isostatic response of the lithosphere to loading. It is roughly equivalent to the depth of the 450 $\pm$ 150 °C isotherm (Watts *et al.* 1980). The 600 °C isotherm is adopted here as corresponding to both the mechanical boundary layer and the elastic thickness (Bergman 1986; Anderson 1991). This layer is probably completely isolated from the convecting mantle and so can then acquire specific isotopic or chemical compositions (McKenzie & Bickle 1988). The thermal boundary layer constitutes a transition between the mechanical boundary layer and the asthenosphere (Fig. 2).

If the conservative view that gives a small elastic strength to the continental crust is true (Grotzinger & Royden 1990), the

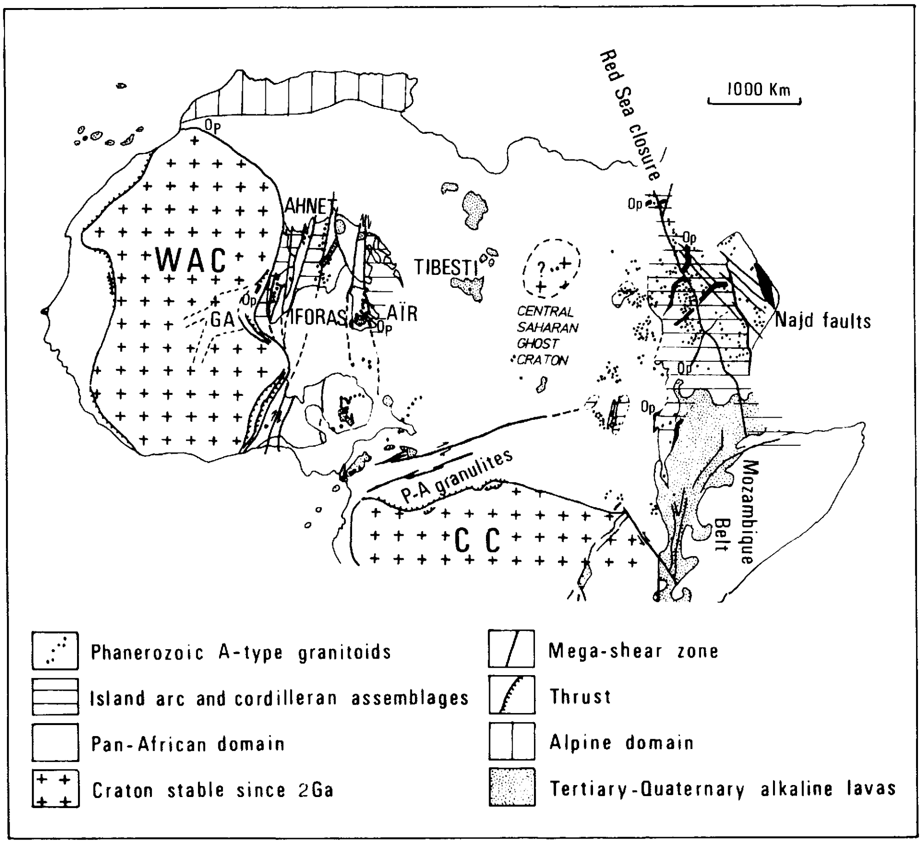

Fig. 1. Schematic map of Northern Africa. WAC, West African craton; CC, Congo craton; GA, Gourma aulacogen; Op, Ophiolite.

elastic thickness corresponds to that of the mechanical boundary layer. It can vary widely, for example, in North America, from 4 km in the extensional Basin and Range area to 130 km in the Precambrian core of the continent (Canadian shield) (Bechtel *et al.* 1990). A craton can be defined as 'a part of the crust which has attained stability and which has not been deformed for a long time' (Bates & Jackson 1980). These characteristics can be linked to a thick and rigid mechanical boundary layer whose growth requires quiet conditions during several hundred Ma (Karner *et al.* 1983). Low heat flow in cratonic areas also induces an absence of a weak zone at the Moho (see Fig. 3 and caption). The heat flow data from Africa (Lucazeau *et al.* 1990; Ballard & Pollack 1987) showing relatively high values for the Pan-African belts (60–70 $mW\,m^{-2}$) compared to cratons (40 $mW\,m^{-2}$) can also be modelled in terms of lithospheric thickness.

When taking into account crustal heat production and the average heat flows of the West African craton (31 ± 10 $mW\,m^{-2}$) and of the Pan-African domain to the east (51 ± 8 $mW\,m^{-2}$), the geotherm defining the base of the lithosphere (1200–1300 °C) lies at 200–250 km beneath the craton and at 100–150 km beneath the Pan-African belts (Roussel & Lesquer 1991). These estimates correspond to those obtained by the study of long period records of surface waves (Hadiouche & Jobert 1991).

Continental lithospheric mantle delamination

Catastrophic continental lithospheric mantle delamination has been proposed by Bird (1979) for the Colorado plateau and by Houseman & McKenzie (1981) for the Tibetan plateau to explain the observed high heat flow, rapid uplift and extension in a compressive belt (see Fig. 4). In the Himalayas, Eurasia with its thin mechanical boundary layer is weaker than cratonic India, whose advance towards the north continues at the rate of 5 cm per year (Minster & Jordan 1978). Eurasia behaves as a wide mobile belt except for the Siberian craton (Angara) and to a lesser extent the small Tarim shield, both Precambrian areas with thick mechanical boundary layers (Molnar & Tapponnier 1981). The model of Fig. 4 implies an increase in rigidity of the former active margin during early collision (B) followed rapidly by strong weakening (C) (Glazner & Bartley 1985). Stage C is accompanied by uplift and both mantle and crustal fusion (McKenna & Walker 1990).

A long-term consequence of continental lithospheric mantle delamination is the existence around cratons of former orogenic areas with relatively thin mechanical boundary layers (Fig. 4D). If stress is applied to the plate, reactivation of pre-existing zones of weakness, localized in the former mobile belts or at their contacts with the craton may trigger off alkaline magmatism by pressure release, (Sykes 1978; Black *et al.* 1985; Liégeois *et al.* 1991). Areas with a thin mechanical boundary layer would be particularly susceptible to intraplate deformation (Pollack & Chapman 1977; Dewey 1988). However, the continental lithospheric mantle of former mobile belts will grow with time by cooling and basal accretion (Fig. 4D) *if quiet conditions prevail.*

In this paper emphasis is laid on collisional tectonics and we will not consider extensional tectonics, which in the Phanerozoic affects many Pan-African zones and may also lead to continental lithospheric delamination (e.g. Afar).

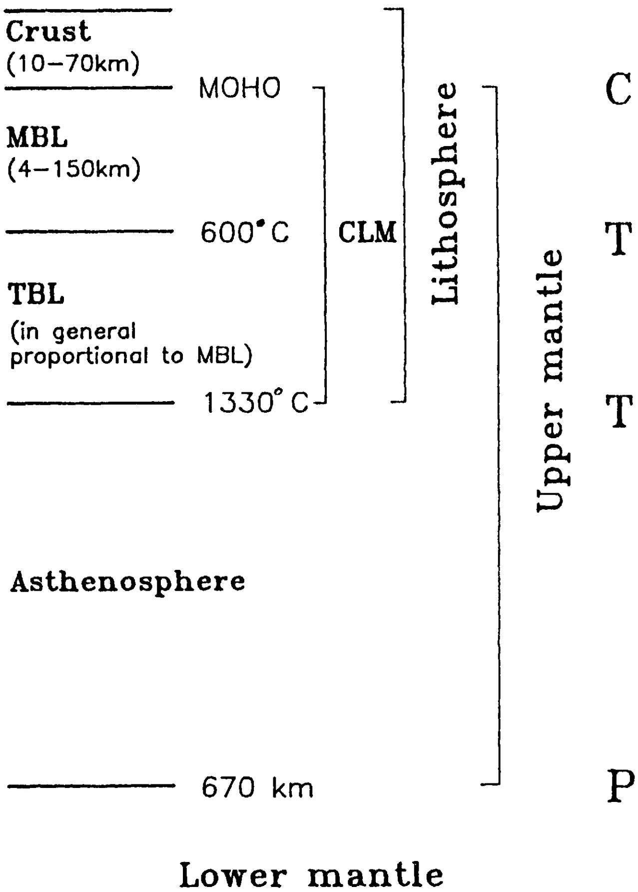

Fig. 2. Nomenclature of the lithosphere used in this paper. CLM, continental lithospheric mantle; MBL, mechanical boundary layer; TBL, thermal boundary layer. C, chemical boundary; T, thermal boundary; P, mineral phase or pressure boundary.

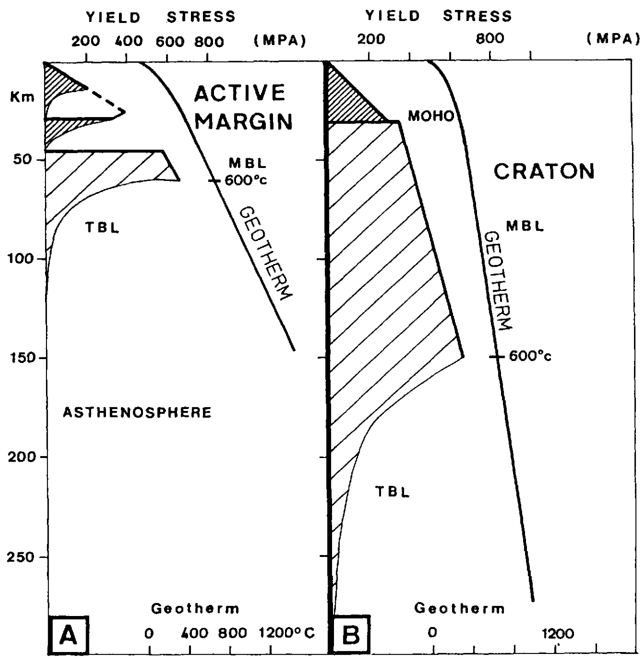

Fig. 3. Rigidity of the lithosphere in the case of an active margin and a craton. Closely spaced hatching, crust; widely spaced hatching, continental lithospheric mantle. Mean geothermal gradient for the entire lithosphere is around $10° km^{-1}$ for active margin and $4° km^{-1}$ for craton, values deduced from elastic thicknesses. Note that geothermal gradients are convex upwards implying higher values for the crust. Greater rigidity of the lithosphere lies in the continental lithospheric mantle, as its major constituent is olivine, in contrast to the crust where it is quartz. Based on the different strength distributions of these two minerals with temperature (Brace & Kolhlstedt 1980), the rigidity of an orogenic plate (crust: 45 km; elastic thickness: 20 km) can be represented as in (**A**) (Dunbar & Sawyer 1988): the greatest rigidity is attained in the mantle part of the lithosphere at *c.* 600 °C, which is equivalent to the elastic thickness and to the lower limit of mechanical boundary layer. Beyond, the curve follows ductile flow laws, zone corresponding to the thermal boundary layer. The crust is much less rigid, particularly if granites are abundant in the middle crust, and presents an almost null value at its base. This induces an important weakness at the Moho. The minimum strength at the base of both crust and continental lithospheric mantle, related to ductile behaviour, are generally marked by horizontal seismic reflectors (Lie *et al.* 1990). (**B**) In the case of an old craton (crust: 30 km, elastic thickness: 130 km), the situation is very different: (1) no minimum strength zone exists at the base of the crust due to dry conditions and the low cratonic geothermal gradient (the frequent absence of a geophysical visible Moho in old shields could be a consequence); (2) the total strength of the lithosphere is much higher, localized essentially in the mechanical boundary layer. These observations explain the much greater rigidity of the cratons. Intermediate figures exist as in quiet conditions continental lithospheric mantle grows continuously, either by simple cooling at the expense of the asthenosphere (Crough & Thompson 1976), or by underplating of much deeper material of various compositions (Jordan 1981; McDonough 1990). In both hypotheses (one does not exclude the other), the continental lithospheric mantle thickening (cratonization) process takes time. Construction of diagram (B) is based on a maximum strength for crust of 400 MPa at 400 °C and for continental lithospheric mantle of 700 MPa at 600 °C (Dunbar & Sawyer 1988).

Continental lithospheric mantle in the framework of the Pan-African orogeny

Kennedy (1964) defined the Pan-African as a thermo-tectonic event at *c.* 500 Ma affecting large areas encircling the West African, Congo and Kalahari cratons. He concluded, in view of the large tracts of reactivated old basement involved in the Pan-African event, that a process of destabilization was at work. This was in opposition to Stille's concept of progressive continental growth around cratons. After a period when purely ensialic models were favoured, the plate tectonic approach became generally accepted following the discovery of true ophiolites, juvenile island arc volcanic assemblages and continental scale collisions (Black 1980; Shackleton *et al.* 1980; Vail 1985; El-Gaby & Greiling 1988 and references therein). Whilst isotopic studies indicate important crustal accretion during the Upper Proterozoic between 950 and 550 Ma (Fleck *et al.* 1976, 1980; Stacey & Stoeser 1983), collisional and major metamorphic events marking Pan-African orogenesis across northern Africa appear to be restricted to the 750–550 Ma time interval (Schandelmeier *et al.* 1990). Now, we should re-examine the meaning we attribute to the concepts of craton and mobile belt.

Cratons within the Gondwana assemblage are generally bounded by thrusts with transport direction towards the craton

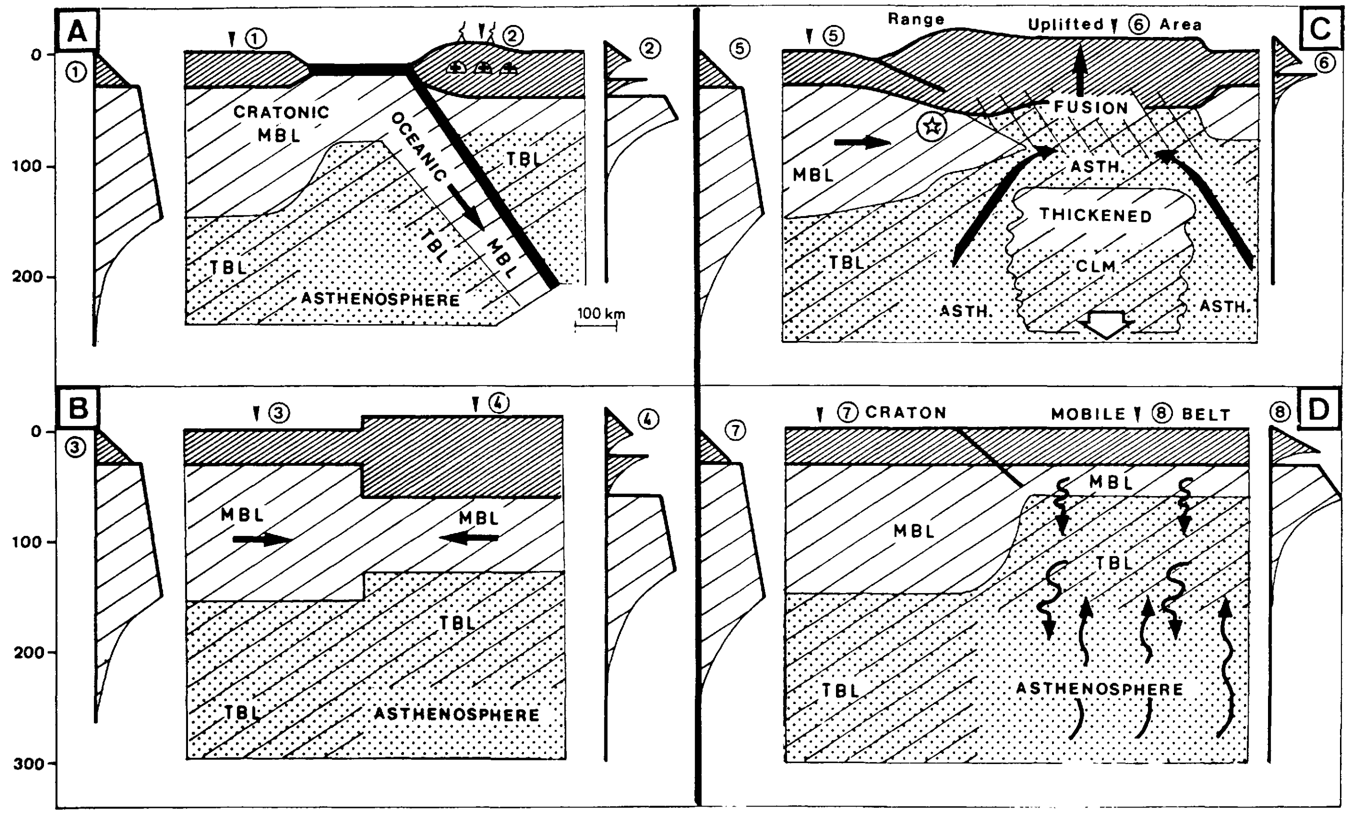

Fig. 4. Idealized sections for a frontal collision of Himalayan-type between a craton and an active margin. Relations between rigidity of the plates and depth are represented left and right of the sections respectively for the craton and the active margin at the vertical of the numbered arrows. (**A**) Classical subduction period. The middle crust of the active margin is weakened by wet intrusions. (**B**) Collision first induces thickening of both crust and mechanical boundary layer of the former active margin which increases rigidity below the Moho. (**C**) Thermal instability (Bird 1979) particularly of the continental lithospheric mantle plunged rapidly into hot asthenosphere, and the weak link with the crust (Dunbar & Sawyer 1988), in contrast with the strong cratonic mechanical boundary layer, induce delamination of this thickened non-cratonic lithospheric mantle, probably entirely (Houseman & McKenzie 1981; England & Houseman 1988). This is accompanied by hot asthenosphere upwelling (Molnar 1988), magmatic underplating (Dewey *et al.* 1988), fusion of lower or middle crust (depending on the availability of fluids), high temperature metamorphism and rapid uplift of the thickened crust due to isostasy. This is probably the case now beneath Tibet: cratonic India with a thick mechanical boundary layer (Lyon-Caen 1986) is little deformed (Molnar & Tapponnier 1981; Molnar 1988), the thickened crust of Tibetan plateau has a thin or no mechanical boundary layer if one accepts the estimated temperature of 750 °C at the Moho (Dewey *et al.* 1988) and has been rapidly uplifted. Similar effects are thought to have occurred in the Colorado plateau (Bird 1979). The asterisk shows the zone (under the high mountain range) where the cratonic lithosphere can lose its rigidity by temperature increase and fluid infiltration at the Moho which, if the intensity of the collision is strong enough, could induce delamination of the continental lithospheric mantle of the craton itself in the late stages of the orogeny. (**D**) After collision, this leads to the juxtaposition (oblique black line indicates suture) of a craton with cold crust and thick mechanical boundary layer and a mobile belt with hotter crust and thin mechanical boundary layer, the latter and the thermal boundary layer growing with time by cooling and by underplating of deeper material (symbolized by the wavy arrows). See Fig. 2 for definition of CLM, MBL and TBL.

and are themselves little affected by the Pan-African event. On the eastern margin of the West African craton, the thrust pile comprises components both of the passive margin of the craton and of the active margin of the eastern continents. Whereas frontal collision (Affaton 1990; Caby 1989) occurred between the Benin-Togo promontory of the West African craton and the Nigerian crustal block leaving little trace of the intervening ocean, further north in the Iforas embayment are preserved ophiolites, an ocean island arc and cordilleran assemblage (Black *et al.* 1980). The suture, as outlined by an alignment of positive gravity anomalies, lies 100 to 150 km to the east of the nappe front. The less well-known northern margin of the Congo craton displays Pan-African granulitic nappes directly resting upon the craton (Pin & Poitevin 1987).

One of the most striking features of the wide Pan-African belt east of the West African craton and north of the Congo craton (Fig. 1) is the spectacular development of mega-shear zones parallel to the craton margins, a characteristic of the Pan-African as a whole (Daly 1986). This pattern is strongly reminiscent of Asia north of the Himalayas (Black 1980) where intraplate lateral expulsion of crustal material occurred after oceanic closure (Molnar & Tapponnier 1976).

Besides juvenile material, the Pan-African includes large amounts of Archaean and lower Proterozoic crust often invaded by Pan-African calc-alkaline granitoids. In southeastern Aïr, 1000 km to the east of the West African craton (Figs 1 & 5; Black *et al.* 1990, 1991), an ophiolitic assemblage at the base of a thrust separates two terranes. The emplacement of late-kinematic high-K calc-alkaline concordant granitoids at 700 ± 20 Ma (Be and Ta in Fig. 5) occurred during a collision

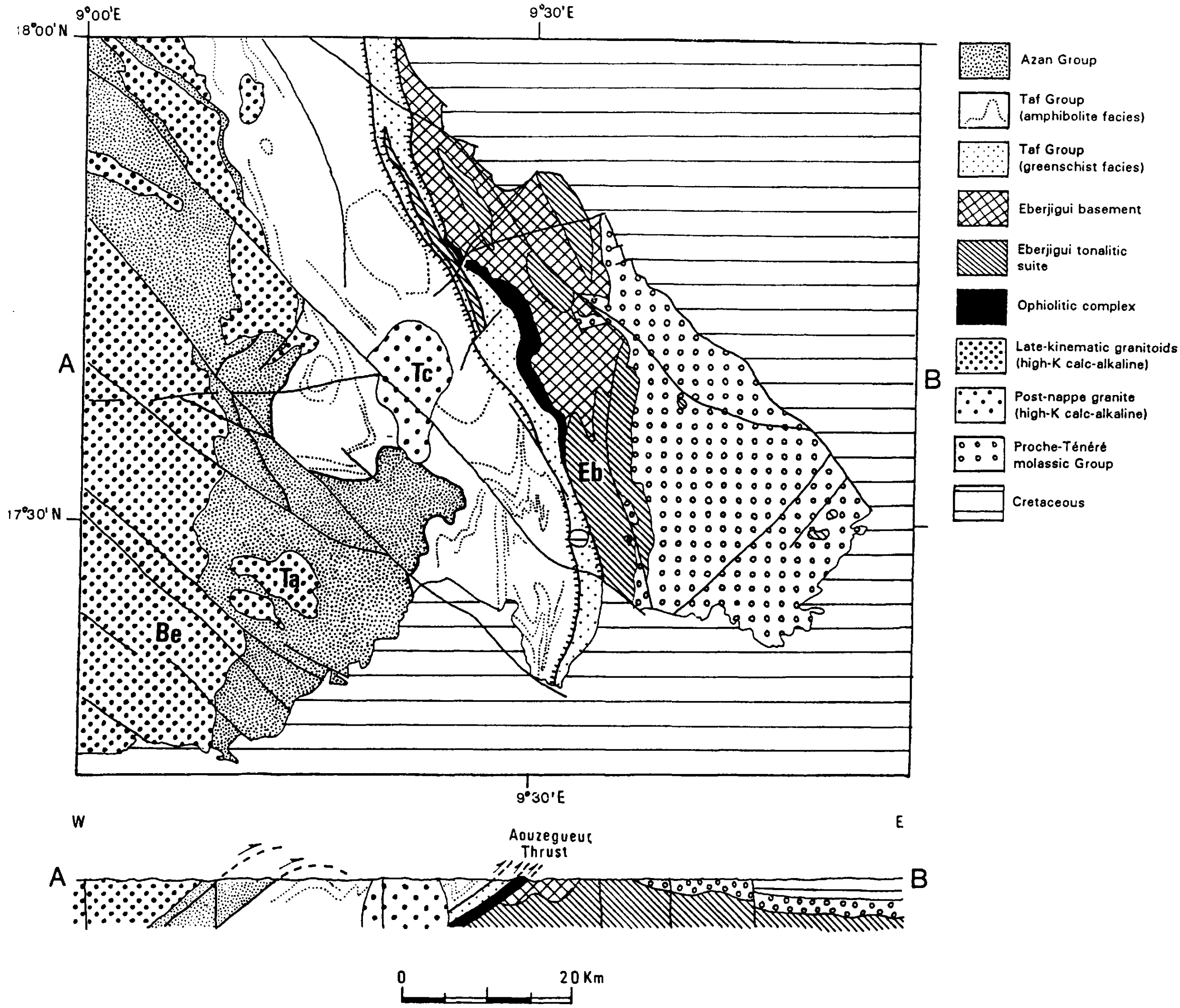

Fig. 5. Simplified geological map of SE Aïr and cross-section (Black *et al.* 1991). Be, Beurhot; Ta, Takarakoum; Tc, Tchebarlare; Eb, Eberjigui.

which had generated upper amphibolite facies metamorphism and ended with the thrusting of SE Aïr, around 680–670 Ma, upon an eastern terrane composed of 730 Ma tonalite–trondhjemite–granodiorite assemblages (Eb in Fig. 5) with amphibolitic country-rock resembling volcanic arc material. A high-level calc-alkaline pluton (Tc in Fig. 5) truncating the folds and nappes and dated at 664 ± 8 Ma, is contemporaneous with the onset of molassic deposition (Proche-Ténéré Group) and marks the end of the collision and of westerly subduction. In Central Aïr, extensive anatectic granites (Renatt type, about 6600 km^2) and associated high temperature-low pressure metamorphism dated at *c.* 670 Ma may be a consequence of regional continental lithospheric mantle delamination (Liégeois, Black *et al.*, work in progress). This major collisional event took place 100 Ma prior to the collision of the Tuareg shield with the West African craton. The absence of a craton to the east, where one might be expected, leads us to propose the destabilization of a Central Saharan craton in the early Pan-African. The Uweynat inlier in Libya (Klerkx & Deutsch 1977), as a surviving segment, and other old terrains as relicts, would represent this dismembered craton (Toteu *et al.* 1990). This is a concept equivalent to the 'palaeoplatform' of Lewry & Collerson (1990), used to describe a large area of reactivated but unsevered old basement. This ghost craton extending from Libya across western Sudan, Chad, Central African Republic down to the Congo craton has often been described as a true craton (Rocci 1965; Kröner 1979; Vail 1990), essentially on the basis of old Nd model ages and some mineral ages. In fact, major Pan-African events are recorded in these areas (Schandelmeier *et al.* 1990; Pin & Poitevin 1987).

We propose that this Central Saharan craton lost its thick mechanical boundary layer during the Pan-African period. Two processes can be envisaged for such a decratonization: (a) the craton acquires an active margin with subduction processes

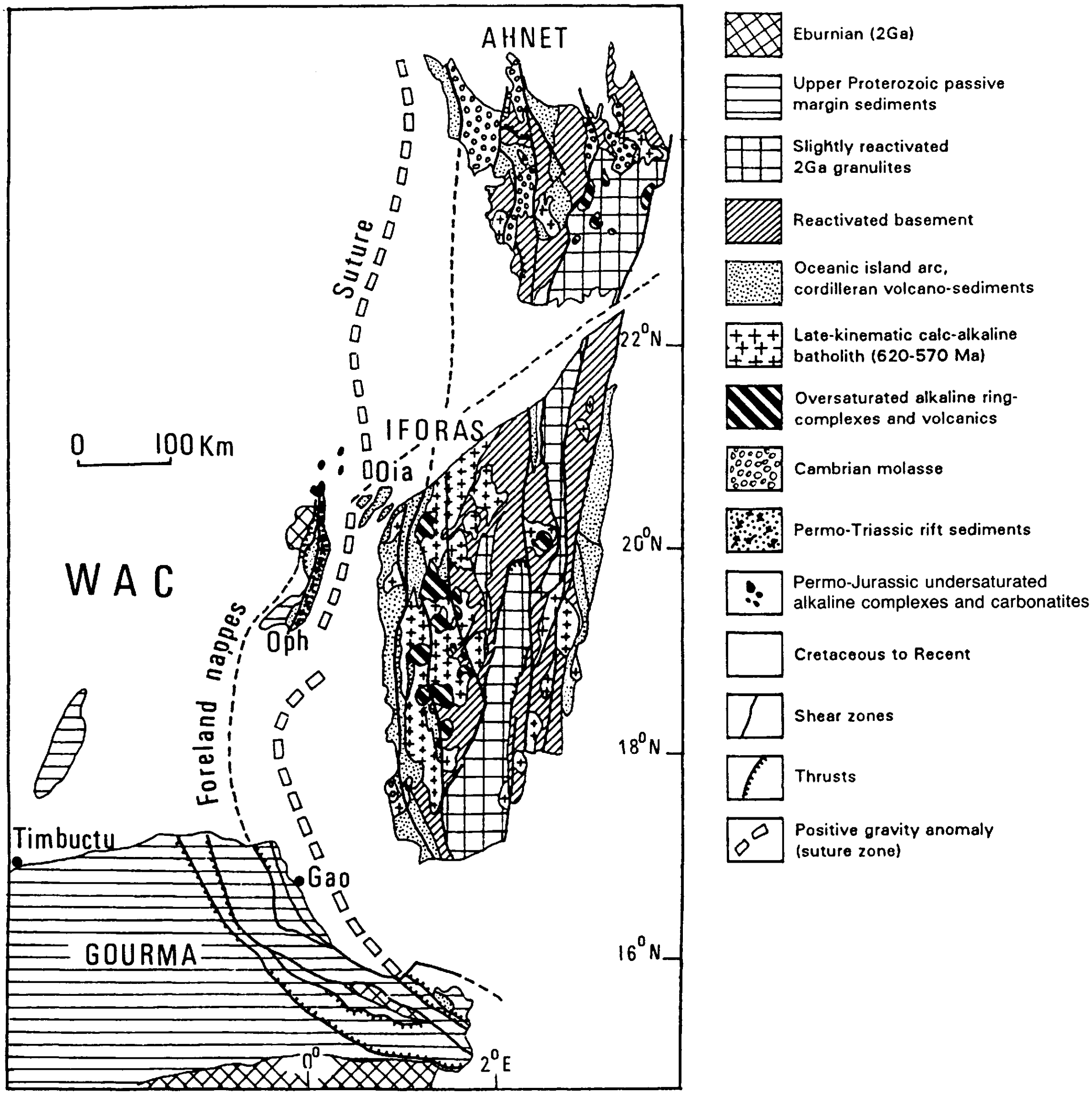

Fig. 6. Simplified geological map of the Ahnet–Iforas–Gourma region on the eastern margin of the West African craton (modified after Fabre *et al.* 1982). The reactivated basement comprises Pan-African-affected Middle and Upper Proterozoic supracrustal formations; Oia, oceanic island arc; Oph, ophiolites.

affecting and destabilizing the continental lithospheric mantle; (b) the craton undergoes an intense and frontal hypercollision. The strong linkage between the mechanical boundary layer and the crust in cratonic areas is due to low temperature and dry conditions at the Moho (Fig. 3B). If the whole lithosphere is overthrust by a huge range such as the Himalayas during a hypercollision, temperature conditions and abundance of fluids can change dramatically allowing a crust-mechanical boundary layer decoupling starting at the asterisk zone in Fig. 4C and propagating along the cratonic Moho, following a process similar to that proposed by Bird (1979). Molnar (1988) has proposed, on the basis of geophysical data, 'the possible detachment of all or much of the crust from India's lithosphere beneath the Himalayan range'. The decratonization of the 'Central Saharan craton' during the 700 Ma event left a very broad weak zone (Fig. 1). This explains why during final Gondwana assembly (630–550 Ma), the Pan-African orogeny was characterized throughout northern Africa by high-K calc-alkaline late-kinematic granitoids, high heat flow, high temperature–low pressure metamorphism and large movements along mega-shear zones which locally produced deformed intracontinental pull-apart basins.

Work in the Adrar des Iforas (Black *et al.* 1979; Caby *et al.* 1981), adjacent to the West African craton (Figs 1 & 6), has identified an oceanic island arc that was active at *c.* 720 Ma (Caby *et al.* 1989), a pre-tectonic tonalite-granodiorite suite and andesites (Chikhaoui *et al.* 1980) related to the cordillera, and a huge high-K calc-alkaline batholith linked to a weak oblique collision between 620 and 570 Ma (Liégeois *et al.* 1987; Liégeois 1988). A rapid switch to silica-oversaturated alkaline magmatism (Ba *et al.* 1985) occurred around 560 Ma, lasted for 20 Ma (Liégeois & Black 1984) and can be related to a reversal

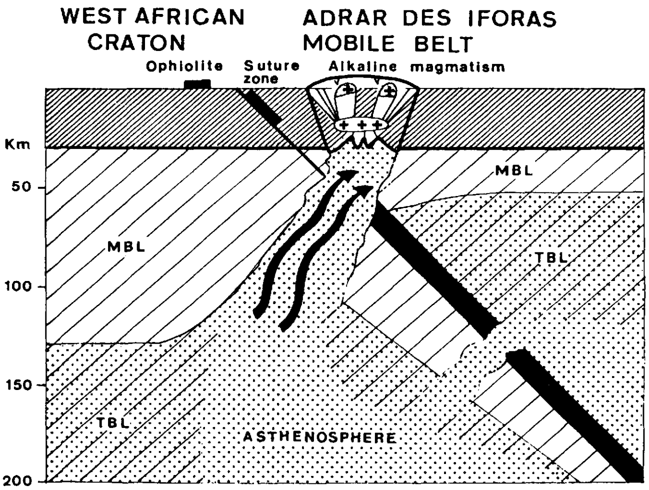

Fig. 7. In the case of gentle oblique collision as in the Adrar des Iforas, continental lithospheric mantle delamination probably does not occur as in intense frontal collision. Detachment of the subducted oceanic plate, sliced by lithospheric mega-shear zones, allows upwelling of the asthenosphene. This induced a short but voluminous post-collision alkaline magmatism (ring-complexes, ignimbritic plateau lavas, dyke swarms) contaminated by lower crust. This configuration preserves pre-collision features such as ophiolitic remnants, oceanic island arc (in the suture zone), cordilleran subduction-related intrusion and volcanic rocks (andesites) and also a simple suture geometry (modified after Liégeois & Black 1987).

of the stress field (Boullier *et al.* 1986) with rejuvenation of N–S mega-shear zones. On the basis of Sr, Nd, Pb isotopes and trace elements (Liégeois 1988), this event was interpreted as reflecting a change from a subduction-related source (lithospheric mantle + subducted oceanic plate fluids) to a shallow, less depleted asthenospheric source. The detachment of the subducted plate, probably provoked by movements along reactivated mega-shear zones, allowed the upwelling of the underlying convective mantle (Fig. 7). This would explain the very rapid calc-alkaline–alkaline transition (Liégeois & Black 1984, 1987). A similar phenomenon took place in Saudi Arabia around 600 Ma and there also it seems to coincide with a reversal in the sense of movement of shear zones (Najd fault system; Agar 1986).

Continental lithospheric mantle and alkaline rocks

The important amounts of continental lithospheric mantle which we believe have been delaminated during the Pan-African can produce megaliths at the upper–lower mantle boundary as conceived by Ringwood (1982). The products of their delayed destabilization can rise as plume material and affect the composition of recent ocean island or intraplate basalts (Fig. 8). A continental lithospheric mantle signature in alkaline intraplate magmatism has often been invoked (Davies *et al.* 1989; Sun & McDonough 1989). Delamination of continental lithospheric mantle during the Pan-African can determine the 'ages' characteristic of the 'ocean-island basalt' source: the Pb-Pb ages, or isolation times, between 1 Ga and 2 Ga are equal to the probable mean age of the continental lithospheric mantle delaminated during the Pan-African; the maximum age of 600 Ma found by Galer & O'Nions (1985) for the residence time of the ocean island basalt Pb in the mid-ocean ridge basalt source is a Pan-African age. An explanation for the Dupal anomaly (Dupré & Allègre 1983; Hart 1984) located in the southern hemisphere and characterized by an extreme isotopic 'exotic' component also may well be provided by Pan-African continental lithospheric mantle delamination during the assembly of the Gondwana supercontinent.

The Pan-African domain of northern Africa is renowned for its numerous provinces of anorogenic alkaline magmatism. The location and genesis of the alkaline magmatism by pressure release and accompanied by volatile input is thought to have been determined by reactivation of deep lithospheric faults (Bailey 1977; Sykes 1978; Black *et al.* 1985). It is thus triggered from above, the source of partial melting being

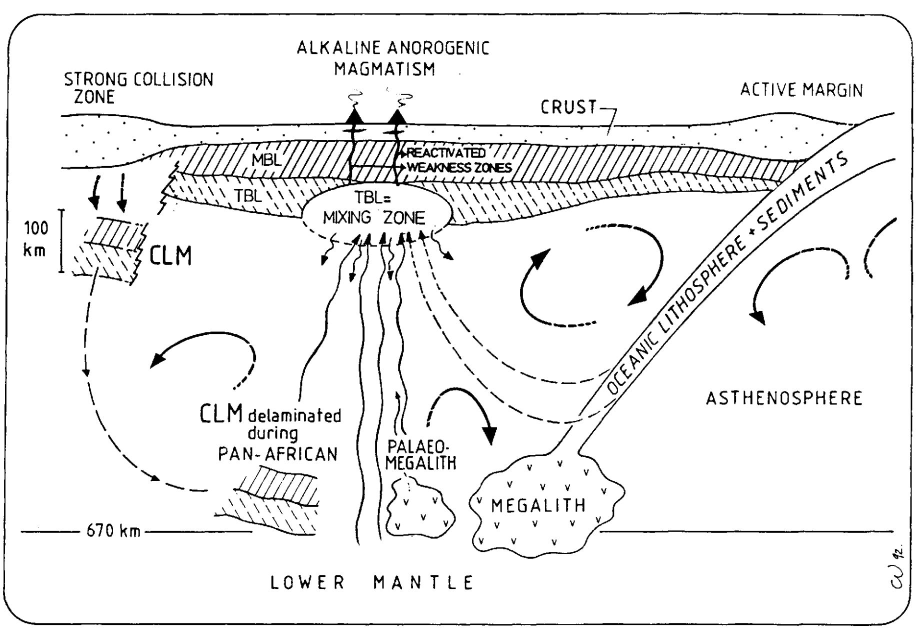

Fig. 8. Schematic view showing: (1) the suggested link between the Pan-African orogeny and Phanerozoic alkaline rocks through continental lithospheric mantle delamination; (2) the major mixing source zone constituted by the thermal boundary layer. This mixing concerns the different source precursors of the alkaline magmatism (subducted oceanic lithosphere and sediments, products of the destabilization of the megaliths (Ringwood 1982) and of the previously delaminated continental lithospheric mantle, lower and asthenospheric products). See Fig. 2 for definition of CLM, MBL and TBL.

located beneath the rigid lithosphere. This apparent contradiction with what has been written in the preceding paragraph and in general with the deep source for ocean island basalt proposed by geochemists (upper–lower mantle boundary or core–mantle boundary) can be removed if the thermal boundary layer acts as a major mixing zone and relay source. The mechanical boundary layer is considered as being isolated from the asthenosphere and the base of the thermal boundary layer as exchanging with the asthenosphere on a short time scale. All the intermediates probably exist in the thermal boundary layer whose shallower levels may eventually be incorporated in the mechanical boundary layer and can then constitute a storage reservoir of rather long life.

Consideration of the thermal boundary layer as a main mixing zone explains several points (Fig. 8).

(1) Variability in the rate of exchange of the thermal boundary layer with the asthenosphere and its storage capacity, allows deeper material (asthenosphere, destabilized megalith or continental lithospheric mantle, lower mantle and recycled subducted sediments, ... or using another vocabulary, MORB, EM1, EM2, HIMU ... as used by Zindler & Hart 1986; Weaver 1991) to penetrate and mix partially, creating all scales of heterogeneities.

(2) The thermal boundary layer can move with the plate as in the case of the Tadhak area (Mali) where undersaturated alkaline magmatism occurs with a similar mantle source (Weis *et al.* 1987) during 100 Ma from early Permian to Jurassic when the plate moved by more than 2500 km (Liégeois *et al.* 1983, 1991). Such a storage over 100 Ma has also been proposed for the Cameroon line (Halliday *et al.* 1990).

(3) Reactivation of zones of weakness in the rigid lithosphere can produce intraplate magmatism through pressure release in the underlying thermal boundary layer.

We cannot agree with the model of Ashwal & Burke (1989) to explain the recent volcanism, where cratons and Pan-African domains differ only in composition as a result of the postulated contrast between the former made by accretion of island arcs, and the latter resulting from Tibetan-style continental collisions which generate instantaneously and permanently the 'alkaline fertile mantle' in the Pan-African domain. We believe that progressive thickening of the mechanical boundary layer in stable areas is essential to an understanding of the Phanerozoic alkaline magmatism in Africa. This alkaline magmatism has been interpreted in a review paper in terms of lithospheric thickness (Black *et al.* 1985). Phanerozoic silica-oversaturated alkaline ring-complexes sometimes associated with anorthosites (Black 1965; Bowden *et al.* 1987; Demaiffe *et al.* 1991) are restricted to Pan-African belts (Black & Girod 1970, Thorpe & Smith 1974), whereas silica-undersaturated complexes, carbonatites and mixed provinces occur either in a cratonic environment (often related to rifting) or in Pan-African domains but only since the Mesozoic. When repeated alkaline magmatism occurs in an area, the undersaturated associations post-date the alkaline granitoids. This can be linked to progressive lithospheric thickening and tapping of a deeper source. On experimental grounds, Bailey (1987) estimated the depth of fusion for transitional basalts which give rise to the pantellerite–comendite–rhyolite association at about 70 km, for basanite (olivine basalt–phonolite–trachyte association) at about 80 to 100 km, for the most strongly undersaturated nephelinite-carbonatite association at 120 to 150 km, and beyond for kimberlites and lamproites. The spectacular development of oversaturated alkaline complexes in Northern Africa (e.g. Vail 1989; Barth *et al.* 1983; Ba *et al.* 1985; Bowden *et al.* 1987) can then be related to a young and warm mechanical boundary layer and the more undersaturated alkaline provinces to an older, thicker and colder mechanical boundary layer. We note however that the shear zones responsible for the location of the alkaline magmatism may have provoked local delamination of the continental lithospheric mantle in particular in more recent times. This could explain the rather thin mechanical boundary layer existing now under the Cenozoic volcanic domes of the Trans-Saharan belt.

Conclusions

Continental lithospheric mantle plays a preponderant role in geology through both its upper brittle part, the mechanical boundary layer, much stronger than the crust and chemically isolated from the asthenosphere, and its lower ductile part, the thermal boundary layer, a hybrid zone above the asthenosphere.

A thick, rigid mechanical boundary layer well attached to the crust characterizes cratons whereas active margins display much thinner continental lithospheric mantle, susceptible to delamination if rapidly thickened tectonically during collision. This is due to Moho weakness at intermediate and high temperatures counteracting the progressive growth of continental lithospheric mantle with time. This explains, at the end of a Wilson cycle after oceanic closure, the continued mobility of the mosaic of terranes wedged between rigid moving cratons as is currently the case of Eurasia indented by cratonic India.

The contrast in Phanerozoic behaviour noted by Kennedy (1964) between inert cratons and Pan-African mobile belts, the loci of abundant alkaline magmatism and of sedimentary basins, depends on the thickness of continental lithospheric mantle.

Cratons can also suffer continental lithospheric mantle delamination during hypercollision. This 'decratonization' leads to the development of extensive 'old reactivated terrains'. We propose that the disappearance of such a craton in Central Sahara (ghost craton) during the early Pan-African events (*c.* 700 Ma) clarifies many features of the wide Pan-African domain of Northern Africa.

Voluminous continental lithospheric mantle delaminated during early Pan-African collisions and recycled into the asthenosphere, can constitute one of the end-member components contributing to the genesis of recent ocean island basalt type rocks. A clear relationship between anorogenic magmatism, lithospheric thickness and reactivation of lithospheric shear zones points to the thermal boundary layer as a major relay source zone for alkaline rocks where mixing of the different mantle ocean island basalt components can occur.

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