UNIVERSITÉ DU QUÉBEC À MONTRÉAL

LE COMPLEXE CRÉTACÉ DU RICHAT (MAURITANIE); UN PROCESSUS ALCALIN PÉRI-ATLANTIQUE

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THE CRETACEOUS RICHAT COMPLEX (MAURITANIA); A PERI-ATLANTIC ALCALINE PROCESS

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RÉSUMÉ

Le Richat est un complexe alcalin Crétacé situé dans la partie mauritanienne du désert du Sahara en Afrique de l'Ouest. Le niveau d'érosion de cette structure a permis, dans un premier temps, de mieux comprendre un sommet de complexe alcalin. Ceux-ci sont rarement documentés et le Richat offrait la possibilité d'étudier les processus de dissolution et de fracturation en sommet de pluton ainsi que l'évolution des différents épisodes intrusifs et extrusifs qu'on y retrouve. L'âge et la position géographique du Richat ont également permis d'investiguer les causes de mise en place d'un complexe isolé et de l'intégrer dans le cadre régional de l'ouverture de l'Océan Atlantique

Des analyses géochimiques, pétrologiques et fractales, ainsi que des datations $^{40}Ar/^{39}Ar$, ont été effectuées sur la brèche centrale du Richat. Les résultats indiquent la présence d'un épisode hydrothermal important à l'origine de la formation de la brèche par dissolution des unités sédimentaires en place et leur effondrement subséquent. Un lien avec la mise en place du pluton est suggéré alors que le bombement et la production de fluide ont conditionné les processus de bréchification et d'altération au sommet du complexe du Richat.

Le Richat présente une grande variété de roches intrusives et éruptives marquant un contraste d'érosion entre la partie centrale et la partie externe du dôme. Plusieurs analyses ont été entreprises sur les diverses roches ignées du Richat et incluent, notamment, des études géochimiques, pétrologiques, d'isotopes stables de carbone et d'oxygène, structurales, de microscope électronique à balayage (MEB) et de diffractométrie des rayons X (DRX). Des données magnétiques et satellitaires viennent également compléter l'information. Nos travaux montrent que le Richat est la superposition d'une série bimodale tholéiitique recoupée par des magmas carbonatitiques et kimberlitiques. Les magmas alcalins proviendraient d'une source asthénosphérique alors que les magmas tholéiitiques seraient issus du manteau sous-continental et apparentés à ceux de l'épisode du CAMP (*Central Atlantic Magmatic Province*) de l'ouest africain. Les anisotropies préexistantes ont servi de conduit à la remontée des magmas asthénosphériques et sous-continentaux permettant la coexistence de magmas alcalins et tholéiitiques au sein d'un même complexe igné. Le contraste d'érosion entre la partie centrale et externe du Richat est expliqué par un effondrement en piston préservant les facies extrusifs centraux ainsi que la brèche hydrothermale karstique.

Afin de trouver une explication à la mise en place du Richat, l'hypothèse du passage d'un point chaud a ensuite été testée. La trajectoire d'un point chaud hypothétique situé sous le Richat à 100 Ma jusqu'à aujourd'hui (0 Ma) a été calculée en utilisant le logiciel PointTracker v4c sur un intervalle de 10 Ma. Les résultats montrent que le Richat ne proviendrait pas du passage de la plaque africaine au-dessus d'une plume mantellique. Pour obtenir des éléments de réponse complémentaires, une compilation régionale des structures, intrusifs et anomalies géophysiques a été effectuée. Un aulacogène Protérozoïque, probablement lié à la phase d'océanisation des Mauritanides, a été identifié et associé spatialement au Richat. Il est proposé que la réactivation de cette structure préexistante lors des rééquilibres tectoniques mid-Crétacé ayant affectés les bordures continentales de la zone péri-Atlantique ait joué un rôle dans la mise en place du Richat.

Le Richat s'inscrit dans le contexte géodynamique particulier du démantèlement de la Pangée, au moment de la séparation finale entre la plaque africaine et sud-américaine. Cette période est caractérisée par une activité anormale alcaline sur les bordures continentales de l'Océan Atlantique, généralement associée au développement de divers points chauds indépendants. Afin de comparer le Richat dans le cadre dynamique de l'ouverture Atlantique, une compilation spatiale et temporelle des manifestations alcalines mésozoïques de toute la bordure Atlantique a été réalisée. Les caractéristiques structurales des différents districts alcalins de même que les critères communément associés aux modèles de point chaud ont également été compilés et traités. Les résultats montrent des liens étroits entre les différents districts alcalins. Leur distribution spatiale et temporelle a permis de définir une Province Alcaline Péri-Atlantique (PAAP) d'âge mid-Crétacé à fort contrôle structural. La PAAP présente peu de caractères propres aux points chauds profonds et semble incompatible avec une origine de plume mantellique. Elle serait plutôt en lien avec des processus de réactivation de faiblesses lithosphériques préexistantes ayant agi comme zones d'accommodation du stress intraplaque lors d'épisodes spécifiques de l'évolution tectonique de l'Océan Atlantique. Une origine superficielle et asthénosphérique est favorisée à une origine profonde. Le Richat serait donc un témoin de cet événement péri-Atlantique.

Mots clés: karst, hydrothermal, brèche, silicification, Afrique, complexe alcalin, caldera, plume mantellique, ouverture Atlantique, Crétacé, réactivation.

ABSTRACT

The Richat structure is a Cretaceous alkaline complex in the Mauritanian part of the Sahara Desert in West Africa. Due to its level of erosion, Richat provides a rare opportunity to study the summit of a complex, particularly its dissolution and fracture processes, and allows its volcanic and magmatic evolution to be established. Its age and geographical location were used to determine the cause of its emplacement and its position within the dynamic context of the Atlantic Ocean opening.

Geochemical, petrological and fractal analyses, as well as $^{40}Ar/^{39}Ar$ age dating, were performed on samples from the central breccia of the Richat structure. Results indicate the presence of a large-scale hydrothermal event responsible for breccia formation by dissolution and collapse of the host sedimentary unit. The breccia core was genetically related to plutonic activity because doming and the production of hydrothermal fluids must have been instrumental in creating a favourable setting for dissolution.

The Richat structure presents a wide variety of extrusive and intrusive rocks from different erosional levels. Several analyses were performed on these igneous rocks, including geochemical, petrological, isotopic (oxygen and carbon), structural, scanning electron microscope (SEM) and X-ray diffraction (XRD) studies. Magnetic and satellite data were also used. Our work shows that Richat represents the superposition of a bimodal tholeiitic series and crosscutting undersaturated carbonatitic and kimberlitic magmas. The alkaline magmas may have been derived from the asthenosphere, whereas the tholeiitic magmas could have come from metasomatized sub-continental lithospheric mantle. In such a scenario, pre-existing anisotropies could have acted as a pathway for ascending asthenospheric and sub-continental melts, allowing for the coexistence of alkaline and tholeiitic magmas within the same igneous complex. The erosional contrast between the central and outer parts of the Richat complex is explained by a piston-like collapse that preserved the central extrusive felsic facies and the hydrothermal karst infilling.

In order to provide a satisfactory explanation for the Richat occurrence, the hot spot hypothesis must be tested. The trajectory of a hypothetical hotspot beneath the Richat structure at 100 Ma until recent time (0 Ma) was calculated with a time interval of 10 Ma using PointTracker v4c. Results show that Richat is unlikely to be related to the passage of the African plate over a mantle plume. To find alternative solutions, a regional compilation of structural features, magmatic intrusions and geophysical anomalies was completed. A Proterozoic aulacogen, probably related to the oceanization stage in the Mauritanides, was identified and recognized as being spatially associated with the Richat structure. It is proposed that the reactivation of such pre-existing structures may have played a role in the emplacement of the Richat complex during mid-Cretaceous tectonic re-equilibrations along continental margins around the peri-Atlantic zone.

The Richat complex was part of the unusual geodynamics processes operating during the breakup of Pangea, specifically at the moment of final separation between the African and South American plates. This period is characterized by abnormal alkaline activity along the continental margins of the Atlantic Ocean, generally associated with the development of various individual hot spots. In order to place the Richat within the dynamic framework of the Atlantic opening, it was necessary to complete a spatial and temporal compilation of Mesozoic alkaline occurrences around the edges of the ocean. The structural characteristics of the various alkaline districts, along with the criteria commonly associated with hot spot models, were also compiled and processed. The results demonstrate that direct links exist between the alkaline districts. Their spatial and temporal distributions were used to define a mid-Cretaceous Peri-Atlantic Alkaline Province (PAAP) with a strong structural control. The PAAP presents few of the classic deep hot spot characteristics, and appears incompatible with the origin of a mantle plume. It is more likely linked to the reactivation of pre-existing lithospheric weaknesses acting as accommodation zones for intraplate stresses during specific episodes in the tectonic evolution of the Atlantic Ocean. A shallow asthenospheric origin is favoured over a deep mantle source. The Richat structure thus bears witness to this peri-Atlantic event.

Key words: karst, hydrothermal, breccia, silicification, Africa, alkaline complex, caldera, mantle plume, Atlantic opening, Cretaceous, reactivation.

INTRODUCTION

L'introduction du modèle de point chaud, il y a plus de 40 ans (Wilson 1963), a joué un rôle important dans les sciences de la Terre modernes. Utilisée initialement pour expliquer le volcanisme des chaînes d'îles océaniques (Morgan 1971), la notion de plume mantellique est maintenant considérée comme un phénomène majeur dans les processus de refroidissement d'une planète (Davies 1999). La cartographie détaillée de la surface de Mars et Vénus lors des missions Pathfinder et Magellan a permis d'observer des images montrant la présence de volcans gigantesques, rifts et bombements lithosphériques. À l'aide des données géophysiques et géomorphologiques, des chercheurs ont démontré que ces phénomènes sont l'expression superficielle de plumes mantelliques tant sur Mars (Fuller et Head 2003; Redmond et King 2004) que sur Vénus (Stofan et al. 1992; Krassilnikov 2002; Ernst et Desnoyer 2003). La nature mantellique des magmas produits dans les points chauds introduit un débat quant à la source du réservoir. Plusieurs idées s'affrontent et le débat persiste actuellement. Certains soutiennent une remontée mantellique provenant de la limite manteau inférieur-noyau externe (Cambell et Griffiths 1990; Thompson et Tackley 1998), d'autres de la zone de transition entre le manteau inférieur et supérieur (White et Mckenzie 1989) ou encore du manteau supérieur (Anderson 1995; King et Anderson 1998; Sheth 1999; King 2007).

Des études sur les isotopes d'hélium, de néon et d'osmium (Walker et al. 1995; Farley et al. 1998) appuient l'hypothèse d'une source très profonde provenant du noyau externe. Cependant, les travaux récents de Schersten et al. (2004), montrent, par les isotopes de tungstène, l'absence de contribution du noyau externe dans les plumes mantelliques et suggèrent plutôt un apport provenant d'une croûte recyclée. Ernst et Desnoyers (2003) arrivent à la même conclusion avec les isotopes d'hafnium alors que différents modèles, s'appuyant sur des mécanismes de fracturation lithosphérique, de contrôles structuraux et d'instabilités convectives peu profonde, supportent une origine purement asthénosphérique (Sykes 1978; McHone 1996, 2000; Anderson 2000; Bailey et Woolley 2005). Courtillot et al.

(2003) proposent pour leur part un modèle unificateur comportant trois différents types de points chauds issus de trois sources de différentes profondeurs.

L'intérêt de l'étude des points chauds n'est pas seulement d'ordre scientifique mais aussi économique. Le magmatisme alcalin qu'ils produisent, présente un grand intérêt économique. Il est responsable de nombreux gisements tels que diamants, niobium, uranium, terres rares, etc. (Pirajno 2004). La compréhension de leur mode de mise en place revêt donc une importance toute aussi significative que leur source mantellique. On sait que la distribution géographique, la répartition dans le temps ainsi que l'association avec certains phénomènes géologiques ne sont pas laissées au hasard dans l'étude des intrusions alcalines (Woolley 1989). Ces derniers se trouvent généralement associés aux tracés de points chauds, sutures de plaques, zones de rifting, bordures de craton, points triples, accidents structuraux importants, etc. Ces zones favorables expliquent d'ailleurs l'aspect "grégaire" des manifestations alcalines. En effet, on les retrouve souvent en amas, alignées ou bien concentrées sur une zone géologique favorable.

L'origine d'un complexe alcalin isolé

Il existe cependant quelques cas où l'on retrouve des complexes alcalins isolés, loin de toute évidence de point chaud et dans un environnement géologique qui ne semble pas propice à leur mise en place. Ces phénomènes sont mal compris et constituent un type d'intrusif rare et peu documenté.

La problématique est donc de connaître la signification et l'origine d'un complexe alcalin isolé. Quels sont les processus de mise en place? Pourquoi se trouve-t-il à cet endroit précis? Quelle est l'origine et l'évolution des magmas? À quel type de point chaud est apparenté ce phénomène (superficiel, intermédiaire, profond)? Répondre à ces questions permettrait de mieux comprendre un certain type de point chaud et entraînerait des conséquences sur la connaissance actuelle de la géodynamique du manteau. La résolution de cette problématique pourrait aussi se reporter aux planètes telluriques telles Mars et Vénus, qui ne possèdent pas de tectonique des plaques et donc peu de caractères propres aux modèles actuellement acceptés sur Terre.

Pour répondre à cette problématique, il s'agit donc d'étudier un complexe alcalin ne comportant pas d'alignement volcanique, n'étant pas lié à une LIP (*Large Igneous Province*), situé dans un environnement cratonique stable et isolé de manifestations alcalines environnantes. Nous avons choisi le site exceptionnel du Guelb er Richat. Par sa simplicité et sa position géologique, la structure du Richat est un endroit unique pour aborder la question. En effet, le Richat est isolé et situé en plein Craton Ouest Africain (Sahara mauritanien). Il est loin de tout accident structural important apparent, il est bien préservé, n'a pas subi de métamorphisme et inclut une série de roches témoignant de profondeurs de fusion différentes (kimberlite, filons de carbonatite, gabbros et roches éruptives à analcime). Le Richat s'avère être ainsi une véritable fenêtre sur la structure profonde du Craton Ouest Africain et un endroit exemplaire pour étudier la question.

L'étude du Richat permet également d'inclure une sous-problématique liée à l'étude des complexes alcalins. Initialement interprétée comme le résultat d'un impact météoritique, la structure du Richat, et principalement la brèche centrale qu'elle renferme, pose une véritable énigme quant à son origine. Plusieurs chercheurs ont étudié cette structure mais sa compréhension demeure faible (Destombes et Plote 1962; Bardossy et al. 1963; Cailleux et al. 1964; Dietz et al. 1969; Fudali, 1973; Monod et Pomerol, 1973; Boussaroque 1975; Woolley et al. 1984). Les phénomènes de dissolution et de fracturation au sommet d'un pluton alcalin sont rarement documentés et ce dôme structural, avec sa brèche centrale et sa surface d'érosion, constitue un endroit extraordinaire pour étudier ces phénomènes. De plus, le Richat présente une rare exposition de phénomènes magmatiques peu profonds où se côtoies des roches intrusives et extrusives indiquant un contraste d'érosion et une mécanique particulière de mise en place. Cette exposition fournie un aperçu unique de l'évolution dynamique volcanique et magmatique d'un complexe alcalin. Résoudre l'énigme du Richat permettrait de mieux comprendre ces structures.

Le contexte géodynamique péri-Atlantique

Des datations sur les carbonatites du Richat ont donné un âge Crétacé (99 Ma) à cet épisode magmatique (Poupeau et al. 1996). La période de mise en place du Richat se situe donc en plein cœur de l'ouverture de l'Océan Atlantique, au moment de la séparation finale entre la plaque africaine et sud-américaine vers ~ 100 Ma (Mascle et al. 1988; Nürnberg et Müller 1991; Torsvik et al. 2006). Cette période, mid-Crétacé, est également le théâtre d'un pulse important de l'activité magmatique alcaline sur les bordures continentales de la zone péri-Atlantique. En effet, l'intrusion simultanée et soudaine d'une multitude de complexes alcalins sur les marges continentales éloignées de l'est de l'Amérique du Nord, d'Iberia, de l'Afrique de l'Ouest, de même que sur les marges accolées de l'Afrique et de l'Amérique du

Sud est reportée par plusieurs auteurs de travaux de synthèse (Marsh 1973; McHone 2000; Woolley 1987; 1989; 2001).

Cependant, les études traitant de ces événements alcalins font l'objet d'interprétations locales où chaque groupement de roches alcalines (le terme « district alcalin » sera utilisé ici) est abordé de façon indépendante (Foland et Faul 1977; Rock 1982; Eby 1984; Rahaman et al. 1984; Byerly 1991; Comin-Chiaramonti et Gomes 1996; Baksi 1997; Bernard-Griffiths et al. 1997; Alberti et al. 1999; Gomes et al. 2004; et autres). Cette approche conduit généralement à l'association directe entre l'éruption d'un district alcalin et le passage au-dessus d'une plume mantellique et conséquemment, à invoquer une multitude de points chauds indépendants et répartis un peu partout autour de l'Océan Atlantique au Crétacé. Toutes ces manifestations de plumes mantelliques s'inscrivent alors au sein d'une période définie comme un grand événement de superplume (*Superplume Event;* 120-80 Ma; Larson, 1991).

Lorsque l'on considère l'étendue de la zone, la multitude de points chauds à invoquer et le court laps de temps en jeu, on peut s'interroger sur la nécessité de faire appel à ce modèle pour expliquer les différentes manifestations péri-Atlantique de ce pulse alcalin et si les conclusions seraient les mêmes en utilisant une approche plus globale. Il y a donc lieu de se poser certaines questions : y a-t-il un lien entre les différents événements dispersés sur la bordure Atlantique? Est-ce que le modèle classique du point chaud issu d'une plume mantellique profonde est aussi convaincant en traitant tous les intrusifs globalement? L'ouverture de l'Atlantique peut-elle avoir joué un rôle? Bref, que s'est-t-il passé aux environs de 100 Ma?

Dans le contexte géodynamique du démantèlement de la Pangée, l'étude indépendante des différents districts alcalins pourrait représenter une vision étriquée des choses et nous allons tester toutes ces questions en prenant le pulse alcalin comme un tout, en l'observant à une autre échelle et en considérant les manifestations alcalines dans leur ensemble.

Objectifs

Quatre objectifs correspondent aux problématiques soulevées précédemment. Ces objectifs seront abordés de façon méthodique sous la forme d'un zoom arrière. Il s'agit donc de comprendre la structure à l'étude puis de l'intégrer au contexte régional de l'époque afin d'identifier les différents moteurs ayant pu jouer sur la mise en place d'un tel système magmatique. En ayant comme point d'origine le cœur de la structure du Richat, les objectifs spécifiques du projet sont:

1- Déterminer la nature de la brèche centrale du Richat.

2- Déterminer l'évolution pétrologique du Richat et le modèle de mise en place du complexe alcalin.

3- Déterminer l'origine du Richat et vérifier l'hypothèse de point chaud.

4- Intégrer le Richat dans le cadre régional de l'ouverture de l'Océan Atlantique.

Présentation

Cette thèse est constituée de quatre chapitres concordant aux objectifs spécifiques définis précédemment. Les chapitres 1, 2 et 4 ont été rédigés en anglais à des fins de publication dans des revues scientifiques internationales et prennent la forme exigée par les comités de lecture où ils sont appelés à paraître.

Le chapitre 1 a été publié dans la revue *Geology* alors que les chapitres 2 et 4 sont actuellement en processus de révision par les revues *Journal of Volcanology and Geothermal research* et *Tectonophysics,* respectivement.

Les autres sections de la thèse sont rédigées en français, conformément aux règles en vigueur concernant les thèses de doctorat à l'Université du Québec à Montréal.

Afin de parvenir à une compréhension globale du phénomène et d'effectuer une cueillette appropriée de données et d'échantillons, trois missions de terrain au Richat ont été nécessaires. En raison de la logistique requise, de l'accessibilité restreinte de la zone à l'étude et des tensions politiques en Mauritanie (p.ex. coup d'état en 2005), les missions de terrain ont dû être espacées. Ceci se traduit par une évolution des idées du chapitre 1 au chapitre 2, rédigés à deux années d'intervalle. Il est donc important de noter que le modèle génétique et l'interprétation structurale sommaire qu'on retrouve au chapitre 1 sont repris au chapitre 2, rédigé à partir des informations cumulées.

Tous les articles ont été rédigés par l'auteur de cette thèse et les idées originales présentées sont les siennes. Le Dr. Michel Jébrak a agi en tant que deuxième auteur dans chaque publication, puisqu'il a pris part aux trois missions de terrain en Mauritanie, qu'il a financé le projet (missions de terrain, production de lames minces, analyses diverses, etc.) et qu'il a été relecteur de chaque article. Le Dr. Jébrak, qui a supervisé la thèse, a également guidé et aidé l'auteur à demeurer focalisé sur les objectifs de l'étude tout au long de l'élaboration de cette thèse. Le Dr. James K.W. Lee est troisième auteur dans l'article publié dans *Geology* qui correspond au premier chapitre de cette thèse. Le rôle du Dr. Lee dans cet article s'est limité à la datation $^{40}Ar/^{39}Ar$ des sédiments internes de la brèche centrale du Richat ainsi qu'à la révision du texte en langue anglaise.

Tous les chapitres de cette thèse ont également fait l'objet de conférences présentées lors des congrès internationaux suivants:

- 2007 Ocean Continent Transition meeting, Académie des sciences, Paris, France;

- 2006 Geological Association of Canada and Mineralogical Association of Canada, Montréal, Canada;

- 2004 72^e^ congrès de l'ACFAS, Montréal, Canada;

- 2004 32^e^ Congrès de géologie africaine, Orléans, France;

- 2002 Geological Society of America, Denver, USA.

Le résumé de ces conférences se trouve en annexe.

Références

Alberti, A., Castorina, F., Censi, P., Comin-Chiaramonti, P., et Gomes, C. B. 1999. Geochemical characteristics of Cretaceous carbonatites from Angola. Journal of African Earth Sciences, v. 29, p. 735-759.

Anderson, D. L. 1995. Lithosphere, asthenosphere, and perisphere. Reviews of Geophysics, v. 33, p. 9-41.

Anderson, D. L. 2000. The thermal state of the upper mantle; no role for mantle plumes. Geophys. Res. Lett., v. 27, p. 3623-3626.

Bailey, D. K. et Woolley, A. R., 2005, Repeated, synchronous magmatism within Africa: Timing, magnetic reversals, and global tectonics. In Foulger, G. R., Natland, J. H., Presnall, D. C., et Anderson, D. L., éd., Plates, plumes, and paradigms. Geological Society of America , Special Paper, v. 388, p. 365-377.

Baksi, A. K., 1997, The timing of Late Cretaceous alkalic igneous activity in the northern Gulf of Mexico basin, southeastern USA. Journal of Geology, vol. 105, Issue 5, p. 629-643.

Bardossy, G., Monod, T., et Pomerol, C. 1963. Découverte d'analcimolites d'origine endogène dans les Richat (Adrar mauritanien). Comptes Rendues de l'Académie des Sciences, v. 256, p. 3934-3936.

Bernard-Griffiths, J., Gruau, G., Cornen, G., Azambre, B., et Mace, J. 1997. Continental lithospheric contribution to alkaline magmatism; isotopic (Nd, Sr, Pb) and geochemical (REE) evidence from Serra de Monchique and Mount Ormonde complexes. Journal of Petrology, v. 38, p. 115-132.

Boussaroque, J.-L. 1975. Étude des analcimolites des Richat (Adrar de Mauritanie). Thèse de doctorat, Paris, Université de Paris VI, 105 p.

Byerly, G. R. 1991. Igneous activity: Boulder, CO, United States, vol. J, p. 91-108.

Cailleux, A., Guillemaut, A., et Pomerol, C. 1964. Présence de coésite, indice de hautes pressions dans l'accident circulaire des Richat (Adrar mauritanien). Comptes Rendues de l'Académie des Sciences, v. 258, p. 5488-5490.

Campbell, I, H., et Griffiths, R. W. 1990. Implications of mantle plume structure for the evolution of flood basalts. Earth and Planetary Sciences Letters, v. 99, p. 79-93.

Comin-Chiaramonti, P. et Gomes, C. B. 1996. Alkaline magmatism in central-eastern Paraguay; relationships with coeval magmatism in Brazil: EDUSP, 458 p.

Courtillot, V., Davaille, A., Besse, J., et Stock, J. 2003. Three distinct types of hotspots in the Earth's mantle. Earth and Planetary Sciences Letters, v. 205, p. 295-308.

Davies, G. F. 1999. Dynamic Earth: Plates, plumes and mantle convection. Cambridge: Cambridge University Press, 458 p.

Destombes, J.-P., et Plote, H. 1962. Une structure aberrante dans le Précambrien du Sahara mauritanien; hypothèse nouvelle sur l'origine du Richat (Adrar mauritanien). Comptes rendue sommaire de la Société Géologique de France, v. 7. p. 202-203.

Dietz, R. S., Fudali, R., et Cassidy, W. A. 1969. Richat and Semsityat Domes Mauritania: Not astroblemes. Geological Society of America Bulletin, v. 80, p. 1367-1372.

Eby, G. N. 1984. Geochronology of the Monteregian Hills alkaline igneous province, Quebec. Geology, v. 12, p. 468-470.

Ernst, R. E., et Desnoyers, D. W. 2004. Lessons from Venus for understanding mantle plumes on Earth. Physics of the Earth and Planetary Interiors, v. 146, p. 195-229.

Farley, K. A., et Neroda, E. 1998. Noble gases in the Earth's mantle. Annual Review of Earth and Planetary Sciences, v. 26, p. 189-218.

Foland, K. A. et Faul, H. 1977. Ages of the White Mountain intrusives; New Hampshire, Vermont, and Maine, USA. American Journal of Science, v. 277, p. 888-904.

Fudali, R. F. 1973. Origin of the analcime-bearing rocks of Richat. In Monod, T., et Pomerol, C., éd., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, p. 97-105.

Fuller, E. R., et Head, J. W. 2003. Olympus Mons, Mars: Detection of extensive preaureole volcanism and implications for initial mantle plume behavior. Geology, v. 31, p. 175-178.

Gomes, C. B., Milan, A., Velazquez, V. F., Riccomini, C., Comin-Chiaramonti, P., Vasconcelos, P. M., et Tassinari, C. C. G. 2004. Magmatismo alcalino na porcao centro-oriental do Paraguay: novos dados geocronologicos para as rochas das provyncias Central e Assuncao. Rev. Fac. Cienc., Univ. Eduardo Mondane, Maputo, Moçambique, Ed. Esp. 1, p. 179-183.

King, S. D. 2007. Hotspots and edge-driven convection. Geology, v. 35, issue 3, p. 223-226.

King, S. D., et Anderson, D. L. 1998. Edge-driven convection. Earth and Planetary Sciences Letters, v. 160, p. 289-296.

Krassilnikov, A. S. 2002. Tectonics and evolution of novae and coronae on Venus: tectonophysical modeling based on gravitational models. Solar System Research, v. 36, p. 293-321.

Larson, R. L. 1991. Latest pulse of Earth: evidence for a mid-cretaceous superplume. Geology, v. 19, p. 547-550.

Marsh, J. S. 1973. Relationships between transform directions and alkaline igneous rock lineaments in Africa and South America. Earth Planet. Sci. Lett., v. 18, p. 317-323.

Mascle, J., Blarez, E., et Marinho, M. 1988. The shallow structures of the Guinea and Ivory Coast-Ghana transform margins: their bearing on the Equatorial Atlantic evolution. Tectonophysics, v. 188, p. 193-209.

McHone, J. G. 1996. Constraints on the mantle plume model for Mesozoic alkaline intrusions in northeastern North America. Canadian Mineralogist, v. 34, p. 325-334.

McHone, J. G. 2000. Non-Plume magmatism and rifting during the opening of the central Atlantic Ocean. Tectonophysics, v. 316, p. 287-296.

Monod, T., et Pomerol, C. 1973. Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, 191 p.

Morgan, W. J. 1971. Convection plumes in the lower mantle. Nature, v. 230. p. 42-43.

Nürnberg, D. et Müller, R. D. 1991. The tectonic evolution of the South Atlantic from Late Jurassic to present. Tectonophysics, v. 191, p. 27-53.

Pirajno, F. 2004. Hotspots and mantle plumes: global intraplate tectonics, magmatism and ore deposits. Mineralogy and petrology, v. 82, p. 183-216.

Poupeau, G., Fabre, J., Labrin, E., Azdimoussa, A., Netto, A. M., et Monod, T. 1996. Nouvelles datations par traces de fission de la structure circulaire des Richat (Mauritanie). Mémoires du Service Géologique de l'Algérie, v. 8, p. 231-236.

Rahaman, M. A., Van Breemen, O., Bowden, P., et Bennet, J. N. 1984. Age migrations of anorogenic ring complexes in Northern Nigeria. Journal of Geology, v. 92, p. 173-184.

Redmond, H. L., et King, S. D. 2004. A numerical study of a mantle plume beneath the Tharsis Rise: Reconciling dynamic uplift and lithospheric support models. Journal of Geophysical research Planets, v. 109, p. 1-14.

Rock, N. M. S. 1982. The Late Cretaceous alkaline igneous province in the Iberian Peninsula, and its tectonic significance. Lithos, v. 15, p. 111-131.

Schersten, A., Elliott, T., Hawkesworth, C., et Norman, M. 2004. Tungsten isotope evidence that mantle plumes contain no contribution from the Earth's core. Nature, v. 427, p. 234-237.

Sheth, H. C. 1999. A historical approach to continental flood basalt volcanism: insights into pre-volcanic rifting, sedimentation, and early alkaline magmatism. Earth and Planetary Science Letters, v. 168, p. 19–26

Stofan, E. R., Sharpton, V. L. Schubert, G., Baer, G., Bindschadler, D. L., Janes, D. M., et Squyres, S. W. 1992. Global distribution and characteristics of coronae and related features on Venus: implications for origin and relation to mantle processes. Journal of Geophysical Research, v. 97, p. 13347-13378.

Sykes, L. R. 1978. Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation. Rev. Geophys. Space Phys., v. 16, p. 621-688.

Thompson, P. F., et Tackley, P. J. 1998. Generation of megaplumes from the core-mantle boundary in a compressible mantle with temperature-dependent viscosity. Geophysical Research Letters, v. 25, p. 1999-2002.

Torsvik, T. H., Müller, R. D., Van der Voo, R., Steinberger, B., et Gaina, C. 2006. Global plate motion frames: toward a unified model. Earth Science Reviews (en revu) – version soumise disponible à: http://www.geodynamics.no/guest/GlobalFrames ESR.pdf

Walker, R. J., Morgan, J. W., et Horan , M. F. 1995. Osmium-187 enrichment in some plumes: evidence for core-mantle interaction? Science, v. 269, p. 819-822.

White, R. S., et McKenzie, D. P. 1989. Magmatism at rift zones: The generation of volcanic continental margins and flood basalts. Journal of Geophysical Research, v. 94, p. 7685-7729.

Wilson, J. T. 1963. A possible origin of the Hawaiian islands. Canadian Journal of Physics, v. 41, p. 863-870.

Woolley, A. R. 1987. Alkaline rocks and carbonatites of the world; Part 1, North and South America. British Museum (Nat. Hist.): London, U. K., 216 p.

Woolley, A. R. 1989. The spatial and temporal distribution of carbonatites. In Bell, K., éd, Carbonatites; genesis and evolution. Unwin Hyman: London, U. K., p. 15-37.

Woolley, A. R. 2001. Alkaline rocks and carbonatites of the world; Part 3, Africa. The Geological Society London, 372 p.

Woolley, A. R., Rankin, A. H., Elliott, C. J., Bishot, A. C., et Niblett, D. 1984. Carbonatite dykes from the Richat dome, Mauritania, and the genesis of the dome. The Indian Mineralogist, p. 189-207.

CHAPITRE I

RESOLVING THE RICHAT ENIGMA: DOMING AND HYDROTHERMAL KARSTIFICATION ABOVE AN ALKALINE COMPLEX

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1. Abstract

The Richat structure (Sahara, Mauritania) appears as a large dome at least 40 km in diameter within an Upper Proterozoic to Ordovician sequence. Erosion created circular cuestas represented by three nested rings dipping outward from the structure. The center of the structure consists of a limestone-dolomite shelf that encloses a kilometer-scale siliceous breccia and is intruded by gabbroic ring dikes, kimberlitic intrusions, carbonatites and felsic volcanic rocks. Several hypotheses have been presented to explain the spectacular Richat structure and breccia, but their origin remains enigmatic. The breccia body is lenticular in shape and irregularly thins at its extremities to only a few meters. The breccia was created during karst dissolution and collapse. Internal sediments fill the centimeter- to meter-scale cavities. Alkaline enrichment and the presence of Cretaceous automorphous neoformed K-feldspar demonstrate the hydrothermal origin of these internal sediments and their contemporaneity with magmatism. A model is proposed in which doming and the production of hydrothermal fluids were instrumental in creating a favorable setting for dissolution. The circular Richat structure and its breccia core thus represent the superficial expression of a Cretaceous alkaline complex with an exceptionally well preserved hydrothermal karst infilling at its summit.

Keywords: karst, hydrothermal, breccia, silicification, Africa

1.1 Introduction

The Richat structure is one of the most spectacular terrestrial features visible from space. Ever since the first *Gemini* missions, its nested circular shape more than 40 km in diameter has become a landmark for space crews (NASA and U.S./Japan ASTER Science Team, 2000). The Richat structure is located in the Mauritanian part of the Sahara Desert, and lies within the Proterozoic Taoudenni Basin that is bounded by the Man and Reguibat basements of Archean age (Monod and Pomerol, 1973).

Early workers cited the crater-like shape and the high relief center with its kilometer-scale breccia as evidence of a meteorite impact (e.g., Cailleux et al., 1964). Other researchers have explained the structure as the result of basement adjustments, which acted like a cylindrical piston causing vertical telescopic movements (Destombes and Plote, 1962). Magmatic evidence and domal deformation of the structure have also been used to support a granitic plutonic hypothesis related to lithospheric doming (Dietz et al., 1969; Boussaroque, 1975).

The origin of the spectacular Richat structure and its breccia therefore remains enigmatic. Our field observations, breccia analyses, and petrological studies, combined with new geophysical and geochemical data allow us to address the problem from a new perspective. We demonstrate that the Richat structure and its breccia core are the superficial expression of an alkaline complex with an exceptionally preserved hydrothermal karst infilling at its summit.

1.2 Geologic setting

The Taoudenni Basin of Proterozoic to Carboniferous age is one of the major structural units of the West African craton and covers about two million square kilometers (Trompette, 1973; Bronner, 1992; Fig. 1). The stratigraphy of the basin begins with a Late Proterozoic (1100–1000 Ma) sequence consisting of sandstone, mudstone, dolomite, and dolomitic limestone. Overlying an angular unconformity, the Proterozoic (650 Ma) to Cambrian-Ordovician sequence is composed of alternating layers of limestone, dolomitic limestone, sandstone, chert, and mudstone. The summital units range in age from Late Ordovician to Carboniferous, and are mostly sandstone with intercalations of mudstone and limestone. Stromatolites have been observed in late Precambrian and Cambrian-Ordovician units.

The Richat structure is a large structural dome located within the Late Proterozoic to Ordovician part of the Taoudenni Basin. Erosion has created circular cuestas that provide a spectacular view from space of three nested rings dipping outwardly at 10°–20° (Fig. 2).

1.3 Magmatism

Four types of magmatic rocks crosscut the Richat dome (Fig. 1). These magmatic rocks are restricted to the structure and are not found in the region beyond the dome.

1. Gabbros occurs as subvertical dikes that appear continuous on recent magnetic surveys (T. Abdivall, 2005, personal commun.) and are interpreted here as two ring dikes almost centered on the circular structure. The external ring dike is ~50 m in width and the internal one ~20 m. They are located respectively ~7-8 km and ~3 km from the center of the structure. The rocks represent microgranular to pegmatitic facies of tholeiitic magma.

2. Extensional en-echelon carbonatite dikes and less common sills occur in the southern and western part of the structure. They are ~300 m long, 2–4 m wide, and oriented N15–N20°. Dikes are mainly composed of dolomite and ankerite and show beforsite affinities (Woolley et al., 1984). Enclaves of underlying sediments, such as sandstone and shale, are abundant. The dikes are silicified and partially ankeritized.

3. Extrusive felsic rocks crosscut the inner gabbroic ring dike in the central part of the structure. Collectively they form a circular plug and a small volcanic basin that outcrop to the SW and to the NE of the breccia, respectively. The former occurrence, previously described as "analcimolite" by Boussaroque (1975), was attributed to the sodic alteration of sediments but is interpreted here as a hydrothermally altered diatreme pipe. The small volcanic basin contains conglomeratic and tuffaceous rocks that dip toward the center of the dome, suggesting that it represents the remnant of a maar system.

4. Kimberlitic rocks in the form of plug and sills were recognized in the northern part of the circular structure. Aeromagnetic surveys do not reveal any notable anomalies except for the kimberlitic intrusions and the two magnetic rings formed by the gabbroic dikes, yet the

presence of carbonatite and analcime-bearing rocks shows that an alkaline complex is located at depth. Fission-track dating of apatite in the carbonatite yielded a Mid-Cretaceous age (99 ± 5 Ma; Poupeau et al., 1996), thus implying a similar age for the Richat structure.

1.4 Structural features

The regional structure is characterized by two major sets of fracture networks oriented approximately NNE-SSW and ENE-WSW (N70°–N90°; Netto et al., 1992). The NNE-SSW fractures are known to be related to the Pan-African event (600 Ma; Poupeau et al., 1996) whereas the ENE-WSW fractures are younger, crosscutting Pan-African structures (Bayer and Lesquer, 1978). Satellite observations have shown that the Richat structure has a slightly elliptical shape. As suggested by Bonin (1995), the noncircular shape of the structure could be the result of symmetrical magmatic doming overprinted by a regional stress field with a major strain oriented NW-SE, suggesting emplacement in a transtensional environment. The Richat structure postdates the Pan-African event but appears to predate the E-W regional event.

The fractures measured in the Richat structure can be classified into two sets of brittle faults trending subparallel to regional fractures. The NNE-SSW faults in the western part of the structure display dextral movements as well as minor vertical offsets. The dextral component of these faults is demonstrated by the displacement of the external ring dike (Fig. 1) and by the en-echelon pattern of the N15–20° carbonatite dikes. The ENE-WSW faults display sinistral movement, clearly shown by the offset of the external cuesta in satellite pictures (Fig. 2). These geometric relationships imply a late, regional-scale, NE-SW–oriented horizontal main stress in a conjugated pattern. Stress inversion related to doming could explain the apparent misfit between the main stress orientations suggested by the structure shape (NW-SE) and the brittle faults (NE-SW).

1.5 Breccia

The center of the structure is occupied by a breccia body ~3 km in diameter and at least 40 m thick. The breccia defines a lenticular shape that thins at its extremeties to only a few meters. This thinning is irregular and creates a finger-like shape.

The breccia body is located within the Proterozoic series lying between partially dolomitized limestone and an overlying sandstone sequence. The breccia is composed of four types of material: coarse fragments, cement, and two types of internal sediment. Matrix-supported textures are dominant but fragment-supported textures appear locally.

Breccia fragments are polymictic and include white to dark gray cherty material, quartz-rich sandstone, diagenetic chert nodules, and silicified stromatolitic limestone. Stromatolite-bearing fragments are restricted to the upper part of the unit. Locally restricted transport distances (usually centimeter- to meter-scale) characterize the fragments displaced inside the breccia. Units show downward dismantling by simple normal faulting, which can usually be reconstructed to their original position. Vugs between breccia fragments range from millimeters to several meters in size and are filled by microcrystalline quartz cement (± calcite, ± ankerite) and internal sediments.

Fragments display various shapes, from angular to rounded. Angular fragment edges are sharp, whereas rounded fragment borders are irregular, corroded, and may show reaction rims and dissolution textures such as embayments and cusps. Fragment shapes were quantified using the boundary fractal analysis method of Bérubé and Jébrak (1999), which measures the complexity of a fragment boundary. Strongly corroded fragments are characterized by large fractal dimensions (Fd), whereas weakly corroded fragments reveal small fractal dimensions. A bimodal distribution of the fractal values was observed, with one group providing an average Fd value of 1.12 (subrounded fragments) and the other, an average value of 1.04 (angular fragments).

The breccia body is intensely silicified. Thin sections reveal that most of the fragments have been replaced by microcrystalline quartz, although rare carbonate remnants are observed. Field examinations of the dolomitized and pyritized limestone at the base of the breccia unit reveal vertical veins up to several centimeters wide containing microcrystalline quartz and marked by silicified selvages up to several centimeters thick. The veins define two sets of fractures oriented N90° and N153° and may represent the basal conduits for silicification through the breccia.

The silicification event was contemporaneous with the main breccia event but also continued afterward, as evidenced by the fact that both fragments and the internal sediments are extensively silicified, and that microcrystalline quartz cement fills the breccia cavities and intergranular spaces. Furthermore, fragments are found within internal sediments in the breccia as clasts dropped downward from the overlying unit (Fig. 3A). The microcrystalline quartz cement is consistently equicrystalline. Such a texture suggests that nucleation sites were evenly distributed and that the level of silica supersaturation in the fluid was high (Sibley and Gregg, 1987).

1.6 Internal sediments

The degree of sediment infilling of any cavity varies depending on the vertical position of the cavity in the system (Sass-Gustkiewicz, 1996). At Richat, two types of internal sediments can be distinguished and display distinct spatial distributions: pink quartzitic sandstone is mainly observed in the upper levels of the breccia where it partially fills cavities, and gray laminated sediments dominate the lower levels where they create a more extensive infilling.

Pink sandstone is partially silicified and composed of quartz grains with a minor clay component, cemented by quartz; grains are subrounded to rounded and vary in size from 0.2 to 1 mm. The finely laminated gray sediments fill cavities of all sizes, the largest being ~3 m high and >7 m wide in the core of the breccia body. Sediment layers are usually horizontal, but some meter-scale folds have been observed and are interpreted as slump features along cavity margins.

Two distinct subunits were observed in the large central cavity as alternating layers <1 m thick (Fig. 3B): (1) The first subunit is banded sediment composed of millimeter-scale microsequences of quartz and K-feldspar. Quartz-dominated basal sequences appear pale gray, whereas K-feldspar–dominated upper sequences are light beige. The K-feldspar content increases up-sequence from 40% to 85%. A dynamic depositional environment in the cavity is indicated by slumps and synsedimentary faults. (2) The second subunit occurs as massive, homogenous layers containing almost equal proportions of quartz and K-feldspar.

Optical and scanning electron microscope (SEM) observations show that the K-feldspar and quartz grains for both subunits are composed of tiny, homogeneous crystals (25–50 μm). K-feldspar grains are automorphous and partially replaced by xenomorphic silica.

Furthermore, the gray sediments are enriched in alkaline elements in comparison to pink quartzitic sandstone (Fig. 4). Two K-feldspar concentrates from the gray sediments were dated using the $^{40}Ar/^{39}Ar$ method. Samples were analyzed at Queen's University, Canada, using an 8W continuous argon-ion laser coupled with a Mass Analyzer Products 216 noble-gas mass spectrometer. All data were corrected for mass discrimination, extraction-line blanks, and neutron-induced interferences. All ages were calculated using the decay constants recommended by Steiger and Jäger (1977) and are reported with 2σ uncertainties; a representative age spectrum is shown in Figure 5. Minor disturbances in the low- and high-temperature portions of the age spectrum are directly attributed to ^{39}Ar recoil artifacts resulting from the fine grain size of the crystals, but the data from both K-feldspar runs yield an integrated Mid-Cretaceous age of 98.2 ± 2.6 Ma, which can be directly related to the emplacement of the carbonatites. It is thus proposed that the origin of this sediment is more consistent with chemical precipitation during hydrothermal circulation of a fluid enriched in alkaline elements, whereas the pink quartzitic sandstone is detrital. Microsequences in subunit 1 therefore represent pulses of variable K-feldspar content in siliceous fluids.

1.7 Discussion

The presence of corrosion textures and reaction rims on fragment boundaries as well as the low level of transportation of the rounded fragments imply a chemical dissolution process rather than physical rounding in the Richat breccia formation. If more material is dissolved than new minerals are formed, cavities will appear and may enhance fluid flow, thus resulting in a positive feedback loop where further dissolution leads to cavity enlargement (Lorilleux et al., 2002). A temporary loss of cohesion of the breccia may occur when the amount of dissolved material becomes very large, inducing gravity control of the fragment organization (Knipe, 1993) and collapse phenomena. Gravitational collapses and normal faults produce mixing of well-rounded and angular fragments and would explain the presence of blocks of the overlying rock units within the breccia and local fragment-supported texture. The collapses occurred in open cavities as rock falls that caused the brittle warping of consolidated layers with concomitant disaggregation and the slumping of unconsolidated chemical sediments. The observed bimodal distribution of fragment complexity (Fd) is directly linked to the dissolution-collapse processes. Low Fd values reflect the noncorroded fragments, fallen from the breccia hanging wall, whereas high Fd values represent the fragments corroded by the dissolution event prior to the collapse. Such relations were observed by Lorilleux et al. (2002) in sandstone-hosted breccias surrounding uranium ore in the Athabasca Basin (Canada).

A magmatic event is dated to 99 ± 5 Ma by fission tracks in apatite in the carbonatite. A corresponding age of 98.2 ± 2.6 Ma is obtained by ^{40}Ar-^{39}Ar dating of the chemical sediment in the breccia. The hydrothermal episode therefore appears related to the magmatic event. Furthermore, the position of the brecciated and silicified zones at the very center of the Richat structure implies a doming-related origin. Doming would have created an intensely fractured and permeable zone, thus favoring fluid infiltration and consequent karstification and silicification. Vertical microcrystalline quartz veins observed at the base of the breccia could reflect the conduits for such fluids. Silicification and brecciation are spatially and temporally coincident, although silicification also continued after brecciation ceased.

The proposed model invokes relatively continual siliceous hydrothermal flow from the underlying pluton, leading to karst formation. The hydrothermal fluids then provoked the precipitation of chemical sediments in the lower levels and pervaded the rest of the karst zone to produced widespread silicification and cementation of the remaining voids. The alkaline complex located at depth acted as a heat source for the hydrothermal system.

1.8 Conclusion

The circular Richat structure appears to be the superficial expression of a buried alkaline complex of Cretaceous age that was affected by cuesta-type erosion. The breccia core is genetically related to plutonic activity, since doming and the production of hydrothermal fluids were instrumental in creating a favorable setting for dissolution. The resulting fluids were also responsible for subsequent silicification and hydrothermal infilling. To the best of our knowledge, karst collapse phenomena at the summit of an alkaline complex are unique but may be more frequent than previously believed.

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1. References cited

Bayer, M., and Lesquer, A., 1978, Les anomalies gravimétriques de la bordure orientale du craton Ouest-africain: Géométrie d'une suture panafricaine: Bulletin de la Société Géologique de France, v. 7, p. 863–876.

Bérubé, D., and Jébrak, M., 1999, High precision boundary fractal analysis for shape characterization: Computers & Geosciences, v. 25, p. 1059–1071.

Bonin, B., 1995, Pétrologie endogène: Paris, Dunod, Collection Géosciences, 336 p.

Boussaroque, J.L., 1975, Étude des analcimolites des Richat (Adrar de Mauritanie) [Ph.D. thesis]: Paris, Université de Paris VI, 105 p.

Bronner, G., 1992, Structure et évolution d'un craton archéen: La dorsale Réguibat occidentale (Mauritanie); tectonique et métallogénie des formations ferrifères: Editions du Bureau de Recherches Géologiques et Minières, 448 p.

Cailleux, A., Guillemaut, A., and Pomerol, C., 1964, Présence de coésite, indice de hautes pressions dans l'accident circulaire des Richat (Adrar Mauritanien): Comptes rendus de l'Académie des sciences, v. 258, p. 5488–5490.

Destombes, J.-P., and Plote, H., 1962, Une structure aberrante dans le Précambrien du Sahara Mauritanien; hypothèse nouvelle sur l'origine du Richat (Adrar Mauritanien): Comptes rendus sommaires de la Société Géologique de France, 7, p. 202–203.

Dietz, R.S., Fudali, R., and Cassidy, W.A., 1969, Richat and Semsityat Domes, Mauritania: Not astroblemes: Geological Society of America Bulletin, v. 80, p. 1367–1372.

Gromet, L.P., Dymek, R.F., Haskin, L.A., and Korotev, R.L., 1984, The 'North American shale composite': Its compilation, major and trace element characteristics: Geochimica et Cosmochimica Acta, v. 48, p. 2469–2482.

Knipe, R.J., 1993, The influence of fault zone processes and diagenesis on fluid flow, *in* Horbury, A.D., and Robinson, A.G., eds., Diagenesis and basin development: American Association of Petroleum Geologists Studies in Geology, v. 36, p. 135–151.

Lorilleux, G., Jébrak, M., Cuney, M., and Baudemont, D., 2002, Polyphase hydrothermal breccias associated with unconformity-related uranium mineralization (Canada): From fractal analysis to structural significance: Journal of Structural Geology, v. 24, p. 323–338.

Monod, T., and Pomerol, C., 1973, Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie): Nancy, Sciences de la Terre, Mémoires 28, 191 p.

NASA/GSFC/METI/ERSDAC/JAROS and U.S./Japan ASTER Science Team, "Richat Structure, Oudane, Mauritania", ASTER Image acquired October 7, 2000, http://asterweb.jpl.nasa.gov/gallery-detail.asp?name=Richat.

Netto, A.M., Fabre, J., Poupeau, G., and Champemmois, M., 1992, Datations par traces de fissions de la structure circulaire des Richats: Comptes rendus de l'Académie des sciences, Paris, v. 314, p. 1179–1186.

Poupeau, G., Fabre, J., Labrin, E., Azdimoussa, A., Netto, A.M., and Monod, T., 1996, Nouvelles datations par traces de fission de la structure circulaire des Richat (Mauritanie): Mémoires du Service Géologique de l'Algérie, v. 8, p. 231–236.

Sass-Gustkiewicz, M., 1996, Internal sediments as a key to understanding the hydrothermal karst origin of the upper Silesian Zn-Pb ore deposits: Society of Economic Geologists Special Publication 4, p. 171–181.

Sibley, D.F., and Gregg, J.M., 1987, Classification of dolomite rock textures: Journal of Sedimentary Petrology, v. 57, no. 6, p. 967–975.

Steiger, R.H., and Jäger, E., 1977, Subcommission on geochronology: Convention on the use of decay constants in geo- and cosmochronology: Earth and Planetary Science Letters, v. 36, p. 359–362.

Trompette, R., 1973, Le Précambrien supérieur et le Paléozoïque inférieur de l'Adrar de Mauritanie: Bordure occidentale du bassin de Taoudenni, Afrique de l'ouest [Ph.D. thesis]: Marseille, Université de Saint Jérôme, 573 p.

Woolley, A.R., Rankin, A.H., Elliott, C.J., Bishot, A.C., and Niblett, D., 1984, Carbonatite dykes from the Richat dome, Mauritania, and the genesis of the dome: The Indian Mineralogist, p. 189-207.

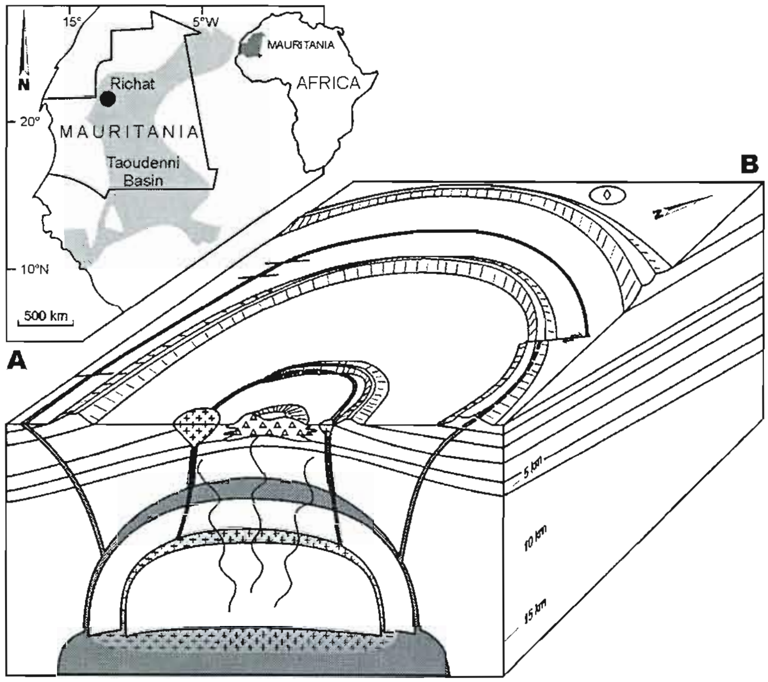

Figure 1. A: Location map for Richat dome. Taoudenni Basin is shaded. B: Block diagram displaying distribution of magmatic phases: mafic (gray), felsic (crosses), and kimberlitic (diamonds). External gabbroic ring dike is displaced by NNE-SSW fault system in northeastern part of structure and is crosscut by carbonatite dikes in southern and western sectors (oblique slashes). Circular plug and small volcanic basin outcrop respectively to SW and NE of breccia. Wavy lines indicate possible flow of hydrothermal fluids.

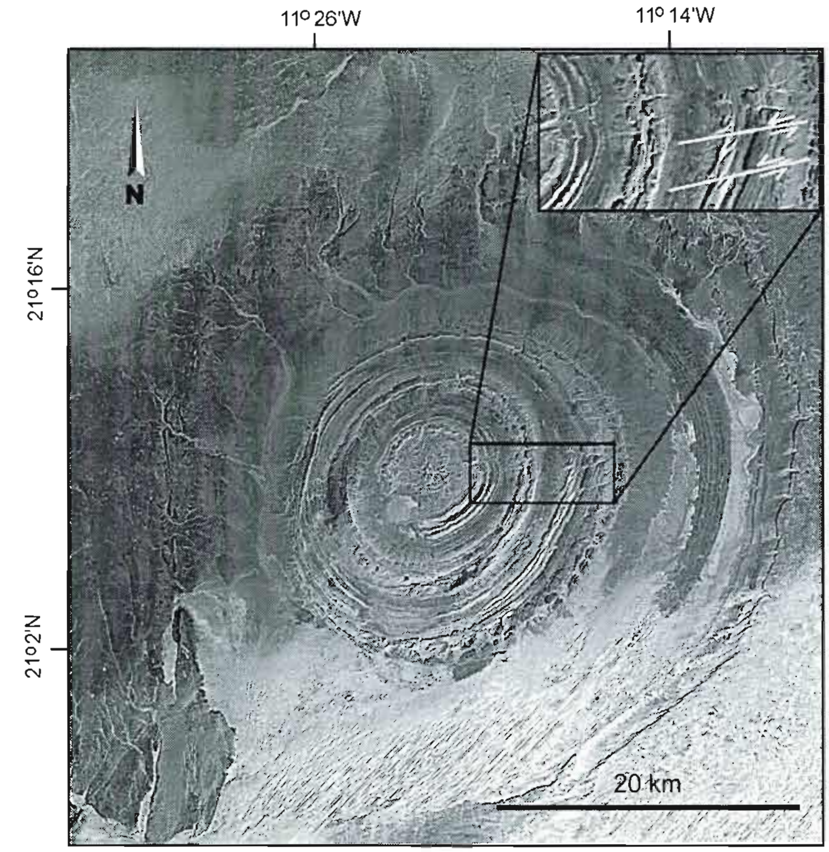

Figure 2. Modified satellite image of Richat structure after NASA and U.S./Japan ASTER Science Team, 2000. ENE-WSW faults display sinistral movement as shown by offset of external cuesta.

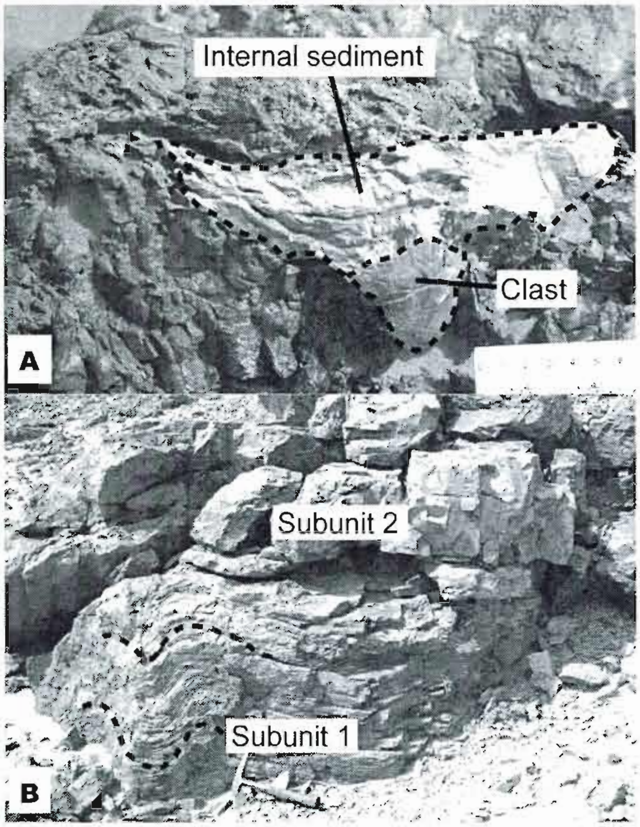

Figure 3. A: Small cavity filled by internal sediment. The limits of internal sediment, cavity, and clasts are outlined (dashes). Fragments are common in sediments that fill breccia cavities. B: Large cavity filled by internal sediment: subunit 1 is banded and displays slump structures (dashes); subunit 2 is massive and homogenous. Both subunits are composed of quartz and K-feldspar. Hammer for scale.

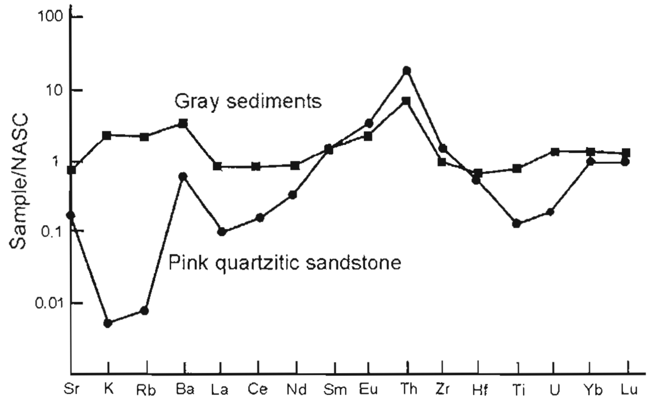

Figure 4. North American Shale Composite (NASC)-normalized spider diagram for pink quartzitic sandstone and gray sediments. NASC values are taken from Gromet et al. (1984).

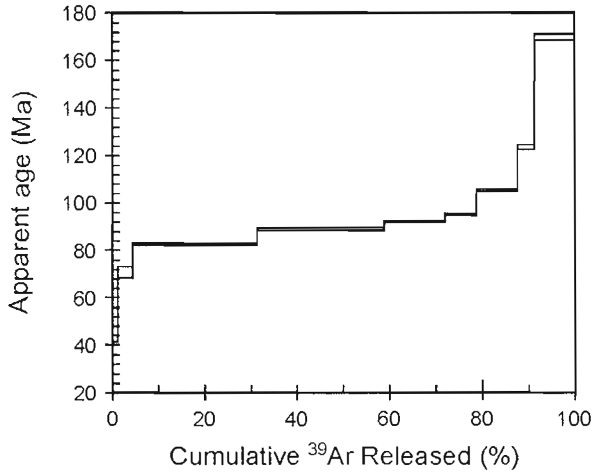

Figure 5; $^{40}Ar/^{39}Ar$ age spectrum of fine-grained K-feldspar (25-50 μm diameter) from gray laminated sediments. Apparent ages increase from 56 Ma in the initial step to a range of ages between 80-95 Ma over 75% of the ^{39}Ar released to ages greater than 100 Ma in the final three steps. The integrated (average) age of two duplicate $^{40}Ar/^{39}Ar$ analyses is 98.2 ± 2.6 Ma.

CHAPITRE II

THE "EYE OF AFRICA" (RICHAT DOME, MAURITANIA): AN ISOLATED CRETACEOUS ALKALINE-HYDROTHERMAL COMPLEX

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2. Abstract

The Richat dome is a spectacular circular structure located in the Mauritanian part of the Sahara Desert. The current erosion level of this alkaline complex presents a wide variety of contrasting extrusive and intrusive rocks from shallow to deep source regions providing insight into the magmatic process responsible for the complex.

The Richat is the superposition of a bimodal tholeiitic series crosscut by undersaturated carbonatitic and kimberlitic magmas. The bimodal series is charaterized by two concentric gabbroic ring dikes and two extrusive rhyolitic centers representing the remnant of two maar systems. Undersaturated magmas occur as carbonatite dikes, a kimberlite plug, and kimberlite sills extruded along the old regional anisotropies filling NNE-SSW dextral strike-slip faults and en-echelon tension gashes. An intense low-temperature hydrothermal event affected the Richat area. It is responsible, notably, for the karst-collapse central mega-breccia, the alteration of the rhyolites, the potassic alteration of the gabbros and the stable isotope enrichment in the carbonatites. A piston-like collapse is proposed to explain the erosional constrast existing between the central and outer part of the Richat.

Structural inheritance played an important role in the history of the Richat complex. Pre-existing anisotropies acted as a pathway for the ascent of asthenospheric and sub-continental melts and allowed the coexistence of alkaline and tholeiitic magmas within the same igneous complex. At the subsurface level, they controlled the development of the tholeiitic magma chamber and the emplacement of the undersaturated magma.

2.1 Introduction

Mauritania hosts the *Guelb er Richat*, a fascinating circular structure also called the "eye of Africa". The Richat complex appears as a large dome, at least 40 km in diameter and presents a rare exposure of volcanic and intrusive rocks from contrasting erosion levels. It consists of gabbroic ring dikes, kimberlitic intrusions, carbonatites dikes and felsic volcanic rocks and encloses a kilometer-scale siliceous breccia in its center.

This complex has been studied by several workers but remains poorly understood. Early workers cited the crater-like shape and the high-relief centre, with its kilometer-scale breccias, as evidence of a meteorite impact (e.g., Cailleux et al., 1964) or explained the structure as the result of basement adjustments and plutonic doming (Destombes and Plote, 1962; Dietz et al., 1969; Boussaroque, 1975). The debate has also focussed on the origin of the igneous rocks. "Analcimolite", a rock containing up to more than 75% analcime, for example, has been variously interpreted as an extrusive volcanic rock with primary analcime (Bardossy et al., 1963), a hot spring deposit (Fabre, in Eureka, 1999), a weathered sediment (Boussaroque, 1975) and a secondary deposit generated by the replacement of rhyolite, rhyolitic tuff, or sediment (Fudali, 1973). Associated gabbroic rocks were also poorly understood and thought to be associated to the Triassic/Jurassic extensive tholeiitic magmatism of West Africa (Trompette, 1973a; Netto et al., 1992) whereas the core megabreccia remained unexplained.

The first comprehensive study of these breccias was presented by Matton et al. (2005) who demonstrated their Cretaceous hydrothermal origin. Here, we present new petrological, structural, geophysical and geochemical data collected from three field missions on the Richat structure. New features are documented, notably the felsic volcanic basins and associated maar systems, the interpretation of the caldera in the centre of the dome and its piston-like mechanism. This paper provides new insight on the wide variety of rocks at Richat and proposes a comprehensive model of the magmatic system. We demonstrate that the Richat magmatism involved a bimodal tholeiitic series crosscut by undersaturated carbonatitic and kimberlitic magmas. Our data underline the importance of the regional structural system on a caldera history.

2.2 Geological and structural setting

The Proterozoic to Carboniferous Taoudenni Basin , is one of the major structural features of the West African craton and covers about two million square kilometers (Trompette, 1973b; Bronner, 1992). The basin is bounded by the Reguibat and Man basements of Archean age to the north and to the south respectively and is bordered on its western side by the Pan-African Mauritanides orogen (Monod and Pomerol, 1973; Fig. 1) and the Pan-African belt to the east. Sedimentary units are composed of continental shelf marine sediments with terrestrial intercalations. The stratigraphy of the basin begins with a Late Proterozoic (~1100–1000 Ma) sequence consisting of sandstone, mudstone, dolomite, and dolomitic limestone (Trompette, 1973b). Overlying an angular unconformity, the Proterozoic (~650 Ma) to Cambrian-Ordovician sequence is composed of alternating layers of limestone, dolomitic limestone, sandstone, chert, and mudstone. The summital units range in age from Late Ordovician to Carboniferous, and are mostly composed of sandstone with intercalations of mudstone and limestone. Stromatolites have been observed in late Precambrian and Cambrian-Ordovician units.

The Richat complex forms a large structural dome more than 40 km in diameter in the northern part of the Taoudenni Basin. Sedimentary rocks range in age from Late Proterozoic towards the center of the dome to Ordovician (Chinguetti Sandstone) on its periphery. The layers dip outward at 10°-20°. Erosion has created high-relief circular cuestas of resistant quartzitic units to form the three concentric ridges visible from space (Fig. 1). In this paper the use of "inner part" refers to the area located inside the innermost ridge (inset B on Fig. 2A). The Richat structure has a slightly elliptical shape. Measures of the ellipticity (length of minor axis / length of major axis) give values ranging from 0.87-0.88 for both the central depression ("inner part") and the entire structure with the major axis oriented NE-SW.

Two major sets of regional fracture networks are recognized. The first one is oriented approximately NNE-SSW and could be related to the Pan-African event (600 Ma; Netto et al., 1992; Poupeau et al., 1996). The other set of fractures strikes ~ ENE-WSW (N70°–N90°)

and crosscuts Pan-African structures (Bayer and Lesquer, 1978). Moreover, a deep narrow NE-SW trough is recognized between the Richat dome and the Mauritanides orogen (Bronner, 1992). The trough corresponds to the greatest thickness (~3500 m) of the Late Proterozoic sequence (1100-650 Ma) in the Taoudenni Basin and is interpreted as an aulacogen structure related to the Pan-African I rift event in the Mauritanides (~700-650 Ma; Villeneuve and Cornée, 1994; Matton and Jébrak, 2006).

At the scale of the dome, the fractures can be classified into three sets; two are planar and trending subparallel to regional fractures, and one appears concentric and limited to the inner part of the Richat (Fig. 1B and Fig. 2B). NNE-SSW faults display dextral movements as well as minor vertical offsets. The dextral component of these faults is demonstrated by the displacement of the external ring dike (Fig. 2C), the en-echelon pattern observed in some carbonatite dikes and by the dextral movement in wall rock striations of N15–20° carbonatite dikes. The ENE-WSW faults display sinistral movement, clearly shown by the offset of the external cuesta in satellite image (Fig. 1B, inset). Since satellite images demonstrate the extension of local fractures outside the dome, we interpret them to be related to the large scale subparallel regional fracture networks.

Usually covered by sand and/or wadis, the central concentric fractures are not directly observed in the field, but subvertical normal displacements are inferred from stratigraphic misfits, repetitions of central sedimentary sequences and duplication of the breccia unit. As seen on figure 2B, the breccia is found in the core and in the periphery of the inner central depression. Cut by previously-described NNE-SSW and ENE-WSW fracture sets, circular faults appear discontinuous and do not form a perfect concentric pattern. Post-doming movements are clearly visible in the Richat structure with the crosscutting and displacement of doming-related cuestas, ring dikes and late-stage felsic volcanic rocks (described below) by the two fracture sets for which reactivation processes are described until the Holocene (Poupeau et al., 1996). Therefore, the Richat dome appears to be characterized by the overprinting of an old regional fracture networks and doming-related structures.

2.3 Petrology

A range of igneous rocks has been reported previously from the Richat and includes quartz-gabbros and dolerites, diorites, aplites and granites (Blanc and Pomerol, 1973). These authors also mentioned abnormal values of up to 8% K_2O for the gabbroic rocks. However, our field works did not identify the more evolved part of this series (aplites and granites). Large-scale hydrothermal potassic enrichment in the Richat dome (see section 4) may have led previous workers to call certain altered rocks "aplites and granites". Four types of igneous rocks are recognized: rhyolitic volcanic rocks, gabbros, carbonatites and kimberlites (Fig. 2)

2.3.1 Rhyolite

Rhyolitic volcanic rocks occupy the inner part of the Richat dome. It is noteworthy that the volcanic rocks are restricted to the central part of the dome and do not extend beyond the inner basaltic ring dike. They consist of lava flows and volcaniclastic rocks spatially associated with two eruptive centers located to the SW and to the NE of the core, respectively (Fig. 2B). The first one is a ~2.5 km diameter volcanic crater that crosscuts the inner basaltic ring dike, forming a circular depression filled by a sebkha. The crater ring is clearly visible from satellite images and is characterized by up to 2 m-thick layer of coarse ejecta at its base, overlain by tuffs and lava flows. The coarse ejecta layer consists of fragmented volcanic material in a completely hematized matrix (Fig. 3A). The second eruptive center forms a small volcanic basin (Fig. 3B) that contains brecciated to conglomeratic rocks, lavas and tuffs (Fig. 3C) that dip toward the center of the Richat dome. Both low relief craters are interpreted as the remnant of two maar systems.

All volcanic rocks lie on an erosional unconformity and appear to postdate all other igneous phases. The stratigraphy of volcanic units is consistent from the southwestern crater area to the northeastern crater but the ejecta ring is only associated with the southwestern one. A typical cross-section of the volcanic rocks includes four main gradational units. Some units may be absent locally due to lateral variations (Fig. 4);

The poorly sorted, matrix-supported basal unit is ~1-4 m thick and ranges in texture from a breccia to a conglomerate. It contains fragments of sedimentary rocks in a tuffaceous matrix. The abundance of country rock clasts and their heterogeneous composition suggest that they were derived from the underlying sedimentary rocks by subsurface explosions. Fragments reach 30 cm in diameter and include sandstone, quartzite, chert and red jasper. Jasper always appears in the form of small rounded fragments, less than 1 cm in diameter, and such clasts have not been found elsewhere in outcrops at Richat; they may represent an underlying sedimentary unit. The roundness of the fragments and the thickness of the unit increase with distance to craters creating a transition from breccia (angular fragments) near the crater to a conglomeratic facies away from it.

Unit 2 is a 3-4 m-thick massive quartz-phyric rhyolite in sharp contact with unit 1 (Fig. 3C) and may represent a distinct lava flow. The texture is uniform with 5-10% euhedral quartz phenocrysts (< 1 mm to 2 mm) in an analcime-rich groundmass. Some quartz crystals appear partially corroded by analcime and calcite. Small (sub-millimeter) vesicles occupy 1-5% of the rock.

Unit 3 is ~5-6 m-thick and composed of up to 70% analcime glomerocrysts and up to 40% large vesicles. Glomerocrysts are frequently characterized by a core of analcime and a rim of nordstrandite + analcime (Fig. 5A) but may also be entirely composed of analcime. An increase is observed in the size and proportion of vesicles from base to top, whereas the glomerocrysts decrease in size but not in proportion. Near the contact with the underlying massive rhyolite, the rock contains ~70% glomerocrysts (1-3 mm) and ~15-20% vesicles (~0.5 mm) respectively. Progressing upward, vesicles reach 1 to 1.5 cm. The upper part of the section contains so many vesicles (~40%) that the rock has a pumiceous aspect (Fig. 3E). Note that euhedral vesicles are found and therefore, some of the cavities may come from the dissolution of an anterior mineral phase. Quartz phenocrysts are present but are less abundant and smaller in size than in the massive rhyolite. Small vesicles from the bottom part of the unit are generally filled by analcime whereas progressing to the upper part, large vesicles show variable degree of filling with botroidal analcime and nordstrandite sometimes giving an "egg shell" texture due to differential weathering (Fig. 3D).

Unit 4 is generally 3 m-thick and is characterized by the presence of large vertically-elongated lithophysae. Lithophysae are up to 1 m high and 5 cm wide. They are always oxidized and appear purple or more rarely yellowish to brown. In contrast to unit 3, porosity does not exceed 10% of the rock and vesicles are much smaller, generally <1 mm. When they are present, vesicles form vertical cavity streaks. Glomerocrysts are present but also less abundant (< 20%) and smaller (<1 mm) than in the underlying unit. In horizontal outcrops, diffusion structures give spectacular purple concentric rings that may reach 25 cm in diameter

(Fig. 3F). Pervasive oxidation also occurs as large purple stains to completely oxidized outcrops. The latter may have developed in more porous rocks than the lithophysae facies. Quartz crystals are rare and always microcrystalline.

Hand specimens show a wide range of colors from cream or light beige through yellow to ochre and purple due to the variable presence of iron oxide. Variable proportion of quartz, analcime and nordstrandite accounts for 90 % of all volcanic rocks, with minor amount of dawsonite, calcite, k-feldspar, anatase and hematite. Some volcanic rocks contain so much analcime (more than 75%) that they were called "analcimolites" by Bardossy et al. (1963). In thin section, analcime appears in two aspectss; (1) fine-grained interstitial filling and veinlets (Fig. 5C) or (2) coarse-grained cavity fills (Fig. 5D) and glomerocrysts. Nordstrandite is microcrystalline and difficult to identify, being interlaced with analcime (Fig. 5A). Close association between analcime and gibbsite or analcime and dawsonite in the Richat has been previously suggested (Fudali, 1973; Boussaroque, 1975). Although the presence of dawsonite and gibbsite has been confirmed locally, X-ray diffraction (XRD) and scanning electron microscope (SEM) analyses show that the main mineral associated with analcime is in fact the $Al(OH)_3$ polymorph of nordstrandite.

2.3.2 Gabbro

Field mapping and aeromagnetic data have shown that gabbroic rocks form two concentric ring dikes (Fig. 2A-2C). The internal ring dike is ~20 m in width and the external one ~50 m. They are located respectively ~3 km and ~7-8 km from the center of the structure. Gabbroic rocks are microgranular (< 1 mm) to coarse grained (≤ 1 cm) and show doleritic textures (Fig. 5H). Plagioclase consists of variable proportions of andesine, anorthite and albite, whereas pyroxenes are augite and diopside. Pyroxenes are partially replaced by amphiboles (actinote, hastingstite, pargasite), magnetite, biotite and chlorite, where chlorite is the last alteration phase, destroying biotite crystals. Minor proportions of quartz and muscovite are also found.

These rocks were previously described as a sill or dikes associated with the extensive Triassic/Jurassic tholeiitic magmatism of West Africa (Trompette, 1973a; Netto et al., 1992) and therefore, not necessary related to the Richat dome. Ring dikes are confined to, and almost centered on, the circular structure of the dome and consequently, more likely related to the Richat dome than part of the widespread West African Late Triassic – Early Jurassic magmatic event (now accepted as part of the Central Atlantic Magmatic Province, the CAMP event; Marzoli, 1999) as previously suggested. Therefore, we suggest a Late Cretaceous age, similar to that of the alkaline complex (see section 6).

2.3.3 Carbonatite

Thirty-two carbonatite intrusions (dikes and less common sills) are recognized in the Richat structure. First identified by Woolley et al. (1984), carbonatite dikes were thought to be radial and synchronous with the doming or emplaced along the radial system of joints and faults generated by it. Even though radial distribution of carbonatite dikes is a common feature in alkaline complexes (i.e. Alnö, Araxa, Agate Mountain, Gross Brukkaros, among others) this assumption is implausible at the Richat since carbonatite dikes are all oriented N15-20° (Fig. 2A) and were emplaced in pre-existing NNE-SSW fracture sets. Further, their occurrence is limited to certain parts of the dome. None of the carbonatites dikes extend beyond the internal depression toward the central part of the Richat.

Dikes are generally ~300 m long and 1–4 m-wide. The carbonatites are massive and are mostly devoid of vesicles. Fresh rocks appear pale grey and weather to a rusty-red to light brown color. Carbonate grain size is variable, ranging from fine to coarse subhedral to euhedral crystal grains sometimes included in less well-defined aphanitic carbonate groundmass. Dikes are mainly composed of dolomite and, in a minor proportion, ankerite. Other components include quartz, apatite, barite, hematite, calcite and traces of pyrite. Values of up to 0.23 % Nb_2O_5 in geochemical analyses strongly suggest the presence of pyrochlore or others Nb-bearing minerals.

Multiple magma injections in dikes are frequent and formed vertical banding. Vertical bands are of different colors according to the quantity of iron oxides and some have breccia envelopes. Breccia envelopes are generally 10-50 cm-wide and include angular to rounded country rock fragments (sandstone, chert, limestone and shale) as well as gabbroic clasts. Fragments are millimetric to decimetric in size and some display corrroded borders. Intensely corroded fragments are preferentially composed of quartzitic sandstone and chert.

Note that enclaves of sediments are abundant in most of the dikes and not limited to the breccia envelopes. Carbonatite dikes are silicified and partially ankeritized. Contacts with country rocks are generally sharps without significant alteration or fenitization although pervasive silicification may be found locally. Tensional veinlets oriented N254° have also been found within the wall rocks of some dikes. Fission-track dating of apatite in the carbonatite yielded a Mid-Cretaceous age (99 ± 5 Ma; Poupeau et al., 1996).

2.3.4 Kimberlite

Kimberlitic rocks in the form of a plug and several sills were recognized in the northern part of the circular structure (Fig. 2A, 2C). The plug does not outcrop but a sample of a sill was recovered in a trench. The sill is highly altered and mainly composed of coarse (~10 mm) phlogopite. Other primary minerals are deeply weathered although pyroxene and olivine remnants are found. X-ray diffraction (XRD) analyses reveal in order of importance: phlogopite, serpentine, smectite, chlorite, magnetite, hematite and amphibole. The kimberlite plug is thought to have an age of about 99 Ma (T. Abdivall, 2005, personal commun.).

2.4 Hydrothermal effects at Richat

The Richat dome is the locus of spectacular hydrothermal features such as a central megabreccia related to hydrothermal dissolution and collapse (Matton et al., 2005), alteration of rhyolites and gabbros, and shifts in stable isotopes. Moreover, several quartz veins were observed, especially in the western and central part of the dome.

2.4.1 Central breccia

A large-scale hydrothermal event occurred at the core of the Richat. The center of the structure is occupied by a karst-collapse siliceous megabreccia body at least ~3 km in diameter and up to 40 m thick (Fig. 3H). The breccia forms a lens that thins at its extremities to only a few meters. This thinning is irregular and creates finger-like shapes. Breccia fragments are polymictic and include white to dark gray cherty material, quartz-rich sandstone, diagenetic cherty nodules, and stromatolitic limestone. Hydrothermal internal layered-sediments fill the breccia. Hydrothermal infillings are composed of quartz and automorphous neoformed K-feldspar and is geochemically characterized by strong enrichment in alkaline elements (Sr, K, Rb and Ba, notably) in comparison to the detritic infilling of the upper part of the breccia (Matton et al., 2005). K-feldspar from the hydrothermal sediment was dated using the $^{40}Ar/^{39}Ar$ method and yielded an integrated age of 98.2 ± 2.6 Ma (Matton et al, 2005).

The breccia body is intensely silicified. The silicification event was contemporaneous with the main breccia event but also continued afterward, as evidenced by the fact that both fragments and the internal sediments are extensively silicified, and that microcrystalline quartz cement fills the breccia cavities and intergranular spaces. The breccia was formed during karst dissolution and collapse where doming and the production of hydrothermal fluids were instrumental in creating a favorable setting for dissolution. Further details can be found in Matton et al. (2005). The scale of this mega-breccia implies that huge quantities of hydrothermal fluids were produced at the very centre of the Richat dome.

2.4.2 Alteration of rhyolitic rocks

Early phases of alteration in the rhyolitic rocks are characterized by the replacement of quartz crystals by calcite and analcime (Fig. 5B, 5F). Neo-formed calcite was then replaced by analcime, accompanied or not by nordstrandite (Fig. 5E). In thin section, analcime and nordstrandite always appears as the last mineral phases. It is plausible that when silica was unavailable in the system, excess aluminium formed nordstrandite. Nordstrandite would then be the last phase of the alteration process. Table 1 shows this sequence of alteration. In massive rhyolites, only automorphous quartz and analcime-rich matrix is found with small amounts of k-feldspar, anatase and hematite. In the altered rhyolites, quartz becomes scarcer and is replaced by calcite and analcime while dawsonite makes is first appearance whereas in the highly altered sections, only analcime and nordstrandite remain with trace amounts of kaolinite, gibbsite, calcite, anatase and hematite.

2.4.3 Alteration of gabbroic rocks

High-K_2O intrusive rocks are found in the northern part of the structure. These rocks differ from the gabbros with up to 6.13% K_2O and SiO_2 content from 58.1% to 60.4% which gives them an apparent syenitic composition. Their occurrence is restricted to an area of about 100 m x 100 m corresponding to the northern part of a major NNE-SSW deformation zone responsible for the dextral shearing of the external gabbroic ring dike (Fig. 2A). "Syenites" have not been found elsewhere in the Richat dome. They are mainly composed of albite, microcline, quartz and chlorite with minor proportions of augite, biotite and magnetite. In thin section, secondary neoformed k-feldspar appears to corrode plagioclase crystals (Fig. 5I). All the feldspars are strongly sericitized and pyroxenes are partly to totally replaced by chlorite and magnetite. Nevertheless, a doleritic texture is still recognizable. We interpret these rocks as hydrothermally altered gabbros. The exposure is poor in that sandy area, but potassic and silica enrichment seems to be related to fracture zones inside the external gabbroic ring dike. These rocks may correspond to the aplites and granites mentioned by Blanc and Pomerol (1973) and may coincide with their reported high-K gabbros.

2.4.4 Stable isotopes

The carbon and oxygen isotope compositions of carbonatite dikes from the Richat are plotted in figure 6. None of the samples lie within the field of the Primary Igneous Carbonatites (PIC). The $\delta^{13}C$ values are close to slightly higher than the PIC range, but the $\delta^{18}O$ values are clustered at much higher levels. It is thus clear that theses carbonatites have not retained their primary isotopic signature. Deines (1989) and Reid and Cooper (1992), documented a positive correlation between $\delta^{13}C$ and $\delta^{18}O$ in the 5 to 15‰ range for the carbonatites. This probably reflects fundamental processes involved in carbonatite formation. Deines (1989) have attributed this process to Rayleigh fractionation but specify, however, that Rayleigh fractionation is unlikely to have produced $\delta^{18}O$ values greater than +17‰. The more extreme $\delta^{18}O$ values are then attributed to secondary processes involving fluid-rock interaction in a low-temperature hydrothermal environment (Deines, 1989 ; Keller and Hoefs, 1995; Castorina et al., 1997; Onuonga et al, 1997). Hydrothermal processes are also supported by the strong presence of barite (up to 10000 ppm Ba) in the carbonatites dikes from the Richat, a typical product of late-stage crystalisation of carbonatites deposited by circulating hydrothermal fluids (Woolley et al., 1984; Mariano, 1989).

In summary, the presence of a large-scale hydrothermal karst-collapse breccia (Matton et al. 2005), stable isotopes anomalies in carbonatites, potassic alteration of the gabbros, the sequence of hydrothermal alteration in the rhyolites, and the presence of barite mineralisation in the carbonatites are several evidences that a significant hydrothermal event occurred at the Richat and particularly in its central part.

2.5 Geochemistry

2.5.1 Rhyolites

Volcanic rocks from the center of the Richat show extreme present-day compositional variations, with SiO_2 values ranging from ~ 42.5 to 73.7% (Fig. 7A). Moreover, K_2O contents appear abnormally low for rhyolites, with all values falling under 0.17 wt % whereas Na_2O concentrations are always high, from 5.7 to 9.2%. However, as described above, volcanic rocks show a suite of secondary minerals (Table 1).

Figure 8 shows the variation of CaO, Na_2O and Al_2O_3 against SiO_2. A decrease in silica appears correlated with an increase in CaO up to a critical point beyond which the CaO decreases abruptly (Fig. 8A). The CaO increase may represent the addition of calcite at the expense of silica in the system whereas the CaO decrease occurs when calcite begins to be replaced by analcime (figure 8B). Sodium also increases with decreasing silica content but decreases in the most silica-depleted rocks (Fig. 8A). This is reflected by the observed replacement of quartz and calcite by analcime, whereas in the advanced stages of alteration, sodium was not sufficient and excess alumina crystallized in nordstrandite (Fig. 8B).

Insoluble alumina thus concentrates as a residual phase throughout the alteration process, since more material is removed than added (Fig. 8A-8B). In highly altered zones, increasing water circulation and alkali depletion may cause nordstrandite to dissolve and destabilize, favouring gibbsite as the most likely $Al(OH)_3$ polymorph to be found (Dani et al., 2001). High gibbsite contents reported by Fudali (1973) may thus correspond to late-stage extremely altered rocks representing the end part of this alteration suite, or misidentified nordstrandite.

The depletion of K_2O in the rhyolites may come from early destabilization of potassium-bearing minerals. Precipitation of analcime from alteration of leucite (leucite + Na + H_2O = Analcime + K) has been known since the experiments of Lemberg (1876). Although leucite was not observed in the volcanic rocks, previous studies have shown that transformation of leucite into analcime by ion exchange with Na-rich fluids is extremely rapid, taking place in a matter of days (Gupta and Fyfe, 1975; Taylor and MacKenzie, 1975; Comin-Chiaramonti et al., 1979). It is thus plausible that leucite, if originally present, has been completely transformed into analcime (Table 1). Below-average values of K_2O and high Na_2O contents could then indicate that even the "freshest" Richat rhyolites have undergone an early stage of potassium destabilization.

Alteration of rhyolitic lavas and tuffs by hydrothermal solutions thus appears the more likely process to produce the extrusive rocks outcropping at Richat, which contain up to 75% analcime. The recognition of an intense low-temperature hydrothermal event, thought to have strongly affected the Richat central area, also favours this origin for the "analcimolites". Gupta and Fyfe (1975) demonstrated that alteration of rhyolite into analcime can be achieved with the involvement of saline fluids at temperatures of ~150°C. The discovery of the two volcanic craters associated with the "analcimolites" also confirms the volcanic origin of these rocks and precludes the involvement of weathered sediments as reported by Boussaroque (1975). It is noteworthy that the depletion of silica and potassium in the rhyolitic rocks may be related to the enrichments of these elements in the breccia (and internal sediments) and the "high-K gabbros".

2.5.2 Gabbros

The mafic ring dikes are relatively primitive melts, spanning a short compositional range of Mg# 56-62. They have low TiO_2 (0.7-1.3 wt%) and moderate MgO (6.3-7.6 wt%) content, and fairly high SiO_2 (46.5-52.2 wt%). No difference in geochemistry is discerned between external and inner ring dikes.

Figures 9A and 9B show the trace element concentrations in mafic dikes from the Richat, normalized to mid-ocean ridge basalts (MORB) and ocean island tholeiites (OIT), respectively. Analysed samples differ significantly from MORB compositions, but figure 9B clearly shows the similarity between these mafic rocks and oceanic-island tholeiites. Nevertheless, the Richat gabbros are enriched in Rb, Ba, Th and Pb and depleted in Ta, Nb, Ti relative to the ocean-island tholeiites. Ta-Nb negative anomaly is as typical feature of subduction-related magmas but may also be the consequence of crustal contamination in continental flood basalt with the relative abundance of Rb and Th, wich are strongly concentrated in the upper crust, indicative of depth of contamination (Cox and Hawkesworth, 1985; Wilson, 1989; Taylor and McLennan, 1995). Comparing multi-element patterns of the Richat gabbros with the average amphibolite (upper crust) and granulite (lower crust) (Fig. 9C) suggests that gabbros from the Richat have been contaminated by upper crustal rocks. This is evidenced by the distinctive Rb and Th enrichment and the relative anomalies against the OIT plot (arrows; Fig. 9B and 9C). These features are confirmed on a Th/Yb versus Ta/Yb diagram, where the Richat gabbros plot near the enriched mantle source, suggesting the involvement of subcontinental lithosphere or OIB source mantle in their petrogenesis (Wilson, 1989) with high Th/Yb ratios which might be indicative of crustal contamination.

When compared with CAMP tholeitic dikes from West Africa (Taoudenni Basin and Guinea; Fig. 9D), Richat gabbros are very similar. Despite their relatively higher anomalies in Pb and Ba, which can be attributed to a greater degree of crustal contamination, Richat dikes are remarkably analogous with those from Low-Ti CAMP basalts, supporting their derivation from a similar mantle source. It is now widely accepted that Early Jurassic CAMP tholeiitic geochemical features are indicative of a heterogeneous source residing in metasomatized portions of the sub-continental lithospheric mantle (Pegram, 1990; Heatherington and Mueller, 1999; De Min et al., 2003; Jourdan et al., 2003; Verati et al., 2005). The younger Richat mafic igneous rocks may come from a similar source with a different extent of crustal contamination.

A bimodal series in the Richat is suggested by the presence of gabbroic ring dikes and rhyolites with no intermediate rocks observed (Fig. 11). The basaltic rocks of the bimodal series has a tholeiitic composition (Fig. 7B). The diagrams in figures 11 and 12 show the variation of major and trace element abundances as a function of the SiO_2 content in the bimodal series from the Richat. Mafic to felsic rocks display variation trends of decreasing MgO, FeO, CaO, TiO_2 and increasing Na_2O with higher SiO_2 content. As mentioned earlier, the K_2O content in the rhyolites appears abnormally low for a rhyolite and therefore, the usual trend of increasing K_2O from mafic to evolved silicic magmas is not observed. This may be due to early destabilisation of K-bearing minerals such as leucite. Abundance of the compatible trace elements (V, Ni, Sc, Sr) are systematically higher in basalt compared to rhyolite (Fig. 12). In contrast, (High Field Strength) incompatible trace elements Nb and Y show little to no variation, whereas Zr abundance is highly variable in the rhyolites. Moreover, rubidium content of rhyolites is generally low (under 20 ppm) and correlation between rubidium and silica is absent which may be an other indication of early destabilisation of K-bearing minerals.

Chondrite-normalized rare earth elements (REE) data for rhyolites and gabbros are plotted on figure 13A. Both rock types show flat pattern of heavy rare earth elements (HREE) with enrichment in light rare earth elements (LREE). Rhyolites display similar HREE concentrations relative to the gabbros but higher LREE contents. Little to no Eu anomaly is present, indicating that fractionation of feldspar in the source is limited. HREE concentrations of more than 10 x chondrite in the gabbros also suggest that garnet is absent from the source (Wilson, 1989). LREE enrichment may either be a consequence of crustal contamination or their derivation from enriched mantle sources.

2.5.3 Undersaturated rocks

Carbonatites from the Richat dome show beforsite (magnesio-carbonatite) affinities (Fig. 13C). They are rich in REE and plot within the field of carbonatites (Fig. 13B). Enrichment is much greater in light REE than heavy REE with La/Yb ratios ranging 80 and 222. This confirms the previous description made by Woolley et al. (1984) on the carbonatitic nature of three dikes from the Richat. Ba and Nb yield very high values, up to 9558 ppm and 1600 ppm respectively. The high Ba content is related to the presence of observed barite, whereas the high Nb values suggest the presence of pyrochlore or others Nb-bearing minerals. The kimberlitic sill falls within the world kimberlitic rock domain and displays relatively steep REE profile with a La/Yb ratio of 47.

2.6 Chronology

Field observations and available radiometric dating allow the following chronology of magmatic events to be proposed:

1- Gabbros;

2-Kimberlite; ~99 Ma (T. Abdivall, 2005, personal commun.);

Carbonatites; 99 ± 5 Ma (Poupeau et al., 1996);

Breccia; 98.2 ± 2.6 Ma (Matton et al, 2005);

3-Rhyolites.

No radiometric ages currently exist for the gabbros and rhyolites. However, gabbros clearly appear as the first magmatic phase in the Richat dome. Ring dikes are crosscut and displaced by the main NNE-SSW and ENE-WSW fracture sets. Moreover, they are crosscut by NNE-SSW carbonatite dikes and thus older than 99 ± 5 Ma. Consequently, a reactivation of the NNE-SSW deep pre-existing lineament (Pan-African) could have occurred around 99 Ma, allowing the carbonatite magmas to rise and fill the fractures. The kimberlite plug has been dated by Asthon Mining Inc. to ~99 Ma. However, information on the age dating method and the analytical error is not available. Moreover, its position in the northern part of the structure does not allow relative chronology with the other magmatic phases. The rhyolites lie on an erosional unconformity and appear to postdate all other igneous phases. Brecciated to conglomeratic rocks, lavas and tuffs have been found dipping toward the center of the dome, suggesting a late, post-doming emplacement. Located in the central part of the structure, they do not show any evidence of silicification and are likely to postdate the large-scale siliceous hydrothermal event that forms the core megabreccia. Taking into account that large bimodal volcanoes are generally active ≤ 5 Ma (i.e. Mont-Dore, Cantal, Siroua, etc.) and that mafic and felsic magmas appear to be respectively the first and the final magmatic events at Richat, we may assume that approximate ages could be constrained to ~ 104-99 Ma for the gabbros and ~ 99-94 Ma for the rhyolites (or 101.5 ± 2.5 Ma and 96.5 ± 2.5 Ma).

2.7 Discussion

2.7.1 Piston caldera

The center of the Richat structure is characterized by circular fractures creating stratigraphic misfits as well as repetitions of sedimentary sequences and duplication of the breccia unit. In addition, it is noteworthy that all intrusive rocks are in the external part of the complex whereas volcanic rocks are exclusively restricted to the central part. This strongly suggests a model of piston-like caldera collapse preserving the central part from erosion.

Measures of the ellipticity (length of minor axis / length of major axis) give values ranging from 0.87-0.88 for both the inner depression and the entire structure with the major axis oriented NE-SW. If, as suggested by Acocella et al. (2002), we accept that the configuration of a caldera in map view reflects the shape of the underlying magma chamber, the elongation at the Richat caldera may be due to two factors:

1- Stretching of an original subcircular caldera (or reservoir) under regional extension (De Chabalier and Avouac, 1994; Bosworth et al., 2000);

2- The presence of a pre-existing NE-SW structure.

If the non-circular shape of the structure is the result of symmetrical magmatic doming overprinted by a regional stress field, the emplacement of the magma chamber and the caldera should have occurred in a transtensional environment with the main stress oriented ~NW-SE. However, the structural faulting implies a regional-scale NE-SW–oriented horizontal main stress in a conjugated pattern (Matton et al., 2005) and therefore, the first hypothesis appears unlikely. Moreover, the Richat structure is located over a pre-existent NE-SW aulacogenous structure (Tagant-Richat-Tiris Through; Bronner, 1992; Villeneuve and Cornée, 1994; Matton and Jébrak, 2006) sub-parallel to the elongation axis of the caldera. This major structural anisotropy was therefore favourably oriented for tensional reactivation and to have controlled the development of the magma chamber and the overlying caldera. This is supported by the NE-SW alignment of the two volcanic craters in the elongation axis (Fig. 2A). The values of ellipticity measured in the Richat (0.87-0.88) also agree with analogue modelling experiments where collapse calderas controlled by pre-existing discontinuities displayed ellipticity values between 0.8 and 0.9 (Acocella et al., 2004). It is thus likely that the NE-SW elongation of the Richat caldera represents the surface expression of the reactivation of the pre-existing Tagant-Richat-Tiris Trough..

2.7.2 Structural implications

The Richat complex appears to be characterized by the overprinting of two structural systems: (1) a regional system defined by old regional fracture networks, mostly oriented NNE-SSW and ENE-WSW (Bayer and Lesquer, 1978; Netto et al., 1992; Poupeau et al., 1996) and (2) a local, doming-related, fracture set. Close relationships between magmatism and these two systems may also be defined. The undersaturated rocks (carbonatites and kimberlite) are controlled by the regional fracture set, filling NNE-SSW dextral strike-slip faults and en-echelon tension gashes. On the other hand, the bimodal series is controlled by the local stress regime as shown by the concentric gabbroic dikes, probably related to annular fracturing, and the rhyolites restricted to a central piston-like collapse.

The relationship between structures and magma types is directly related to the emplacement processes and generation depth of the magmas. For the bimodal series, eruptions were controlled by the underlying magma chamber and shallow mechanisms of caldera collapse and resurgence, whereas for the carbonatites/kimberlite, emplacement followed deep structural weaknesses that acted as pathways for magma ascent through the lithosphere. Obviously, gabbroic magmas which may originate from the sub-continental lithospheric mantle (section 5) may have followed the same regional weaknesses to rise into the crust before the formation of the magma chamber at a shallow level. Moreover, no difference in geochemistry is known between the external and the inner gabbroic ring dikes. This suggests that both ring dikes were generated from the same magma chamber and that the time interval between both dikes was small and did not allow differentiation.

2.7.3 Emplacement history

Taking account the petrological, structural, geochemical, geophysical and field data, the following stages of emplacement history is proposed (Fig. 14):

Stage 1 (Pre-100 Ma): Magma derived from the sub-continental lithospheric mantle rises in the crust, following pre-existing anisotropies, and produces doming and propagation of ring and radial fractures. A magma chamber is created and is affected by upper crustal contamination. Increasing pressure in the magma chamber results in eruptions of basaltic magmas along concentric fractures to form the gabbroic ring dikes.

Stage 2 (~100 Ma): Reactivation of the pre-existent aulacogenous structure at deep lithospheric levels allows rapid ascension of kimberlitic and carbonatitic magmas, which crosscutt previous gabbroic ring dikes. Fusion of the crust and formation of rhyolites are initiated. A convective low-temperature hydrothermal system is developed, with the magma chamber acting as a heat source. Hydrothermal fluids are channelized into the intensely fractured and permeable central zone created by the doming. Dissolution of the limestone unit leads to karst-collapse brecciation.

Stage 3 (post-100 Ma): Felsic magma rises up along the conical fractures and erupts in violent explosions, probably due to interaction with caldera-lake water. Reopening of ring fractures associated with felsic volcanism led to piston-like collapse along circular fractures.

Stage 4 (post volcanic to present day): Erosion creates circular cuestas represented by nested quartzitic sandstone rings dipping outward. Piston-like collapse preserves the central extrusive felsic facies as well as the hydrothermal karst infilling. The external part of the complex is characterised by deeper basaltic, carbonatitic and kimberlitic intrusive facies.

2.9 Conclusion

The nature and significance of the rocks from the Richat complex has been the subject of prolonged debate and interrogation. Several studies focussed on the description and the understanding of the central breccia, the "analcimolites", the gabbroic rocks or the carbonatites but previous studies have never led to a global understanding of these rocks and the establishment of their possible relationships into a comprehensive model.

In this paper, we showed that the Richat complex is composed of a bimodal tholeiitic series crosscut by undersaturated carbonatitic and kimberlitic dikes and sills. The bimodal series is charaterized by two concentric gabbroic ring dikes and two extrusive rhyolitic centers.

The gabbroic rocks are not part of the Triassic-Jurassic tholeiitic event (CAMP), as previously thought, but are confined to the Cretaceous Richat structure. However, a similar metasomatized sub-continental lithospheric mantle source, affected by subsequent upper crustal contamination, is likely for the origin of the Richat mafic igneous rocks. The remarkable geochemical compatibility with Low-Ti CAMP basalts of West Africa and the fact that both eruptions (CAMP and Richat emplacement) are separated by ~100 Ma also preclude plume intervention in the genesis of the Richat.

The extrusive rhyolitic centers represent the remnant of two volcanic craters aligned along a NE-SW trend, in the elongation axis of the caldera. Previously described as "analcimolites" these rocks appear to be rhyolites and rhyolitic tuffs, altered by low temperature hydrothermal solutions.

Hydrothermal fluids played an important role in the evolution of the Richat complex. They are responsible, notably, for the karst-collapse central megabreccia, the alteration sequence of the rhyolites, the potassic alteration of the gabbros and the stable isotope enrichment in the carbonatites.

The erosional contrast between the central and outer parts of the Richat complex is explained by a piston-like collapse which preserved the central extrusive felsic facies as well as the hydrothermal karst infilling, whereas the external part of the complex is characterised by deeper basaltic, carbonatitic and kimberlitic intrusive facies.

Coexistence of intrusive alkaline and tholeiitic rocks within the same igneous complex shows the importance of the regional structural system on the caldera history. Pre-existing deep structural weaknesses responsible for the elliptical shape of the Richat caldera acted as a pathway for the ascent of magmas and allowed the rise of carbonatitic and kimberlitic melts from the asthenosphere, and basaltic magma from the sub-continental lithospheric mantle. At subsurface levels, eruption of the tholeitiic series was controlled by the underlying magmatic chamber and shallow mechanisms of caldera collapse and resurgence, whereas the carbonatites/kimberlite magmas were emplaced along the old regional anisotropies.

The Richat complex thus provides a rare exposure of magma collected at several distinct depths from the asthenospheric undersaturated alkaline magmas to shallow crustal rhyolites (Fig. 15). Chronology of the magmatic phases suggests a downward propagation of a permeable zone under tensional intra-plate stresses. Direction of vertical propagation during reactivation of lithospheric anisotropies is rarely documented and should be more precisely tested with analogue modelling or age dating on complexes providing magmas from different depths.

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2. References cited

Acocella, V., Funiciello, R., Marotta, E., Orsi, G., and de Vita, S., 2004, The role of extensional structures on experimental calderas and resurgence. Journal of Volcanology and Geothermal Research, v. 129, p. 199-217.

Acocella, V., Korme, T., Salvini, F., and Funiciello, R., 2002, Elliptic calderas in the Ethiopian Rift; control of pre-existing structures. Journal of Volcanology and Geothermal Research, v. 119, p. 189-203.

Bardossy, G., Monod, T., and Pomerol, C., 1963, Decouverte d'analcimolites d'origine endogene dans les Richat (Adrar mauritanien). Comptes Rendus, Acad. Sci. v. 256, p. 3934-3936.

Bayer, M. and Lesquer, A., 1978, Les anomalies gravimétriques de la bordure orientale du craton Ouest-africain. Géométrie d'une suture panafricaine. Bulletin de la Société Géologique de France, v. 7, p. 863-876.

Blanc, P. and Pomerol, C., 1973, Étude pétrographique des roches magmatiques, des filons carbonatés et des analcimolites des Richat. In: Monod, T., Pomerol, C., eds, Contributions a l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy, Sciences de la Terre, Mémoires 28, p. 107-121.

Bosworth, W., Burke, K., and Strecker, M., 2000, Magma chamber elongation as an indicator of intraplate stress field orientation; 'borehole breakout mechanism' and examples from the late Pleistocene to Recent Kenya Rift valley. In: Jessell, M.W., Urai, J. L., eds, Stress, strain and structure; a volume in honor of W. D. Means. Journal of the Virtual Explorer online, v. 2

Boussaroque, J. L., 1975, Étude des analcimolites des Richat (Adrar de Mauritanie) [Ph.D. thesis]. Paris, Université de Paris VI, 105 p.

Bronner, G., 1992, Structure et évolution d'un craton archéen: La dorsale Réguibat occidentale (Mauritanie); tectonique et métallogénie des formations ferrifères. Éditions du Bureau de Recherches Géologiques et Minières, 448 p.

Cailleux, A., Guillemaut, A., and Pomerol, C., 1964, Presence de coesite, indice de hautes pressions, dans l'accident circulaire des Richat (Adrar mauritanien). Compte Rendus Hebdomadaires des Seances de l'Académie des Sciences, v. 258, p. 5488-5490.

Carmichael, I. S. E., 1964, The petrology of Thingmuli, a Tertiary volcano in eastern Iceland. Journal of Petrology, vol. 5, pp.435-460

Castorina, F., Censi, P., Comin-Chiaramonti, P., Piccirillo, E. M., Alcover Neto, A., Gomes, C. B., Ribeiro de Almeida, T. I., Speziale, S., and Toledo, M. C. M., 1997, Carbonatites from eastern Paraguay and genetic relationships with potassic magmatism; C, O, Sr and Nd isotopes. Mineralogy and Petrology, v. 61, p. 237-260.

Comin-Chiaramonti, P., Meriani, S., Mosca, R., and Sinigoi, S., 1979, On the occurrence of analcime in the northeastern Azerbaijan volcanics (northwestern Iran). Lithos, v. 12, p. 187-198.

Cox, K. G. and Hawkesworth, C. J., 1985, Geochemical stratigraphy of the Deccan Traps at Mahabaleshwar, western Ghats, India, with implications for open system magmatic processes. Journal of Petrology, v. 26, p. 355-377.

Dani, N., Formoso, M. L. L., Decarreau, A., and Meunier, A., 2001, Nordstrandite in bauxite derived from phonolite, Lages, Santa Catarina, Brazil. Clays and Clay Minerals, v. 49, p. 216-226.

De Min, A., Piccirillo, E. M., Marzoli, A., Bellieni, G., Renne, P. R., Emesto, M., and Marques, L. S., 2003, The Central Atlantic Magmatic Province (CAMP) in Brazil-petrology, geochemistry, $^{40}Ar/^{39}Ar$ ages, paleomagnetism and geodynamic implications. In: Hames, W. E., McHone, J. G., Renne, P. R., Ruppel, C., eds, The Central Atlantic Magmatic Province: Insights from Fragments of Pangea. AGU, Geophysical Monograph, vol. 136, p. 91-128.

Deckart, K., Bertrand, H., and Liegeois, J.-P., 2005, Geochemistry and Sr, Nd, Pb isotopic composition of the Central Atlantic Magmatic Province (CAMP) in Guyana and Guinea. Lithos, v. 82, p. 289-314.

Deines, P., 1989, Stable isotope variations in carbonatites. In: Bell, K., ed, Carbonatites; genesis and evolution. Unwin Hyman, London, U. K., p. 301-359.

Destombes, J.-P. and Plote, H., 1962, Une structure aberrante dans le Précambrien du Sahara mauritanien; hypothèse nouvelle sur l'origine du Richat (Adrar mauritanien). C. R. somm. Soc. géol. Fr. v. 7, p. 202-203.

Dietz, R. S., Fudali, R., and Cassidy, W., 1969, Richat and Semsiyat domes (Mauritania); not astroblemes. Geological Society of America Bulletin, v. 80, p. 1367-1372.

Eby, G. N., 1975, Abundance and distribution of the rare-earth elements and yttrium in the rocks and minerals of the Oka carbonatite complex, Quebec. Geochimica et Cosmochimica Acta, v. 39, p. 597-620.

Fudali, R. F., 1973, Origin of the analcime-bearing rocks of Richat, In: Monod, T., Pomerol, C., eds, Contributions a l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy, Sciences de la Terre, Mémoires 28, p. 97-105.

Gupta, A. K. and Fyfe, W. S., 1975, Leucite survival; the alteration to analcime. Canadian Mineralogist, v. 13, p. 361-363.

Heatherington, A. L. and Mueller, P. A., 1999, Lithospheric sources of North Florida, USA tholeiites and implications for the origin of the Suwannee Terrane. Lithos, v. 46, p. 215-233.

Irvine, T. N. and Baragar, W. R. A., 1971, A guide to the chemical classification of the common volcanic rocks. Canadian Journal of Earth Sciences, v. 8, p. 523-548.

Jourdan, F., Marzoli, A., Bertrand, H., Cosca, M., and Fontignie, D., 2003, The northernmost CAMP; $^{40}Ar/^{39}Ar$ age, petrology and Sr-Nd-Pb isotope geochemistry of the Kerforne Dike, Brittany, France, In: Hames, W. E., McHone, J. G., Renne, P. R., Ruppel, C., eds, The Central Atlantic Magmatic Province: Insights from Fragments of Pangea. AGU, Geophysical Monograph, vol. 136, p. 209-226.

Keller, J. and Hoefs, J., 1995, Stable isotope characteristics of natrocarbonatites from Oldoinyo Lengai. In: Bell, K., Keller, J., eds, Carbonatite Volcanism: Oldoinyo Lengai and the Petrogenesis of Natrocarbonatites. IAVCEI Proceedings in Volcanology, vol. 4, p. 113–123.

Le Bas, M. J., Le Maitre, R. W., Streckeisen, A., and Zanettin, B. A., 1986, Chemical classification of volcanic rocks based on the total alkali-silica diagram. Journal of Petrology, v. 27, p. 745-750.

Lemberg, J., 1876, Ueber Silicatumwandlungen. Z. Deut. Geol. Ges., v. 28, p. 519-621.

Loubet, M., Bernat, M., Javoy, M., and Allegre, C. J., 1972, Rare earth contents in carbonatites. Earth and Planetary Science Letters, v. 14, p. 226-232.

Mariano, A. N., 1989, Nature of economic mineralisation in carbonatites and related rocks. In: Bell, K., ed, Carbonatites; genesis and evolution. Unwin Hyman, London, U. K., p. 149-176.

Marzoli, A., Renne, P. R., Piccirillo, E. M., Ernesto, M., Bellieni, G., and De Min, A., 1999, Extensive 200-Millon-Year-Old Continental flood basalts of the central Atlantic Magmatic Province. Science, v. 284, p. 616-618.

Matton, G. and Jebrak, M., 2006, A Monteregian-style intrusion in occidental Sahara? Geological and mineralogical associations of Canada annual meeting (GAC-MAC), v. 31, p. 97.

Matton, G., Jebrak, M., and Lee, J. K. W., 2005, Resolving the Richat enigma: Doming and hydrothermal karstification above an alkaline complex. Geology, v. 33, p. 665-668.

McDonough, W. F., Sun, S. S., Ringwood, A. E., Jagoutz, E., and Hofmann, A. W., 1992, Potassium, rubidium, and cesium in the Earth and Moon and the evolution of the mantle of the Earth. Geochimica and Cosmochimica Acta. v. 56, p. 1001-1012.

Mitchell, R. H., 1995, Kimberlites, orangeites, and related rocks. Plenum Press, New York, USA, 410 p.

Monod, T. and Pomerol, C., 1973, Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy, Sciences de la Terre, Mémoires 28, 191 p.

NASA/GSFC/METI/ERSDAC/JAROS and U.S./Japan ASTER Science Team, "Richat Structure, Oudane, Mauritania", ASTER Image acquired October 7, 2000, http://asterweb.jpl.nasa.gov/gallery-detail.asp?name=Richat.

Netto, A. M., Fabre, J., Poupeau, G., and Champemmois, M., 1992, Datations par traces de fissions de la structure circulaire des Richats. Comptes rendus de l'Académie des sciences, v. 314, p. 1179-1186.

Onuonga, I. O., Fallick, A. E., and Bowden, P., 1997, The recognition of meteoric-hydrothermal and supergene processes in volcanic carbonatites, Nyanza Rift, western Kenya, using carbon and oxygen isotopes. Journal of African Earth Sciences, v. 25, p. 103-113.

Pearce, J. A., 1983, Role of sub-continental lithosphere in magma genesis in active continental margins. In: Hawkesworth, C.J., Norry, M.J., eds, Continental basalts and mantle xenoliths, Shiva, Amsterdam, p. 230–249.

Pegram, W. J., 1990, Development of continental lithospheric mantle as reflected in the chemistry of the Mesozoic Appalachian tholeiites, U.S.A. Earth and Planetary Science Letters, v. 97, p. 316-331.

Fabre, J., 1999, In: Quinze jours avec Théodore Monod, Eurêka, v. 41, p. 36-47.

Poupeau, G., Fabre, J., Labrin, E., Azdimoussa, A., Netto, A. M., and Monod, T., 1996, Nouvelles datations par traces de fission de la structure circulaire des Richat (Mauritanie). Mémoires du Service Géologique de l'Algérie, v. 8, p. 231-236.

Reid, D. L. and Cooper, A. F., 1992, Oxygen and carbon isotope patterns in the Dicker Willem carbonatite complex, southern Namibia. Chemical Geology, v. 94, p.2293-2405.

Schmincke, H.-U., 1977, Eifel-Vulkanismus oestlich des gebietes Rieden-Mayen. Fortschritte der Mineralogie, v. 55, p. 1-31.

Sun, S. S. and McDonough, W. F., 1989, Chemical and isotopic systematics of oceanic basalts; implications for mantle composition and processes. In: Saunders, A. D., Norry, M. J., eds, Magmatism in the ocean basins, Geological Society Special Publications, v. 42, p.313-345.

Taylor, D. and MacKenzie, W. S., 1975, A contribution to the pseudoleucite problem. Contributions to Mineralogy and Petrology, v. 49, p. 321-333.

Taylor, S. R. and McLennan, S. M., 1995, The geochemical evolution of the continental crust. Reviews of Geophysics, v. 33, p. 241-265.

Thompson, R. N., Morrison, M. A., Hendry, G. L., and Parry, S. J., 1984, An assessment of the relative roles of crust and mantle in magma genesis: an elemental approach. Phil.

Trans. Royal Soc. Lond., v. A310, p. 549–590.

Trompette, R., 1973a, Les dolérites de l'Adrar de Mauritanie et le problème des analcimolites des Richat. In : Monod, T., Pomerol, C., eds., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie): Nancy, Sciences de la Terre, Mémoires 28, p.83-95.

Trompette, R., 1973b, Le Précambrien supérieur et le Paléozoique inférieur de l'Adrar de Mauritanie: Bordure occidentale du bassin de Taoudenni, Afrique de l'ouest [Ph.D. thesis]. Marseille, Université de Saint Jérôme, 573 p.

Verati, C., Bertrand, H., and Feraud, G., 2005, The farthest record of the Central Atlantic magmatic province into West Africa Craton; precise $^{40}Ar/^{39}Ar$ dating and geochemistry of Taoudenni Basin intrusives (northern Mali). Earth and Planetary Science Letters, v. 235, p. 391-407.

Villeneuve, M. and Cornee, J. J., 1994, Structure, evolution and palaeogeography of the West African Craton and bordering belts during the Neoproterozoic. Precambrian Research, v. 69, p. 307-326.

Weaver, B. L. and Tarney, J., 1981, Lewisian gneiss geochemistry and Archaean crustal development models. Earth and Planetary Science Letters, v. 55, p. 171-180.

Wilson, M., 1989, Igneous petrogenesis; a global tectonic approach, Unwin Hyman, London, United Kingdom, 466 p.

Woolley, A. R. and Kempe D.R.C., 1989, Carbonatites: nomenclature, average chemical compositions and element distiribution. In: Bell, K., ed, Carbonatites; genesis and evolution, Unwin Hyman, London, U. K., p. 1-14.

Woolley, A. R., Rankin, A. H., Elliott, C. J., Bishot, A. C., and Niblett, D., 1984, Carbonatite dykes from the Richat dome, Mauritania, and the genesis of the dome. The Indian Mineralogist, p. 189-207.

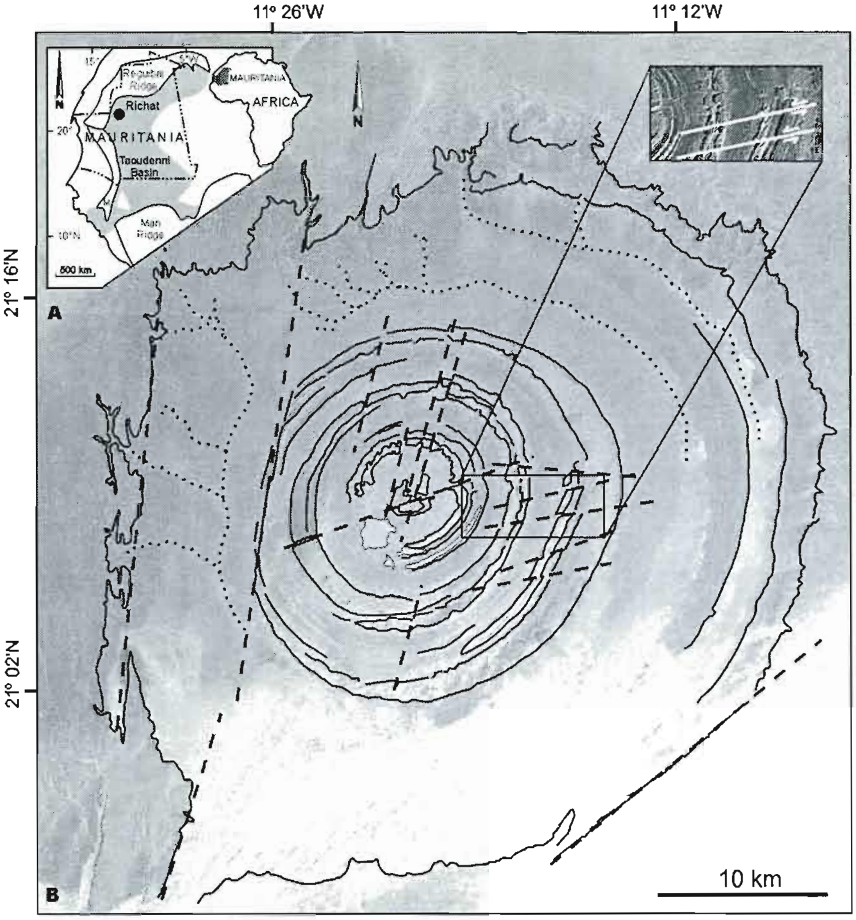

Figure 1. A: Location map for Richat dome. Taoudenni Basin is shaded; (M) Mauritanides. B: Modified satellite image of Richat structure after NASA and U.S./Japan ASTER Science Team, 2000. ENE-WSW faults display sinistral movement as shown by offset of external cuesta (white arrows). Dashed lines represent major faults and fractures; dot-stippled lines, endoreic fluvial streams. The irregular patch in the southwestern central area is the locus of a sebkha.

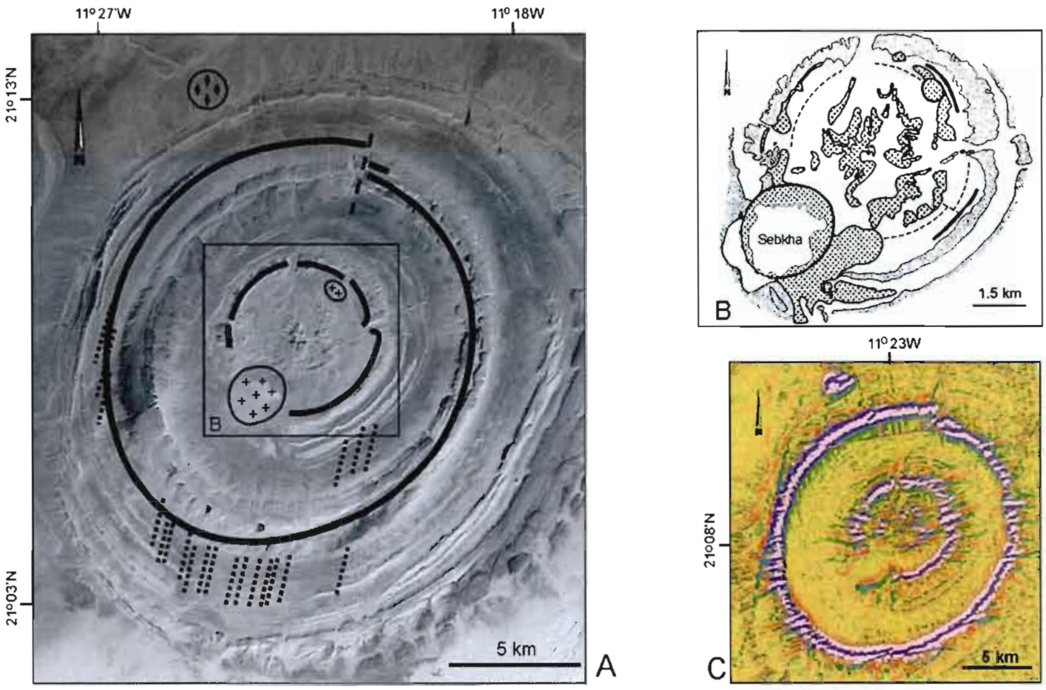

Figure 2. A: Schematic map displaying distribution of magmatic phases: gabbroic ring dikes (thick black lines), volcanic craters (crosses), kimberlitic plug and sills (diamond) and carbonatite dikes (dot-stippled lines). Satellite image after NASA and U.S./Japan ASTER Science Team, 2000. B: Central part of Richat structure. Rhyolitic volcanic basins (crosses) and associated diatremes outcrop to SW and NE of breccia (triangles). Internal gabbroic ring dike (thick black lines) only outcrops in the eastern central part of the Richat. Circular faulting appears in dashed lines. Ring faults are inferred from stratigraphic considerations, repetitions of geological units, satellite imagery and from previous work (Monod and Pomerol, 1973; Boussaroque, 1975). The inner circular ridge is composed of sandstone and represented by a dotted pattern. C: Magnetic survey of the Richat structure showing the kimberlitic plug (northern part) and the gabbroic rings. Note the gap in the southwestern part of the inner gabbroic dike due to the maar system and the displacement of the external gabbroic ring dike by the NNE-SSW fault system in the northeastern part of the map. Magnetic survey courtesy of Ashton Mining Inc.

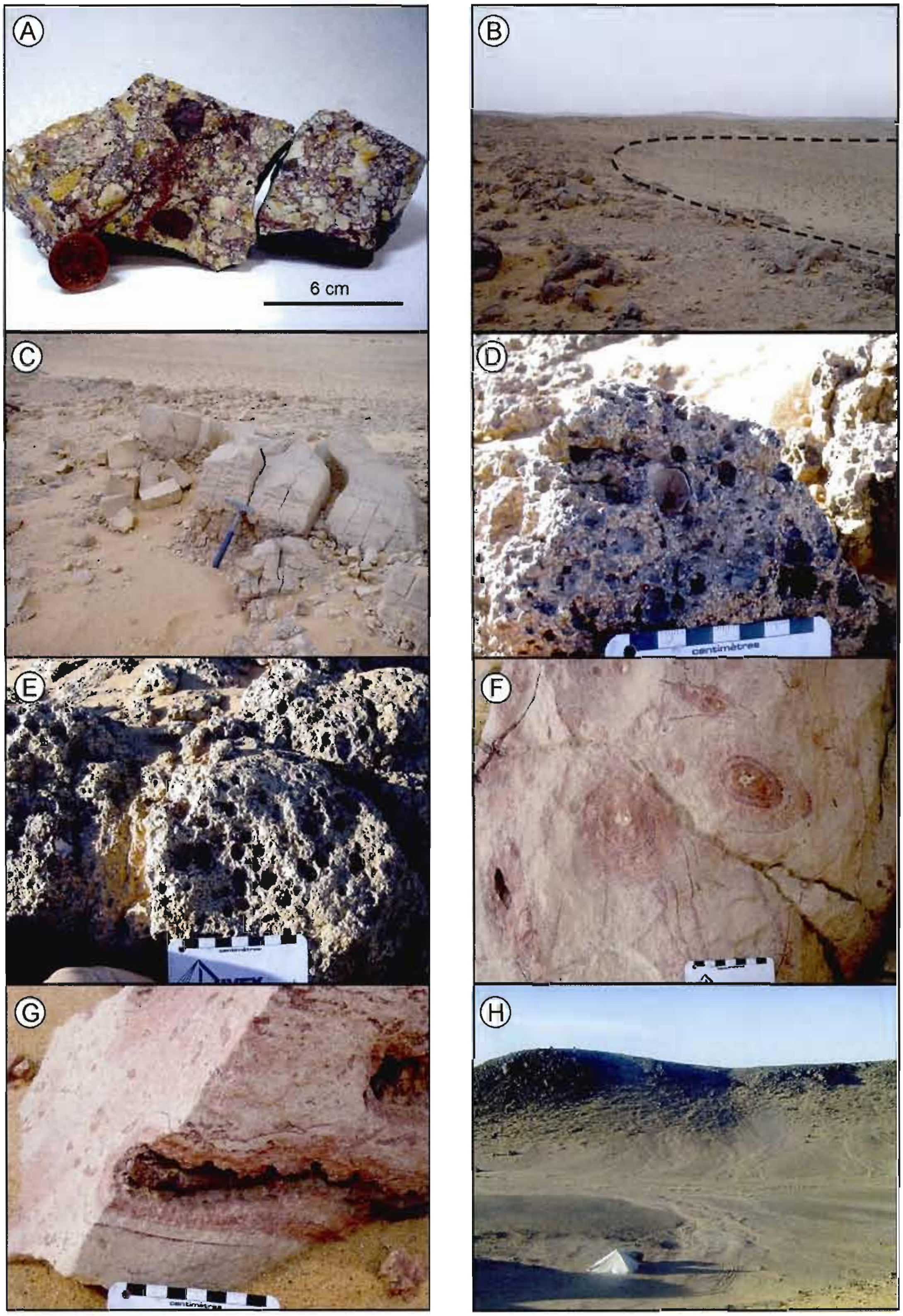
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Figure 3. Photographs of different features from the volcanic rocks of the Richat dome. Scale bar is calibrated in cm. A: Hand specimen of the basal ejecta layer from the southwestern crater rim. B: View looking southwest of the northeastern crater. Dashed line highlights the central depression. C: Sharp contact between the brecciated-conglomeratic unit and the overlying massive quartz-phyric rhyolite (unit 2) at the northeastern crater rim. D: Hand specimen showing the "egg-shell" texture (center) and the analcime glomerocrysts-rich unit (glomerocrysts appear in cream - light beige color on the picture; unit 3). E: Vesicles-rich specimen from the upper part of unit 3. F: Outcrop showing diffusion structures in unit 4. G: Large lithophysae characterize the summital unit. H: The core of the Richat structure is occupied by the high-relief karst-collapse siliceous megabreccia. Tent in foreground for scale.

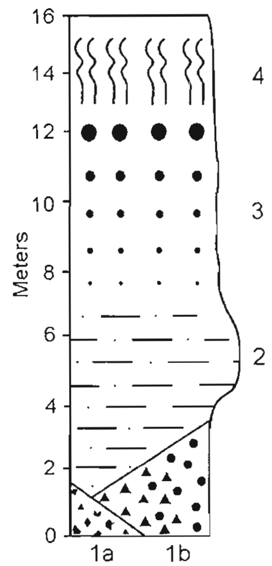
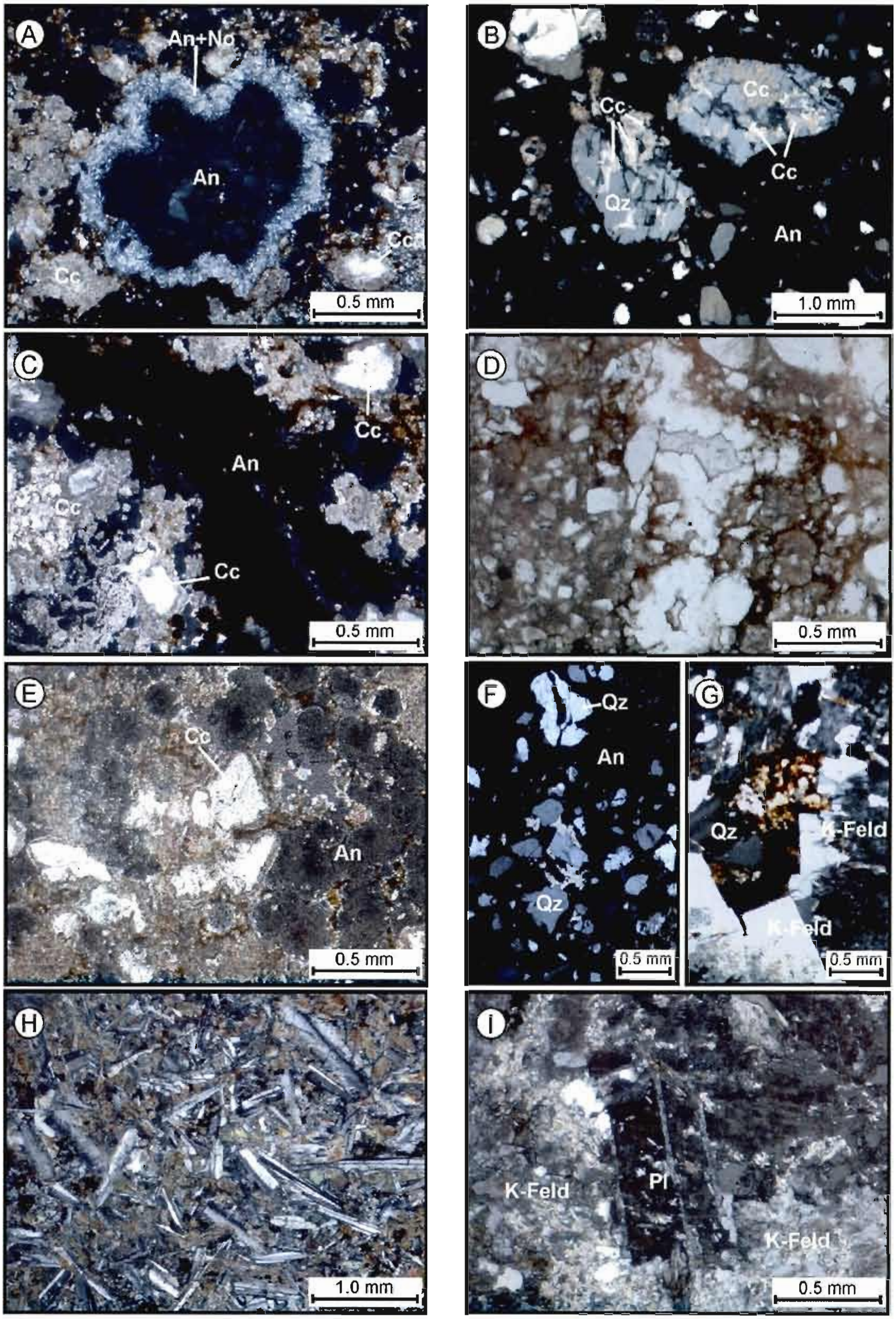


Figure 4. Stratigraphy of volcanic units. 1a: ejecta, 1b: breccia to conglomeratic facies, 2: massive porphyritic quartz rhyolite, 3: "glomerules-supported" unit with vesicles increasing in size and quantity from base to top, 4: lithophysae facies.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K-Feld
0.5 mm

Figure 5. Photomicrographs of features from different Richat lithologies. A: Glomerocrysts of analcime characterized by a core of analcime and a rim of nordstrandite interlaced with analcime. XPL. B: Quartz crystals corroded by calcite in rhyolites. The groundmass is highly anisotropic due to analcime content. XPL. C: Intersticial filling and veinlet of analcime and iron oxide in rhyolite. Remnants of calcite are present. XPL. D: Coarse-grained cavity filling of euhedral analcime. PPL. E: Calcite crystals corroded by analcime in highly altered rhyolites. XPL. F: Quartz crystals corroded by analcime in rhyolites. XPL. G: Secondary neoformed k-feldspar along veinlet wall in altered gabbros. XPL. H: Doleritic texture of the gabbros. Note the lath-shaped sub-euhedral plagioclase crystals with interstitial pyroxene. XPL. I: Lath-shaped plagioclase crystal corroded by sericitized k-feldpars in altered gabbros. XPL. Abbreviations: (An) analcime; (No) nordstrandite; (Qz) quartz; (Cc) Calcite; (K-Feld) k-feldspar; (Pl) Plagioclase; (PPL) Plane-polarized light; (XPL) Cross-polarized light.

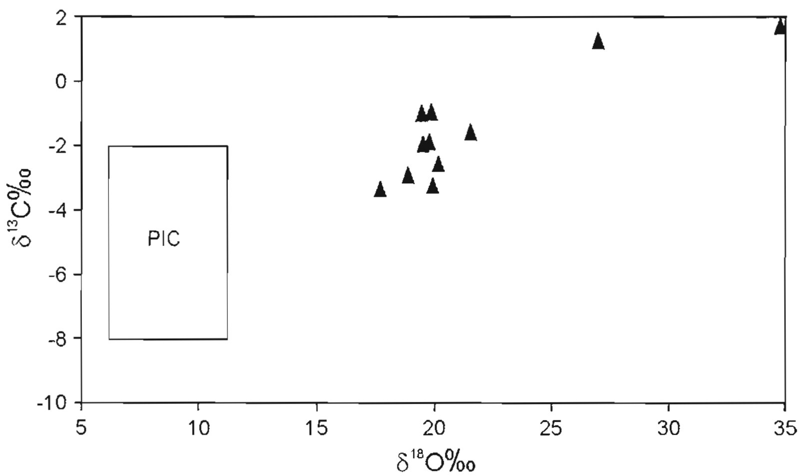

Figure 6. Plot of $\delta^{13}C$ vs $\delta^{18}O$ composition of carbonatite dikes from the Richat dome. PIC: box of primary igneous carbonatites (Deines, 1989).

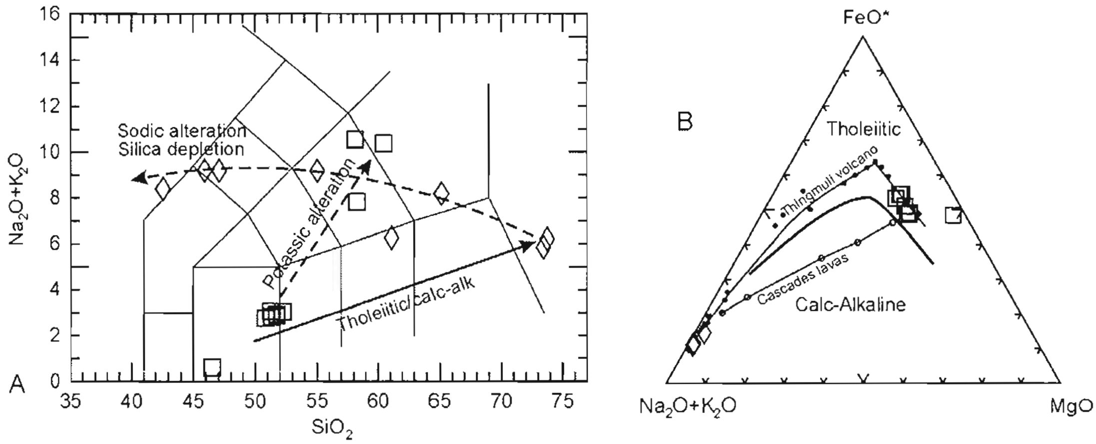

Figure 7. A: Total alkali-silica classification diagram for mafic (squares) to felsic (diamonds) rocks from the Richat complex (after Le Bas et al., 1986). Solid line and dashed lines represent magmatic series and alteration sequences respectively. B: AFM diagram for unaltered gabbros and the freshest rhyolites. Also shown are lava compositions and trends (faint lines) for a typical tholeiitic sequence (thingmuli volcano, Iceland – shown as filled circles – from Carmichael, 1964) and a typical calc-alkaline trend (the average composition of Cascades lavas – shown as open rings – from Carmichael, 1964). The boundary between the calc-alkaline field and the tholeiitic field (thick black line) is from Irvine and Baragar, 1971. FeO*: FeO + Fe_2O_3

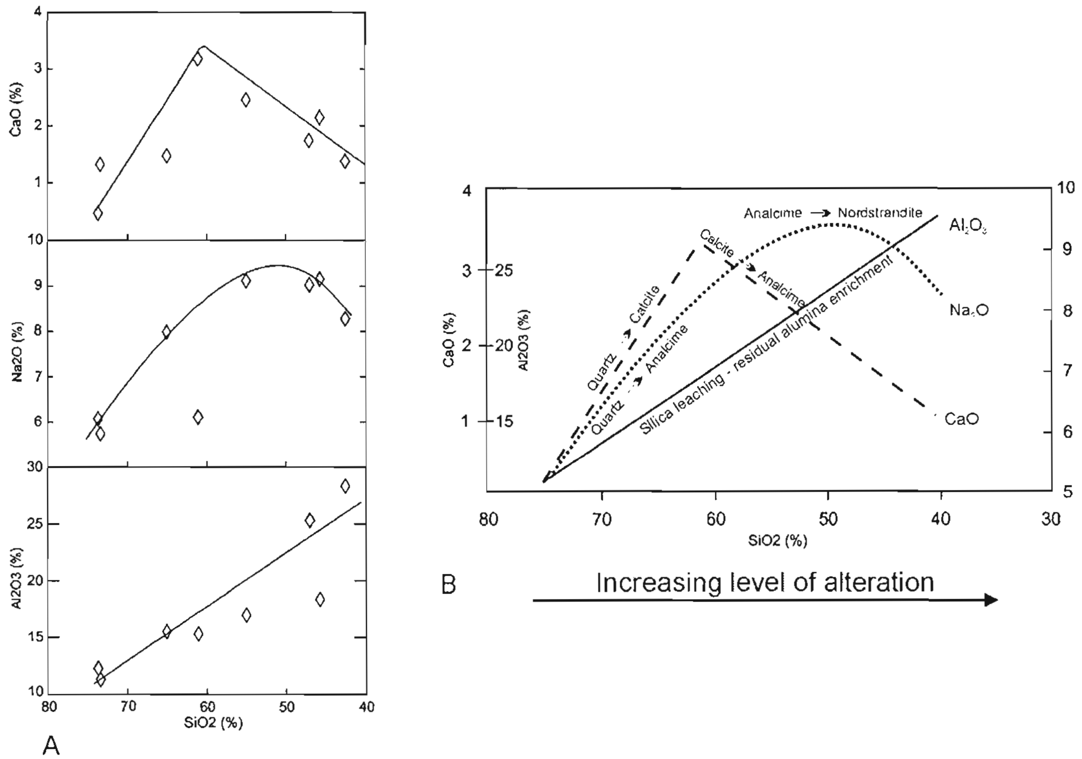

Figure 8. Geochemical plots of volcanic rocks showing CaO, Na_2O and Al_2O_3 against SiO_2. Increasing levels of alteration are shown by decreasing SiO_2.

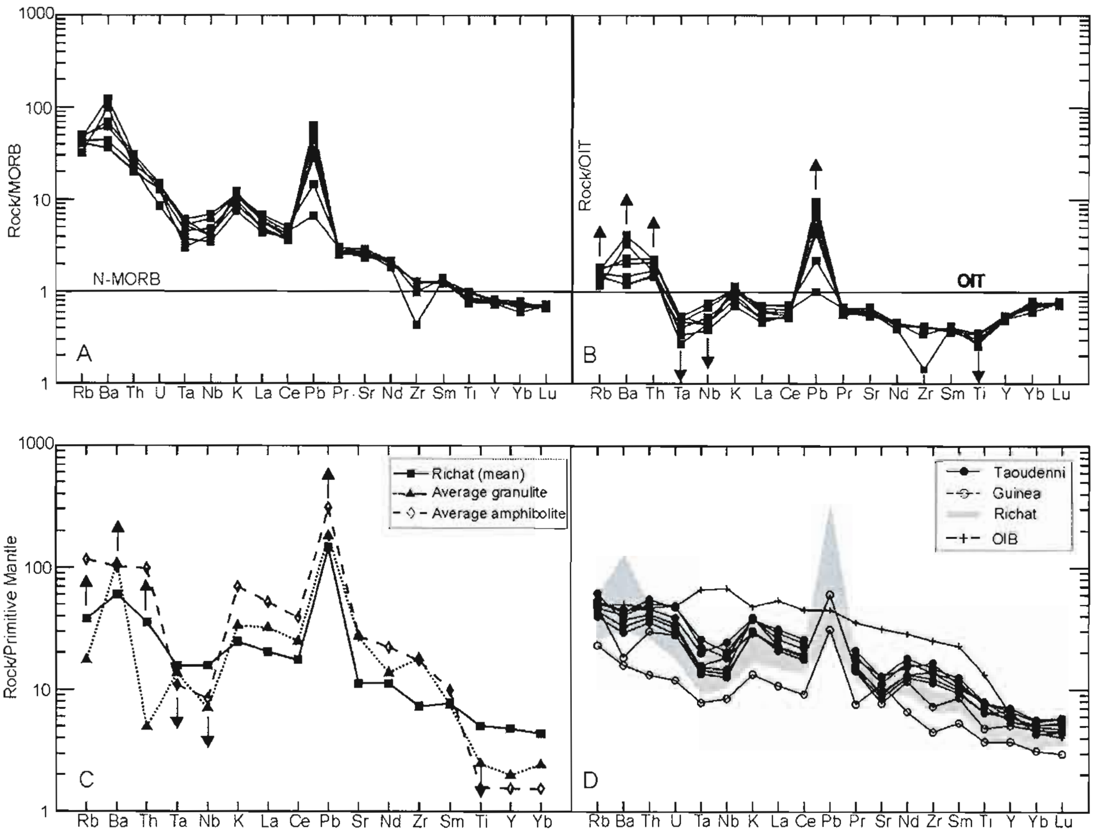

Figure 9. Trace element plots for Richat ring dikes, normalized to (A) Mid-Ocean Ridge Basalts (after Sun and McDonough, 1989) and (B) Ocean Island Tholeiites (after Thompson et al., 1984), respectively. Arrows indicate main enrichments and depletions relative to the ocean-island tholeiites. C: Primitive mantle normalized multi-element diagrams for Richat tholeiites and average amphibolites (upper crust) and granulites (lower crust) (after Weaver and Tarney, 1981). D: Primitive mantle-normalized (after McDonough et al., 1992) multi-element diagrams for Richat tholeiites and CAMP representatives from West Africa (Taoudenni dikes and sills from Mali after Verati et al., 2005 and tholeiitic samples from Guinea after Deckart et al. 2005).

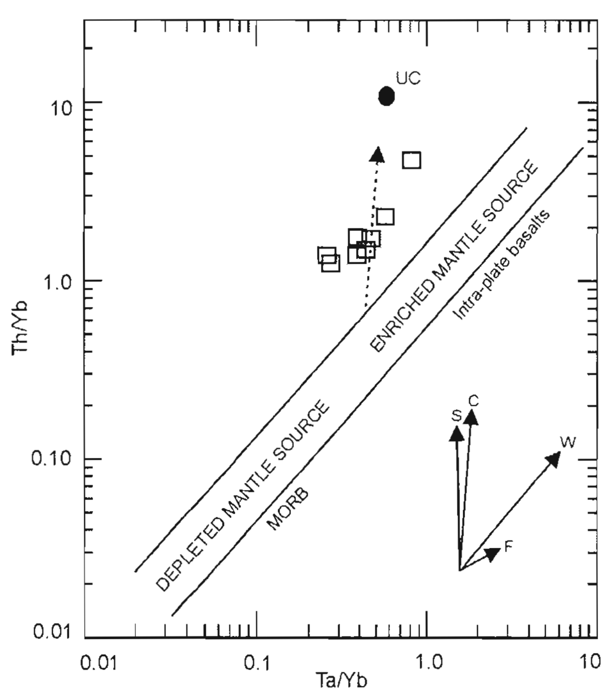

Figure 10. Th/Yb vs Ta/Yb diagram (after Pearce, 1983 and inspired from Wilson, 1989) for Richat gabbros. Vectors shown indicate the influence of subduction components (S), crustal contamination (C), within-plate enrichment (W) and fractional cristalisation (F). UC: Upper crustal composition after Weaver and Tarney (1981). Uncontamined intracontinental plate basalts should plot in the enriched mantle region (Wilson, 1989). Dashed line arrow indicates possible path of Richat gabbros contamination by the upper crust.

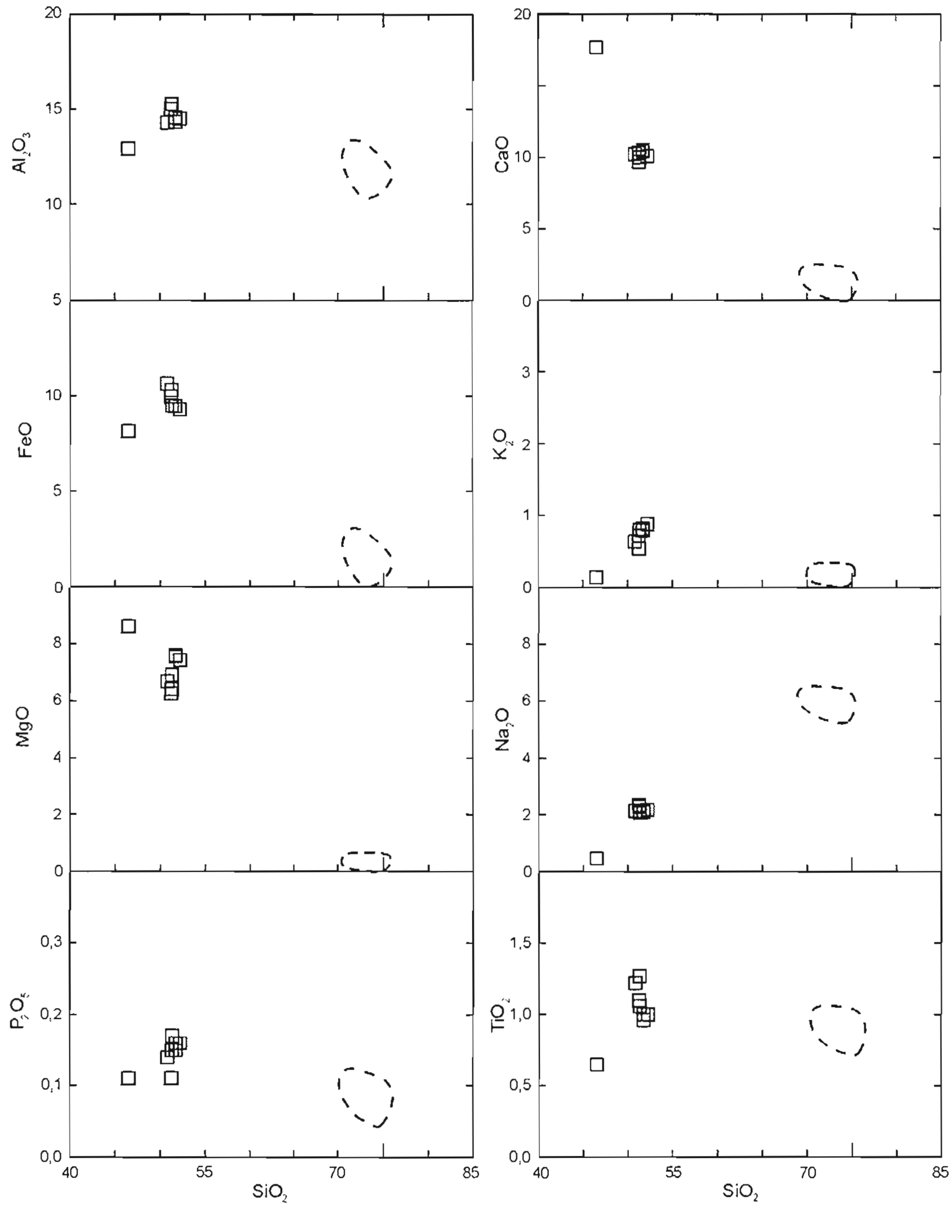

Figure 11. Harker diagrams for gabbros and rhyolites from the Richat complex. Gabbros are represented by squares whereas freshest rhyolites define the dashed line area.

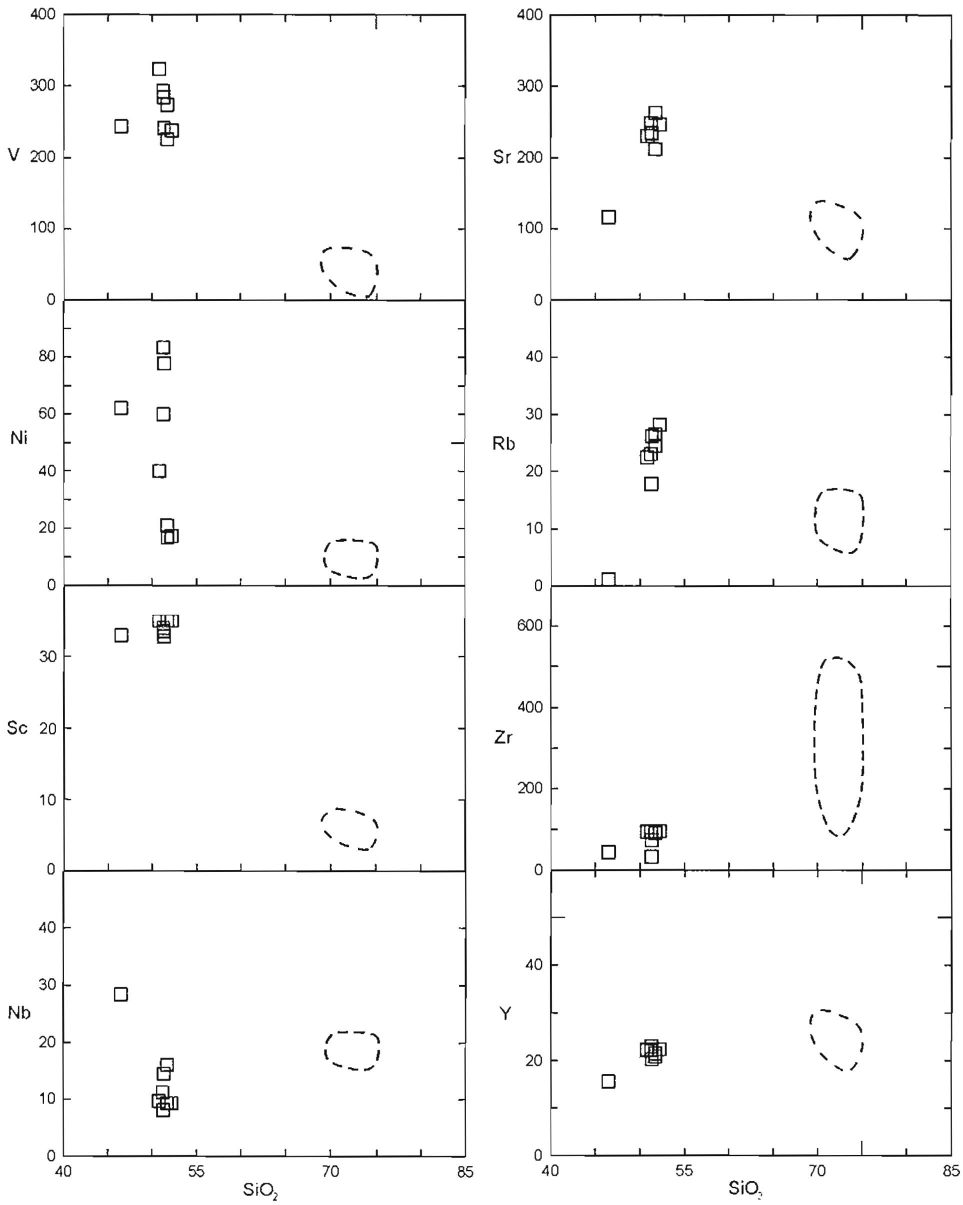

Figure 12. Compatible and incompatible trace elements versus SiO_2 for gabbros and rhyolites from the Richat complex. Symbols as in figure 11.

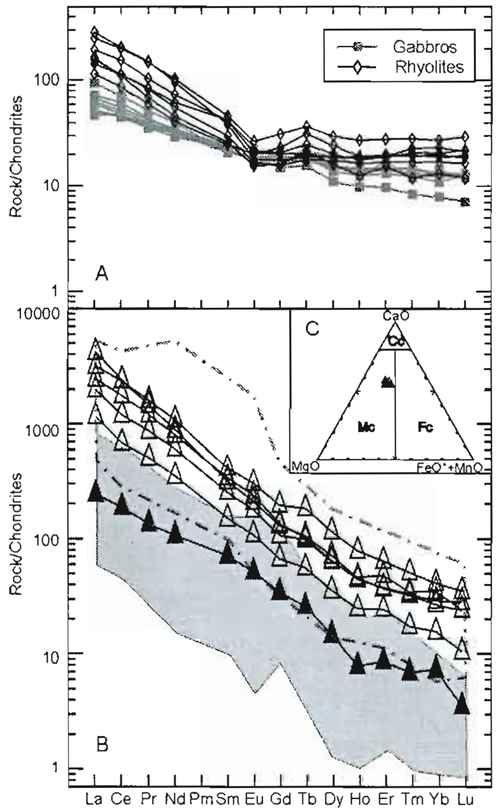

Figure 13. Rare earth elements (REE) normalized to chondritic composition (Sun and McDonough, 1989) for of (A) gabbros and rhyolites and (B) kimberlites and carbonatites. The shaded carbonatite field (pale grey – dashed line) is based on data from Loubet et al. (1972) and Eby (1975) and the shaded field of kimberlites (dark grey) is from Mitchell (1995). Carbonatites are represented by open triangles, kimberlite by solid triangles. (C) Chemical classification of carbonatites using wt % oxides (after Woolley and Kempe, 1989). Abbreviations: (Cc) Calcio-carbonatite; (Mc) Magnesio-carbonatite; (Fc) Ferro-carbonatite. FeO*: $FeO+Fe_2O_3$

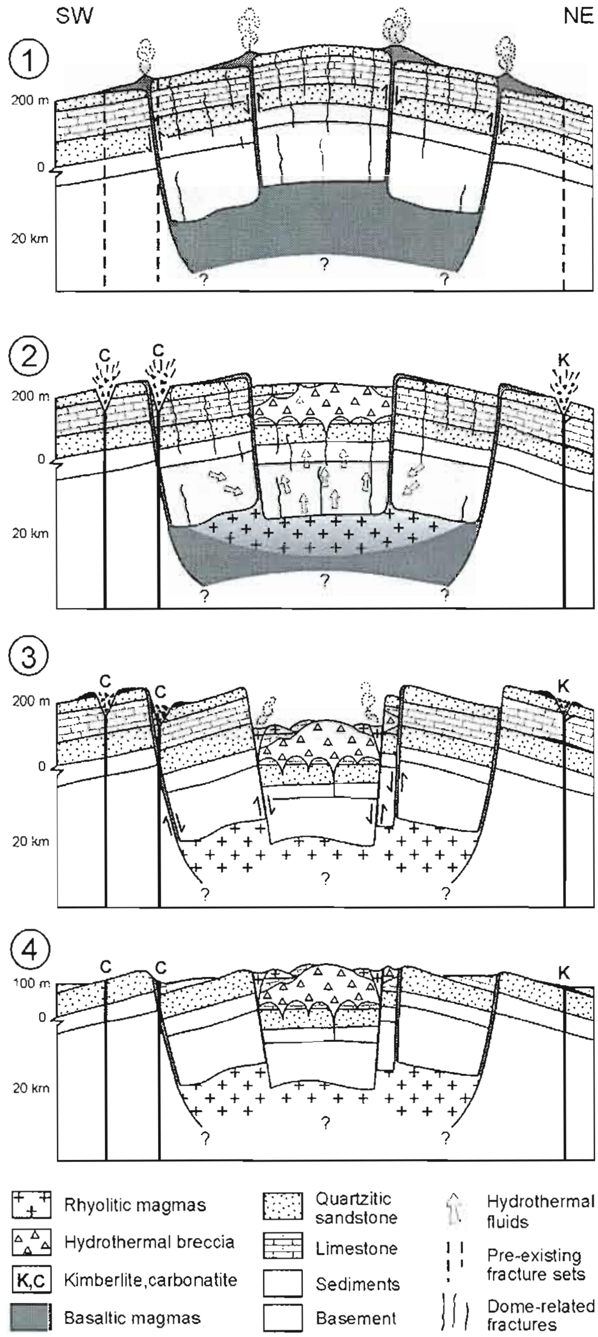

Figure 14. Emplacement history of the Richat complex. See text for explanation.

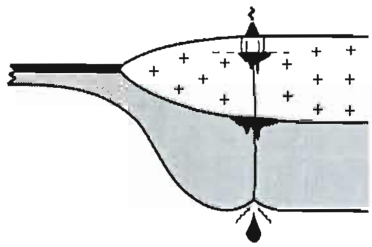

Figure 15. Schematic cross section of the West African craton showing a proposed model for melts generated at different depths and their rise along a deep pre-existing structural anisotropy. Crosses represent continental crust; thick black line, oceanic crust; shaded areas, oceanic and sub-continental lithosphere; white, asthenosphere; dashed line, brittle-ductile transition.

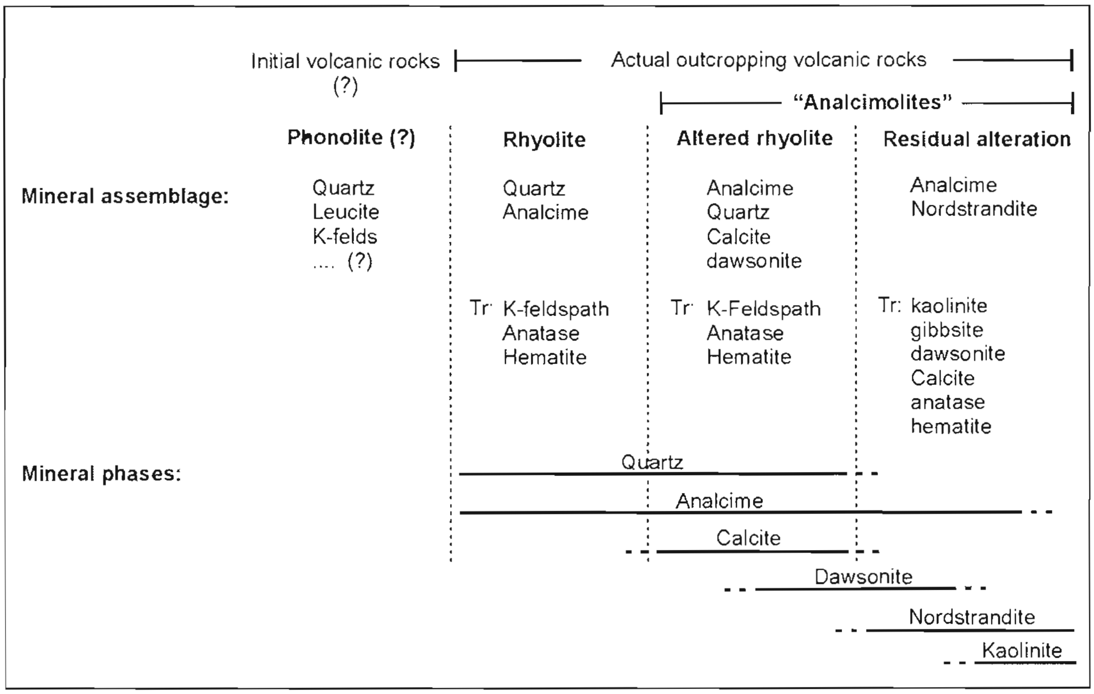

Table 1. Paragenetic mineral sequence of volcanic rocks of the Richat complex.

CHAPITRE III

L'ORIGINE DU RICHAT ET L'HYPOTHÈSE DU POINT CHAUD

3.1 Introduction

Les premiers chercheurs à s'intéresser à la structure du Richat, au début des années soixante, l'ont tout d'abord interprété comme étant le résultat d'un impact météoritique (e.g. Cailleux et al., 1964). Plusieurs études ont par la suite été menées sur le Richat mais elles se limitèrent essentiellement à la description et l'interprétation des différentes roches ignées qu'on y retrouve (Bardossy et al., 1963; Fudali, 1973; Trompette, 1973a; Boussaroque, 1975; Netto, 1992; Poupeau et al., 1996). Très peu de travaux ont été orientés vers l'origine de ce dôme magmatique. Si l'hypothèse de l'astroblème est maintenant écartée (Dietz et al., 1969; Matton et al., 2005), une seule publication suggère actuellement une hypothèse quant à l'origine du Richat; il résulterait du passage d'un point chaud (Piboule et al. dans Eureka, 1999). Ce chapitre a donc pour objectif d'évaluer les différentes options possibles selon les informations disponibles à ce jour.

Dans la première section de ce chapitre, nous allons vérifier l'hypothèse du passage d'un point chaud. Une alternative sera ensuite discutée dans la deuxième section.

3.2 Hypothèse du point chaud

Selon l'hypothèse du point chaud, la région du Richat aurait été située, au Crétacé, au-dessus d'un panache mantellique aujourd'hui localisé sous l'Océan Atlantique et proche des îles du Cap-Vert au sud-ouest de la Mauritanie (Piboule et al. dans Eureka, 1999). Le Richat serait alors l'expression de surface de cette plume mantellique dont l'activité récente aurait entrainé l'émergence des îles du Cap-Vert. Afin de valider cette hypothèse et sachant que l'âge du Richat est établi à environ de 100 Ma (Poupeau et al., 1996; Matton et al., 2005), nous avons calculé la trajectoire d'un point chaud situé sous le Richat à 100 Ma jusqu'à aujourd'hui (0 Ma). La reconstruction du positionnement a été effectuée sur un intervalle de 10 Ma en utilisant le logiciel PointTracker v4c (Scotese, 2002) puis reportée sur un système d'information géographique dans MapInfo Professional 8.0 SCP (Pitney Bowes MapInfo Corporation 2007). La figure 1 montre le résultat de cette reconstruction ainsi que l'hypothèse initiale de Piboule et al. (Eureka, 1999).

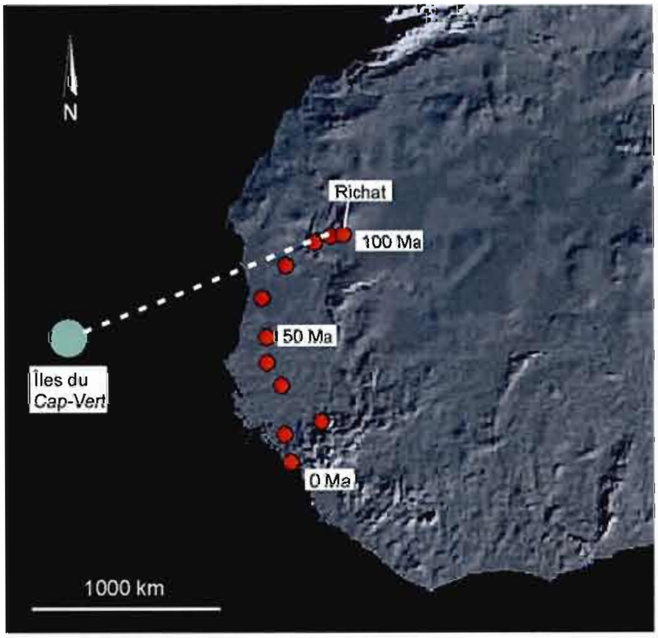

Figure 1. Tracé du passage de l'Afrique de l'Ouest au-dessus d'une plume mantellique selon l'hypothèse d'une migration du Richat au Cap-Vert (tirets blancs). Tracé théorique selon le déplacement calculé de la plaque africaine suivant des intervalles de 10 Ma (cercles rouges). Image satellite Endeavour (STS-99) de l'Afrique de l'Ouest modifiée de NASA/SRTM.

Selon le déplacement calculé de la plaque africaine, un point chaud se situant sous le Richat il y a 100 Ma se trouverait aujourd'hui au niveau de la Guinée. Il apparaît donc évident que la trajectoire calculée n'est pas compatible avec le déplacement d'un point chaud vers le Cap-Vert (figure 1). De plus, la présence d'une plume mantellique actuellement localisée sous la Guinée semble improbable puisqu'aucune plume ou manifestation de plume n'est mentionnée dans la littérature à cet endroit du globe. Conséquemment, le Richat ne proviendrait pas d'une plume mantellique actuellement située sous le Cap-Vert.

3.3 Alternative possible

Si la mise en place du Richat ne provient pas d'un impact météoritique ni du passage d'une plume mantellique, quel processus a pu permettre la montée de magmas alcalins au travers de la lithosphère? Afin d'obtenir des éléments de réponse, nous avons fait un recensement et une compilation régionale des structures, intrusifs, anomalies géophysiques, etc., dans le but de déceler une faiblesse lithosphérique et/ou de trouver une explication à la mise en place du Richat.

Les données proviennent d'images satellites, des cartes géologiques disponibles et d'une revue détaillée de la littérature du secteur. Il est à noter que l'information géologique disponible de la région est pauvre et qu'elle provient principalement des rapports du Bureau de Recherche Géologique et Minière (BRGM) consultés à la Société Géologique de France (SGF) à Paris. Afin de compléter cette information avec des données récentes, des discussions avec des compagnies minières ayant œuvré en Mauritanie et sur le Craton Reguibat ont également été entreprises. Les compagnies ayant acceptées d'échanger de l'information sont Rex Diamond Mining Corporation (Dr. Luc Rombouts) et Ashton Mining Inc. (Dr. Taleb Abdivall).

Les aulacogènes sont des structures de rifts avortés qui s'étendent sur des dizaines, voire des centaines de kilomètres à l'intérieur de continents en voie de séparation (Burke et Wilson, 1976). Ils sont généralement caractérisés par de longues fosses se prolongeant dans les continents à partir de ceintures plissées et contiennent des accumulations de sédiments pouvant couramment atteindre plus de trois fois l'épaisseur des sédiments des séquences contemporaines avoisinantes (Burke, 1977).

La couverture sédimentaire du bassin de Taoudenni présente des formations dont les âges varient du Précambrien Supérieur au Carbonifère (Trompette 1973b). Dans la partie Ouest du bassin, ces formations sédimentaires sont regroupées au sein de trois Supergroupes. Le Supergroupe 1, s'échelonne de ~1000-1100 à ~650 Ma; le Supergroupe 2, de ~650 à ~490 Ma et le Supergroupe 3 de ~490 à ~420 Ma (Trompette 1973b). Chacun de ces Supergroupes reposent en discordance sur le précédent. L'épaisseur moyenne des sédiments du bassin de Taoudenni est de 1000-1500 m (Bronner, 1992). Toutefois, un épaississement maximal des sédiments est observé à l'intérieur d'une fosse étroite caractérisée par plus de 4000 m de dépôt entre la région du Tagant, du Richat et du Tiris (Fig. 2A). Cette fosse se termine abruptement vers le sud-ouest au contact de l'orogène des Mauritanides à l'intérieur desquelles une grande suture est discernable par gravimétrie où un bout d'asthénosphère est resté coincé lors de l'orogénèse Pan-africaine (Guetat, 1981; Le Page et Lécorché, 1991; Villeneuve, 1991). Plusieurs particularités ont été retrouvées le long de cette fosse ainsi que son prolongement dans le socle Reguibat (Fig. 2B). Du sud vers le nord on retrouve notamment;

(1) Le complexe Crétacé du Richat;

(2) Un système de filons annulaires mafiques de 10 km de diamètre. Celui-ci est non-affleurant et a été intercepté par forage (L. Rombouts, 2006, commun. pers.);

(3) Des systèmes de grabens. Le Supergroupe 1, qui remplit la partie inférieure de la fosse, est recoupé par une succession de failles normales actives durant la sédimentation (Bronner, 1992). Ces failles forment une succession de horsts et grabens dans le socle dans la région de Zouerate (Rocci, 1991) et se prolongent dans l'axe de la fosse où des zones de grabens ont également été confirmées par géophysique (L. Rombouts, 2006, commun. pers.);

(4) Une unité structurale énigmatique, la Kédiat Ijil dont on ne connaît ni l'âge, ni le lieu d'origine (Rocci, 1991). Cette unité structurale est caractérisée par une brèche carbonatée contenant de grandes poches d'hématites pures. Un processus de métasomatose suivi d'une désilicification complète expliqueraient la minéralogie de cette brèche, ce qui entraîna certains auteurs à soutenir l'hypothèse d'une cheminée bréchique (*breccia pipe*; Blanchot et Weppe, 1951). La nature bréchique de cette unité, le contenu élevé en hématite, de même que les processus métasomatiques carbonatés avec lessivage de silice intense reportés, nous laissent supposer l'implication d'un système magmatique alcalin;

(5) L'accident Sud-Ijilien. C'est une suture très ancienne coupant le Reguibat en deux et sur laquelle on retrouve la Kedia Ijil à son extrémité sud-est (Bronner, 1992). Cet accident serait Archéen tardif à rejeu post-Protérozoïque inférieur.

(6) Un axe gravimétrique lourd. Une anomalie de gravité positive est reportée dans le Reguibat (Bronner, 1992). Cette anomalie se prolonge sous le couvert du bassin de Taoudenni dans l'axe de la fosse.

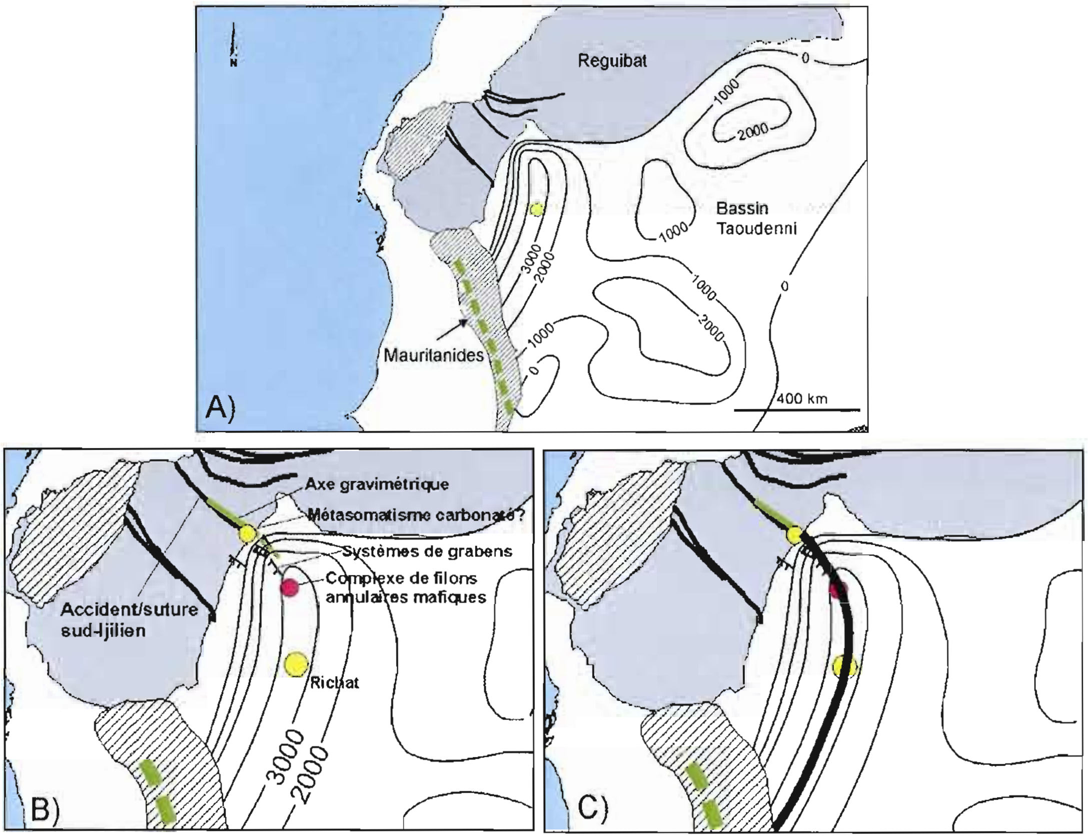

Figure 2. A: Carte des isobathes du socle (en mètres) du bassin de Taoudenni. L'anomalie gravimétrique des Mauritanides est représentée en tirets verts et le Richat par le cercle jaune. B: Agrandissement de la figure 2A illustrant les particularités retrouvées le long de la fosse Tiris-Richat-Tagant. C: Interprétation de la position de l'aulacogène (trait noir épais). Isobathes tirés du rapport No 201, BRGM (Bureau de Recherche Géologique et Minière) 1992.

L'épaississement des séries sédimentaires et la majorité des failles normales sont limités au Supergroupe 1 du bassin de Taoudenni (Fig. 3). En effet, à partir du Supergroupe 2 (650 – 490 Ma) les séries sédimentaires, qui reposent en discordance sur le Supergroupe 1, deviennent régulières et ne présentent aucun épaississement significatif dans la fosse. La période active de subsidence de la fosse est donc restreinte à un âge antérieur à 650 Ma.

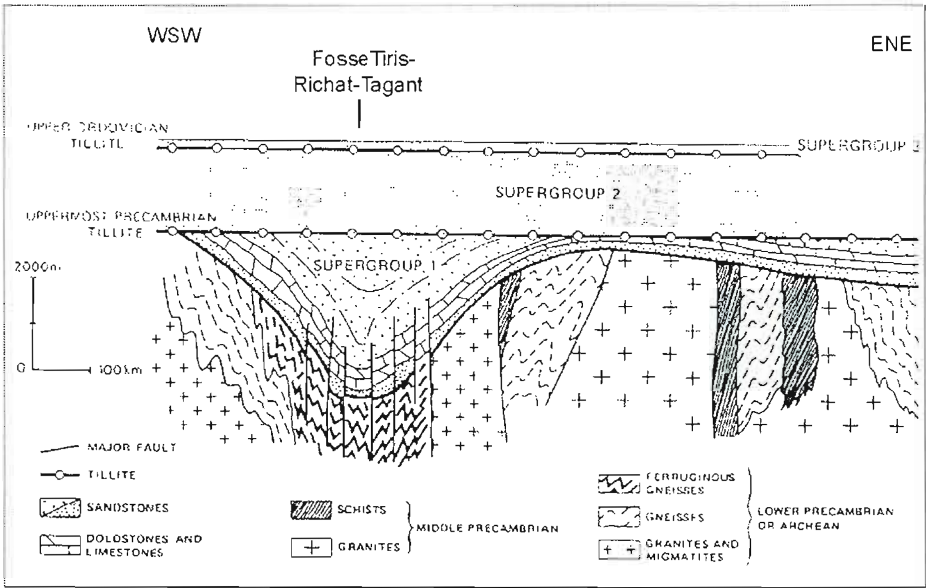

Figure 3. Section schématique de la bordure nord-ouest du bassin de Taoudenni. Modifié de Bronner, 1992.

3.4 Discussion

Notre reconstruction montre que le Richat n'est pas associé à une plume mantellique aujourd'hui localisée sous les îles du Cap-Vert, car le tracé de cette plume l'aurait plutôt située au niveau de la Guinée. De plus, le Richat n'est pas associé à des basaltes de plateaux, une LIP (*Large Igneous Province*), un bombement lithosphérique régional ou des structures radiales de rift, et présente donc peu de critère compatible avec le modèle classique de point chaud (Morgan, 1971; Courtillot et al., 2003; Condie, 2001). Finalement, la géochimie du Richat suggère une origine des magmas, probablement de source sub-lithosphérique métasomatisée (chapitre 2), incompatible avec une plume mantellique profonde.

Le Richat est cependant associé géographiquement à une fosse profonde présentant plusieurs évidences d'un aulacogène. Avec un épaississement des sédiments atteignant les 4000 m, cette fosse contient près de quatre fois l'épaisseur moyenne des séries sédimentaires du bassin de Taoudenni (1000-1500 m; Bronner, 1992). De plus, les anomalies positives de gravimétrie reportées par Bronner (1992) sont typiques des phases initiales de rifting souvent caractérisées par la mise en place de matériel lourd, généralement basaltique (Burke et Wilson, 1976). Le tracé de cet axe gravimétrique est clairement défini dans le Reguibat et sous le couvert du Taoudenni dans la partie nord de la fosse (Fig. 2B). Toutefois, l'auteur ne précise pas si cette anomalie se poursuit sur la portion située entre le Richat et les Mauritanides. La présence de l'accident Sud-Ijilien dans le prolongement de la fosse est également une particularité que l'on retrouve fréquemment associée à la propagation des aulacogènes (et des rifts) favorisant généralement les zones de faiblesse préexistantes telles que les anciennes sutures lithosphériques et les accidents majeurs du socle (Sykes, 1978). La fin abrupte de la fosse au contact d'une ceinture plissée, celle des Mauritanides, est une autre caractéristique classique relative à un aulacogène (Burke, 1977). Dans ce cas-ci, on pourrait supposer un lien entre la période d'océanisation des Mauritanides (*Mauritanides sea*; Villeneuve et Cornée, 1994) et cette fosse qui représenterait alors la partie avortée d'un rift à trois branches.

Le synchronisme entre les processus tectoniques et sédimentaires demeure l'élément le plus caractéristique du développement d'un aulacogène (Chamov et al., 2003). Les travaux de corrélation de Bertrand-Sarfati et al. (1987) définissent la période du rifting Pan-Africain postérieure au Groupe d'Atar (~ 998 – 694 Ma) et considèrent le groupe de l'Assabet el Hassiane (~694-650 Ma), discordant sur ce dernier, comme déposé durant cette période. Il est important de constater que c'est précisément le groupe sédimentaire de l'Assabet el Hassiane qui marque l'épaississement de la fosse Tagant-Richat-Tiris (Bronner, 1992). Nous interprétons donc la fosse du Tagant-Richat-Tiris comme le résultat d'un aulacogène (Fig. 2C).

Puisque Villeneuve et Cornée (1994) contraignent l'épisode du rifting Pan-Africain dans les Mauritanides avant 660 Ma et que les premiers dépôts de l'Assabet el Hassiane se mettent en place vers 694 Ma, l'âge de formation de cet aulacogène serait vraisemblablement situé entre ~694-660 Ma. L'aulacogène du Tagant-Richat-Tiris serait donc Protérozoïque tardif. Il est à noter que dans leurs travaux sur l'évolution et la paléogéographie du Craton Ouest Africain et des ceintures avoisinantes, Villeneuve et Cornée (1994) semblent également considérer la possibilité de la présence d'un aulacogène ouest africain. En effet, une de leur carte synthèse montre un système de bassins subsidents à trois branches dont une des branches passerait par le Richat (dans Villeneuve et Cornée 1994, Fig. 5). On ignore toutefois les raisons qui ont motivé la géométrie de la zone délimitée.

Les intrusions que l'on retrouve le long de la fosse (Richat et filons annulaires mafiques; Fig. 2B) et les grabens détectés par géophysique (L. Rombouts, 2006, commun. pers.) recoupent les séquences sédimentaires de remplissage. Il est donc probable que la zone de faiblesse lithosphérique que constitue l'aulacogène du Tagant-Richat-Tiris ait été réactivée tardivement au cours de son histoire. D'ailleurs, le magmatisme alcalin (carbonatites et kimberlite) que l'on retrouve au Richat témoigne d'une source profonde impliquant un tel processus (l'hypothèse de la plume mantellique étant écartée). Cette zone d'anisotropie préexistante aurait donc servi de conduit pour la remontée du magma sous-saturé au travers la lithosphère.

La mise en place du Richat lors du Crétacé correspond à une période de grandes perturbations tectoniques. Cette période est marquée, notamment, par l'ouverture de l'Océan Atlantique Sud, par une réorganisation majeure des mouvements de plaque et de l'activité des dorsales, par une variation généralisée de l'orientation du stress le long des bordures de plaques péri-Atlantiques, par l'ouverture de l'Océan Atlantique Équatorial, etc., (Klitgord and Schouten, 1986; Savostin et al., 1986; Fairhead and Binks, 1991; Binks and Fairhead, 1992; Guiraud et al., 1992; Janssen et al., 1995; Torsvik et al. 2006). Il s'agit donc d'une période de rééquilibres tectoniques importants impliquant un bouleversement majeur des contraintes. Dans un tel contexte, il est largement documenté que les zones de faiblesse lithosphérique préexistantes agissent comme zone d'accommodation du stress intraplaque (Romine et al., 2000; Spicakova et al., 2000; Cortes et al., 2003; Jacques, 2003; Jelsma et al., 2004; Edel et al., 2007). Cela peut se traduire par des phénomènes de réactivation épisodiques permettant l'ascension de magmas mantelliques le long des anisotropies (Jelsma et al., 2004). Le contexte géologique de mise en place et l'âge du Richat permettent donc de le considérer comme un marqueur magmatique probable de la réactivation tectonique de l'aulacogène Tagant–Richat–Tiris.

3.5 Conclusion

Le Richat n'est pas le résultat d'une plume mantellique aujourd'hui localisée sous les îles du Cap-Vert. De plus, il présente peu d'affinité en relation avec une origine de point chaud. Le complexe alcalin du Richat est cependant situé sur une fosse sédimentaire profonde interprétée comme un aulacogène dont l'âge serait compris entre 694-660 Ma et probablement lié à la phase d'océanisation des Mauritanides. La mise en place du Richat serait issue de la réactivation de l'aulacogène Tagant-Richat-Tiris qui pourrait avoir constitué une zone d'accommodation de stress intraplaque lors des rééquilibres tectoniques mid-Crétacé ayant affectés les bordures continentales de la zone péri-Atlantique. Cette issue sera traitée plus en détail dans le chapitre 4.

3. Références

Bardossy, G., Monod, T., et Pomerol, C. 1963. Découverte d'analcimolites d'origine endogène dans les Richat (Adrar mauritanien). Comptes Rendues de l'Académie des Sciences, v. 256, p. 3934-3936.

Bertrand-Sarfati, J., Moussine-Pouchkine, A., et Caby, R. 1987. Les corrélations du Protérozoïque au Cambrien en Afrique de l'Ouest; nouvelle interprétation géodynamique. Bulletin de la Société Géologique de France, v. 3, p. 855-865.

Binks, R. M., et Fairhead, J. D. 1992. A plate tectonic setting for Mesozoic rifts of West and Central Africa. Tectonophysics, v. 213, p. 141-151.

Blanchot, A., et Weppe, M. 1951. La brèche d'Idjil en Mauritanie et sa genèse probable. Compte Rendu Sommaire des Séances de la Société Géologique de France, v. 11-12, p. 185-187.

Boussaroque, J.-L. 1975. Étude des analcimolites des Richat (Adrar de Mauritanie). Thèse de doctorat, Paris, Université de Paris VI, 105 p.

Bronner, G. 1992. Structure et évolution d'un craton archéen: la dorsale Réguibat occidentale (Mauritanie); tectonique et métallogénie des formations ferrifères. Orléans : BRGM, 448 p.

Burke, K. 1977. Aulacogens and continental breakup. Ann. Rev. Earth Planet. Sci., v. 5, p. 371-396.

Burke, K., et Wilson, J. T. 1976. Hot spots on the Earth's surface. Sci. Amer., v. 235, p. 46-57.

Cailleux, A., Guillemaut, A., et Pomerol, C. 1964. Présence de coésite, indice de hautes pressions dans l'accident circulaire des Richat (Adrar mauritanien). Comptes Rendues de l'Académie des Sciences, v. 258, p. 5488-5490.

Chamov, N. P., Kostyleva, V. V., Veis, A. F., et Gorbachev, V. I. 2003. Late Riphean sedimentation in the central russian aulacogen. Lithology and Mineral Resources, v. 38, p. 458-467.

Condie, K. C. 2001. Mantle plumes and their record in earth history. Cambridge: Cambridge University Press, 305 p.

Cortes, A. L., Soriano, M. A., Maestro, A., et Casas, A. M. 2003. The role of tectonic inheritance in the development of recent fracture systems, Duero Basin, Spain. International Journal of Remote Sensing, v. 24, p. 4325-4345.

Courtillot, V., Davaille, A., Besse, J., et Stock, J. 2003. Three distinct types of hotspots in the Earth's mantle. Earth and Planetary Sciences Letters, v. 205, p. 295-308.

Dietz, R. S., Fudali, R., et Cassidy, W. A. 1969. Richat and Semsityat Domes Mauritania: Not astroblemes. Geological Society of America Bulletin, v. 80, p. 1367-1372.

Edel, J.-B., Schulmann, K., et Rotstein, Y. 2007. The Variscan tectonic inheritance of the Upper Rhine Graben; evidence of reactivations in the Lias, late Eocene-Oligocene up to the Recent. Geologische Rundschau, v. 96, p. 305-325.

Fairhead, J. D., et Binks, R. M. 1991, Differential opening of the Central and South Atlantic Oceans and the opening of the West African Rift System. Tectonophysics, v. 187, p. 191-203.

Fudali, R. F. 1973. Origin of the analcime-bearing rocks of Richat. In Monod, T., et Pomerol, C., éd., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, p. 97-105.

Guetat, Z. 1981. Étude gravimétrique de la bordure occidentale du craton ouest-africain. Thèse de doctorat, Montpellier, Université de Montpellier, 178 p.

Guiraud, R., Binks, R. M., Fairhead, J. D., et Wilson, M. 1992. Chronology and geodynamic setting of Cretaceous-Cenozoic rifting in West and Central Africa. Tectonophysics, v. 213, p. 227-234.

Jacques, J. M. 2003. A tectonostratigraphic synthesis of the Sub-Andean basins: inferences on the position of South American intraplate accommodation zones and their control on South Atlantic opening. Journal of the Geological Society, v. 160, p. 703-717.

Janssen, M. E., Stephenson, R. A., et Cloetingh, S. 1995. Temporal and spatial correlations between changes in plate motions and the evolution of rifted basins in Africa. Geological Society of America Bulletin, v. 107, p. 1317-1332.

Jelsma, H. A., de Wit, M. J., Thiart, C., Dirks, P. H. G. M., Viola, G., Basson, I. J. et Anckar, E. 2004. Preferential distribution along transcontinental corridors of kimberlites and related rocks of Southern Africa. South African Journal of Geology, v. 107, p. 301-324.

Klitgord, K. D., et Schouten, H. 1986. Plate kinematics of the Central Atlantic. In Tucholke, B. E., and Vogt, P. P., éd., The Western North Atlantic region. (The Geology of North America). Geological Society of America, v. M, p. 351-378.

Le Page, A., et Lécorché, J.-P. 1991. La chaîne des Mauritanides. Un exemple d'orogène polyphasé, pan-africain à hercynien, en Mauritanie. In Caruba, R., et Dars, R., éd., Géologie de la Mauritanie, Nice : CRDP, p.79-94.

Matton, G., Jebrak, M., et Lee, J. K. W. 2005. Resolving the Richat enigma: Doming and hydrothermal karstification above an alkaline complex. Geology, v. 33, p. 665-668.

Morgan, W. J. 1971. Convection plumes in the lower mantle. Nature, v. 230, p. 42-43.

Netto, A. M., Fabre, J., Poupeau, G., et Champemmois, M. 1992. Datations par traces de fissions de la structure circulaire des Richats. Comptes Rendus de l'Académie des Sciences, v. 314, p. 1179-1186.

Piboule, M., Fabre, J., et Monod, T. 1999. In Eureka, v. 41, p. 36-47.

PointTracker: PointTracker pour Windows, version 4c. Scotese, C. R., PALEOMAP Project, Arlington, Texas, 2002.

Poupeau, G., Fabre, J., Labrin, E., Azdimoussa, A., Netto, A. M., et Monod, T. 1996. Nouvelles datations par traces de fission de la structure circulaire des Richat (Mauritanie). Mémoires du Service Géologique de l'Algérie, v. 8, p. 231-236.

Rocci, G. 1991. Le socle ancien de Mauritanie : un élément de la croûte continentale stable du craton ouest-africain. In Caruba, R., et Dars, R., éd., Géologie de la Mauritanie, Nice : CRDP, p. 23-38.

Romine, K., Pryer, L., Stuart-Smith, P., Thomas, P., Wakefield, L., et Blevin, J. 2000. Impact of basement architecture on the evolution of the petroleum systems of the Canning Basin, Western Australia. AAPG Bulletin, v. 84, p. 1482-1483.

Savostin, L. A., Sibuet, J.-C., Zonenshain, L. P., Le Pichon, X., et Roulet, M.-J. 1986. Kinematic evolution of the Tethys belt from the Atlantic Ocean to the Pamirs since the Triassic. Tectonophysics, v. 123, p. 1-35.

Spicakova, L., Ulicny, D., et Koudelkova, G. 2000. Tectonosedimentary evolution of the Cheb Basin (NW Bohemia, Czech Republic) between late Oligocene and Pliocene; a preliminary note. Studia Geophysica et Geodetica, v. 44, p. 556-580.

Sykes, L. R. 1978. Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation. Rev. Geophys. Space Phys., v. 16, p. 621-688.

Torsvik, T. H., Müller, R. D., Van der Voo, R., Steinberger, B., et Gaina, C. 2006. Global plate motion frames: toward a unified model. Earth Science Reviews (en revu) – version soumise disponible à: http://www.geodynamics.no/guest/GlobalFrames ESR.pdf

Trompette, R. 1973a. Les dolérites de l'Adrar de Mauritanie et le problème des analcimolites des Richat. In Monod, T., et Pomerol, C., éd., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, p. 83-95.

Trompette, R. 1973b. Le Précambrien supérieur et le Paléozoique inférieur de l'Adrar de Mauritanie: Bordure occidentale du bassin de Taoudenni, Afrique de l'ouest. Thèse de doctorat, Marseille, Université de Saint Jérôme, 573 p.

Villeneuve, M. 1991. Géophysique en Mauritanie. In Caruba, R., et Dars, R., éd., Géologie de la Mauritanie, Nice : CRDP, p. 151-174.

Villeneuve, M., et Cornee, J. J. 1994. Structure, evolution and paleogeography of the West African Craton and bordering belts during the Neoproterozoic. Precambrian Research, v. 69, p. 307-326.

CHAPITRE IV

THE CRETACEOUS PERI-ATLANTIC ALKALINE PROVINCE (PAAP): DEEP MANTLE PLUME ORIGIN OR SHALLOW LITHOSPHERIC BREAK-UP?

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Soumis à *Tectonophysics*

4. Abstract

Although the Early Jurassic initial phases of Atlantic rifting were marked by the tholeiitic basalt flows and dikes of the Central Atlantic Magmatic Province (CAMP), the Cretaceous Era was characterised by widespread alkaline igneous activity on widely separated peri-Atlantic continental margins. Nearly half the peri-Atlantic alkaline rocks dated between 250 to 50 Ma fall within the 125-80 Ma range. This period of intense alkaline activity during the mid-Cretaceous formed the Peri-Atlantic Alkaline Province (PAAP). The mechanisms for the opening of the Atlantic Ocean, as well as for the CAMP and later PAAP magmatic events, have been attributed either to deep mantle plumes or to a combination of tensional forces, lithospheric rifting and structural controls. We describe the PAAP event for the first time and test the plume and structural models.

The PAAP displays evidence of strong structural control and appears to be related to reactivation of pre-existing lithospheric structures throughout the peri-Atlantic zone. Early stages of continental separation and abrupt changes in plate motion were the most likely mechanisms for the reactivation of favourably oriented pre-existing structures. Mid-Cretaceous pulses in alkaline magmatic activity were closely associated with major Atlantic tectonic events. The opening of the South Atlantic, the change in intensity and direction of both seafloor spreading and plate motion, and the consequent fluctuation in intra-plate stresses were responsible for two spikes in peri-Atlantic alkaline magmatism at ~125 Ma and ~85-80 Ma.

The general lack of evidence for a deep hotspot associated with the peri-Atlantic zone seems incompatible with a mantle plume origin. On the other hand, the spatial distribution of the PAAP along passive continental margins suggests an ideal setting for the development of edge-driven convection. We propose that the periodic reactivation of deep-seated pre-existing zones of weakness during the major stages of Atlantic tectonic evolution, combined with coeval asthenospheric upwelling due to edge-driven convection and continental insulation flow, enhanced the ascent of alkaline magmas. This mechanism of shallow, small-scale upwelling during periodic structural reactivation provides a more likely explanation for the PAAP than the commonly proposed deep mantle plume scenario.

4.1 Introduction

The non-uniform distribution of alkaline magmatism in space and time have been linked to specific geological features, such as extensional zones, plate sutures, lithospheric weaknesses and mantle plumes (e.g., Morgan, 1971; Sykes, 1978; Anderson, 1995, 2000; McHone, 2000). Several workers noted that the intense alkaline activity beginning in the Jurassic to Early Cretaceous could be related to the break-up of Pangea (Marsh, 1973; Woolley, 1989; Bailley, 1992). One of the major stages during the dismantling of Pangea was the rifting and opening of the Atlantic Ocean. The initial phase of Atlantic rifting is marked by the 199 Ma large tholeiitic basalt dikes and flows that crop out mainly in eastern North America, western Africa, northeastern South America, and Europe (Marzolli et al., 1999). Marzolli et al. (1999) grouped these occurrences of contemporaneous magmatic activity into the Central Atlantic Magmatic Province (CAMP). After a hiatus of ~70 Ma, numerous igneous complexes developed along the widely separated continental margins regions of eastern North America, Iberia and western Africa (McHone, 2000), as well as the merged margins of western South Africa and eastern South America (Marsh, 1973; Woolley, 1987; 1989; 2001). This magmatism occurred mainly as clusters of plutons and stocks, usually along major lineaments, and it differed from the Triassic-Jurassic magmatic bodies in its alkaline nature (Marsh, 1973; McHone and Butler, 1984; Bédard, 1985; Byerly, 1991; McHone, 1996; McHone, 2000). Nearly half the peri-Atlantic alkaline rocks dated between 250 and 50 Ma fall within the 125–80 Ma interval, and we propose that this intense mid-Cretaceous alkaline activity be defined as the Peri-Atlantic Alkaline Province event (PAAP; Fig. 1).

The 125 – 80 Ma interval coincides with several global events. According to Larson (1991a), a mid-Cretaceous superplume event was responsible for a 50 to 75% increase in the rates of ocean crust formation between 120 and 80 Ma. Moreover, 60% of the world's known oil was generated between the Albian and Turonian Stages (112 to 88 Ma). Increases in sea level, temperature, atmospheric CO_2 concentration, and black shale deposition have also been noted (Larson, 1991a, 1991b). The 120–80 Ma interval also correlates with the Cretaceous normal (CN) superchron. Even if Larson's superplume model is still a matter of debate

(Anderson, 1994; King and Anderson, 1998; Lay, 2005), it is now widely accepted that the mid-Cretaceous is marked by several abnormal events.

The mechanisms for the opening of the Atlantic, in addition to the CAMP and later PAAP events, have been attributed to either deep mantle plumes (Morgan, 1971; Burke and Dewey, 1973; Wilson, 1997; Leitch et al., 1998, and others), or tensional forces, lithospheric rifting and structural controls (Sykes, 1978; Bédard, 1985; McHone, 1996, 2000; Anderson, 2000; Bailey and Woolley, 2005). In the mantle plume model, continental separation and anorogenic magmatism are the results of deep mantle upwelling. Depth of the plume source varies among authors, but upwelling may have originated as deep as the lowermost mantle. Igneous activity would have been the result of adiabatic melting of mantle material, creating hotspots as the superficial expressions of mantle plumes. In the rift model, continental separation and anorogenic magmatism are produced by crustal tectonic mechanisms. The geometry and location of continental rifting would be controlled by the distribution of regional stresses and inherited lithospheric weaknesses. Permissive anorogenic magmatism would thus have been a direct consequence of the attenuation of the lithosphere due to tensional stresses and the structural reactivation of weak zones.

In this paper, we describe the PAAP event for the first time and test the contrasting structural and plume models that attempt to explain this Cretaceous pulse of alkaline magmatism. The spatial and temporal distribution of alkaline rocks will be presented in the geodynamic context of the opening of the Atlantic Ocean to determine whether specific external events were acting on the timing of magmatic activity. We will demonstrate that the peri-Atlantic zone experienced abnormal alkaline activity in the mid-Cretaceous, and that the PAAP displays a strong structural control, apparently related to the widespread reactivation of pre-existing crustal structures during supercontinent break-up. It is thus unnecessary to invoke a deep mantle plume origin.

4.2 Methodology

Occurrences of alkaline rocks were mainly compiled from Woolley (1987; 2001), but additional data were incorporated when available in the literature. The alkaline igneous plutonic/volcanic complexes compiled for this study include carbonatites, syenites, quartz syenites, nepheline syenites (phonolites), ijolites (nephelinites), basanites and feldspathoid-bearing gabbroic rocks. Kimberlitic rocks were excluded because they are mostly limited to deep lithospheric zones that are not uniformly distributed across the PAAP. Geographical coordinates at 100 Ma were calculated using PointTracker v4c (Scotese, 2002). Paleolocations were plotted in a geographical information system on a 100-Ma reconstruction map obtained from Plate Tracker v2.0 (Eldridge et al., 1998 PALEOMAP project) and modified for the purposes of this study. In order to precisely correlate spatial and temporal distributions, we limited the study to alkaline igneous rocks that have been dated and used the best available age data. Dates for North America were mainly taken from Foland and Faul (1977; K-Ar), Eby (1984; Rb-Sr and fission-track), Foland et al, (1986; Ar-Ar), and Baksi (1997; Ar-Ar). Dates for South America were taken from Amaral et al. (1967; K-Ar), Hasui and Cordani (1968; K-Ar), Comin-Chiaramonti and Gomes (1996; fission-track), Velazquez et al. (2004; Ar-Ar), and Gomes et al. (2004; Ar-Ar). Dates for Africa were taken from Van Breemen et al. (1975; Rb-Sr), Rahaman et al. (1984; Rb-Sr), Allsopp and Hargraves (1985; Rb-Sr), and Milner et al. (1995; Ar-Ar; Rb-Sr), and dates for Iberia were taken from Macintyre and Berger (1982; K-Ar) and Solé (2003; K-Ar; Ar-Ar). Whenever more recent or more precise data were available, they took precedence over the sources mentioned above.

Spatial and structural characteristics for the ~125–80 Ma peri-Atlantic alkaline igneous rocks are presented in Table 1. Since hotspot models are often proposed to explain occurrences of alkaline igneous rocks, criteria commonly associated with hotspots are also included (Morgan, 1971; Condie, 2001; Courtillot et al., 2003). If it was possible to associate a hotspot with an alkaline district, or if such an association had been previously suggested, the hotspot was classified according to the categories proposed by Courtillot et al. (2003): primary (originating from the deep D" layer), secondary (originating from the bottom of the transition zone between the lower and upper mantle), or tertiary (more superficial, possibly linked to the asthenosphere as a passive response to lithospheric break-up). Courtillot's categories are defined by the degree to which they meet the following five criteria: (1) a long-lived hotspot track; (2) flood basalts at the origin of the track; (3) a buoyancy flux in excess of 10^3 kg s^{-1}; (4) a consistently high He-isotope ratio; and (5) a significant low shear wave velocity in the underlying mantle. Primary or deep plume-related hotspots must satisfy at least three criteria, secondary hotspots less than three, and hotspots that meet little to no criteria are classified as tertiary hotspots.

4.3 Spatial distribution

The spatial distribution of the PAAP is shown in Figure 1. The number of mid-Cretaceous peri-Atlantic alkaline complexes in the northern and southern hemispheres are almost equal, with 25 and 26 occurrences respectively, although they are not evenly distributed. In North America, they occur mainly in two districts: the New England-Quebec (NEQ) Province in northeastern North America, and the Arkansas Alkaline Province (AAP) in southeastern North America. In South America, the alkaline igneous rocks are found mostly in southeast Brazil and Paraguay, whereas in Africa, the occurrences within the selected timeframe of 125–80 Ma are clustered along the western border of Angola and Namibia, with isolated bodies elsewhere (Nigeria, Senegal and Mauritania). The Late Cretaceous Iberian Alkaline Province (IAP) in western Iberia is also included in this study.

Almost all the intrusive bodies in Table 1 tend to be concentrated along pre-existing lithospheric weaknesses, such as rift systems (aulacogens, grabens, normal faults, etc.) and/or lineaments. These structures represent older features that are not related to Cretaceous phenomena. For the African part of the PAAP, pre-existing structures are mainly related to the Pan-African event. Their orientations are consistently NE-SW, with only slight variations (E-W) among the Guineo-Nubians lineaments. In the Angola-Damaraland Province, alkaline intrusions are concentrated within the Moçamedes Arch and the Damara Belt along major crustal anisotropies such as the Autseib fault, the Omaruru lineament and the Moçamedes lineaments (Fig. 1).

In North America, the alkaline districts (AAP and NEQ) are located along pre-existing structures related to the opening of the Iapetus Ocean during the Late Proterozoic and Early Cambrian. In the AAP and surrounding areas, older structures are mainly oriented NE-SW and NW-SE. Alkaline complexes form a northeast trend along the faults bordering the NE-SW-oriented Reelfoot Rift near or at their intersection with Iapetus-related NW-SE-trending transform faults, such as the Alabama-Oklahoma transform fault (Fig. 1). In the NEQ Province, most mid-Cretaceous alkaline complexes are confined to the WNW-ESE-oriented Ottawa-Bonnechere graben and its extension, although some New-England intrusives are located on less well-defined N-S structures.

In South America, major structural anisotropies provide zones of weakness oriented NE-SW and NW-SE to WNW-ESE. Both sets of lineaments are related to Precambrian basement structures, although the NE-SW features predate the others (Jacques, 2003). Mid-Cretaceous alkaline igneous rocks are found in Paraguay and southeastern Brazil, with one occurrence in eastern Uruguay, and they are collectively referred to herein as the Southeastern Brazil-Paraguay Province. This province displays a strong spatial relationship to the Parana-Chacos Deformation Zone (PCDZ) – a broad zone of intraplate accommodation – and major NW-SE transcontinental faults (Fig. 1). The alkaline occurrences are associated with several structural features, notably the Ponta Grossa Arch (Curitiba Maringa fault zone), which lies in the approximate centre of the PCDZ and is considered to be the symmetrical NW-SE expression of the Moçamedes Arch (Alberti et al., 1999; Comin-Chiaramonti, 1999), as well as the NW-SE Piquiri lineament and the Asunciòn-Sapucai-Villarica graben.

In western Iberia (IAP), zones of weakness are commonly represented by basement-related Variscan lineaments (Late Carboniferous–Early Permian; Wilson et al., 1989; Pinheiro et al., 1996). Alkaline magmatism shows a strong association with offshore fault systems and their extensions onto the continent. Major anisotropies include the NE-SW-oriented Nazare, Tagus and Odemira-Avila faults, as well as the NNW-SSE Sintra-Sines-Monchique fault (Wilson et al., 1989; Pinheiro et al., 1996; Alves et al., 2003).

Alkaline activity at zones of deep-seated weakness was not restricted to the Cretaceous Period. In Namibia and Angola, repeated episodes of alkaline magmatism along deep lineaments occurred from the Proterozoic to Late Cretaceous (Lapido-Loureiro, 1973; Woolley, 1989, 2001; Clemson et al., 1997, 1999), whereas in northeast North America, Triassic and Jurassic intrusive complexes (the White Mountain magma series) are overlapped by the Early Cretaceous NEQ Province (McHone, 1996). Apatite fission track analyses by Hegenberger (1988), Clemson et al. (1997, 1999) and Raab et al. (2002) demonstrated that Namibia's major lineaments were reactivated during the Triassic, Early Cretaceous and Late Cretaceous, suggesting a close relationship between reactivation processes along structural lineaments and coeval magmatic pulses. A long history of reactivation along major zones of weakness has also been recognised in the Southeastern Brazil-Paraguay Province (NE-SW and NW-SE Precambrian lineaments; Jacques, 2003), the Arkansas Alkaline Province and surrounding areas (i.e., the Reelfoot Rift and NW-SE faults; Thomas, 2006; Garrote, 2006), the Iberia Alkaline Province (reactivation of structures within the Variscan basement until Miocene time; Pinheiro et al., 1996; Alves et al., 2003), the Los Island region (Guineo-Nubians lineaments; Guiraud et al., 1985, 1987), the Benue Trough (Guiraud et al., 1992; Binks and Fairhead, 1992), and possibly the Richat area (Matton and Jébrak, 2006). Repeated reactivation of the main PAAP-related structural features thus appears to be a common characteristic among the alkaline districts.

Several processes related to the evolution of the Atlantic Ocean have been proposed to account for such systematic reactivation, including multi-stage opening and rifting of the Atlantic, a shift in the pole of rotation of the peri-Atlantic plates, a change in the rates or geometry of seafloor spreading, a change in plate motions, differential motion between the Central and South Atlantic, and transform directions (see Table 1 for references).

A genetic relationship between mid-Cretaceous peri-Atlantic alkaline districts and fixed deep mantle plumes is difficult to establish. Most of the districts do not satisfy the classic criteria commonly related to deep-seated hotspots, such as radial rift structures, flood or plateau basalts, regional uplift, etc., (Table 1). In fact, Angola-Damaraland and Southeastern Brazil-Paraguay provinces appear to be the only districts that may be associated with a deep hotspot (Tristan da Cunha) owing to the presence of the Early Cretaceous Parana-Angola-Etendeka (PAE) flood basalts and regional uplifts. However, the age progression of magmatic intrusions is inconsistent with a hotspot origin and there are no radial rift structures or fractures. Magmatism appears to be controlled by a set of pre-existing parallel lineaments rather than a radial trend. Moreover, continental margins were already far apart when alkaline igneous bodies formed along each side of the South Atlantic during the Late Cretaceous. Tristan da Cunha's influence, if any, must be limited to Early Cretaceous alkaline magmatism in the Angola-Damaraland and Southeastern Brazil-Paraguay provinces when the margins of Africa and South America were closer or merged together. Furthermore, of the known hotspots with a spatial association to the PAAP igneous complexes or for which a genetic link has been suggested in the literature, only Tristan da Cunha qualifies as primary (deep plume-related) according to the Courtillot et al. (2003) hotspot classification. The others – Great Meteor, Bermuda, Trindade and St. Helena – are tertiary, and thus fairly superficial, more likely a response to lithospheric break-up (Table 1).

4.4 Temporal distribution

To determine whether specific external events were acting on crustal reactivation processes and the timing of magmatic activity, we constructed histograms of the ages of peri-Atlantic alkaline occurrences between 50 Ma and 250 Ma in Figure 2. The heterogeneities in the temporal distributions are striking. The period characterised by initial rifting of the Central Atlantic, which is marked by the CAMP event at ~199 Ma (Marzolli et al., 1999) and initiation of seafloor spreading (180-190 Ma; Foland and Faul, 1977; Klitgord and Schouten, 1986; Torsvik et al., 2006), appears to be almost free of alkaline igneous activity (Fig. 2A). One would expect such initial rifting to be accompanied by several alkaline occurrences due to lithospheric fracturing processes or plume activity, but significant alkaline magmatic activity only occurred in the Early Cretaceous during the initial phase of the Peri-Atlantic Alkaline Province (PAAP). Almost half (46%) of the peri-Atlantic alkaline rocks dated between 250 and 50 Ma fall into the 45-Ma PAAP interval. Two major peaks occur during the PAAP event (Fig. 2B): the first at about 125 Ma with the number of occurrences diminishing until ~90 Ma, and the second at about 85 Ma, again decreasing until the end of the Cretaceous.

To better understand the possible correlation between igneous alkaline episodes and the evolution of the Atlantic Ocean, the intrusives were grouped into populations that underwent the same tectonic evolution. It is well documented that the opening of the South Atlantic lagged behind the Central Atlantic by ~60 Ma (Klitgord and Schouten, 1986; Nürnberg and Müller, 1991, Torsvik et al., 2006), and that the two sections evolved separately because the Equatorial thicker, colder and, therefore, stronger than normal continental lithosphere acted as a "locked zone" (Bonatti, 1996; see Fig. 1). Any attempt to establish links between magmatism and tectonic control should thus distinguish populations north and south of the Equatorial Atlantic.

When these populations are presented separately (Figs. 2C, D), the distributions for both northern and southern alkaline occurrences peak between 110 and 130 Ma, but the number of northern igneous bodies tapers off after 110 Ma, and only the southern PAAP experienced a second pulse of alkaline activity at ~80 Ma.

4.5 Correlations with Atlantic tectonic evolution

Several phases mark the evolution of the Atlantic Ocean. The early stages of continental break-up are thought to have occurred at approximately 199 Ma and were marked by the CAMP event (Marzolli, 1999). Seafloor spreading began in the Central Atlantic soon after that, sometime between 180 to 190 Ma (Foland and Faul, 1977; Klitgord and Schouten, 1986; Torsvik et al., 2006). This was followed by the opening of the South Atlantic and a major reorganisation of plate motion and ridge activity in the Central and South Atlantic sections from about 125 to 130 Ma (Kliford and Schouten, 1986; Fairhead and Binks, 1991; Nürnberg and Müller, 1991; Binks and Fairhead, 1992; Torsvik et al., 2006). The Early Cretaceous is also marked by the propagation of rifting into the North Atlantic and the separation of Greenland and Iberia from North America (Kliford and Schouten, 1986). The opening of the Equatorial Atlantic occurred at ~119 Ma during the early Aptian Stage, and the final separation between Africa and South America was complete by ~100 Ma at the end of the Albian (Mascle et al., 1986; Nürnberg and Müller, 1991; Torsvik et al., 2006). This was followed by another major plate reorganisation event around 85–80 Ma (Rabinowitz and Labrecque, 1979; Kliford and Schouten, 1986; Fairhead and Binks, 1991; Binks and Fairhead, 1992).

Although the CAMP event and the initiation of spreading in the Central Atlantic are not associated with any appreciable pulse of alkaline magmatism, two major tectonic events coincide with the distribution of peri-Atlantic alkaline events. The first peak in Figure 3 marks the opening of the South Atlantic in the Early Cretaceous and represents a period of multiple ocean-continent tectonic perturbations. The opening of the South Atlantic and subsequent seafloor spreading caused ridge reorganisation in the Central Atlantic, a change in plate motion in the Central and South Atlantic, and widespread variations in stress orientations along plate boundaries (Klitgord and Schouten, 1986; Savostin et al., 1986; Fairhead and Binks, 1991; Binks and Fairhead, 1992; Guiraud et al., 1992; Janssen et al., 1995). According to Torsvik et al. (2006), the rate of seafloor spreading peaked in the Central Atlantic at 125 Ma. Several major features were reactivated during that time, including pre-existing Iapetus-related rift zones and aulacogens in northeast North America (NEQ Province), Precambrian lineaments and rift systems in South America, and several Pan-African structures in Africa. Major rift zones developed synchronously over large areas in Africa and South America (Chang et al., 1992; Binks and Fairhead, 1992; Guiraud and Maurin, 1992). Both the Central and South Atlantic zones were affected by this episode (Fig. 2C).

After ~40 Ma of spreading activity, the second Peri-Atlantic peak is marked by another major plate reorganisation (Fig. 3). This event was characterised by a shift in the rotation pole between the South American and African plates, and by rapid changes in plate motion in the Central, Equatorial and South Atlantic around 80 Ma (Marsh, 1973; Prins, 1981; Klitgord and Schouten, 1986; Almeida, 1991; Fairhead and Binks, 1991; Binks and Fairhead, 1992). This second event mainly affected the South Atlantic zone of the PAAP (Fig. 2C), and was caused by the reactivation of major Precambrian lineaments in eastern South America and Pan-African lineaments in Africa.

4.6 Discussion

Almost all PAAP intrusives tend to be localised along zones of pre-existing lithospheric weakness, such as rift systems (aulacogens, grabens, normal faults, etc.) and/or lineaments. Those structures are mainly older features unrelated to Cretaceous phenomena. This indicates that a favourable pre-existing geological setting was in place before the PAAP magmatic event, and also implies that a major event must have occurred to suddenly reactivate widely dispersed structures several hundred million years after their formation.

Reactivation processes along old structural lineaments and coeval pulses in alkaline magmatic activity appear to be closely linked to distinct periods during the evolution of the Atlantic. This relationship may be linked to crustal permeability (a static feature related to the presence of large lithospheric-scale discontinuities) and/or to dynamic processes related to the opening of dilatational zones within the lithosphere. Early stages of continental separation are frequently accompanied by the reactivation of old zones of weakness (Sykes, 1978).

Rifting and the subsequent opening of an ocean basin would lead to the reactivation of fault zones, suture zones and other tectonic boundaries, particularly those near continental margins, thus providing possible pathways for the ascent of magmas.

It is possible to establish the relationships between magmatism and changes in plate motion, as suggested by Bellion and Crovela (1991) and Janssen et al. (1995). In a passive oceanic margin setting, the compressive stress is attributed to ridge push and basal drag induced by decoupling of the lithosphere and the asthenosphere in a direction essentially parallel to absolute plate motion and nearly perpendicular to the accretion axis (Zoback et al., 1986). Changes in the velocity and geometry of spreading would thus imply changes in the intensity and orientation of the stresses acting on plate boundaries. This is supported by Bailey (1992), who argued that the strongest candidate for a periodic reactivation mechanism would be abrupt changes in plate motion or configuration. Early stages of continental separation and abrupt changes in plate motion are therefore likely to reactivate favourably oriented pre-existing anisotropies.

For the peri-Atlantic zone, the opening of the South Atlantic, the change in intensity and direction of both seafloor spreading and plate motion, and the consequent fluctuation in intra-plate stresses were responsible for widespread reactivation processes at ~125 Ma and ~85–80 Ma (Fig. 3). The long and complex history of reactivation of the main PAAP-related structures indicates that these weak zones acted as long-lived intraplate accommodation zones. In the Cretaceous, they played an important role in focusing intraplate stress during the opening of the South Atlantic and subsequent plate motion changes, and in exercising a strong control on the siting of sub-lithospheric mantle melts. The effects of North Atlantic rift propagation on plate interactions within the PAAP study zone are less well-defined in the literature, and we therefore cannot rule out the possibility that such rifting was responsible for certain reactivation processes in the northernmost part of the PAAP. For example, Faure (1996) demonstrated that tensional stresses related to the onset of rifting between Labrador and Greenland were favourably oriented to cause extension and magmatism along the Ottawa-Bonnechere graben during the Early Cretaceous.

In North America, favourably oriented zones of weakness include the Iapetus-related WNW-ESE Ottawa-Bonnechere graben in the NEQ Province, the NE-SW Reelfoot rift system, and NW-SE transform faults in the AAP and its surroundings. In southeast Brazil and Paraguay, intraplate movement was accommodated at a high angle to the main rift axis and concentrated in the NW-SE Parana-Chacos deformation zone (Jacques, 2003). In Africa, reactivated structures were also oriented at high angles to the main rift axis and comprised the NE-SW-trending Pan-African deep crustal lineaments, whereas in western Iberia, the main onshore lineaments associated with magmatism are oriented NE-SW and subparallel to old transform faults.

Initial rifting of the Central Atlantic Ocean appears to be almost free of alkaline magmatism in contrast to the major spike of occurrences observed during the initial rifting of the South Atlantic Ocean (Figure 3). This may be due to the fact that alkaline magmatism in the South Atlantic zone is preferentially focussed along old zones of weakness near the ends of major oceanic transform faults that were active during the early stages of ocean opening (Sykes, 1978). Moreover, stepwise separation of the continents was facilitated by the presence of weak zones oriented at high angles to the South Atlantic spreading centre (Jacques, 2003), which was not the case for weak zones in the northern PAAP where the margins lacked favourably oriented crustal lineaments to accommodate large-scale, lateral block movements.

4.6.1 Are hotspots needed to explain the PAAP?

Although the alkaline nature of the magmas generated during the PAAP event could be interpreted as an indicator of hot spot activity, the high number of synchronous eruptions at widely dispersed sites would require several hotspots to account for such broad-scale adiabatic melting of upwelling mantle material. However, few of the mid-Cretaceous peri-Atlantic alkaline districts, display the criteria typically related to deep hotspots (age progression, flood basalts, uplift, radial structures; Table 1). Further, structural controls and a strong correlation with plate tectonic events are common to all of them, and repeated magmatic activity along old zones of structural weakness on moving plates presents a serious problem for any hotspot model.

As mentioned earlier, Tristan da Cunha appears to be the only clearly plume-related candidate that could have played a role in the formation of the PAAP, but its influence would have been limited to the Early Cretaceous alkaline rocks in the Angola-Damaraland and Southeastern Brazil-Paraguay provinces. Invoking a mantle plume as the origin of the southern PAAP in the Early Cretaceous seems questionable, however, because a second pulse in the Late Cretaceous occurred along the same lineaments and produced an even greater number of intrusions in the absence of any deep-seated hotspot (Figs. 2C, D: second peak). Moreover, several other arguments also rule out Tristan da Cunha's involvement:

- The location of alkaline igneous rocks appears to be controlled by a set of pre-existing parallel lineaments and not related to radial structures;
- Time progressions for carbonatites on either side of the South Atlantic are not coherent (Woolley, 1989);
- Carbonatite activity in Angola and Namibia is repeated at least twice during Earth's history. A spatial association of Proterozoic and Cretaceous occurrences can be observed. (Woolley, 1989; Raab et al., 2002; Bailey and Woolley, 2005);
- Almost all of the alkaline intrusives along the South Atlantic margins are concentrated along tectonic lineaments that were active since at least the Early Mesozoic (Comin-Chiaramonti and Gomes, 1996; Alberti et al., 1999; Raab et al., 2002; Jacques, 2003);
- The Walvis Ridge and Rio Grande Rise, commonly cited as being genetically related to the Tristan da Cunha hotspot track, may instead represent two volcanic lineaments resulting from shear motions arising from a periodic release of stress during the evolution of the South Atlantic (Fairhead and Wilson, 2005);
- The Parana-Angola-Etendeka flood basalt can be explained by continental insulation flow (King and Anderson, 1995).

4.6.2 Superplume or superevent?

This study demonstrates that the mid-Cretaceous (125-80 Ma) period was characterised by abnormal alkaline activity in the peri-Atlantic zone defining the PAAP, and that intrusions were associated with the reactivation of pre-existing structures. Strong structural control on emplacement suggests that magmas were generated in the upper mantle via focussed decompression melting, generating chains of alkaline intrusions (Turcotte and Oxburgh, 1978; Bédard, 1985; Bailey, 1992). A structurally-controlled origin also explains repeated magmatic activity along the same lineaments, the absence of time progression in the magmatic record, the absence of radial structures, and the lack of widespread flood basalts in the PAAP. Moreover, Bailey and Woolley (2005) observed two coeval spikes in kimberlite distribution in Africa at 120 and 80 Ma. Those spikes mark the start and end of the CN superchron, emphasizing the unusual core conditions that prevailed during this period (Bailey and Woolley, 2005). It is thus clear that the mid-Cretaceous marks a time of critical geodynamic perturbations.

Because mantle plumes are not needed to account for the PAAP, it is unlikely that the magma is derived from material rising from the D" layer, as per the superplume model of Larson (1991a). A more plausible explanation – at least for the peri-Atlantic zone – is that of permissive anorogenic magmatism related to asthenospheric upwelling and controlled by lithospheric fracturing. As a response to the multi-stage opening of the Atlantic during the worldwide dismantling of Pangea, PAAP magmatism was more likely the result of a "superevent" than a superplume.

4.6.3 Proposed model

The PAAP is characterised by a combination of strong structural control, an asthenosphere-related origin, and lithospheric break-up phenomena. Any spatially associated hotspots are generally tertiary according to the classification by Courtillot et al. (2003), and thus more likely linked to the asthenosphere. Moreover, the geographical position of the PAAP – along the old continental edges of passive margins – seems to be an ideal setting for the edge-driven convection model (EDC; King and Anderson, 1998, King and Ritsema, 2000). According to these authors, EDC is an instability that occurs at the boundary between thick stable lithosphere and thinner lithosphere. It requires a thick continental or cratonic root adjacent to a thinner plate, such as an oceanic plate. The convection is driven by the vertical temperature variation along the boundary of the thicker continental or cratonic root where small convective cells are created and confined to the upper mantle (King and Anderson, 1998). King and Anderson (1998) were able to demonstrate via simulations that EDC reaches a peak velocity after about 80-100 Ma. Applying this to the Central Atlantic, and assuming that EDC starts when an initial contrast between continental and adjacent oceanic plates is first established, EDC would have started between approximately 200 and 180-190 Ma, which represent the initial break-up and initiation of oceanic spreading respectively. A peak convective velocity would then have occurred between 120 and 80-90 Ma for the region, which corresponds well with the spike of alkaline magmatism for the northern PAAP in the Early Cretaceous (Figure 2C, D). Reactivated lithospheric structures along the edge of continents during that period could thus have acted as pathways for the ascent of magmas produced by EDC.

The upper mantle beneath the southern part of the PAAP could not have undergone EDC during the Early Cretaceous because the Southern Atlantic Ocean was just beginning to open at that time, and instabilities due to a contrast in lithospheric thickness had not yet developed. Nevertheless, the opening of the South Atlantic occurred within a supercontinent break-up context, and while it is widely accepted that insulation by a supercontinent allows for the accumulation of mantle heat (Anderson, 1994, Veevers, 1995), King and Anderson (1995; 1998) have shown that lateral variations in upper mantle temperatures due to supercontinent insulation can also drive convective flow patterns. According to the authors, upwelling during the initial phase of rifting should occur as warm, insulated mantle flows beneath the supercontinent and rises toward the craton edges. This type of continental insulation flow pattern has been suggested to explain the development of the Parana-Angola-Etendeka flood basalt (King and Anderson, 1995), and may account for the spike in alkaline activity observed in this study (Figs. 2C, D). In this scenario, PAAP alkaline magmas would have been generated by the escape of accumulated heat via focused decompression melting at zones of reactivated lithospheric discontinuities. According to the model of King and Anderson (1998), once the continental "edge" migrated some distance from the rift (~1000km), the system could have switched from a continental insulation flow pattern to EDC. This situation would have been achieved offshore of the southern PAAP at about 90 Ma, as calculated using the plate reconstruction software of Eldridge et al., (1998). This may explain the second spike observed in the Late Cretaceous for the southern PAAP distribution (Figure 2D). It is thus proposed that the reactivation of deep-seated pre-existing zones of weakness during the major tectonic stages of Atlantic evolution, combined with Cretaceous asthenospheric upwelling due to EDC and continental insulation flow, enhanced the ascent of alkaline magma to the surface.

The differential opening of the Central and South Atlantic may therefore have allowed the two steps of small-scale asthenospheric upwelling to have simultaneously developed within the same ocean during the PAAP event. During the Early Cretaceous, advanced-stage EDC was peaking in the Central Atlantic while early-stage continental insulation flow was taking place in the South Atlantic (Fig. 4).

4.7 Conclusions

The spatial and temporal distributions of peri-Atlantic alkaline igneous occurrences reveal intense alkaline activity in the mid-Cretaceous, which defines the Peri-Atlantic Alkaline Province (PAAP). PAAP magmatism shows strong structural control and appears to be related to widespread reactivation of pre-existing crustal structures. Repeated magmatic activity along the same lineaments is common. The reactivated weak zones mainly include Pan-African-related structures in Africa, Iapetus-related rift zones and aulacogens in eastern North America, Precambrian lineaments and rift systems in South America, and Variscan-related lineaments in western Iberia.

Two spikes in magmatic activity correlate with two major tectonic events in the evolution of the Atlantic Ocean. The first one at ~125 Ma corresponds to the opening of the South Atlantic, related changes in plate motion, and a peak in seafloor spreading rates in the Central Atlantic. The second spike at ~80 Ma corresponds to a major plate reorganisation and a shift in the pole of rotation between the South American and African plates. Alkaline magmatism thus appears to be closely linked to the geodynamic evolution of the Atlantic Ocean.

The general lack of evidence for deep hotspots associated with the peri-Atlantic zone seems incompatible with a mantle plume origin. The magmas was more likely the result of permissive anorogenic magmatism related to the asthenosphere and controlled by lithospheric fracturing, rather than plume generation at the base of the mantle. We propose that the periodic reactivation of deep-seated pre-existing zones of weakness during the major stages of Atlantic tectonic evolution, combined with coeval asthenospheric upwelling due to edge-driven convection and continental insulation flow, enhanced the ascent of alkaline magmas. This mechanism of shallow, small-scale upwelling during periodic structural reactivation along continental edges provides a more plausible explanation for the PAAP than the commonly proposed deep mantle plume scenario.

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4. References cited

Alberti, A., Castorina, F., Censi, P., Comin-Chiaramonti, P., and Gomes, C. B., 1999, Geochemical characteristics of Cretaceous carbonatites from Angola: Journal of African Earth Sciences, v. 29, p. 735-759.

Allsopp, H. L. and Hargraves, R. B., 1985, Rb-Sr ages and palaeomagnetic data for some Angolan alkaline intrusives: Transactions of the Geological Society of South Africa, v. 88, p. 295-299.

Almeida, F. M., 1991, O alinhamento magmatico de Cabo Frio: Simposio de Geologia do Sudeste, v. 2, p. 423-428.

Alves, T. M., Gawthorpe, R. L., Hunt, D. W., and Monteiro, J. H., 2003, Cenozoic tectono-sedimentary evolution of the western Iberian margin: Marine Geology, v. 195, p. 75-108.

Amaral, G., Bushee, J., Cordani, U. G., Kawashita, K., and Reynolds, J. H., 1967, Potassium-argon ages of alkaline rocks from southern Brazil: Geochimica et Cosmochimica Acta, v. 31, p. 117-142.

Anderson, D. L., 1994, Superplumes or supercontinents?: Geology, v. 22, p. 39-42.

Anderson, D. L., 1995, Lithosphere, asthenosphere and perisphere: Rev. Geophys., v. 33, p. 125-149.

Anderson, D. L., 2000, The thermal state of the upper mantle; no role for mantle plumes: Geophys. Res. Lett., v. 27, p. 3623-3626.

Bailey, D. K., 1992, Episodic alkaline igneous activity across Africa; implications for the causes of continental break-up, *in* Storey, B.C., Alabaster, T. and Pankhurst, R.J., eds., Magmatism and the causes of continental break-up: Geological Society of America , Special Paper, v. 68, p. 91-98.

Bailey, D. K. and Woolley, A. R., 2005, Repeated, synchronous magmatism within Africa: Timing, magnetic reversals, and global tectonics, *in* Foulger, G.R., Natland, J.H., Presnall, D.C., and Anderson, D.L., eds., Plates, plumes, and paradigms: Geological Society of America , Special Paper, v. 388, p. 365-377.

Baksi, A. K., 1997, The timing of Late Cretaceous alkalic igneous activity in the northern Gulf of Mexico basin, southeastern USA: Journal of Geology, 1997, vol. 105, Issue 5, p.629-643.

Bédard, J. H., 1985, The opening of the Atlantic, the Mesozoic New England igneous province, and mechanisms of continental break-up: Tectonophysics, v. 113, p. 209-232.

Bellion, Y. and Crevola, G., 1991, Cretaceous and Cainozoic magmatism of the Senegal Basin (West Africa); a review, *in* Kampunzu, A. B., and Lubala, R. T., eds., Magmatism in extensional structural settings; the Phanerozoic African Plate: Springer-Verlag, Berlin, p. 189-208.

Bernard-Griffiths, J., Gruau, G., Cornen, G., Azambre, B., and Mace, J., 1997, Continental lithospheric contribution to alkaline magmatism; isotopic (Nd, Sr, Pb) and geochemical (REE) evidence from Serra de Monchique and Mount Ormonde complexes: Journal of Petrology, v. 38, p. 115-132.

Binks, R. M. and Fairhead, J. D., 1992, A plate tectonic setting for Mesozoic rifts of West and Central Africa: Tectonophysics, v. 213, p. 141-151.

Bonatti, E., 1996, Anomalous opening of the Equatorial Atlantic due to an equatorial mantle thermal minimum: Earth and Planetary Sciences Letters, v. 143, p. 147-160.

Bosworth, W., 1994, A model for the three-dimensional evolution of continental rift basins, North-east Africa: Geologische Rundschau, v. 83, p. 671-688.

Bronner, G., 1992, Structure et évolution d'un craton archéen: La dorsale Réguibat occidentale (Mauritanie); tectonique et métallogénie des formations ferrifčres: Editions du Bureau de Recherches Géologiques et Miničres, 448 p.

Burke, K. and Dewey, J. F., 1973, Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks: Journal of Geology, v. 81, p. 406-433.

Byerly, G. R., 1991, Igneous activity: Boulder, CO, United States, vol. J, p. 91-108.

Chang, H. K., Kowsmann, R. O., Figueiredo, A. M. F., and Bender, A. A., 1992, Tectonics and stratigraphy of the East Brazil Rift system: an overview: Tectonophysics, v. 213, p. 97-138.

Clemson, J., Cartwright, J. A., and Booth, J., 1997, Structural segmentation and the influence of basement structure on the Namibian passive margin: Journal of the Geological

Society of London, v. 154, p. 477-482.

Clemson, J., Cartwright, J., and Swart, R., 1999, The Namib Rift; a rift system of possible Karoo age, offshore Namibia, *in* Cameron, N. R., and Bate, R. H., eds, The oil and gas habitats of the South Atlantic: Geological Society of London, Special Publications, v. 153, p. 381-402.

Cobbold, P. R., Meisling, K. E., and Mount, V. S., 2001, Reactivation of an obliquely rifted margin, Campos and Santos basins, southeastern Brazil: AAPG Bulletin, v. 85, p. 1925-1944.

Comin-Chiaramonti, P. and Gomes, C. B., 1996, Alkaline magmatism in central-eastern Paraguay; relationships with coeval magmatism in Brazil: EDUSP, 458 p.

Comin-Chiaramonti, P., Cundari, A., DeGraff, J. M., Gomes, C. B., and Piccirillo, E. M., 1999, Early Cretaceous-Tertiary magmatism in eastern Paraguay (western Parana Basin); geological, geophysical and geochemical relationships: Journal of Geodynamics, v. 28, p. 375-391.

Condie, K. C., 2001, Mantle plumes and their record in Earth history: Cambridge, U.K. Cambridge University Press, 306 p.

Cornen, G., 1982, Petrology of the alkaline volcanism of Gorringe Bank (Southwest Portugal): Marine Geology, v. 47, p. 101-130.

Courtillot, V., Davaille, A., Besse, J., and Stock, J., 2003, Three distinct types of hotspots in the Earth's mantle: Earth and Planetary Sciences Letters, v. 205, p. 295-208.

Daly, M. C., Chorowicz, J., and Fairhead, J. D., 1989, Rift basin evolution in Africa; the influence of reactivated steep basement shear zones: Geological Society Special Publications, v. 44, p. 309-334.

Eby, G. N., 1984, Geochronology of the Monteregian Hills alkaline igneous province, Quebec: Geology, v. 12, p. 468-470.

Eldridge, J., Walsh, C.B., and Scotese, C.R., 1998, Plate Tracker for Windows version 2.0, PALEOMAP Project, Arlington, Texas.

Fairhead, J. D. and Binks, R. M., 1991, Differential opening of the Central and South Atlantic Oceans and the opening of the West African Rift System: Tectonophysics, v. 187, p. 191-203.

Fairhead, J. D. and Wilson, M., 2005, Plate tectonic processes in the South Atlantic Ocean; do we need deep mantle plumes? *in* Foulger, G.R., Natland, J.H., Presnall, D.C., and Anderson, D.L., eds., Plates, plumes, and paradigms: Geological Society of America , Special Paper, v. 388, p. 537-553.

Faure, S., Tremblay, A., and Angelier, J., 1996, State of intraplate stress and tectonism of northeastern America since Cretaceous times, with particular emphasis on the New England-Quebec igneous province: Tectonophysics, v. 255, p. 111-134.

Foland, K. A. and Faul, H., 1977, Ages of the White Mountain intrusives; New Hampshire, Vermont, and Maine, USA: American Journal of Science, v. 277, p. 888-904.

Foland, K. A., Gilbert, L. A., Sebring, C. A., and Chen J-F., 1986, $^{40}Ar/^{39}Ar$ ages for plutons of the Monteregian Hills, Quebec; evidence for a single episode of Cretaceous magmatism: Geological Society of America Bulletin, v. 97, p. 966-974.

Francheteau, J. and Le Pichon, X., 1972, Marginal Fracture Zones as Structural Framework of Continental Margins in South Atlantic Ocean: American Association of Petroleum Geologists Bulletin, v. 56, p. 991-1007.

Frese, R. R. B., 1982, Regional anomalies of the Mississippi River aulacogen: Society of Exploration Geophysicists Abstracts with Programs, v. 1, p.295-297.

Garrote, J., Cox, R. T., Swann, C., and Ellis, M., 2006, Tectonic geomorphology of the southeastern Mississippi Embayment in northern Mississippi, USA: Geological Society of America Bulletin, v. 118, p. 1160-1170.

Gomes, C. B., Milan, A., Velazquez, V. F., Riccomini, C., Comin-Chiaramonti, P., Vasconcelos, P. M., and Tassinari, C. C. G., 2004, Magmatismo alcalino na porcao centro-oriental do Paraguay: novos dados geocronologicos para as rochas das provyncias Central e Assuncao: Rev. Fac. Cienc., Univ. Eduardo Mondane, Maputo, Moçambique, Ed. Esp. 1, p. 179-183.

Guiraud, R., Issawi, B., and Bellion, Y., 1985, Les lineaments guineo-nubiens; un trait structural majeur a l'echelle de la plaque africaine: Comptes Rendus de l'Academie des Sciences, v. 300, p. 17-20.

Guiraud, R., Bellion, Y., Benkhelil, J., and Moreau, C., 1987, Post-Hercynian tectonics in northern and western Africa, *in* Bowden, P., and Kinnaird, J., eds, African Geology Reviews: John Wiley and Sons, London, p.433-466.

Guiraud, R., Binks, R. M., Fairhead, J. D., and Wilson, M., 1992, Chronology and geodynamic setting of Cretaceous-Cenozoic rifting in West and Central Africa: Tectonophysics, v. 213, p. 227-234.

Guiraud, R. and Maurin, J.C., 1992, Early Cretaceous rifts of Western and Central Africa: an overview: Tectonophysics, v. 213, p. 153-168.

Hames, W. E., McHone, J. G., Renne, P. R., and Ruppel, C., 2003, The Central Atlantic Magmatic Province; insights from fragments of Pangea: American Geophysical Union, Geophysical Monograph, Vol. 136, 267 p.

Hasui, Y. and Cordani, U. G., 1968, Idades potassio-argonio de rochas do macicos intrusivos mesozoicos do oeste Mineiro e sul de Goias: Congr. Brasil. Geol., 22d, pp. 5.

Hegenberger, W., 1988, Karoo sediments of the Erongo Mountains; their environmental setting and correlation: Communications of the Geological Survey of South West Africa/Namibia, v. 4, p. 51-57.

Jacques, J. M., 2003, A tectonostratigraphic synthesis of the Sub-Andean basins: inferences on the position of South American intraplate accommodation zones and their control on South Atlantic opening: Journal of the Geological Society, v. 160, p. 703-717.

Janssen, M. E., Stephenson, R. A., and Cloetingh, S., 1995, Temporal and spatial correlations between changes in plate motions and the evolution of rifted basins in Africa: Geological Society of America Bulletin, v. 107, p. 1317-1332.

King, S. D. and Anderson, D. L., 1995, An alternative mechanism of flood basalt formation: Earth and Planetary Sciences Letters, v. 136, p. 269-279.

King, S. D. and Anderson, D. L., 1998, Edge-Driven Convection: Earth and Planetary Sciences Letters, v. 160, p. 289-296.

King, S. D. and Ritsema, J., 2000, African hot spot volcanism; small-scale convection in the upper mantle beneath cratons: Science, v. 290, p. 1137-1140.

Klitgord, K. D. and Schouten, H., 1986, Plate kinematics of the Central Atlantic, in Tucholke, B.E. and Vogt, P.P., eds, The Western North Atlantic region. (The Geology of North America): Geological Society of America, v. M, p. 351-378.

Kumarapeli, P. S., 1985, Vestige of Iapetan Rifting in the Craton West of the Northern Appalachians: Geoscience Can., v. 12, p. 54-59.

Kumarapeli, P. S. and Saull, V. A., 1966, The St. Lawrence Valley system: A North American equivalent of the East African Rift Valley system: Canadian Journal of Earth Sciences, v. 3, p. 639-658.

Lapido-Loureiro, F. E., 1973, Carbonatitos de Angola: Memorias e Trabalhos do Instituto de Investigacao Cientifica de Angola, v. 11, 242 p.

Larson, R. L., 1991a, Latest pulse of Earth: Evidence for a mid-Cretaceous superplume: Geology, v. 19, no. 6, p. 547-550.

Larson, R. L., 1991b, Geological consequences of superplumes: Geology, v. 19, no. 10, p. 963-966.

Lay, T., 2005, The deep mantle thermo-chemical boundary layer: the putative mantle plume source, *in* Foulger, G.R., Natland, J.H., Presnall, D.C., and Anderson, D.L., eds., Plates, plumes, and paradigms: Geological Society of America, Special Paper, v. 388,

p. 193-205.

Le Pichon, X. and Hayes, H. E., 1971, Marginal Offsets, Fracture Zones, and the Early Opening of the South Atlantic: Journal of Geophysical Research, v. 76, p. 6283-6293.

Leitch, A. M., Davies, G. F., and Wells, M., 1998, A plume head melting under a rifting margin: Earth and Planetary Sciences Letters, v. 161, p. 161-177.

Liegeois, J. P., Sauvage, J. F., and Black, R., 1991, The Permo-Jurassic alkaline province of Tadhak, Mali; geology, geochronology and tectonic significance: Lithos, v. 27, p. 95-105.

Macintyre, R. M. and Berger, G. W., 1982, A note on the geochronology of the Iberian Alkaline Province: Lithos, v. 15, p. 133-136.

Marquis, R. and Kumarapeli, P. S., 1993, An Early Cambrian deltaic-fluvial model for an Iapetan rift-arm drainage system, southeastern Quebec: Canadian Journal of Earth Sciences, v. 30, p. 1254-1261.

Marsh, J. S., 1973, Relationships between transform directions and alkaline igneous rock lineaments in Africa and South America: Earth and Planetary Sciences Letters, v. 18, p. 317-323.

Marzoli, A., Renne, P. R., Piccirillo, E. M., Ernesto, M., Bellieni, G., and De Min, A., 1999, Extensive 200-Millon-Year-Old Continental flood basalts of the central Atlantic Magmatic Province: Science, v. 284, p. 616-618.

Mascle, J., Marinho, M., and Wannesson, J., 1986, The structure of the Guinean continental margin: implications for the connection between the Central and the South Atlantic Oceans: Geol. Rundsch., v. 75, p. 57-70.

Mascle, J., Blarez, E., and Marinho, M., 1988, The shallow structures of the Guinea and Ivory Coast-Ghana transform margins: their bearing on the Equatorial Atlantic evolution: Tectonophysics, v. 188, p. 193-209.

Matton, G., Jebrak, M., and Lee, J. K. W., 2005, Resolving the Richat enigma: Doming and hydrothermal karstification above an alkaline complex: Geology, v. 33, p. 665-668.

Matton, G. and Jebrak, M., 2006, A Monteregian-style intrusion in occidental Sahara?: Geological and mineralogical associations of Canada annual meeting, v. 31, p. 97.

McHone, J. G., 1996, Constraints on the mantle plume model for Mesozoic alkaline intrusions in northeastern North America: Canadian Mineralogist, v. 34, p. 325-334.

McHone, J. G., 2000, Non-Plume magmatism and rifting during the opening of the central Atlantic Ocean: Tectonophysics, v. 316, p. 287-296.

McHone, J. G. and Butler, J. R., 1984, Mesozoic igneous provinces of New England and the opening of the North Atlantic Ocean: Geol. Soc. Am. Bull., v. 95, p. 757-765.

McHone, J. G., Ross, M. E., and Greenough, J. D., 1987, Mesozoic dykes swarms of eastern North America, *in* Halls, H. C., and Fahrig, W. F., eds, Mafic Dyke Swarms: Geological Society of America, Special Paper, v. 34, p. 279-288.

Milner, S. C., Le Roex, A. P., and O'Connor, J. M., 1995, Age of Mesozoic igneous rocks in northwestern Namibia, and their relationship to continental break-up: Journal of the Geological Society of London, v. 152, p. 97-104.

Moreau, C., Ohnenstetter, D., Demaiffe, D., and Robineau, B., 1996, The Los Archipelago nepheline syenite ring-structure; a magmatic marker of the evolution of the Central and Equatorial Atlantic: Canadian Mineralogist, v. 34, p. 281-299.

Morgan, W. J., 1971, Convection plumes in the lower mantle: Nature, v. 230, p. 42-43.

Nürnberg, D. and Müller, R. D., 1991, The tectonic evolution of the South Atlantic from Late Jurassic to present: Tectonophysics, v. 191, p. 27-53.

Pindell, J. and Dewey, J. F., 1982, Permo-Triassic reconstruction of western Pangea and the evolution of the Gulf of Mexico/Caribbean region: Tectonics, v. 1, p. 179-211.

Pinheiro, L. M., Wilson, R. C. L., Pena dos Reis, R., Whitmarsh, R. B., and Ribeiro, A., 1996, The western Iberia Margin: a geophysical and geological overview, *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G., eds, Proceedings of the Ocean Drilling Program, Scientific Results, v. 149, p.1-23.

Poupeau, G., Fabre, J., Labrin, E., Azdimoussa, A., Netto, A. M., and Monod, Th., 1996, Nouvelles datations par traces de fission de la structure circulaire des Richat (Mauritanie): Mémoires du Service Géologique de l'Algérie, v. 8, p. 231-236.

Prins, P., 1981, The geochemical evolution of the alkaline and carbonatite complexes of the Damaraland igneous province, South West Africa: Annall Universiteit van Stellenbosch, Serie A, v. 3, p. 145-278.

Raab, M. J., Brown, R. W., Gallagher, K., Carter, A., and Weber, K., 2002, Late Cretaceous reactivation of major crustal shear zones in northern Namibia; constraints from apatite fission track analysis: Tectonophysics, v. 349, p. 75-92.

Rabinowitz, P. D. and LaBrecque, J. L., 1979, The Mesozoic South Atlantic Ocean and evolution of its continental margins: Journal of Geophysical Research, v. 84, p. 5973-6002.

Rahaman, M. A., Van Breemen, O., Bowden, P., and Bennet, J. N., 1984, Age migrations of anorogenic ring complexes in Northern Nigeria: Journal of Geology, v. 92, p. 173-184.

Rock, N. M. S., 1982, The Late Cretaceous alkaline igneous province in the Iberian Peninsula, and its tectonic significance: Lithos, v. 15, p. 111-131.

Salvador, A., 1991, Triassic-Jurassic, *in* Salvador, A., ed, The Gulf of Mexico Basin: The geology of North America, Geological Society of America, v. J, p. 131-180.

Savostin, L. A., Sibuet, J.-C., Zonenshain, L. P., Le Pichon, X., and Roulet, M.-J., 1986, Kinematic evolution of the Tethys belt from the Atlantic Ocean to the Pamirs since the Triassic: Tectonophysics, v. 123, p. 1-35.

Scotese, C. R., 2002, PointTracker for Windows version 4c, PALEOMAP Project, Arlington, Texas.

Sims, P. K., 2002, The Mesoproterozoic; a time of change in tectonic style and magma types in the North American continent, Geological Society of America, Abstracts with Programs, v. 34, no. 6, p.43.

Sole, J., Pi, T., and Enrique, P., 2003, New geochronological data on the Late Cretaceous alkaline magmatism of the northeast Iberian Peninsula: Cretaceous Research, v. 24, p. 135-140.

Sykes, L. R., 1978, Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation. Rev. Geophys. Space Phys., v. 16, p. 621-688.

Thomas, W. A., 2006, Tectonic inheritance at a continental margin: GSA Today, v. 16, p. 4-11.

Torsvik, T. H., Müller, R.D., Van der Voo, R., Steinberger, B., and Gaina, C., 2006, Global plate motion frames: toward a unified model, Earth Science Reviews (in review) - submitted version available at: http://www.geodynamics.no/guest/GlobalFrames ESR.pdf

Tremblay, A., Long, B., and Masse, M., 2003, Supracrustal faults of the St. Lawrence rift system, Quebec; kinematics and geometry as revealed by field mapping and marine seismic reflection data: Tectonophysics, v. 369, p. 231-252.

Turcotte, D. L. and Oxburgh, E. R., 1978, Intra-plate volcanism: Philos. Trans. R. Soc. London, Ser. A, v. 288, p. 561-579.

Van Breemen, O., Hutchinson, J., and Bowden, P., 1975, Age and origin of the Nigerian Mesozoic granites; a Rb-Sr study: Contributions to Mineralogy and Petrology, v. 50, p. 157-172.

Veevers, J. J., 1995, Emergent, long-lived Gondwanaland vs. submergent, short-lived Laurasia; supercontinental and Pan-African heat imparts long-term buoyancy by mafic underplating: Geology, v. 23, p. 1131-1134.

Velazquez, V. F., Gomes, C. B., Riccomini, C., Comin-Chiaramonti, P., Brumatti, M., Vasconcelos, P. M., and Tassinari, C. C. G., 2004, Caracterysticas isotopicas e idades $^{40}Ar/^{39}Ar$ para as rochas alcalinas da Provyncia Misiones, Paraguai meridional.: Rev. Fac. Cienc., Univ. Eduardo Mondane, Maputo, Moçambique, Ed. Esp. 1, p. 198-200.

Villeneuve, M. and Cornee, J. J., 1994, Structure, evolution and palaeogeography of the West African Craton and bordering belts during the Neoproterozoic: Precambrian Research, v. 69, p. 307-326.

Williams, H., 1979, Appalachian Orogen in Canada: Canadian Journal of Earth Sciences, v. 16, p. 797-807.

Wilson, M., 1997, Thermal evolution of the Central Atlantic passive margins: continental break-up above a Mesozoic super-plume: J. Geol. Soc. London, v. 154, p. 491-495.

Wilson, R. C. L., Hiscott, R. N., Willis, M. G., and Gradstein, F. M., 1989, The Lusitanian Basin of west-central Portugal; Mesozoic and Tertiary tectonic, stratigraphic, and subsidence history, *In* Tankard, A. J., ed., Extensional tectonics and stratigraphy of the North Atlantic margins: AAPG Memoir, v. 46, p. 341-361.

Wilson, M. and Guiraud, R., 1992, Magmatism and rifting in Western and Central Africa, from Late Jurassic to Recent times: Tectonophysics, v. 213, p. 203-225.

Withjack, M. O., Schlische, R. W., and Olsen, P. E., 1998, Diachronous Rifting, Drifting, and Inversion on the Passive Margin of Central Eastern North America: An Analog for Other Passive Margins: AAPG Bulletin, v. 82, p. 817-835.

Woolley, A. R., 1987, Alkaline rocks and carbonatites of the world; Part 1, North and South America: Br. Mus. (Nat. Hist.), London, U. K., 216 p.

Woolley, A. R., 1989, The spatial and temporal distribution of carbonatites, *in* Bell, K., ed, Carbonatites; genesis and evolution: Unwin Hyman, London, U. K., p. 15-37.

Woolley, A. R., 2001, Alkaline rocks and carbonatites of the world; Part 3, Africa: The Geological Society London, 372 p.

Zoback, M. L., Nishenko, S. P., Richardson, R. M., Hasegawa, H. S., and Zoback, M. D., 1986, Mid-plate stress, deformation, and seismicity, *in* Vogt, P.R. and Tuckholke, B.E., eds, Western North Atlantic Region: The geology of North America, Geological Society of America, v. M, p. 297-312.

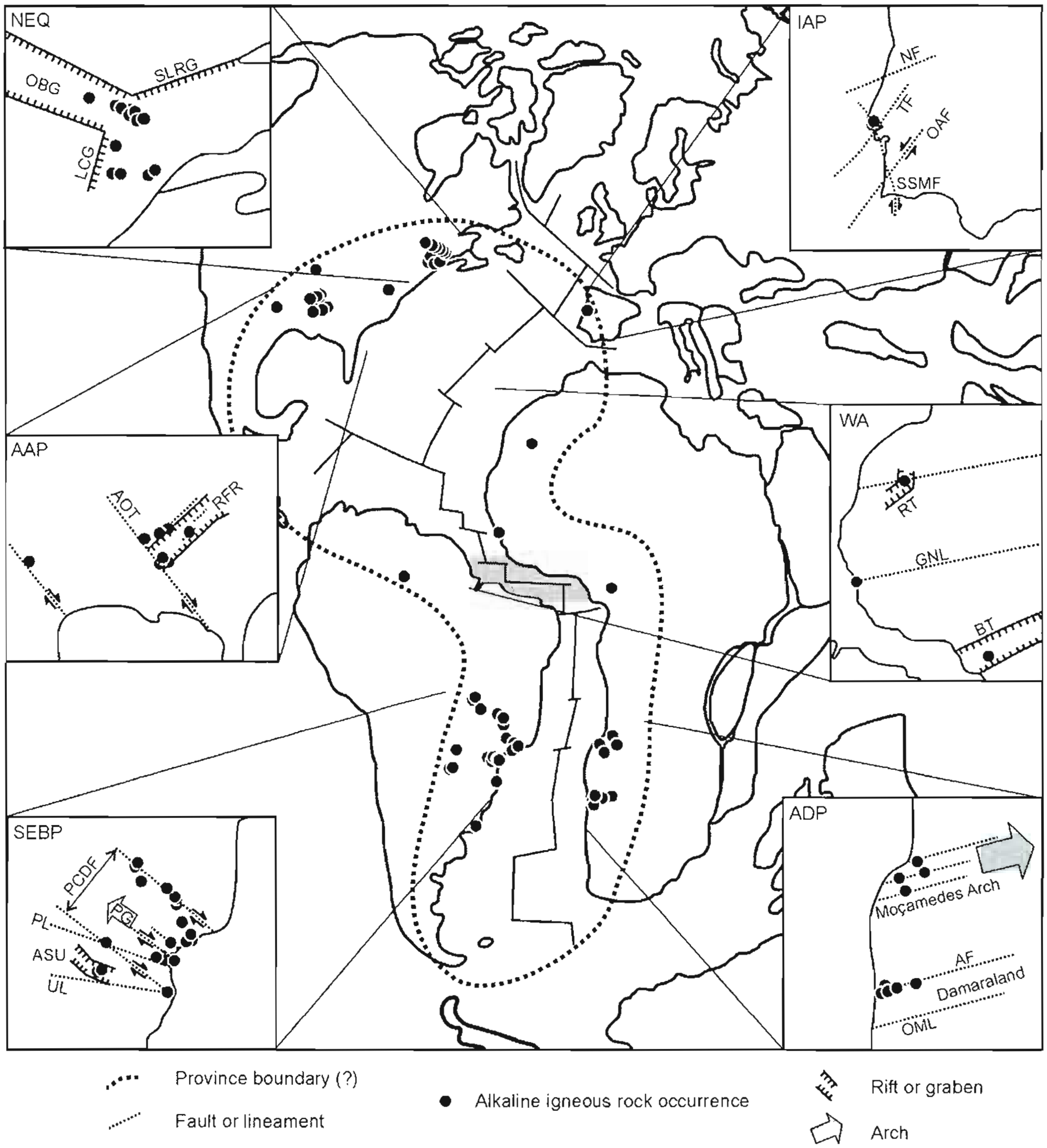

Figure 1 Distribution of the Peri-Atlantic Alkaline Province (125–80 Ma) at 100 Ma. Reconstruction modified from PlateTracker v2.0 (Eldridge et al., 1998). Occurrences of alkaline rocks were mainly taken from Woolley (1987; 2001). Paleolocations were obtaining using PointTracker v4c (Scotese, 2002). The Equatorial "locked zone" of Bonatti (1996) is shaded. Insets show details of the districts. Note the strong spatial correlation between alkaline complexes and structural features. Southeastern Brazil-Paraguay Province (SEBP) inset: (UL) Uruguay Lineament; (PL) Piquiri Lineament; (ASU) Asunciòn-Sapukai-Villarica graben; (PG) Ponta Grossa Arch; (PCDF) Parana-Chacos Deformation Zone. Angola-Damaraland Province (ADP) inset: (OML) Omaruru Lineament; (AF) Autseib Fault. Arkansas Alkaline Province and surrounding (AAP) inset: (RFR) Reelfoot Rift; (AOT)

Alabama-Oklahoma transform fault. New England-Quebec Province (NEQ) inset: (OBG) Ottawa-Bonnechere Graben; (SLRG) St. Lawrence River Graben; (LCG) Lake Champlain Graben. West Africa (WA) inset: (RT) Richat Trough; (GNL) Guineo-Nubian Lineament; (BT) Benue Trough. Iberia Alkaline Province (IAP) inset: (NF) Nazare Fault; (TF) Tagus Fault; (OAF) Odemira-Avila Fault; (SSMF) Sintra-Sines-Monchique Fault.

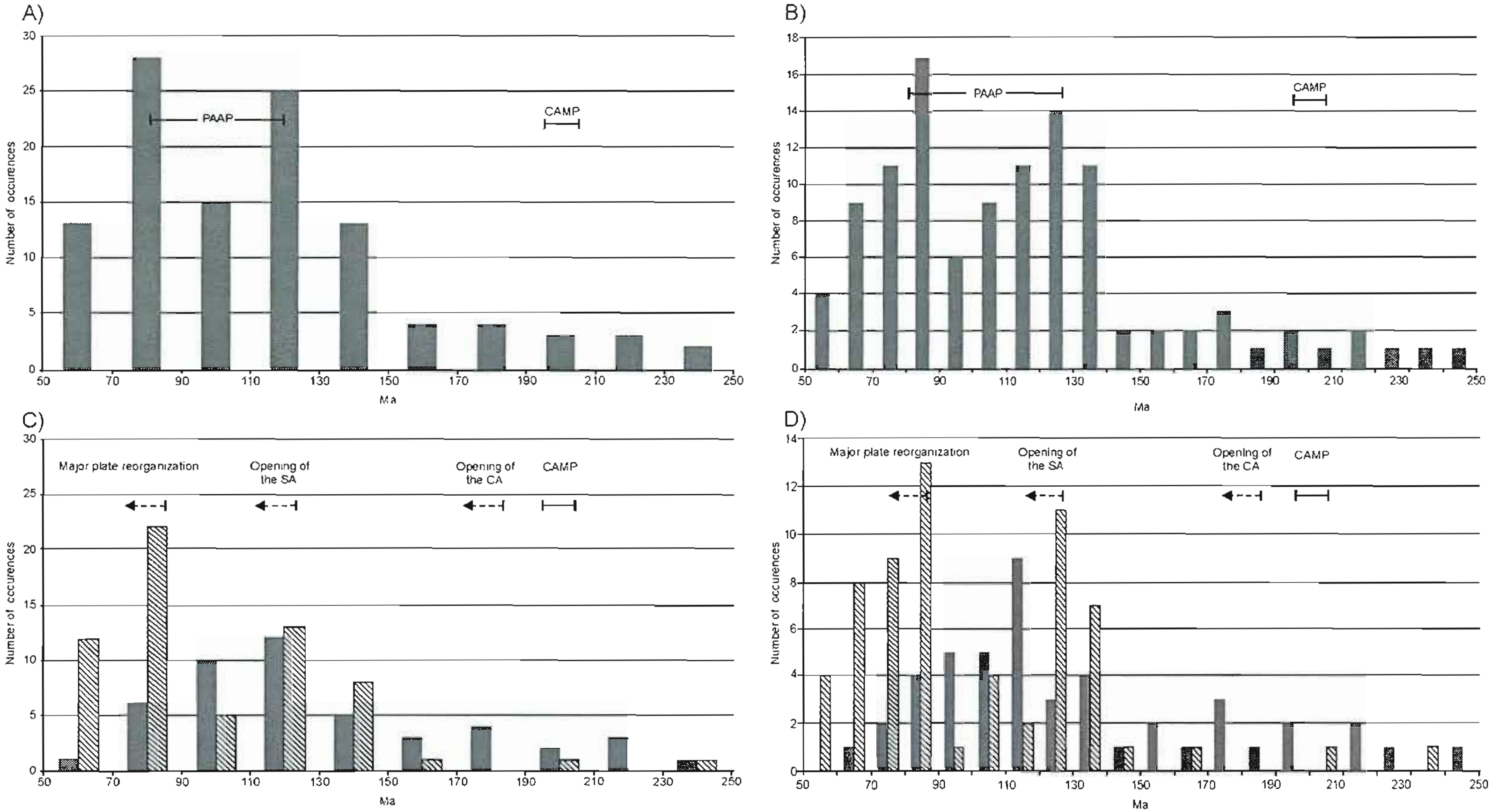

Figure 2. Frequency versus age histograms of alkaline activity from 250 to 50 Ma across the Peri-Atlantic Alkaline Province (PAAP) using 20-Ma (A) and 10-Ma ranges (B). In (C) and (D), the same 20-Ma and 10-Ma distributions are presented respectively, but with the data separated into two distinct populations north (shaded) and south (hatched) of the Equatorial Atlantic. Abbreviations: (SA) South Atlantic; (CA) Central Atlantic; (CAMP) Central Atlantic Magmatic Province.

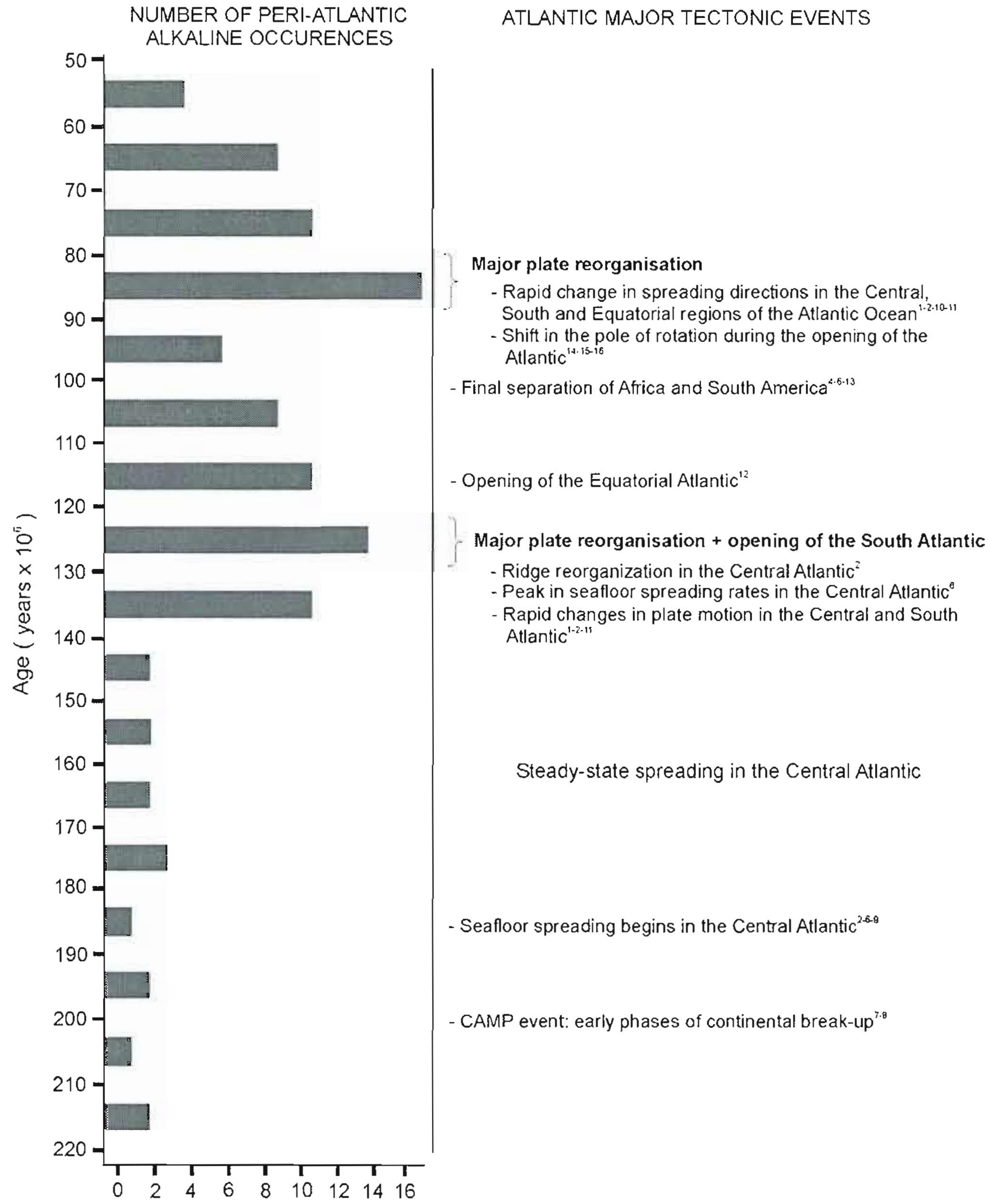

Figure 3. Correlation between alkaline activity (number of occurrences) in the peri-Atlantic zone and Mesozoic Atlantic rifting and spreading events.
Sources: [1]Binks and Fairhead (1992); [2]Kliford and Schouten (1986); [3]Janssen et al. (1995); [4]Nürnberg and Müller (1991); [5]Wilson and Guiraud (1992); [6]Torsvik et al., (2006); [7]Marzolli et al. (1999); [8]Hames et al. (2003); [9]Foland and Faul (1977); [10]Rabinowitz and Labrecque (1979); [11]Fairhead and Binks (1991); [12]Mascle et al. (1986); [13]Mascle et al. (1988); [14]Marsh (1973); [15]Prins (1981); [16]Almeida (1991).

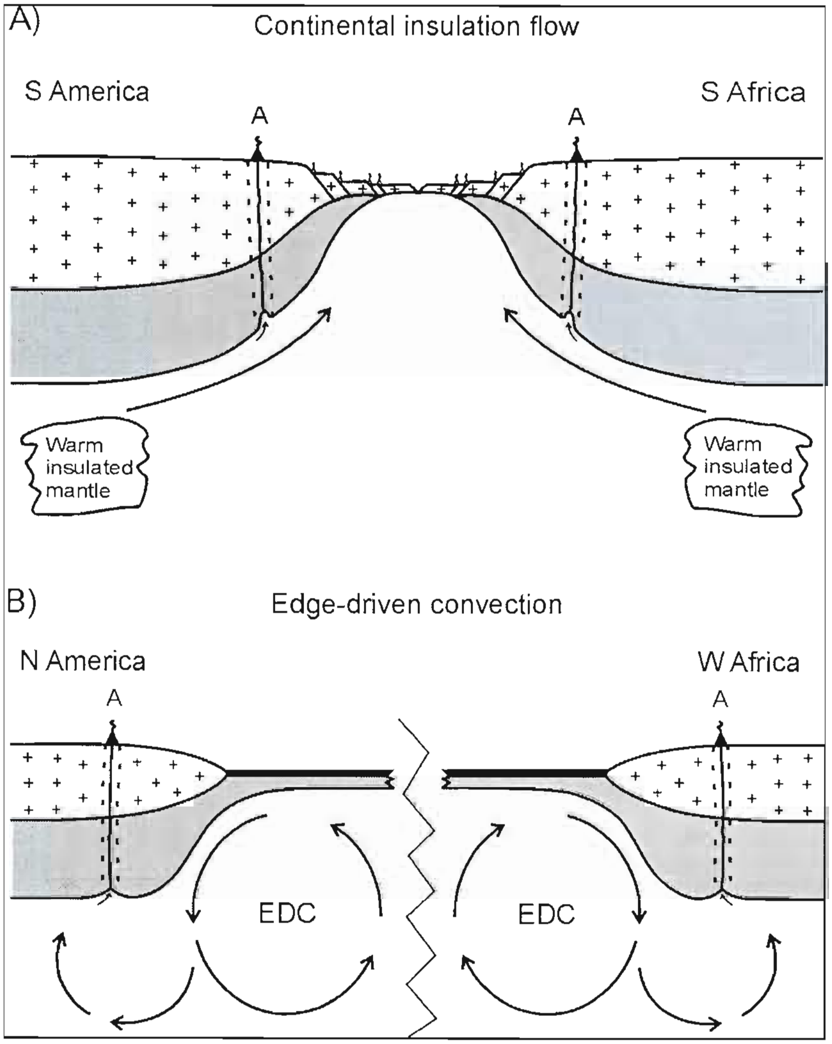

Figure 4. Two stages of small-scale asthenospheric upwelling developed simultaneously in the Atlantic Ocean during the Early Cretaceous. A: Continental insulation flow-driven magmatism in the South Atlantic during supercontinent break-up. B: Edge-driven convection (EDC) on separated continental margins in the Central Atlantic. Once the continental margin (edge) had migrated some distance from the rift zone in the South Atlantic, the system could have switched from continental insulation flow to EDC, which may explain the second magmatic peak during the Late Cretaceous. Thin black lines fringed with dots represent reactivated structures; crosses, continental crust; thick black lines, oceanic crust; shaded areas, oceanic and sub-continental lithosphere; white, asthenosphere.

Table 1. Spatial characteristics of the 125-80 Ma peri-Atlantic alkaline igneous rocks

	New England-Quebec Province	Arkansas Alkaline Province and surroundings	Southeastern Brazil-Paraguay Province and surroundings	Iberia Alkaline Province
Located on zone of weakness	YES	YES	YES	YES
Pre-existing structures	St. Lawrence rift system[1,2,3,4,5,6,8,9,10,22]	Rift system[13,14,49,51] - transform faults[51]	Lineaments[17,19,23,24,26,47] - rift system[21]	Faults/transform faults[52,53,54,55,56,57,58]
Main structural features	Ottawa-Bonnechere graben[1,2,3,4,5,8,10]	Reelfoot Rift[13,14,16,49]	Parana-Chacos Deformation Zone[50]	Nazare fault[55,56,58]
		Alabama-Oklahoma transform fault[51]	Ponta Grossa Arch[19,20,21,23,26]	Tagus fault[55,56,58]
		Mississippi Embayment	Asuncion-Sapucai-Villarica graben[25]	Odemira-Avila fault[55,56,59]
Orientation of main structure	WNW - ESE	NE - SW	NE - SW	NE-SW
		NW - SE	NW - SE / WNW - ESE	NNW-SSE
Structure inherited from	Opening of lapetus (L-P - E-C)[2,3,5,10]	Opening of lapetus[13,51] (L-P - E-C)[16,49,51]	pC[22,50]	L-Carb - E-Perm[55,56]
Reactivation processes	Multiple	Multiple	Multiple	Multiple
Cause of reactivation	Rifting in NA[7]	Plate reorganisation[14,49]	Atlantic opening[17,18]	Rifting in NA[53,54,57]
	Shift in pole of rotation[1,8]	Isostatic adjustment[14,15]	Shift in the pole of rotation[17,18,24,47]	Opening of the Bay of Biscay[53]
	SFS change (rates[7,8] or geometry[1])[4,11,28]		Accomodation plate movements[25,50]	Transform directions* [52]
	Transform directions* [8]		Transform directions* [17,18,47,50]	SFS change (geometry)[54]
	Plate motion[6]			
Radial rift structures	NO	NO	NO	NO
Regional uplift	NO	Somewhat	YES	NO
Flood/plateau basalts	NO	NO	YES	NO
Age progression	NO	Somewhat	NO	NO
Possible hotspot assocation	Great Meteor	Bermuda	Trindade (L-K) Tristan da Cunha (E-K)	None
Courtillot's classification **	0+ (tertiary)	0+ (tertiary)	Trindade: 0+ (tertiary) Tristan da Cunha: 3 (primary)	

Table 1. (continued)

	Angola-Damaraland Province and surroundings	Los Island (offshore Guinea)	Richat (Mauritania)	Benue (Nigeria)
Located on zone of weakness	YES	YES	YES	YES
Pre-existing structures	Lineaments[17,18,19,29,30,31,34,47]	Lineaments[27,35,36]	Aulacogen[39,40] - lineaments[39]	Rift system [28,33,44]
Main structural features	Moçamedes Arch[19,26,30] Omaruru lineament[32] Autseib fault[32]	Guineo-Nubians lineaments[27,35,36]	Richat Trough[39,40,41] Fracture zone[39,42]	Benue Trough[44,45]
Orientation of main structure	NE-SW	ENE-WSW	NE-SW ENE-WSW	NE-SW
Structure inherited from	Pan-African[27,28,32,33]	Pre-Mes[48]	Pan-African[40,39] / Perm[43]	Pan-African[28,33,46]
Reactivation processes	multiple	multiple	multiple	multiple
Cause of reactivation	SA Opening[17,18] Shift in the pole of rotation[17,18,47] Plate reorganisation[32] Change in SA spreading geometry[28,32] Transform directions* [17,18,47]	Plate motion change[37] Opening of the EA[35]	Atlantic drifting relationships[39,42]	Atlantic opening[28] Differential motion in the SA[28,44,45]
Radial rift structures	NO	NO	NO	NO
Regional uplift	YES	NO	Somewhat	NO
Flood/plateau basalts	YES	NO	NO	NO
Age progression	NO	NO	NO	NO
Possible hotspot assocation	Tristan da Cunha	None	None	St. Helena
Courtillot's classification **	3 (primary)			1 (tertiary)

Table 1.

* The continental expressions of transform faults. Term introduced by Francheteau and Le Pichon (1972).

** Numbers according to the classification of Courtillot et al. (2003).

Abbreviations: (pC) Precambrian; (P) Proterozoic; (L-P) Late-Proterozoic; (Pal) Paleozoic; (E-C) Early Cambrian; (Perm) Permian; (E-Perm) Early Permian; (L-Carb) Late Carboniferous; (Mes) Mesozoic; (J) Jurassic; (E-K) Early Cretaceous; (L-K) Late Cretaceous; (NA) North Atlantic; (EA) Equatorial Atlantic; (SA) South Atlantic; (SFS) Seafloor spreading.

Sources: [1]Bédard (1985); [2]Kumarapeli and Saull (1966); [3]Marquis and Kumarapeli (1993); [4]Tremblay et al. (2003); [5]Kumarapeli (1985); [6]McHone and Butler (1984); [7]Faures et al. (1996); [8]Foland and Faul (1977); [9]Sykes (1978); [10]Williams (1979); [11]McHone et al. (1987); [12]Bailey (1992); [13]Salvador (1991); [14]Byerly (1991); [15]Baksi (1997); [16]Sims (2002); [17]Marsh (1973); [18]Prins (1981); [19]Woolley (1989); [20]Cobbold et al. (2001); [21]Chang et al. (1992); [22]Eby (1984); [23]Comin-Chiaramonti and Gomes (1996); [24]Almeida (1991); [25]Fairhead and Wilson (2005); [26]Comin-Chiaramonti et al. (1999); [27]Guiraud et al. (1985); [28]Guiraud et al. (1992); [29]Lapido-Loureiro (1973); [30]Alberti et al. (1999); [31]Woolley (2001); [32]Raab et al. (2002); [33]Binks and Fairhead (1992); [34]Bailley (1992); [35]Moreau et al. (1996); [36]Guiraud et al. (1987); [37]Bellion and Crevola (1991); [38]Matton et al. (2005); [39]Matton and Jébrak (2006); [40]Villeneuve and Cornee (1994); [41]Bronner (1992); [42]Poupeau et al. (1996); [43]Liegeois et al. (1991); [44]Pindell and Dewey (1982); [45]Fairhead and Binks (1991); [46]Daly et al. (1989); [47]Le Pichon and Hayes (1971); [48]Wilson and Guiraud (1992); [49]Frese (1982); [50]Jacques (2003), [51]Thomas (2006), [52]Cornen (1982), [53]Rock (1982), [54]Macintyre and Berger (1982), [55]Wilson et al. (1989), [56]Pinheiro et al. (1996), [57]Bernard-Griffiths et al. (1997), [58]Alves et al. (2003).

CONCLUSIONS GÉNÉRALES

Au cours de cette thèse, quatre objectifs spécifiques représentant une échelle d'observation différente ont été traités sous la forme d'un zoom arrière. Il s'agissait d'abord de comprendre la structure à l'étude, de l'intégrer au contexte régional de l'époque et d'identifier les différents moteurs ayant pu jouer sur la mise en place d'un tel système magmatique. Le but de cette approche était de répondre à plusieurs questions reliées aux problématiques principales. Tout d'abord, il s'agissait de mieux comprendre un sommet de complexe alcalin. Ceux-ci sont rarement documentés et le Richat offrait la possibilité d'étudier les processus de dissolution et de fracturation en sommet de pluton ainsi que la mécanique de mise en place des différents épisodes intrusifs et extrusifs qu'on y retrouve. Ensuite, il s'agissait de comprendre la signification d'un complexe alcalin isolé et des questions qui s'y rattachaient; Quels sont les processus de mise en place? Pourquoi se trouve-t-il à cet endroit précis? Quelle est l'origine et l'évolution des magmas? À quel type de point chaud est apparenté ce phénomène (superficiel, intermédiaire, profond)? Finalement la dernière problématique visait à étudier le pulse alcalin péri-Atlantique Crétacé de façon globale, comme un tout, à une autre échelle. Cette dernière approche avait pour but de fournir des éléments de réponse à des questions telles que : y a-t-il un lien entre les différents événements dispersés de la bordure Atlantique? Est-ce que le modèle classique de point chaud issu d'une plume mantellique profonde est aussi convaincant en traitant tous les intrusifs globalement? L'ouverture de l'Atlantique peut-elle avoir joué un rôle? Que s'est-t-il passé aux environs de 100 Ma? Voici les principales contributions de cette thèse en lien aux grandes questions soulevées dans les problématiques de l'introduction.

Un sommet de complexe alcalin

Les complexes alcalins peuvent être le siège de phénomènes hydrothermaux de grande ampleur. Au Richat, ils se traduisent par une méga-brèche d'effondrement karstique, des anomalies en isotopes stables dans les carbonatites, l'altération des roches gabbroïques et rhyolitiques, de même que des minéralisations en barite. Le regroupement des événements hydrothermaux au centre de la structure implique un lien avec le bombement. En effet, la concentration de fractures à l'apex d'un dôme crée une zone perméable favorisant la canalisation et l'infiltration des fluides hydrothermaux centrées sur la structure. La dissolution des unités sédimentaires contribue également à augmenter la perméabilité par rétroaction positive jusqu'à la perte de cohésion du système et son effondrement gravitaire. Le bombement et la production de fluides hydrothermaux ont donc conditionné les processus de bréchification et d'altération au sommet du complexe du Richat.

Le Richat présente également une exposition de phénomènes magmatiques où se côtoies des roches intrusives et extrusives indiquant un contraste d'érosion et une mécanique particulière de mise en place. Ce contraste d'érosion entre la partie centrale et externe du Richat est expliqué par un effondrement en piston qui a préservé les facies extrusifs centraux ainsi que la brèche hydrothermale karstique.

La structure circulaire du Richat et sa brèche centrale représente donc l'expression de surface d'un complexe alcalin Crétacé incluant un remplissage karstique exceptionnellement bien préservé en son sommet.

La signification d'un complexe isolé

Le complexe du Richat serait lié à la présence d'un aulacogène Protérozoïque ayant contrôlé le développement de la chambre magmatique et la caldera sus-jacente. La réactivation de cette structure profonde a permis la remontée de magmas issus de profondeurs différentes (fusion étagée) expliquant la coexistence de magmas alcalins et tholéiitiques au sein d'un même complexe. Les magmas alcalins seraient issus d'une source asthénosphérique alors que les magmas tholeiitiques proviendraient du manteau sous-continental et apparentés à ceux de l'épisode du CAMP (*Central Atlantic Magmatic Province*) de l'ouest africain. En subsurface, l'éruption de la série tholéiitique est contrôlée par la chambre magmatique et les mécanismes peu profonds d'effondrement-résurgence de caldera, alors que les magmas carbonatitiques et kimberlitiques se mettent en place le long des anisotropies préexistantes. Le patron structural hérité a donc joué un rôle prépondérant dans l'évolution de ce complexe.

Les succès du modèle de plume mantellique pour expliquer les chaînes d'îles océaniques ont mené à la proposition voulant que les plumes soient indispensables à la production de magmas alcalins autant continentaux qu'océaniques (Burke et Wilson, 1976). Cependant, le complexe du Richat est incompatible avec le modèle de point chaud associé à une plume mantellique. C'est un phénomène superficiel relié à des mécanismes de réactivation lithosphérique probablement associés aux perturbations tectoniques survenues lors de l'ouverture et l'expansion de l'Océan Atlantique. Il serait donc apparenté au troisième type de point chaud de Courtillot et al. (2003) dont l'origine serait peu profonde, vraisemblablement asthénosphérique.

De plus, il est peu probable que les complexes isolés, ponctuels ou répétitifs soient reliés à une plume mantellique. Ceux-ci seraient plutôt des marqueurs du troisième type de point chaud. C'est d'ailleurs le seul modèle pouvant expliquer la répétition du magmatisme alcalin à travers le temps sur des plaques en mouvement. Les complexes alcalins isolés pourraient donc se manifester lors de conditions exceptionnelles de grands réajustements tectoniques. Ils constitueraient alors des témoins de ces événements et pourraient agir à titre de traceurs temporels de ces phénomènes.

Le pulse alcalin péri-Atlantique Crétacé – une nouvelle approche

L'étude des intrusifs alcalins péri-Atlantique, selon une approche globale, a révélé des liens étroits entre ces événements dispersés de la bordure Atlantique. Leur distribution spatiale et temporelle a permis de définir une Province Alcaline Péri-Atlantique (PAAP) d'âge mid-Crétacé à fort contrôle structural. Les intrusifs du PAAP sont localisés sur des zones de faiblesse lithosphérique anciennes où la répétition du magmatisme et le rejeu des structures sur de longues périodes sont fréquents. Les pics de distribution du magmatisme alcalin semblent correspondre à la réactivation généralisée de ces structures préexistantes agissant comme zones d'accommodation du stress intraplaque lors d'épisodes spécifiques de l'évolution tectonique de l'Océan Atlantique. Le manque d'affinité avec les points chauds profonds des différents districts alcalins semble incompatible avec une origine issue de plumes mantelliques. Une origine structurale superficielle (asthénosphérique) est favorisée plutôt qu'une origine profonde (D'').

Ces résultats sont en accord avec un article récemment publié dans *Geology* (Jagoutz et al., décembre 2007) faisant état d'observations similaires en milieu océanique. Les auteurs de cet article rapportent un double pulse alcalin Crétacé dans les bassins océaniques conjugués d'Iberia et de Terre-Neuve. Ils décrivent en milieu océanique à l'échelle régionale deux phases d'intrusions alcalines 1- au moment de l'ouverture océanique et 2- lors des réajustements tectoniques résultants de l'expansion océanique. Ces épisodes de réajustements sont décrits comme des événements soudains, isolés et de courte durée durant l'expansion océanique, et marqueraient un « *tectonic spreading* ».

Notre étude montre que le « *tectonic spreading* » de Jagoutz et al. (2007) se développe largement sur la marge des continents. Puisque la déformation se développe plus facilement au sein d'une croûte océanique juvénile qu'à l'intérieur d'une croûte continentale rigide, l'accommodement du stress en milieu continental lors de ces épisodes serait facilité et focalisé au niveau des zones d'anisotropies préexistantes.

Dans le contexte géodynamique du démantèlement de la Pangée, l'étude indépendante des différents districts alcalins semble donc représenter une vision étriquée des choses. Lorsque prise individuellement, la théorie du point chaud classique, est attrayante, mais prise globalement elle s'avère beaucoup moins convaincante. Les intrusifs alcalins issus d'une source peu profonde et liés à des processus purement tectoniques sont probablement plus importants que l'on ne pense, surtout lors de périodes de réajustements tectoniques majeurs. Il serait important de considérer cette issue au même titre que les modèles classiques de plume mantellique automatiquement évoqués pour expliquer tout alignements alcalins.

Pistes de recherche

L'étude du Richat offrait une quantité phénoménale de possibilités de recherche. Tel que mentionné par un des jurys du devis de recherche, plusieurs aspects de cette structure auraient, a eux seuls, mérité un sujet de thèse. Des choix ont donc été faits afin de répondre le plus adéquatement aux problématiques soulevées dans le cadre réalisable d'une thèse de doctorat. Voici différentes recommandations de pistes de recherches dans la poursuite de cette étude.

Étude isotopique Sm-Nd, Rb-Sr

Afin de mieux contraindre la source des magmas, une étude des isotopes Sm-Nd et Rb-Sr serait requise. Celle-ci a d'ailleurs été entamée lors de cette thèse mais a dû être abandonnée en cours de route. La raison principale de ce choix est liée aux résultats décevants obtenus lors des analyses préliminaires. Les échantillons ne semblaient pas favorables à ce type d'analyse dû à l'altération hydrothermale subie.

Un projet de recherche entier pourrait être consacré à une telle étude incluant, au minimum, une mission de terrain portant exclusivement sur la récolte d'échantillons prévus pour ce type d'analyse en s'assurant de la quantité et de la qualité du matériel. Il faudrait aussi tenir compte des coûts importants de transport d'une grande quantité d'échantillons requise à ces fins, une contrainte d'envergure rencontrée lors des travaux de terrain reliés à cette thèse. De plus, puisque nos échantillons de filons de carbonatite semblent tous avoir subi une altération marquée et qu'ils aient été échantillonnés systématiquement par le présent auteur, un doute persiste quant aux chances d'obtenir des échantillons non-affectés par l'altération hydrothermale.

La nature du fluide hydrothermal

Dans le but d'obtenir une meilleure compréhension du fluide silicificateur responsable de la formation de la brèche, une étude sur les isotopes stables d'oxygène a été effectuée sur la matrice. Les résultats, extrêmement variés, ne furent pas très concluants. Les valeurs obtenues varient de + 20‰ à + 27‰ (SMOW). Cette variation énorme ne permet pas d'utiliser ces valeurs scientifiquement (C.-H. Marcel, commun. pers.). Le problème se situe possiblement au niveau de l'échantillonnage de la matrice de la brèche. En effet, il est très difficile de prendre uniquement la matrice, sans inclure des petits fragments microscopiques de roche encaissante. Les variations isotopiques sont donc probablement liées à l'incorporation de quantités plus ou moins grandes de fragments bréchiques.

Suite à la découverte des sédiments hydrothermaux à la base du karst, les mêmes analyses ont été effectuées sur ces derniers. Ces sédiments sont directement liés au fluide de remplissage et ne contiennent pas ou très peu de fragments bréchiques. Les valeurs obtenues sont homogènes et de l'ordre de +24-25‰ (SMOW). En supposant que le liquide était initialement de l'eau de mer, ces valeurs donnent une température cohérente avec celle de phénomènes hydrothermaux de basses températures, soit autour de 100-120°C (C.-H. Marcel, commun. pers.). Cependant, si le liquide initial correspond à de l'eau météorique, la température pourrait être de 18°C. Un mélange de ces deux liquides offrirait aussi tout une gamme de température passant de superficiel à hydrothermal. Pour pouvoir quantifier correctement la température du fluide de remplissage, il faudrait donc connaitre le liquide initial, ce que l'on ignore. Des études plus étendues devraient donc inclure les inclusions de fluides dans les sédiments internes afin d'avoir une idée plus précise du liquide initial. Dans l'hypothèse d'un apport magmatique au fluide, les inclusions de fluides seraient également un outil de caractérisation efficace.

Modèle quantitatif

Il a été démontré qu'une Province alcaline mid-Crétacé pouvait être individualisée sur les bordures continentales ceinturant l'Atlantique. Un modèle ayant pu favoriser la production et l'infiltration de magma alcalin dans la lithosphère a également été proposé pour la remontée asthénosphérique. Si le contrôle structural et la corrélation du PAAP avec des épisodes spécifiques de l'évolution tectonique de l'Océan Atlantique sont bien définis, les modèle du *Edge Driving Convection* et du *Continental insulation flow* (King et Anderson, 1998, King et Ritsema, 2000) ne sont pas démontrés quantitativement pour soutenir l'hypothèse de la remontée asthénosphérique. Idéalement un modèle numérique viendrait appuyer cette proposition. Il s'agirait donc d'utiliser les paramètres les plus réalistes du manteau supérieur soumis à des contrastes de température, soient issus d'une différence d'épaisseur lithosphérique, soient de l'effet d'isolation d'un supercontinent selon les différents emplacements et l'âge de mise en place des districts alcalins du PAAP. Un modèle utilisant les flux de chaleur pourrait également être considéré.

Références

Burke, K., et Wilson, J. T. 1976. Hot spots on the Earth's surface. Sci. Amer., v. 235, p. 46-57.

Courtillot, V., Davaille, A., Besse, J., et Stock, J. 2003. Three distinct types of hotspots in the Earth's mantle. Earth and Planetary Sciences Letters, v. 205, p. 295-308.

Jagoutz, O., Müntener, O., Manatschal, G., Rubatto, D., Péron-Pinvidic, M., G., Turrin, B. D. et Villa, I. M. 2007. The rift-to-drift transition in the North Atlantic: A stuttering start of the MORB machine? Geology, v. 35, no. 12, p. 1087-1090.

King, S. D., et Anderson, D. L. 1998. Edge-driven convection. Earth and Planetary Sciences Letters, v. 160, p. 289-296.

King, S. D. et Ritsema, J. 2000. African hot spot volcanism; small-scale convection in the upper mantle beneath cratons. Science, v. 290, p. 1137-1140.

APPENDICE A

RÉSUMÉS DE CONFÉRENCES

THE CRETACEOUS PERI-ATLANTIC ALKALINE PROVINCE (PAAP): DEEP MANTLE PLUME ORIGIN OR SHALLOW LITHOSPHERIC BREAK-UP?

2007- Ocean Continent Transition meeting, Académie des sciences, Paris, France

Although the Early Jurassic initial phases of Atlantic rifting were marked by the tholeiitic basalt flows and dikes of the Central Atlantic Magmatic Province (CAMP), the Cretaceous Era was characterised by widespread alkaline igneous activity on widely separated peri-Atlantic continental margins. Nearly half the peri-Atlantic alkaline rocks dated between 250 to 50 Ma fall within the 125-80 Ma range. This period of intense alkaline activity during the mid-Cretaceous formed the Peri-Atlantic Alkaline Province (PAAP). The mechanisms for the opening of the Atlantic Ocean, as well as for the CAMP and later PAAP magmatic events, have been attributed either to deep mantle plumes or to a combination of tensional forces, lithospheric rifting and structural controls. We describe the PAAP event for the first time and test the plume and structural models.

The PAAP displays evidence of strong structural control and appears to be related to reactivation of pre-existing lithospheric structures throughout the peri-Atlantic zone. Early stages of continental separation and abrupt changes in plate motion were the most likely mechanisms for the reactivation of favourably oriented pre-existing structures. Mid-Cretaceous pulses in alkaline magmatic activity were closely associated with Atlantic major tectonic events. The opening of the South Atlantic, the change in intensity and direction of both seafloor spreading and plate motion, and the consequent fluctuation in intra-plate stresses were responsible for two spikes in peri-Atlantic alkaline magmatism at ~125 Ma and ~85-80 Ma.

The general lack of evidence for a deep hotspot associated with the peri-Atlantic zone seems incompatible with a mantle plume origin. On the other hand, the spatial distribution of the PAAP along passive continental margins suggests an ideal setting for the development of edge-driven convection. We propose that the periodic reactivation of deep-seated pre-existing zones of weakness during the major stages of Atlantic tectonic evolution, combined with coeval asthenospheric upwelling due to edge-driven convection and continental insulation flow, enhanced the ascent of alkaline magmas. This mechanism of shallow, small-scale upwelling during periodic structural reactivation provides a more likely explanation for the PAAP than the commonly proposed deep mantle plume scenario.

A MONTEREGIAN-STYLE INTRUSION IN OCCIDENTAL SAHARA?

2006 - Geological Association of Canada and Mineralogical Association of Canada, Montréal, Canada.

The Richat structure (Sahara, Mauritania) appears as a large dome at least 40 km in diameter within an Upper Proterozoic to Ordovician sequence. The center of the structure consists of a limestone-dolomite shelf that encloses a kilometer-scale siliceous breccia and is intruded by basaltic ring dikes, kimberlitic intrusions, and alkaline volcanic rocks. The breccia was created during karst dissolution and collapse. Cretaceous hydrothermal internal sediments fill the centimeter- to meter-scale cavities. The circular Richat structure and its breccia core represent the superficial expression of a Cretaceous alkaline complex with an exceptionally well-preserved hydrothermal karst infilling at its summit. The significance of the position of the structure is however still a matter of debate. The meaning of deep-seated magmatic rocks such as carbonatites and kimberlites in that part of the West African Craton is unknown. Several hypotheses (hot spot, oceanic transform faults, intersection of fracture's network, etc,) have been presented to explain the Richat position and will be discussed. We propose a new aulacogens zone located in northern Mauritania and it's reactivation through Cretaceous times.

KARST HYDROTHERMAL AU SOMMET D'UNE INTRUSION ALCALINE, SAHARA, MAURITANIE

2004 - 72e congrès de l'ACFAS, Montréal, Canada

Le complexe du Richat est une des structures terrestres les plus spectaculaires vue de l'espace. Situé dans des sédiments du Protérozoïque au Cambro-Ordovicien du Sahara mauritanien, le complexe du Richat forme un dôme de 40 km de diamètre. Le centre de la structure consiste en une plate-forme de calcaire dolomitique recoupée par une brèche de chert kilométrique. Plusieurs théories ont été proposées sur l'origine de la structure dont un impact météoritique, du plutonisme et des phénomènes intraformationnels.

La distribution spatiale de la brèche est définie par une forme lenticulaire montrant un épaississement dans la zone centrale et un pincement sur les périphéries où elle se trouve limitée à quelques mètres d'épaisseur. La brèche est composée de fragments de chert anguleux à arrondis provenant d'une silicification hydrothermale polyphasée dans une matrice siliceuse. Les analyses et l'observation au MEB montrent un remplissage de cavités par des sédiments internes hydrothermaux. La formation du dôme a créé une zone favorable à la dissolution des séquences sédimentaires et l'infiltration de solutions hydrothermales. Nous démontrons que la structure du Richat avec son coeur de brèche est l'expression superficielle d'un complexe alcalin avec un remplissage karstique hydrothermal exceptionnel à son sommet.

LE RICHAT, UN KARST HYDROTHERMAL AU SOMMET D'UNE INTRUSION ALCALINE, MAURITANIE

2004 - 32e Congrès de géologie africaine, Orléans, France

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FRACTAL ANALYSIS AND ORIGIN OF A CHERT MEGABRECCIA IN THE RICHAT DOME, MAURITANIA, AFRICA

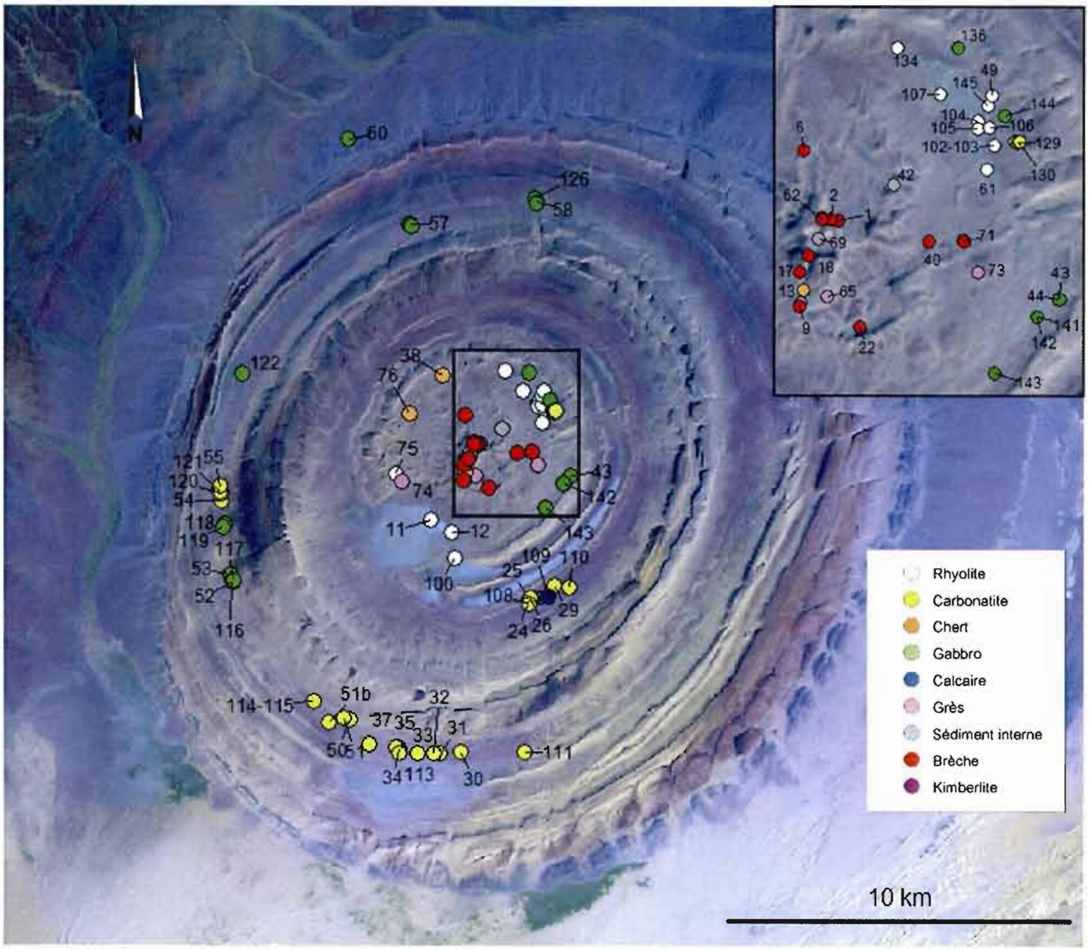
2002 - Geological Society of America, Denver, USA

The Richat complex is one of the most spectacular terrestrial circular structures visible from space. Located in Proterozoic-Cambrian sediments of the Mauritanian part of the Sahara desert, the Richat complex forms a domal structure of 40 km in diameter. The center of the structure consists of a limestone-dolomite shelf intruded by a kilometric scale chert megabreccia. Numerous theories have been proposed for the origin of this mysterious breccia including meteoritic impact, plutonism and intraformational processes.

The stratigraphy of the Richat dome includes limestone, dolomite, sandstone, mudstone and chert. Field and petrologic studies indicate that the breccia is composed of angular and rounded heterolithic chert fragments in a silica cement. Silica dissolution and replacement were observed in the breccia whereas dolomitization and sulphidation were observed in the limestones. Early ductile deformation in the chert indicates an early development of silicification. The chert fragments display a bimodality in the boundary fractal dimension (Euclidian distance mapping) that implies two phases of breccia formation. Field observations suggest that the breccia was form by a process of karst-dissolution collapse. The most likely interpretation is that the Richat chert was formed during a shelf high-stand and persisted in a sub-emergent setting from the Proterozoic up to the Cretaceous. Thus, the Richat chert appears to be an indication of a long term high-standing structural position.

APPENDICE B

ANALYSES GÉOCHIMIQUES ROCHES TOTALES



Localisation des échantillons prélevés sur le site du Richat.

Analyses géochimiques roche totale par ICP-MS (fusion Bore-Lithium)

eléments	unité	30 carbo-natite dike	35 carbo-natite dike	110 carbo-natite dike	111 carbo-natite dike	114 carbo-natite dike	56 carbo-natite Si++ dike	60 kimber-lite sill	44 gabbro ring dike	57 gabbro ring dike	58 gabbro K+, Si+ dike	127 gabbro K+, Si+ dike	126 gabbro K+, Si+ dike	143 gabbro ring dike	144 gabbro ring dike	116 gabbro ring dike	118 gabbro ring dike	122 gabbro ring dike	136 gabbro ring dike
Na2O	wt%	0,34	0,38	0,31	0,18	0,39	0.16	0,09	2,06	2,27	4,44	4,25	4,26	2.18	2,09	2,35	0,48	2,14	2,12
MgO	wt%	12,00	12,52	14,62	13,85	11,66	0,15	16,23	6,91	6,42	2,49	2,73	0,92	7.42	7,55	6,26	8,61	6,66	7,59
Al2O3	wt%	1,89	2,27	0,75	0,92	1,21	1,40	5,01	15,27	15,25	20,20	18,36	17,46	14,49	14,56	15	12,93	14,3	14,33
SiO2	wt%	1,44	2,35	0,56	5,83	12,92	83,12	29,81	51,33	51,28	58,11	68,27	60,42	52.21	51,72	51,24	46,49	50,8	51,71
K 2O	wt%	0,04	0,05	0,04	0,07	<.04	0,02	1,49	0,80	0,54	6,12	3,58	6,13	0,88	0,79	0,72	0,14	0,64	0,82
CaO	wt%	25,49	25,41	30,45	27,26	26,15	3,22	11,78	10,00	9,70	1,51	3,78	1,7	10.09	10,45	10,32	17,68	10,23	10,52
MnO	wt%	0,70	0,55	0,58	0,81	0,8	0,06	0,17	0,18	0,20	0,06	0,04	0,04	0.15	0,15	0,17	0,15	0,18	0,16
TiO2	wt%	0,62	0,77	0,12	0,09	0,14	0,10	3,17	1,06	1,27	1,14	1,47	0,88	1	1	1,1	0,65	1,22	0,96
FeOtot	wt%	8,55	7,44	7,32	8,18	7,39	5,70	10,12	9,50	10,32	4,69	4,57	4,13	9.29	9,46	9,96	8,16	10,63	9,45
P2O5	wt%	1,41	1,12	3,09	0,85	3,33	1,98	0,57	0,17	0,15	0,11	0,11	0,13	0.16	0,15	0,11	0,11	0,14	0,16
Cr2O3	wt%	0,008	0,011	0,001	0.002	0,003	0,002	0,090	0,036	0,019	0.013	0,02	0,013	0,048	0,047	0,011	0,04	0,014	0,058
LOI	wt%	45,03	44,64	40,2	39.1	33,4	3,08	18,55	1,91	1,60	0,90	2,3	3,4	1	0,8	1,6	3,5	1,8	1
Total	wt%	97,51	97,51	98,14	97,14	97,39	99,00	97,09	99,23	99,02	99,78	99,48	99,48	98.91	98,76	98,84	98,94	98,77	98,88
V	ppm	176,0	179,0	119,0	175.0	152,0	74,0	206,0	242,0	285,0	154.0	178,0	104,0	238,0	226,0	293,0	244,0	324,0	274,0
Co	ppm	11,4	15,0	7.7	4.1	7,1	4,6	74,4	40,8	45,7	17 1	15,3	11,5	42,2	41,8	45,1	42,0	45,1	46,1
Ni	ppm	17,6	64,1	12,5	11.3	7.7	7,6	662,2	77,8	83,3	44 1	46,3	27,2	17,2	16,7	60,0	62,0	40,0	20,8
As	ppm	32,2	31,2	4,0	11 4	22,2	11,0	7,2	3,8	5,6	8.9	3,7	2,4	1,3	0,8	2,7	19,8	3.4	16,8
Rb	ppm	1,7	3,4	1,4	2 0	1,4	0,8	110,2	26,2	17,8	166.5	97,9	169,3	28,2	24,5	23.1	1,1	22,5	26.5
Sr	ppm	3199,0	1475,0	3240,4	2134.9	5195,2	3494,0	527,0	234,0	235,0	446.0	261,2	263,0	246,8	212,0	248,6	116,3	230,4	263,2
Zr	ppm	186,9	174,5	142,3	49,3	249.8	<0,2	346,2	73,1	32,1	29.6	271,2	253,4	95,0	93,2	95,1	43,2	94,2	90,2
Bi	ppm	0,2	0,2	785,4	306.0	0,1	0,2	0,1	< 0.04	< 0,04	< 0,04	0,6	0,5	0,3	0,3	<.1	<.1	0,1	<.1
Sb	ppm	0,6	0,5	0,4	0.2	0,3	0,2	0,2	0.1	0,2	0.3	0,1	0,1	0,2	0,1	<.1	0,1	<.1	<.1
Ga	ppm	2,8	3,5	4,2	4.7	5,2	4,0	10,6	16,5	18,7	23.7	21,2	18,7	16,4	16,3	18,1	14,1	18,3	16.0
Nb	ppm	84,6	104,0	816,4	112 4	1596,1	39,5	76.0	14,5	8,1	16.9	18,7	14,8	9,3	9,3	11,2	28,4	9,7	16,1
Cd	ppm	0,3	0,3	0.2	0 1	0,3	0,0	0,2	0,1	0,1	<0,02	<.1	<.1	0,2	0,1	0.1	<.1	0,1	< 1
Cs	ppm	0,3	0,4	0,9	0.1	1,3	0,2	1,6	0,9	0,5	0.6	0.4	0,6	1,2	1,3	1,0	0,4	1,0	1,7
Ba	ppm	479,0	864,0	205,6	9558.2	6271,6	284,0	3350,0	781,0	631,0	2091.0	1015,7	2200,0	388,7	276,7	226,4	163,9	228.7	439,9
Hf	ppm	2,2	2,6	1.8	0.9	3.5	0,0	9,0	2,3	1,3	1.6	7.8	7,2	2,5	2,6	2,9	1,3	2.6	2.5
Ta	ppm	0,8	0,6	47,5	4.5	51.1	0,5	5.3	0,7	0,5	1,2	1,3	1,3	0,8	0,7	0,6	0,9	0.4	0,8
Pb	ppm	26,8	13.4	9,6	12.5	6.6	27,1	18,5	4,4	13,1	22,3	8,0	17,1	15,6	8,6	19.0	1,2	10,1	2,0
Zn	ppm	81,8	71,3	260,0	77.0	132,0	185,6	79,6	80,2	107,8	50.8	19,0	28,0	50,0	38,0	49,0	10,0	44.0	21,0
Ag	ppb	1034,0	645,0	<.1	< 1	0,2	202,0	180,0	134,0	86,0	122,0	< 1	<.1	0,1	< 1	<.1	<.1	< 1	<.1
Mo	ppm	0,4	0,4	0,4	0.4	0,3	0,8	0,3	0,6	0.4	0 3	0,2	0,3	1,5	0,8	0,3	0,1	0,5	0,4
Cu	ppm	9,0	11,8	9,0	2.2	34,0	7,9	112,4	89,0	48,8	4,3	8,1	2,1	88,8	88,8	126,8	15,6	127.7	93,8
Sn	ppm	1,2	1,1	4.0	<1	4,0	2,5	2,0	1,0	0.9	1,3	1,0	1,0	<1	<1	1.0	1,0	1.0	1,0
W	ppm	6,9	4,9	2,3	1.5	2,1	0,7	1,0	0,6	0,8	1.5	1,8	1,8	0,8	0,4	0,5	0,3	0,4	0,2
Au	ppm	<0,1	<0,1	1.2	5,1	3.1	<0,1	<0,1	<0,1	<0,1	<0,1	1,2	1,2	2,1	2,5	2,7	2,9	0.6	1,2
Th	ppm	59,6	29,7	95,7	155.2	258,3	94,3	23.1	3,7	2,7	21.0	14,4	16,7	3,4	2,8	2,4	5,7	2,5	3,7
Be	ppm	1,0	1,0	1,0	2.0	3,0	4,0	2.0	1.0	1,0	3.0	3,0	2,0	1,0	1,0	1,0	<1	1.0	<1
Y	ppm	75,9	42,2	79,5	84.3	147,7	66,2	17,8	20,3	23,0	33 7	39.9	43,1	22,4	21,5	22,1	15,6	22,3	20,9
Sc	ppm	6,7	6,9	16,0	22.0	17,0	18,2	17,0	32,8	33,5	20.4	20,0	18,0	35,0	35,0	34,0	33,0	35,0	35,0
U	ppm	9,0	4,4	30,4	1.5	60,8	10,5	2,4	0,7	0,4	1.9	2,1	2,1	0,6	0,7	0,6	1,2	0,6	0,7
La	ppm	546,0	319,0	681,5	1157.2	872,8	526,0	66,0	17,0	11,0	65.0	39,8	57,3	14,9	14,2	12,3	22,3	12,0	16,0
Ce	ppm	849,8	482,4	1254,1	1706,4	1681.1	859,3	130,9	37,8	29,2	128.1	89,2	117,5	31,1	29,0	28,0	43,7	27.2	33,6
Pr	ppm	95,5	55,6	132.6	160.0	176.7	96,9	14,6	4,1	3,4	15.3	10,9	13,4	3,8	3,6	3,5	4,9	3,4	4,0
Nd	ppm	327,7	188,4	434,1	481.1	588.8	356,7	57,3	16,1	13,5	57,9	42.1	49,0	15,2	15,9	14,6	18,5	15.6	16,0
Sm	ppm	44,2	26,4	59.6	55.7	72.4	49,1	12,5	3,7	3,6	10.7	8,8	9,2	3,4	3,4	3,2	3,2	3,3	3,5
Eu	ppm	11,8	7,2	15,8	13.9	20,2	13,1	3.4	1,3	1,2	2.0	1,6	1,6	1,1	1,1	1,1	1,0	1,1	1,0
Gd	ppm	25,3	15,8	31,9	30 7	45.9	27,4	8,1	3,8	4,1	8,4	7,5	7,6	4,0	3,7	3,8	3,1	3,7	3,5
Tb	ppm	4,2	2,4	4,4	4.6	7,8	4,1	1,1	0,7	0,8	1.4	1,3	1,3	0,7	0,6	0,7	0,6	0,7	0,7
Dy	ppm	18,4	10,3	18,8	20.7	33,6	16,6	4.2	4,1	4,8	7.6	7,0	7,7	3,9	3,7	4,0	2,8	3,8	3,5
Ho	ppm	3,0	1,6	2,9	3.0	5,1	2,4	0,5	0,7	0,9	1.3	1,5	1,6	0,8	0,8	0,8	0,6	0,8	0,7
Er	ppm	9,1	4,6	7,1	7.0	12.1	7,1	1,6	2,8	2,8	4.4	4,2	4,8	2,4	2,3	2,3	1,6	2,3	2,1
Tm	ppm	1,0	0,5	1.0	0.9	1.5	0,7	0,2	0,3	0,3	0.5	0,6	0,7	0,4	0,4	0,3	0,2	0,3	0,3
Yb	ppm	6,8	3,0	5,6	5.2	8,0	4,7	1,4	2,4	2,4	3.8	4,0	4,6	2,2	2,1	2,1	1,3	2,0	1,8
Lu	ppm	0,7	0,3	0,8	0.7	1.0	0,5	0,1	0,3	0,3	0.4	0,6	0,7	0,3	0,3	0,3	0,2	0,3	0,3

Analyses géochimiques roche totale par ICP-MS (fusion Bore-Lithium)

éléments	unité	48 volcanite analcime	61 volcanite analcime	100 volcanite analcime	103 volcanite analcime	104 volcanite analcime	134 volcanite analcime	145 volcanite analcime	12 volcanite analcime	74 volcanite analcime	75 volcanite analcime	107 volcanite analcime	10 gres	38 chert	41 calcaire	68 sédiment Interne	67 sédiment Interne	69 sédiment Interne	101 quartzite
Na2O	wt%	9,11	5,74	4,15	8,27	6,07	7,98	6,39	7,87	6,11	9,15	9,02	0,12	0,06	0,13	0,10	0,09	0,09	0,34
MgO	wt%	0,27	0,17	11,09	0,35	0,09	0,14	0,12	8,17	0,26	0,33	0,26	0,01	1,8	2,56	0,03	0,01	< .01	0,07
Al2O3	wt%	16,95	11,30	8,19	28,33	12,27	15,5	19,05	14,34	15,31	18,03	25,28	0,2	0,2	0,77	10,01	8,35	9,04	0,64
SiO2	wt%	55,01	73,41	30,87	42,54	73,71	65,1	39,98	27,99	61,09	45,74	47,05	98,18	89,02	7,37	75,88	80,60	80,21	97,33
K 2O	wt%	0,07	0,12	0,39	0,12	0,15	0,2	0,09	0,09	0,16	0,09	0,17	0,02	0,05	0,24	9,03	7,36	8,22	0,04
CaO	wt%	2,46	1,33	15,63	1,38	0,46	1,47	0,3	13,34	3,18	2,15	1,74	0,15	2,76	48,17	0,27	0,54	0,34	0,37
MnO	wt%	0,01	0,00	0,06	0,01	0,01	0,01	0,01	0,08	0,01	0,03	0,01	0	0,04	0,02	< .01	< .01	< .01	0,01
TiO2	wt%	1,20	0,93	0,47	1,44	0,87	0,94	1,13	0,43	1,04	0,85	1,29	0,09	0	0,04	0,52	0,49	0,53	0,04
FeOtot	wt%	0,94	1,03	3,16	2,08	0,78	0,93	19,85	0,79	3,58	17,02	1,82	0,64	1,90	0,73	4,41	1,42	0,54	0,40
P2O5	wt%	0,11	0,07	0,07	0,12	0,09	0,12	0,25	0,13	0,18	0,31	0,15	0,03	0,01	0,06000	0,16	0,39	0,28	0,02
Cr2O3	wt%	0,009	0,004	0,014	0,035	0,011	0,014	0,017	0,021	0,011	0,015	0,019	0,002	0,003	0,000	0,003	0,002	0,002	0,002
LOI	wt%	13,30	5,50	25,5	15	5,3	7,4	10,5	26,5	8,4	6,01	12,9	0,70	4,05	37,91	1,20	1,20	0,70	0,7
Total	wt%	99,45	99,60	99,59	99,67	99,81	99,80	97,69	99,75	99,33	99,71	99,71	100,14	99,89	98,00	101,61	100,45	99,95	99,96
V	ppm	40,0	33,0	38,0	172,0	43,0	137,0	362,0	84,0	47,0	224,0	38,0	6,0	5,0	7,0	31,0	28,0	31,0	6,0
Co	ppm	0,3	1,0	9.7	1,0	1,9	2,8	2,6	3,0	5,1	21,6	2,3	1,5	1,2	1,2	15,6	7,2	2,6	0,8
Ni	ppm	8,2	6,4	23.5	13,4	10,1	10,9	30,7	5,2	15,3	40,3	15,9	0,0	0,0	0,0	20,6	9,3	2,1	2,4
As	ppm	6,0	9,8	2,9	3,6	1,2	1,0	8,9	47,1	7,2	55,6	7,2	0,0	9,9	0,8	30,3	18,5	9,8	1,2
Rb	ppm	1,5	8,9	41,6	8,5	12,6	19,6	14,1	4,9	7,2	4,8	15,5	1,0	1,8	7,6	269,0	246,0	268,6	1,5
Sr	ppm	175,0	94,0	284,5	301,7	105,6	125,0	105,8	218,7	156,1	98,5	220,0	23,6	35,2	226,1	139,2	96,9	81,2	33,4
Zr	ppm	154,2	110,5	195,8	334,6	477,7	493,2	457,1	140,9	773,4	464,0	400,5	264,6	6,3	9,8	183,9	205,1	180,2	41,1
Bi	ppm	0,3	0,2	>2000	1,0	0,0	0,4	259,0	<.1	0,2	0,1	0,1				< .1	0,1	< .1	2,3
Sb	ppm	1,1	1,1	0,7	0,6	0,3	0,2	0,6	0,5	0,2	1,0	0,1	0,3	0,6	0,1	2,2	1,1	0,3	0,3
Ga	ppm	25,0	14,1	8,6	37,5	14,3	22,1	29,6	6,9	15,0	18,9	31,1	1,1	2,6	1,7	15,6	12,5	14,9	1,0
Nb	ppm	22,1	17,4	20,3	40,9	19,6	18,3	27,3	33,8	25,4	24,3	29,5	8,7	0,7	1,2	68,0	38,1	34,4	2,6
Cd	ppm	0,1	0,0	0,4	<.1	<.1	0,1	<.1	0,1	0,1	0,1	<.1	0,1	0,3	0,0	0,2	< .1	< .1	0,1
Cs	ppm	29,4	76,1	54,3	73,8	94,7	24,8	26,1	78,2	45,0	97,3	40,7	0,6	0,2	0,6	1,1	1,0	1,1	2,0
Ba	ppm	105,0	181,0	140,7	277,7	468,8	482,7	72,8	318,4	228,0	404,0	310,4	393,0	494,0	82,0	3526,0	1505,0	1292,0	415,6
Hf	ppm	3,8	3,2	4,4	8,3	12,2	12,3	12,3	3,6	20,4	11,4	11,1	3,3	0,0	0,0	4,3	4,1	3,9	0,9
Ta	ppm	1,2	1,0	4,9	2,0	1,3	1,5	5,7	0,7	1,6	1,2	1,9	0,0	0,0	0,0	0,7	0,7	0,7	0,1
Pb	ppm	25,6	16,9	45,2	5,2	6,9	12,2	11,8	3,1	8,1	8,5	2,5	13,1	12,7	2,9	3,4	4,4	2,5	7,9
Zn	ppm	21,2	20,6	24,0	13,0	26,0	35,0	34,0	9,0	27,0	68,0	24,0	8,0	25,0	6,0	62,0	16,0	1,0	14,0
Ag	ppb	<20	<20	< 1	0,1	<.1	0,1	< 1	< 1	<.1	0,1	0,3				< 1	0,1	< .1	<.1
Mo	ppm	0,5	0,8	1,3	0,2	0,5	0,4	1,4	0,4	0,5	2,9	0,2	8,5	2,9	3,4	2,5	0,6	0,2	0,3
Cu	ppm	11,4	7,5	16,3	9,0	5,9	7,3	43,2	73,5	169,1	113,8	57,8	9,2	1,7	3,6	18,0	11,1	4,1	6,2
Sn	ppm	2,7	2,3	2.0	4,0	2,0	3,0	5,0	1,0	3,0	3,0	4,0	<1	<1	<1	2,0	2,0	2,0	<1
W	ppm	28,2	11,9	7.9	37,1	18,2	36,5	52,0	3,5	31,3	38,2	36,6	4,7	0,3	0,3	19,0	21,0	13,3	1,1
Au	ppm	<0,1	<0,1	<.5	1,5	<.5	1,0	2,7	6,3	4,9	4,0	2,7	<,5	<,5	<,5	< .5	1,6	< .5	0.6
Th	ppm	47,2	37,9	299,4	29,5	19,2	34,1	25,3	122,3	25,0	49,4	30,6	220,9	2,0	0,2	105,4	100,0	50,3	34,9
Be	ppm	1,0	1,0	3,0	1,0	2,0	1,0	2,0	2,0	1,0	2,0	<1		.	.				<1
Y	ppm	11,9	20,5	75,8	29,0	25,4	29,8	28,6	137,5	46,1	31,4	28,0	49,6	4,3	4,3	61,1	80,7	49,7	13,9
Sc	ppm	8,3	4,6	18,0	14,0	6,0	6,0	18,0	24,0	11,0	11,0	12,0	2,0	1,0	1,0	5,0	2,0	2,0	1,0
U	ppm	1,1	1,5	1,1	3,3	1,7	1,8	2,4	6,7	3,2	2,6	2,2	0,5	0,3	0,4	3,5	4,0	3,3	0,3
La	ppm	23,0	39,0	21,0	67,5	37,2	46,2	52,5	19,0	34,0	27,3	59,5	3,0	0,0	5,3	25,4	25,6	21,9	5,9
Ce	ppm	49,5	67,3	45,5	125,9	69,5	94,9	94,6	47,1	68,8	52,8	122,2	9,6	0,9	9,7	56,9	57,0	48,1	21,0
Pr	ppm	5,3	6,6	6,1	14,4	7,9	9,9	10,7	4,7	8,1	5,7	14,1	1,7	0,1	1,2	6,0	6,2	5,2	1,6
Nd	ppm	18,9	21,3	28.1	45,1	26,1	34,6	32,6	21,8	29,1	19,1	48,5	8,8	0,5	4,2	25,4	26,0	20,8	6,5
Sm	ppm	3,0	4,0	11,0	5,9	4,6	6,4	4,4	10,0	6,5	3,8	7,1	8,4	0,4	0,9	8,6	8,9	6,8	1,5
Eu	ppm	0,7	0,9	3,7	1,0	1,0	1,2	0,8	4,8	1,4	1,2	1,3	3,7	0,2	0,2	2,9	3,1	2,1	0,3
Gd	ppm	2,1	3,3	15.0	3,6	3,3	4,2	3,2	20,0	6,4	4,8	4,3	14,2	0,7	0,7	11,4	12,3	8,5	1,8
Tb	ppm	0,4	0,8	2,8	0,7	0,7	0,8	0,8	5,5	1,4	1,2	0,9	2,5	0,2	0,1	2,2	2,6	1,7	0,4
Dy	ppm	2,7	4,0	15,3	4,7	4,4	4,7	4,7	29,4	7,5	6,1	5,1	11,5	0,9	0,7	11,4	14,4	9,2	2,4
Ho	ppm	0,5	0,7	2,7	1,0	0,9	1,1	1,0	5,3	1,6	1,1	1,0	1,8	0,2	0,1	2,2	2,8	1,7	0,5
Er	ppm	1,9	2,5	7,1	3,2	2,8	3,2	3,3	12,4	4,6	3,3	3,1	3,9	0,0	0,3	5,4	6,9	4,4	1,3
Tm	ppm	0,3	0,3	0,9	0,6	0,4	0,5	0,5	1,5	0,7	0,5	0,5	0,5	0,0	0,0	0,7	0,8	0,6	0,2
Yb	ppm	2,0	2,2	5,5	3,9	2,9	3,4	3,5	7,9	4,6	3,2	3,5	3,1	0,3	0,3	4,2	5,0	3,2	0,8
Lu	ppm	0,2	0,3	0,7	0,6	0,4	0,5	0,5	1,1	0,8	0,5	0,6	0,4	0,0	0,0	0,5	0,7	0,5	0,1

APPENDICE C

ANALYSES ISOTOPES STABLES

Description	Échantillons	$\delta^{13}C$	$\delta^{18}O$
Filon de carbonatite	24	-3,27	19,95
Brèche carbonatée	27	1,74	35,05
Filon de carbonatite	28	-1,98	19,81
Filon de carbonatite	30	-2,98	18,91
Filon de carbonatite	33	-1,62	21,56
Filon de carbonatite	34	-1,02	19,48
Filon de carbonatite	35	-2,63	20,17
Filon de carbonatite	50	-1,97	19,53
Sill de carbonatite	51	-1,02	19,86
Carbonatite ankéritisée	51b	-3,37	17,70
Filon de carbonatite	55	1,23	26,96

Analyses des isotopes stables de carbone et d'oxygène sur les carbonatites du Richat. Les valeurs isotopiques en carbone et oxygène sont exprimées en ‰ du ratio standard PDB (*Pee Dee belemnite*) et SMOW (*Standard Mean Ocean Water*) respectivment.

Échantillon	$\delta^{18}O$
Bx Chert 1	25,0
Bx Chert 1x	25,8
Bx Chert 1y	25,2
Bx Chert 2	20,3
Bx Chert 4	27,0
Bx Chert 6	25,7
Bx Chert 9	25,9
Bx Chert 9x	26,5
Bx Chert 10	20,1
Bx Chert 13	26,7
Bx Chert 17	23,0
Bx chert 18	23,9
Bx chert 22	14*
Bx chert 38	25,1
Bx chert 40	24,6
Bx chert 42	22,5
Séd. Interne 67	24,08
Séd. Interne 68	24,81
Séd. Interne 69	23,78

*Faible pourcentage de réaction

Composition isotopique de l'oxygène exprimée en ‰ du ratio standard SMOW (*Standard Mean Ocean Water*) sur les brèches et sédiments internes du Richat.

LISTE DES RÉFÉRENCES COMPLÈTES

Acocella, V., Korme, T., Salvini, F., et Funiciello, R. 2002. Elliptic calderas in the Ethiopian Rift; control of pre-existing structures. Journal of Volcanology and Geothermal Research, v. 119, p. 189-203.

Acocella, V., Funiciello, R., Marotta, E., Orsi, G., et de Vita, S. 2004. The role of extensional structures on experimental calderas and resurgence. Journal of Volcanology and Geothermal Research, v. 129, p. 199-217.

Alberti, A., Castorina, F., Censi, P., Comin-Chiaramonti, P., et Gomes, C. B. 1999. Geochemical characteristics of Cretaceous carbonatites from Angola. Journal of African Earth Sciences, v. 29, p. 735-759.

Allsopp, H. L. et Hargraves, R. B. 1985. Rb-Sr ages and palaeomagnetic data for some Angolan alkaline intrusives. Transactions of the Geological Society of South Africa, v. 88, p. 295-299.

Almeida, F. M. 1991 O alinhamento magmatico de Cabo Frio. Simposio de Geologia do Sudeste, v. 2, p. 423-428.

Alves, T. M., Gawthorpe, R. L., Hunt, D. W., et Monteiro, J. H. 2003. Cenozoic tectono-sedimentary evolution of the western Iberian margin. Marine Geology, v. 195, p. 75-108.

Amaral, G., Bushee, J., Cordani, U. G., Kawashita, K., et Reynolds, J. H. 1967. Potassium-argon ages of alkaline rocks from southern Brazil. Geochimica et Cosmochimica Acta, v. 31, p. 117-142.

Anderson, D. L. 1994. Superplumes or supercontinents? Geology, v. 22, p. 39-42.

Anderson, D. L. 1995. Lithosphere, asthenosphere, and perisphere. Reviews of Geophysics, v. 33, p. 9-41.

Anderson, D. L. 2000. The thermal state of the upper mantle; no role for mantle plumes. Geophys. Res. Lett., v. 27, p. 3623-3626.

Bailey, D. K. 1992. Episodic alkaline igneous activity across Africa; implications for the causes of continental break-up. In Storey, B. C., Alabaster, T. et Pankhurst, R. J., éd., Magmatism and the causes of continental break-up: Geological Society of America, Special Paper, v. 68, p. 91-98.

Bailey, D. K. et Woolley, A. R., 2005, Repeated, synchronous magmatism within Africa: Timing, magnetic reversals, and global tectonics. In Foulger, G. R., Natland, J. H., Presnall, D. C., et Anderson, D. L., éd., Plates, plumes, and paradigms. Geological Society of America , Special Paper, v. 388, p. 365-377.

Baksi, A. K., 1997, The timing of Late Cretaceous alkalic igneous activity in the northern Gulf of Mexico basin, southeastern USA. Journal of Geology, v. 105, Issue 5, p. 629-643.

Bardossy, G., Monod, T., et Pomerol, C. 1963. Découverte d'analcimolites d'origine endogène dans les Richat (Adrar mauritanien). Comptes Rendues de l'Académie des Sciences, v. 256, p. 3934-3936.

Bayer, M. et Lesquer, A. 1978. Les anomalies gravimétriques de la bordure orientale du craton Ouest-africain. Géométrie d'une suture panafricaine. Bulletin de la Société Géologique de France, v. 7, p. 863-876.

Bédard, J. H. 1985. The opening of the Atlantic, the Mesozoic New England igneous province, and mechanisms of continental break-up. Tectonophysics, v. 113, p. 209-232.

Bellion, Y. et Crevola, G. 1991. Cretaceous and Cainozoic magmatism of the Senegal Basin (West Africa); a review. In Kampunzu, A. B., et Lubala, R. T., éd., Magmatism in extensional structural settings; the Phanerozoic African Plate. Berlin: Springer-Verlag, p. 189-208.

Bernard-Griffiths, J., Gruau, G., Cornen, G., Azambre, B., et Mace, J. 1997. Continental lithospheric contribution to alkaline magmatism; isotopic (Nd, Sr, Pb) and geochemical (REE) evidence from Serra de Monchique and Mount Ormonde complexes. Journal of Petrology, v. 38, p. 115-132.

Bertrand-Sarfati, J., Moussine-Pouchkine, A., et Caby, R. 1987. Les corrélations du Protérozoïque au Cambrien en Afrique de l'Ouest; nouvelle interprétation géodynamique. Bulletin de la Société Géologique de France, v. 3, p. 855-865.

Bérubé, D., et Jébrak, M. 1999. High precision boundary fractal analysis for shape characterization. Computers & Geosciences, v. 25, p. 1059–1071.

Binks, R. M., et Fairhead, J. D. 1992. A plate tectonic setting for Mesozoic rifts of West and Central Africa. Tectonophysics, v. 213, p. 141-151.

Blanc, P. et Pomerol, C. 1973. Étude pétrographique des roches magmatiques, des filons carbonatés et des analcimolites des Richat. In Monod, T. et Pomerol, C., éd., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy : Sciences de la Terre, Mémoires 28, p. 107-121.

Blanchot, A., et Weppe, M. 1951. La brèche d'Idjil en Mauritanie et sa genèse probable. Compte Rendu Sommaire des Séances de la Société Géologique de France, v. 11-12, p. 185-187.

Bonatti, E. 1996. Anomalous opening of the Equatorial Atlantic due to an equatorial mantle thermal minimum. Earth and Planetary Sciences Letters, v. 143, p. 147-160.

Bonin, B. 1995. Pétrologie endogène. Paris: Dunod, Collection Géosciences, 336 p.

Bosworth, W. 1994. A model for the three-dimensional evolution of continental rift basins, North-east Africa. Geologische Rundschau, v. 83, p. 671-688.

Bosworth, W., Burke, K., et Strecker, M. 2000. Magma chamber elongation as an indicator of intraplate stress field orientation; 'borehole breakout mechanism' and examples from the late Pleistocene to Recent Kenya Rift valley. In Jessell, M. W. et Urai, J. L., éd., Stress, strain and structure; a volume in honor of W. D. Means. Journal of the Virtual Explorer online, v. 2

Boussaroque, J.-L. 1975. Étude des analcimolites des Richat (Adrar de Mauritanie). Thèse de doctorat, Paris, Université de Paris VI, 105 p.

Bronner, G. 1992. Structure et évolution d'un craton archéen: la dorsale Réguibat occidentale (Mauritanie); tectonique et métallogénie des formations ferrifères. Orléans : BRGM, 448 p.

Burke, K. 1977. Aulacogens and continental breakup. Ann. Rev. Earth Planet. Sci., v. 5, p. 371-396.

Burke, K. et Dewey, J. F. 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. Journal of Geology, v. 81, p. 406-433.

Burke, K., et Wilson, J. T. 1976. Hot spots on the Earth's surface. Sci. Amer., v. 235, p. 46-57.

Byerly, G. R. 1991. Igneous activity: Boulder, CO, United States, v. J, p. 91-108.

Cailleux, A., Guillemaut, A., et Pomerol, C. 1964. Présence de coésite, indice de hautes pressions dans l'accident circulaire des Richat (Adrar mauritanien). Comptes Rendues de l'Académie des Sciences, v. 258, p. 5488-5490.

Campbell, I, H., et Griffiths, R. W. 1990. Implications of mantle plume structure for the evolution of flood basalts. Earth and Planetary Sciences Letters, v. 99, p. 79-93.

Carmichael, I. S. E. 1964. The petrology of Thingmuli, a Tertiary volcano in eastern Iceland. Journal of Petrology, v. 5, p.435-460.

Castorina, F., Censi, P., Comin-Chiaramonti, P., Piccirillo, E. M., Alcover Neto, A., Gomes, C. B., Ribeiro de Almeida, T. I., Speziale, S., et Toledo, M. C. M. 1997. Carbonatites from eastern Paraguay and genetic relationships with potassic magmatism; C, O, Sr and Nd isotopes. Mineralogy and Petrology, v. 61, p. 237-260.

Chamov, N. P., Kostyleva, V. V., Veis, A. F., et Gorbachev, V. I. 2003. Late Riphean sedimentation in the central russian aulacogen. Lithology and Mineral Resources, v. 38, p. 458-467.

Chang, H. K., Kowsmann, R. O., Figueiredo, A. M. F., et Bender, A. A. 1992. Tectonics and stratigraphy of the East Brazil Rift system: an overview. Tectonophysics, v. 213, p. 97-138.

Clemson, J., Cartwright, J. A., et Booth, J. 1997. Structural segmentation and the influence of basement structure on the Namibian passive margin. Journal of the Geological Society of London, v. 154, p. 477-482.

Clemson, J., Cartwright, J., et Swart, R. 1999. The Namib Rift; a rift system of possible Karoo age, offshore Namibia. In Cameron, N. R., et Bate, R. H., éd., The oil and gas habitats of the South Atlantic. Geological Society of London, Special Publications, v. 153, p. 381-402.

Cobbold, P. R., Meisling, K. E., et Mount, V. S. 2001. Reactivation of an obliquely rifted margin, Campos and Santos basins, southeastern Brazil. AAPG Bulletin, v. 85, p. 1925-1944.

Comin-Chiaramonti, P., Meriani, S., Mosca, R., et Sinigoi, S. 1979. On the occurrence of analcime in the northeastern Azerbaijan volcanics (northwestern Iran). Lithos, v. 12, p. 187-198.

Comin-Chiaramonti, P. et Gomes, C. B. 1996. Alkaline magmatism in central-eastern Paraguay; relationships with coeval magmatism in Brazil: EDUSP, 458 p.

Comin-Chiaramonti, P., Cundari, A., DeGraff, J. M., Gomes, C. B., et Piccirillo, E. M. 1999. Early Cretaceous-Tertiary magmatism in eastern Paraguay (western Parana Basin); geological, geophysical and geochemical relationships. Journal of Geodynamics, v. 28, p. 375-391.

Condie, K. C. 2001. Mantle plumes and their record in earth history. Cambridge: Cambridge University Press, 305 p.

Cornen, G. 1982. Petrology of the alkaline volcanism of Gorringe Bank (Southwest Portugal). Marine Geology, v. 47, p. 101-130.

Cortes, A. L., Soriano, M. A., Maestro, A., et Casas, A. M. 2003. The role of tectonic inheritance in the development of recent fracture systems, Duero Basin, Spain. International Journal of Remote Sensing, v. 24, p. 4325-4345.

Courtillot, V., Davaille, A., Besse, J., et Stock, J. 2003. Three distinct types of hotspots in the Earth's mantle. Earth and Planetary Sciences Letters, v. 205, p. 295-308.

Cox, K. G. et Hawkesworth, C. J. 1985. Geochemical stratigraphy of the Deccan Traps at Mahabaleshwar, western Ghats, India, with implications for open system magmatic processes. Journal of Petrology, v. 26, p. 355-377.

Daly, M. C., Chorowicz, J., et Fairhead, J. D. 1989. Rift basin evolution in Africa; the influence of reactivated steep basement shear zones. Geological Society Special Publications, v. 44, p. 309-334.

Dani, N., Formoso, M. L. L., Decarreau, A., et Meunier, A. 2001. Nordstrandite in bauxite derived from phonolite, Lages, Santa Catarina, Brazil. Clays and Clay Minerals, v. 49, p. 216-226.

Davies, G. F. 1999. Dynamic Earth: Plates, plumes and mantle convection. Cambridge: Cambridge University Press, 458 p.

De Min, A., Piccirillo, E. M., Marzoli, A., Bellieni, G., Renne, P. R., Emesto, M., et Marques, L. S. 2003. The Central Atlantic Magmatic Province (CAMP) in Brazil-petrology, geochemistry, $^{40}Ar/^{39}Ar$ ages, paleomagnetism and geodynamic implications. In Hames, W. E., McHone, J. G., Renne, P. R., et Ruppel, C., éd., The Central Atlantic Magmatic Province: Insights from Fragments of Pangea. AGU, Geophysical Monograph, v. 136, p. 91-128.

Deckart, K., Bertrand, H., et Liegeois, J.-P. 2005. Geochemistry and Sr, Nd, Pb isotopic composition of the Central Atlantic Magmatic Province (CAMP) in Guyana and Guinea. Lithos, v. 82, p. 289-314.

Deines, P. 1989. Stable isotope variations in carbonatites. In Bell, K., ed, Carbonatites; genesis and evolution. London: Unwin Hyman, U. K., p. 301-359.

Destombes, J.-P., et Plote, H. 1962. Une structure aberrante dans le Précambrien du Sahara mauritanien; hypothèse nouvelle sur l'origine du Richat (Adrar mauritanien). Comptes rendue sommaire de la Société Géologique de France, v. 7. p. 202-203.

Dietz, R. S., Fudali, R., et Cassidy, W. A. 1969. Richat and Semsityat Domes Mauritania: Not astroblemes. Geological Society of America Bulletin, v. 80, p. 1367-1372.

Eby, G. N. 1975. Abundance and distribution of the rare-earth elements and yttrium in the rocks and minerals of the Oka carbonatite complex, Quebec. Geochimica et Cosmochimica Acta, v. 39, p. 597-620.

Eby, G. N. 1984. Geochronology of the Monteregian Hills alkaline igneous province, Quebec. Geology, v. 12, p. 468-470.

Edel, J.-B., Schulmann, K., et Rotstein, Y. 2007. The Variscan tectonic inheritance of the Upper Rhine Graben; evidence of reactivations in the Lias, late Eocene-Oligocene up to the Recent. Geologische Rundschau, v. 96, p. 305-325.

Eldridge, J., Walsh, C.B., et Scotese, C.R. 1998. Plate Tracker for Windows version 2.0, PALEOMAP Project, Arlington, Texas.

Ernst, R. E., et Desnoyers, D. W. 2004. Lessons from Venus for understanding mantle plumes on Earth. Physics of the Earth and Planetary Interiors, v. 146, p. 195-229.

Fabre, J. 1999. In Quinze jours avec Théodore Monod. Eurêka, v. 41, p. 36-47.

Fairhead, J. D., et Binks, R. M. 1991, Differential opening of the Central and South Atlantic Oceans and the opening of the West African Rift System. Tectonophysics, v. 187, p. 191-203.

Fairhead, J. D. et Wilson, M. 2005. Plate tectonic processes in the South Atlantic Ocean; do we need deep mantle plumes? In Foulger, G.R., Natland, J.H., Presnall, D.C., et Anderson, D.L., éd., Plates, plumes, and paradigms. Geological Society of America, Special Paper, v. 388, p. 537-553.

Farley, K. A., et Neroda, E. 1998. Noble gases in the Earth's mantle. Annual Review of Earth and Planetary Sciences, v. 26, p. 189-218.

Faure, S., Tremblay, A., et Angelier, J. 1996. State of intraplate stress and tectonism of northeastern America since Cretaceous times, with particular emphasis on the New England-Quebec igneous province. Tectonophysics, v. 255, p. 111-134.

Foland, K. A. et Faul, H. 1977. Ages of the White Mountain intrusives; New Hampshire, Vermont, and Maine, USA. American Journal of Science, v. 277, p. 888-904.

Foland, K. A., Gilbert, L. A., Sebring, C. A., et Chen J-F. 1986. $^{40}Ar/^{39}Ar$ ages for plutons of the Monteregian Hills, Quebec; evidence for a single episode of Cretaceous magmatism. Geological Society of America Bulletin, v. 97, p. 966-974.

Francheteau, J. et Le Pichon, X. 1972. Marginal Fracture Zones as Structural Framework of Continental Margins in South Atlantic Ocean. American Association of Petroleum Geologists Bulletin, v. 56, p. 991-1007.

Frese, R. R. B. 1982. Regional anomalies of the Mississippi River aulacogen. Society of Exploration Geophysicists, Abstracts with Programs, v. 1, p.295-297.

Fudali, R. F. 1973. Origin of the analcime-bearing rocks of Richat. In Monod, T., et Pomerol, C., éd., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, p. 97-105.

Fuller, E. R., et Head, J. W. 2003. Olympus Mons, Mars: Detection of extensive preaureole volcanism and implications for initial mantle plume behavior. Geology, v. 31, p. 175-178.

Garrote, J., Cox, R. T., Swann, C., et Ellis, M. 2006. Tectonic geomorphology of the southeastern Mississippi Embayment in northern Mississippi, USA. Geological Society of America Bulletin, v. 118, p. 1160-1170.

Gomes, C. B., Milan, A., Velazquez, V. F., Riccomini, C., Comin-Chiaramonti, P., Vasconcelos, P. M., et Tassinari, C. C. G. 2004. Magmatismo alcalino na porcao centro-oriental do Paraguay: novos dados geocronologicos para as rochas das provyncias Central e Assuncao. Rev. Fac. Cienc., Univ. Eduardo Mondane, Maputo, Moçambique, ed. Esp. 1, p. 179-183.

Gromet, L.P., Dymek, R.F., Haskin, L.A., et Korotev, R.L. 1984. The 'North American shale composite': Its compilation, major and trace element characteristics. Geochimica et Cosmochimica Acta, v. 48, p. 2469–2482.

Guetat, Z. 1981. Étude gravimétrique de la bordure occidentale du craton ouest-africain. Thèse de doctorat, Montpellier, Université de Montpellier, 178 p.

Guiraud, R., Issawi, B., et Bellion, Y. 1985. Les linéaments guineo-nubiens; un trait structural majeur à l'échelle de la plaque africaine. Comptes Rendus de l'Académie des Sciences, v. 300, p. 17-20.

Guiraud, R., Bellion, Y., Benkhelil, J., et Moreau, C. 1987. Post-Hercynian tectonics in northern and western Africa. In Bowden, P., et Kinnaird, J., éd., African Geology Reviews. London: John Wiley and Sons, p.433-466.

Guiraud, R. et Maurin, J.C. 1992. Early Cretaceous rifts of Western and Central Africa: an overview. Tectonophysics, v. 213, p. 153-168.

Guiraud, R., Binks, R. M., Fairhead, J. D., et Wilson, M. 1992. Chronology and geodynamic setting of Cretaceous-Cenozoic rifting in West and Central Africa. Tectonophysics, v. 213, p. 227-234.

Gupta, A. K. et Fyfe, W. S. 1975. Leucite survival; the alteration to analcime. Canadian Mineralogist, v. 13, p. 361-363.

Hames, W. E., McHone, J. G., Renne, P. R., et Ruppel, C. 2003. The Central Atlantic Magmatic Province; insights from fragments of Pangea. American Geophysical

Union, Geophysical Monograph, v. 136, 267 p.

Hasui, Y. et Cordani, U. G. 1968. Idades potassio-argonio de rochas do macicos intrusivos mesozoicos do oeste Mineiro e sul de Goias. Congr. Brasil. Geol., v. 22 d, p. 5.

Heatherington, A. L. et Mueller, P. A. 1999. Lithospheric sources of North Florida, USA tholeiites and implications for the origin of the Suwannee Terrane. Lithos, v. 46, p. 215-233.

Hegenberger, W. 1988. Karoo sediments of the Erongo Mountains; their environmental setting and correlation. Communications of the Geological Survey of South West Africa/Namibia, v. 4, p. 51-57.

http://asterweb.jpl.nasa.gov/gallery-detail.asp?name=Richat.

Irvine, T. N. et Baragar, W. R. A. 1971. A guide to the chemical classification of the common volcanic rocks. Canadian Journal of Earth Sciences, v. 8, p. 523-548.

Jacques, J. M. 2003. A tectonostratigraphic synthesis of the Sub-Andean basins: inferences on the position of South American intraplate accommodation zones and their control on South Atlantic opening. Journal of the Geological Society, v. 160, p. 703-717.

Jagoutz, O., Müntener, O., Manatschal, G., Rubatto, D., Péron-Pinvidic, M., G., Turrin, B. D. et Villa, I. M. 2007. The rift-to-drift transition in the North Atlantic: A stuttering start of the MORB machine? Geology, v. 35, no. 12, p. 1087-1090.

Janssen, M. E., Stephenson, R. A., et Cloetingh, S. 1995. Temporal and spatial correlations between changes in plate motions and the evolution of rifted basins in Africa. Geological Society of America Bulletin, v. 107, p. 1317-1332.

Jelsma, H. A., de Wit, M. J., Thiart, C., Dirks, P. H. G. M., Viola, G., Basson, I. J. et Anckar, E. 2004. Preferential distribution along transcontinental corridors of kimberlites and related rocks of Southern Africa. South African Journal of Geology, v. 107, p. 301-324.

Jourdan, F., Marzoli, A., Bertrand, H., Cosca, M., et Fontignie, D. 2003. The northernmost CAMP; $^{40}Ar/^{39}Ar$ age, petrology and Sr-Nd-Pb isotope geochemistry of the Kerforne Dike, Brittany, France. In Hames, W. E., McHone, J. G., Renne, P. R., et Ruppel, C., éd., The Central Atlantic Magmatic Province: Insights from Fragments of Pangea. AGU, Geophysical Monograph, v. 136, p. 209-226.

Keller, J. et Hoefs, J. 1995. Stable isotope characteristics of natrocarbonatites from Oldoinyo Lengai. In Bell, K., et Keller, J., éd., Carbonatite Volcanism: Oldoinyo Lengai and the Petrogenesis of Natrocarbonatites. IAVCEI Proceedings in Volcanology, v. 4, p. 113–123.

King, S. D. 2007. Hotspots and edge-driven convection. Geology, v. 35, issue 3, p. 223-226.

King, S. D. et Anderson, D. L. 1995. An alternative mechanism of flood basalt formation. Earth and Planetary Sciences Letters, v. 136, p. 269-279.

King, S. D., et Anderson, D. L. 1998. Edge-driven convection. Earth and Planetary Sciences Letters, v. 160, p. 289-296.

King, S. D. et Ritsema, J. 2000. African hot spot volcanism; small-scale convection in the upper mantle beneath cratons. Science, v. 290, p. 1137-1140.

Klitgord, K. D., et Schouten, H. 1986. Plate kinematics of the Central Atlantic. In Tucholke, B. E., et Vogt, P. P., éd., The Western North Atlantic region. (The Geology of North America). Geological Society of America, v. M, p. 351-378.

Knipe, R.J. 1993. The influence of fault zone processes and diagenesis on fluid flow. In Horbury, A.D., et Robinson, A.G., éd., Diagenesis and basin development. American Association of Petroleum Geologists Studies in Geology, v. 36, p. 135–151.

Krassilnikov, A. S. 2002. Tectonics and evolution of novae and coronae on Venus: tectonophysical modeling based on gravitational models. Solar System Research, v. 36, p. 293-321.

Kumarapeli, P. S. 1985. Vestige of Iapetan Rifting in the Craton West of the Northern Appalachians. Geoscience Can., v. 12, p. 54-59.

Kumarapeli, P. S. et Saull, V. A. 1966. The St. Lawrence Valley system: A North American equivalent of the East African Rift Valley system. Canadian Journal of Earth Sciences, v. 3, p. 639-658.

Lapido-Loureiro, F. E. 1973. Carbonatitos de Angola. Memorias e Trabalhos do Instituto de Investigacao Cientifica de Angola, v. 11, 242 p.

Larson, R. L. 1991a. Latest pulse of Earth: evidence for a mid-cretaceous superplume. Geology, v. 19, p. 547-550.

Larson, R. L. 1991b. Geological consequences of superplumes. Geology, v. 19, no. 10, p. 963-966.

Lay, T. 2005. The deep mantle thermo-chemical boundary layer: the putative mantle plume source. In Foulger, G.R., Natland, J.H., Presnall, D.C., et Anderson, D.L., éd., Plates, plumes, and paradigms. Geological Society of America, Special Paper, v. 388, p. 193-205.

Le Bas, M. J., Le Maitre, R. W., Streckeisen, A., et Zanettin, B. A. 1986. Chemical classification of volcanic rocks based on the total alkali-silica diagram. Journal of Petrology, v. 27, p. 745-750.

Le Page, A., et Lécorché, J.-P. 1991. La chaîne des Mauritanides. Un exemple d'orogène polyphasé, pan-africain à hercynien, en Mauritanie. In Caruba, R., et Dars, R., éd., Géologie de la Mauritanie, Nice : CRDP, p.79-94.

Le Pichon, X. et Hayes, H. E. 1971. Marginal Offsets, Fracture Zones, and the Early Opening of the South Atlantic. Journal of Geophysical Research, v. 76, p. 6283-6293.

Leitch, A. M., Davies, G. F., et Wells, M. 1998. A plume head melting under a rifting margin. Earth and Planetary Sciences Letters, v. 161, p. 161-177.

Lemberg, J. 1876. Ueber Silicatumwandlungen. Z. Deut. Geol. Ges., v. 28, p. 519-621.

Liegeois, J. P., Sauvage, J. F., et Black, R. 1991. The Permo-Jurassic alkaline province of Tadhak, Mali; geology, geochronology and tectonic significance. Lithos, v. 27, p. 95-105.

Lorilleux, G., Jébrak, M., Cuney, M., et Baudemont, D. 2002. Polyphase hydrothermal breccias associated with unconformity-related uranium mineralization (Canada): From fractal analysis to structural significance. Journal of Structural Geology, v. 24, p. 323–338.

Loubet, M., Bernat, M., Javoy, M., et Allegre, C. J. 1972. Rare earth contents in carbonatites. Earth and Planetary Science Letters, v. 14, p. 226-232.

Macintyre, R. M. et Berger, G. W. 1982. A note on the geochronology of the Iberian Alkaline Province. Lithos, v. 15, p. 133-136.

Mariano, A. N. 1989. Nature of economic mineralisation in carbonatites and related rocks. In Bell, K., éd., Carbonatites; genesis and evolution. London: Unwin Hyman, U. K., p. 149-176.

Marquis, R. et Kumarapeli, P. S. 1993. An Early Cambrian deltaic-fluvial model for an Iapetan rift-arm drainage system, southeastern Quebec. Canadian Journal of Earth Sciences, v. 30, p. 1254-1261.

Marsh, J. S. 1973. Relationships between transform directions and alkaline igneous rock lineaments in Africa and South America. Earth and Planetary Sciences Letters, v. 18, p. 317-323.

Marzoli, A., Renne, P. R., Piccirillo, E. M., Ernesto, M., Bellieni, G., et De Min, A. 1999. Extensive 200-Millon-Year-Old Continental flood basalts of the central Atlantic Magmatic Province. Science, v. 284, p. 616-618.

Mascle, J., Marinho, M., et Wannesson, J. 1986. The structure of the Guinean continental margin: implications for the connection between the Central and the South Atlantic Oceans. Geol. Rundsch., v. 75, p. 57-70.

Mascle, J., Blarez, E., et Marinho, M. 1988. The shallow structures of the Guinea and Ivory Coast-Ghana transform margins: their bearing on the Equatorial Atlantic evolution. Tectonophysics, v. 188, p. 193-209.

Matton, G., Jebrak, M., et Lee, J. K. W. 2005. Resolving the Richat enigma: Doming and hydrothermal karstification above an alkaline complex. Geology, v. 33, p. 665-668.

Matton, G. et Jebrak, M. 2006. A Monteregian-style intrusion in occidental Sahara? Geological and mineralogical associations of Canada annual meeting (GAC-MAC), v. 31, p. 97.

McDonough, W. F., Sun, S. S., Ringwood, A. E., Jagoutz, E., et Hofmann, A. W. 1992. Potassium, rubidium, and cesium in the Earth and Moon and the evolution of the mantle of the Earth. Geochimica and Cosmochimica Acta. v. 56, p. 1001-1012.

McHone, J. G. 1996. Constraints on the mantle plume model for Mesozoic alkaline intrusions in northeastern North America. Canadian Mineralogist, v. 34, p. 325-334.

McHone, J. G. 2000. Non-Plume magmatism and rifting during the opening of the central Atlantic Ocean. Tectonophysics, v. 316, p. 287-296.

McHone, J. G. et Butler, J. R. 1984. Mesozoic igneous provinces of New England and the opening of the North Atlantic Ocean. Geol. Soc. Am. Bull., v. 95, p. 757-765.

McHone, J. G., Ross, M. E., et Greenough, J. D. 1987. Mesozoic dykes swarms of eastern North America. In Halls, H. C., et Fahrig, W. F., éd., Mafic Dyke Swarms. Geological Society of America, Special Paper, v. 34, p. 279-288.

Milner, S. C., Le Roex, A. P., et O'Connor, J. M. 1995. Age of Mesozoic igneous rocks in northwestern Namibia, and their relationship to continental break-up. Journal of the Geological Society of London, v. 152, p. 97-104.

Mitchell, R. H. 1995. Kimberlites, orangeites, and related rocks. New York: Plenum Press, USA, 410 p.

Monod, T., et Pomerol, C. 1973. Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, 191 p.

Moreau, C., Ohnenstetter, D., Demaiffe, D., et Robineau, B. 1996. The Los Archipelago nepheline syenite ring-structure; a magmatic marker of the evolution of the Central and Equatorial Atlantic. Canadian Mineralogist, v. 34, p. 281-299.

Morgan, W. J. 1971. Convection plumes in the lower mantle. Nature, v. 230. p. 42-43.

NASA/GSFC/METI/ERSDAC/JAROS and U.S./Japan ASTER Science Team, "Richat

Netto, A. M., Fabre, J., Poupeau, G., et Champemmois, M. 1992. Datations par traces de fissions de la structure circulaire des Richats. Comptes Rendus de l'Académie des Sciences, v. 314, p. 1179-1186.

Nürnberg, D. et Müller, R. D. 1991. The tectonic evolution of the South Atlantic from Late Jurassic to present. Tectonophysics, v. 191, p. 27-53.

Onuonga, I. O., Fallick, A. E., et Bowden, P. 1997. The recognition of meteoric-hydrothermal and supergene processes in volcanic carbonatites, Nyanza Rift, western Kenya, using carbon and oxygen isotopes. Journal of African Earth Sciences, v. 25, p. 103-113.

Pearce, J. A. 1983. Role of sub-continental lithosphere in magma genesis in active continental margins. In Hawkesworth, C.J., et Norry, M.J., éd., Continental basalts and mantle xenoliths. Amsterdam: Shiva publications, p. 230–249.

Pegram, W. J. 1990. Development of continental lithospheric mantle as reflected in the chemistry of the Mesozoic Appalachian tholeiites, U.S.A. Earth and Planetary Science Letters, v. 97, p. 316-331.

Piboule, M., Fabre, J., et Monod, T. 1999. In Eureka, v. 41, p. 36-47.

Pindell, J. et Dewey, J. F. 1982. Permo-Triassic reconstruction of western Pangea and the evolution of the Gulf of Mexico/Caribbean region. Tectonics, v. 1, p. 179-211.

Pinheiro, L. M., Wilson, R. C. L., Pena dos Reis, R., Whitmarsh, R. B., et Ribeiro, A. 1996. The western Iberia Margin: a geophysical and geological overview. In Whitmarsh, R.B., Sawyer, D.S., Klaus, A., et Masson, D.G., éd., Proceedings of the Ocean Drilling Program, Scientific Results, v. 149, p.1-23.

Pirajno, F. 2004. Hotspots and mantle plumes: global intraplate tectonics, magmatism and ore deposits. Mineralogy and petrology, v. 82, p. 183-216.

PointTracker: PointTracker pour Windows, version 4c. Scotese, C. R., PALEOMAP Project, Arlington, Texas, 2002.

Poupeau, G., Fabre, J., Labrin, E., Azdimoussa, A., Netto, A. M., et Monod, T. 1996. Nouvelles datations par traces de fission de la structure circulaire des Richat (Mauritanie). Mémoires du Service Géologique de l'Algérie, v. 8, p. 231-236.

Prins, P. 1981. The geochemical evolution of the alkaline and carbonatite complexes of the Damaraland igneous province, South West Africa. Annall Universiteit van

Stellenbosch, Serie A, v. 3, p. 145-278.

Raab, M. J., Brown, R. W., Gallagher, K., Carter, A., et Weber, K. 2002. Late Cretaceous reactivation of major crustal shear zones in northern Namibia; constraints from apatite fission track analysis. Tectonophysics, v. 349, p. 75-92.

Rabinowitz, P. D. et LaBrecque, J. L. 1979. The Mesozoic South Atlantic Ocean and evolution of its continental margins. Journal of Geophysical Research, v. 84, p. 5973-6002.

Rahaman, M. A., Van Breemen, O., Bowden, P., et Bennet, J. N. 1984. Age migrations of anorogenic ring complexes in Northern Nigeria. Journal of Geology, v. 92, p. 173-184.

Redmond, H. L., et King, S. D. 2004. A numerical study of a mantle plume beneath the Tharsis Rise: Reconciling dynamic uplift and lithospheric support models. Journal of Geophysical research Planets, v. 109, p. 1-14.

Reid, D. L. et Cooper, A. F. 1992. Oxygen and carbon isotope patterns in the Dicker Willem carbonatite complex, southern Namibia. Chemical Geology, v. 94, p.2293-2405.

Rocci, G. 1991. Le socle ancien de Mauritanie : un élément de la croûte continentale stable du craton ouest-africain. In Caruba, R., et Dars, R., éd., Géologie de la Mauritanie, Nice : CRDP, p. 23-38.

Rock, N. M. S. 1982. The Late Cretaceous alkaline igneous province in the Iberian Peninsula, and its tectonic significance. Lithos, v. 15, p. 111-131.

Romine, K., Pryer, L., Stuart-Smith, P., Thomas, P., Wakefield, L., et Blevin, J. 2000. Impact of basement architecture on the evolution of the petroleum systems of the Canning Basin, Western Australia. AAPG Bulletin, v. 84, p. 1482-1483.

Salvador, A. 1991. Triassic-Jurassic. In Salvador, A., ed, The Gulf of Mexico Basin: The geology of North America. Geological Society of America, v. J, p. 131-180.

Sass-Gustkiewicz, M. 1996. Internal sediments as a key to understanding the hydrothermal karst origin of the upper Silesian Zn-Pb ore deposits. Society of Economic Geologists Special Publication, v. 4, p. 171–181.

Savostin, L. A., Sibuet, J.-C., Zonenshain, L. P., Le Pichon, X., et Roulet, M.-J. 1986. Kinematic evolution of the Tethys belt from the Atlantic Ocean to the Pamirs since the Triassic. Tectonophysics, v. 123, p. 1-35.

Schersten, A., Elliott, T., Hawkesworth, C., et Norman, M. 2004. Tungsten isotope evidence that mantle plumes contain no contribution from the Earth's core. Nature, v. 427, p. 234-237.

Schmincke, H.-U. 1977. Eifel-Vulkanismus oestlich des gebietes Rieden-Mayen. Fortschritte der Mineralogie, v. 55, p. 1-31.

Scotese, C. R. 2002. PointTracker for Windows version 4c, PALEOMAP Project, Arlington, Texas.

Sheth, H. C. 1999. A historical approach to continental flood basalt volcanism: insights into pre-volcanic rifting, sedimentation, and early alkaline magmatism. Earth and Planetary Science Letters, v. 168, p. 19–26

Sibley, D.F., et Gregg, J.M. 1987. Classification of dolomite rock textures. Journal of Sedimentary Petrology, v. 57, no. 6, p. 967–975.

Sims, P. K. 2002. The Mesoproterozoic; a time of change in tectonic style and magma types in the North American continent. Geological Society of America, Abstracts with Programs, v. 34, no. 6, p.43.

Sole, J., Pi, T., et Enrique, P. 2003. New geochronological data on the Late Cretaceous alkaline magmatism of the northeast Iberian Peninsula. Cretaceous Research, v. 24, p. 135-140.

Spicakova, L., Ulicny, D., et Koudelkova, G. 2000. Tectonosedimentary evolution of the Cheb Basin (NW Bohemia, Czech Republic) between late Oligocene and Pliocene; a preliminary note. Studia Geophysica et Geodetica, v. 44, p. 556-580.

Steiger, R.H., et Jäger, E. 1977. Subcommission on geochronology: Convention on the use of decay constants in geo- and cosmochronology. Earth and Planetary Science Letters, v. 36, p. 359–362.

Stofan, E. R., Sharpton, V. L. Schubert, G., Baer, G., Bindschadler, D. L., Janes, D. M., et Squyres, S. W. 1992. Global distribution and characteristics of coronae and related features on Venus: implications for origin and relation to mantle processes. Journal of Geophysical Research, v. 97, p. 13347-13378.

Sun, S. S. et McDonough, W. F. 1989. Chemical and isotopic systematics of oceanic basalts; implications for mantle composition and processes. In Saunders, A. D., et Norry, M. J., éd., Magmatism in the ocean basins. Geological Society Special Publications, v. 42, p.313-345.

Sykes, L. R. 1978. Intraplate seismicity, reactivation of pre-existing zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation. Rev. Geophys. Space Phys., v. 16, p. 621-688.

Taylor, D. et MacKenzie, W. S. 1975. A contribution to the pseudoleucite problem. Contributions to Mineralogy and Petrology, v. 49, p. 321-333.

Taylor, S. R. et McLennan, S. M. 1995. The geochemical evolution of the continental crust. Reviews of Geophysics, v. 33, p. 241-265.

Thomas, W. A. 2006. Tectonic inheritance at a continental margin. GSA Today, v. 16, p. 4-11.

Thompson, P. F., et Tackley, P. J. 1998. Generation of megaplumes from the core-mantle boundary in a compressible mantle with temperature-dependent viscosity. Geophysical Research Letters, v. 25, p. 1999-2002.

Thompson, R. N., Morrison, M. A., Hendry, G. L., et Parry, S. J. 1984. An assessment of the relative roles of crust and mantle in magma genesis: an elemental approach. Phil. Trans. Royal Soc. Lond., v. A310, p. 549–590.

Torsvik, T. H., Müller, R. D., Van der Voo, R., Steinberger, B., et Gaina, C. 2006. Global plate motion frames: toward a unified model. Earth Science Reviews (en revu) – version soumise disponible à: http://www.geodynamics.no/guest/GlobalFrames ESR.pdf

Tremblay, A., Long, B., et Masse, M. 2003. Supracrustal faults of the St. Lawrence rift system, Quebec; kinematics and geometry as revealed by field mapping and marine seismic reflection data. Tectonophysics, v. 369, p. 231-252.

Trompette, R. 1973a. Les dolérites de l'Adrar de Mauritanie et le problème des analcimolites des Richat. In Monod, T., et Pomerol, C., éd., Contributions à l'étude de l'accident circulaire des Richat (Adrar de Mauritanie). Nancy: Sciences de la Terre, Mémoires 28, p. 83-95.

Trompette, R. 1973b. Le Précambrien supérieur et le Paléozoique inférieur de l'Adrar de Mauritanie: Bordure occidentale du bassin de Taoudenni, Afrique de l'ouest. Thèse de doctorat, Marseille, Université de Saint Jérôme, 573 p.

Turcotte, D. L. et Oxburgh, E. R. 1978. Intra-plate volcanism. Philos. Trans. R. Soc. London, Ser. A, v. 288, p. 561-579.

Van Breemen, O., Hutchinson, J., et Bowden, P. 1975. Age and origin of the Nigerian Mesozoic granites; a Rb-Sr study. Contributions to Mineralogy and Petrology, v. 50, p. 157-172.

Veevers, J. J. 1995. Emergent, long-lived Gondwanaland vs. submergent, short-lived Laurasia; supercontinental and Pan-African heat imparts long-term buoyancy by mafic underplating. Geology, v. 23, p. 1131-1134.

Velazquez, V. F., Gomes, C. B., Riccomini, C., Comin-Chiaramonti, P., Brumatti, M., Vasconcelos, P. M., et Tassinari, C. C. G. 2004. Caracterysticas isotopicas e idades $^{40}Ar/^{39}Ar$ para as rochas alcalinas da Provyncia Misiones, Paraguai meridional. Rev.

Fac. Cienc., Univ. Eduardo Mondane, Maputo, Moçambique, ed. Esp. 1, p. 198-200.

Verati, C., Bertrand, H., et Feraud, G. 2005. The farthest record of the Central Atlantic magmatic province into West Africa Craton; precise 40Ar/39Ar dating and geochemistry of Taoudenni Basin intrusives (northern Mali). Earth and Planetary Science Letters, v. 235, p. 391-407.

Villeneuve, M. 1991. Géophysique en Mauritanie. In Caruba, R., et Dars, R., éd., Géologie de la Mauritanie, Nice : CRDP, p. 151-174.

Villeneuve, M., et Cornee, J. J. 1994. Structure, evolution and paleogeography of the West African Craton and bordering belts during the Neoproterozoic. Precambrian Research, v. 69, p. 307-326.

Walker, R. J., Morgan, J. W., et Horan , M. F. 1995. Osmium-187 enrichment in some plumes: evidence for core-mantle interaction? Science, v. 269, p. 819-822.

Weaver, B. L. et Tarney, J. 1981. Lewisian gneiss geochemistry and Archaean crustal development models. Earth and Planetary Science Letters, v. 55, p. 171-180.

White, R. S., et McKenzie, D. P. 1989. Magmatism at rift zones: The generation of volcanic continental margins and flood basalts. Journal of Geophysical Research, v. 94, p. 7685-7729.

Williams, H. 1979. Appalachian Orogen in Canada. Canadian Journal of Earth Sciences, v. 16, p. 797-807.

Wilson, J. T. 1963. A possible origin of the Hawaiian islands. Canadian Journal of Physics, v. 41, p. 863-870.

Wilson, M. 1989. Igneous petrogenesis; a global tectonic approach. London: Unwin Hyman, United Kingdom, 466 p.

Wilson, M. 1997. Thermal evolution of the Central Atlantic passive margins: continental break-up above a Mesozoic super-plume. J. Geol. Soc. London, v. 154, p. 491-495.

Wilson, M. et Guiraud, R. 1992. Magmatism and rifting in Western and Central Africa, from Late Jurassic to Recent times. Tectonophysics, v. 213, p. 203-225.

Wilson, R. C. L., Hiscott, R. N., Willis, M. G., et Gradstein, F. M. 1989. The Lusitanian Basin of west-central Portugal; Mesozoic and Tertiary tectonic, stratigraphic, and subsidence history. In Tankard, A. J., éd., Extensional tectonics and stratigraphy of the North Atlantic margins. AAPG Memoir, v. 46, p. 341-361.

Withjack, M. O., Schlische, R. W., et Olsen, P. E. 1998. Diachronous Rifting, Drifting, and Inversion on the Passive Margin of Central Eastern North America: An Analog for

Other Passive Margins. AAPG Bulletin, v. 82, p. 817-835.

Woolley, A. R. 1987. Alkaline rocks and carbonatites of the world; Part 1, North and South America. British Museum (Nat. Hist.): London, U. K., 216 p.

Woolley, A. R. 1989. The spatial and temporal distribution of carbonatites. In Bell, K., éd., Carbonatites; genesis and evolution. Unwin Hyman: London, U. K., p. 15-37.

Woolley, A. R. 2001. Alkaline rocks and carbonatites of the world; Part 3, Africa. The Geological Society London, 372 p.

Woolley, A. R., Rankin, A. H., Elliott, C. J., Bishot, A. C., et Niblett, D. 1984. Carbonatite dykes from the Richat dome, Mauritania, and the genesis of the dome. The Indian Mineralogist, p. 189-207.

Woolley, A. R. et Kempe D.R.C. 1989. Carbonatites: nomenclature, average chemical compositions and element distiribution. In Bell, K., éd., Carbonatites; genesis and evolution. London: Unwin Hyman, U. K., p. 1-14.

Zoback, M. L., Nishenko, S. P., Richardson, R. M., Hasegawa, H. S., et Zoback, M. D. 1986. Mid-plate stress, deformation, and seismicity. In Vogt, P.R. et Tuckholke, B.E., éd., Western North Atlantic Region: The geology of North America. Geological Society of America, v. M, p. 297-312.