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Mafic dikes at Kahel Tabelbala (Daoura, Ougarta Range, south-western Algeria): New insights into the petrology, geochemistry and mantle source characteristics

Abderrahmane Mekkaoui [a,b], Nacéra Remaci-Bénaouda [a,*], Khadidja Graïne-Tazerout [c]

[a] Département des sciences de la Terre, Laboratoire « Géoressources, environnement et risques naturels », FSTU, Université d'Oran 2, Mohamed Ben Ahmed, Oran, Algeria
[b] Département de l'hydraulique, Université de Béchar, Béchar, Algeria
[c] Laboratoire de métallogénie, USTHB, Alger, Algeria

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ABSTRACT

New petrological, geochemical and Sr–Nd isotopic data of the Late Triassic and Early Jurassic Kahel Tabelbala (KT) mafic dikes (south-western Algeria) offer a unique opportunity to examine the nature of their mantle sources and their geodynamic significance. An alkaline potassic Group 1 of basaltic dikes displaying relatively high MgO, TiO_2, Cr and Ni, $La/Yb_N \sim 15$, coupled with low $^{87}Sr/^{86}Sr_i \sim 0.7037$ and relatively high $\varepsilon_{Nd}(t) \sim +3$, indicates minor olivine and clinopyroxene fractionation and the existence of a depleted mantle OIB source. Their parental magma was generated from partial melting in the garnet–lherzolite stability field. A tholeiitic Group 2 of doleritic dikes displaying low MgO, Cr and Ni contents, $La/Yb_N \sim 5$, positive Ba, Sr and Pb anomalies, the absence of a negative Nb anomaly coupled with moderate $^{87}Sr/^{86}Sri \sim 0.7044$ and low $\varepsilon_{Nd}(t) \sim 0$ (BSE-like), indicates a contamination of a mantle-derived magma that experienced crystal fractionation of plagioclase and clinopyroxene. This second group, similar to the low-Ti tholeiitic basalts of the Central Atlantic Magmatic Province (CAMP), was derived from partial melting in the peridotite source within the spinel stability field. Lower Mesozoic continental rifting could have been initiated by a heterogeneous mantle plume that supplied source components beneath Daoura, in the Ougarta Range.

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1. Introduction

Mafic dike swarms are a common expression of mantle-derived magma generation in continental setting characterized by extensional tectonic regimes in post-collisional or intraplate extensional settings (Rock, 1991). They resulted from a considerable extension of the continental lithosphere (Halls, 1982; Tarney and Weaver, 1987). They can thus provide important information for understanding not only the mantle source of the magmas, but also the tectonic evolution of the orogenic belt (Gorring and Kay, 2001; Yang et al., 2007).

Mafic dikes, distributed in southwestern Algeria, outcrop in the Ougarta Range, the Tindouf basin, the Eglab, the Reggane and Abadla regions. These occurrences have been widely investigated, especially in Morocco (Bensalah et al., 2011; Bertrand and Westphal, 1977, Choubert and Faure-Muret, 1974; El Aouli et al., 2001; El Maidani et al., 2013; Hafid et al., 1998; Hollard, 1973; Leblanc, 1973; Mahmoudi and Bertrand, 2007; Verati et al.,

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[*] Corresponding author.
E-mail address: remaci.nacera@lycos.com (N. Remaci-Bénaouda).

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2007). In contrast with Morocco, a few studies have been carried out in southern and southwestern Algeria (Boissière, 1971; Chabou et al., 2007; Chabou et al., 2010; Conrad, 1972; Djellit et al., 2006; Fabre, 1976; Mekkaoui, 2015; Mekkaoui and Remaci-Bénaouda, 2014; Menchikoff, 1930; Sebai et al., 1991).

This paper presents a new study of the petrology and geochemistry of these Mesozoic mafic dikes in the Ougarta Range, southwestern Algeria. New major and trace elements analyses as well as Sr–Nd isotopic data of mafic-dike rocks were collected from representative sections in Kahel Tabelbala (KT). These data allow us to identify the magma types, investigate their mantle sources and petrogenesis, and understand the evolution of the Mesozoic lithospheric mantle beneath KT, Ougarta Range.

2. Geological setting

The Ougarta Range corresponds to an imposing series of reliefs arising within the Saharan platform (Fig. 1a). These reliefs are confined to the Saoura and Daoura basins (Menchikoff, 1930). The KT, which is a major orogenic entity within the Daoura, is aligned along a Ougartian direction from N130° to N140°. It is bounded to the north by the "Hamada of Mande", to the east by Erg er Raoui, and to the south by the Erg Chech, and to the west, by the Iguidi and El Aâtchane ergs (Fig. 1a and b). The Ougarta Range belonging to the northern boundary of the West African Craton (Fig. 1c) is involved in the Late Pan-African history (Ennih and Liégeois, 2001; Kurek and Preidl, 1987). The continental collision between Gondwana and Laurasia in the Late Carboniferous was originally a major uprising and

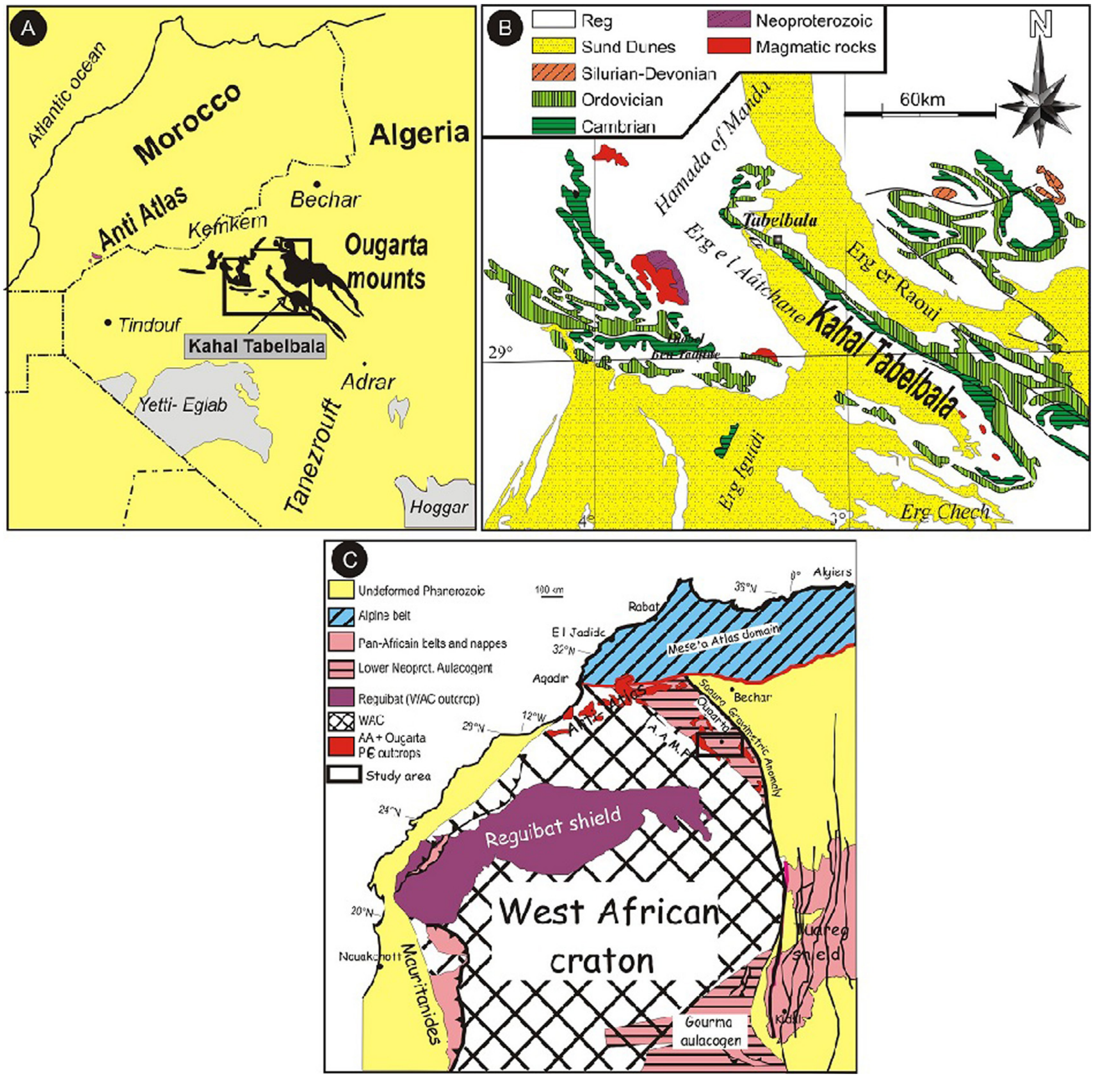

Fig. 1. Map of the studied area, showing (A) the location of Ougarta Range (Michard, 1976), a simplified geological map of the studied area, Kahel Tabelbala (B), a WAC sketch map placing the Ougarta range into its regional context (C) (modified after Fabre, 1976; Ennih and Liégeois, 2001).

overlaps in the northwestern part of the North African platform by folding and inversion in associated intraplate areas. The Ougarta Range was originated in a subsiding area that developed during the Cambrian along an old oceanic suture bordering the West African Craton. The Central Atlantic Oceanic Opening during the Trias–Lias, and the simultaneous separation of the Apulian terrane of northeastern Africa, followed a major extension accompanied by volcanic emissions throughout the North African platform (Bensalah et al., 2011; Bertrand, 1991; Chabou et al., 2007; Chabou et al., 2010; Marzoli et al., 1999; Meddah et al., 2007; Mekkaoui, 2015; Mekkaoui and Remaci-Bénaouda, 2014; Sebai et al., 1991; Verati et al., 2007). A second major extensive Mesozoic stage took place during the Cretaceous, followed by the opening of the south equatorial Atlantic Ocean, which generated the development of a series of aborted rift systems in the northern and central African platform (Gomez et al., 2000; Piqué et al., 1998). The rifting of the northern part of the North Atlantic during the Cretaceous was originally a sudden change in the movements of the European plate.

Intrusions of mafic dikes of Mesozoic age along the KT structure are widespread, occurring as NW–SE-trending swarms, orientations determined also in Morocco (Bensalah et al., 2011; Bertrand, 1991; Mahmoudi and Bertrand, 2007; Verati et al., 2007). These rocks cut across the Neoproterozoic and Cambrian-Ordovician sedimentary stacks (Bouima and Mekkaoui, 2003; Chikhaoui, 1974; Columb and Donzeau, 1974; Fabre, 1976, 2005; Mekkaoui, 2015; Nédjari, 2007). Two areas are briefly described in this structure (Mekkaoui, 2015; Mekkaoui and Remaci-Bénaouda, 2014): (i) at Guelb Berrezouk, the volcanic occurrences also show a remarkable growth, spreading as dikes and sills (Figure is listed in electronic annex as A1a). They are referred to as the Jebel Guelb Berrezouk group (Group 1). At the foot of Jebel, a 0.5- to 1-m-thick dike, visible over 150 m, punctuates the Ougartian fault separating the Neoproterozoic from the Cambrian. The top of this Jebel is crowned by two sills, from 0.5 m to 0.8 m in thickness. In the reverse side of this peak, a system of dikes and sills outcrops in the Cambrian sandstones. The largest dike is up to 5 m in thickness. These rocks cross-cut the Cambrian and, in some localities, the Ordovician as well; (ii) in the southeastern part of the periclinal closure, it is affected by an axial doleritic dike accompanied, from either side, by other satellite dikes (Figure is listed in electronic annex as A1b). This axial dike is visible over a distance of few hundred meters and its width can exceed 150 m. The field observations reveal a medium-to-coarse-grained core underlined, on both sides, by blackish borders with a fine-grained texture at the close contact with the Cambrian. This dike and its satellites constitute the periclinal closure group (Group 2).

3. Sampling and analytical techniques

Electron microprobe analyses of minerals were obtained using CAMECA SX 100 (CNRS-UMR 6524, Blaise Pascal University, Clermont-Ferrand, France). The accelerating voltage was 15 kV and the electron beam current 15 nA, with a beam diameter of 1 mm. The counting time was 10 s.

A suite of 11 fresh samples were selected for major and trace element analysis, using X-ray fluorescence (XFR spectrometer Phillips PW 1404) at the University of Lyon. The precision was 1–2% for major elements and 10–15% for trace elements. Seven of these samples were analyzed for rare-earth elements (REE) using inductively coupled plasma mass spectrometry (ICP–MS) (Jobin Yvon JY38 P) at the "École des mines", Saint-Étienne, France. The analyses were realized on solid solutions, after HF dissolution to 130 °C, evaporation of the fluorides was resumed using HCl 2 N.

Sr and Nd isotopes were measured by TIMS at the GET (Observatoire Midi-Pyrénées, Toulouse-3 University, France). 100 mg of whole-rock powder were weighed in a Teflon beaker and dissolved in a 2:1 HF/HNO_3 mixture. Nd and Sr were extracted from the matrix using a combination of Sr-Spec, Thru-spec and Ln-Spec resins. An equivalent of 500 ng Sr and 150 ng Nd were run on a MAT261 Finnigan mass spectrometer. NBS987 and La Jolla isotopic standards were regularly run during the measurements. The standard reproducibility values are 0.510855 ± 18 ($n = 50$) for La Jolla and 0.710250 ± 35 ($n = 70$) for NBS987. Typical blanks were 30 pg for Nd and 150 pg for Sr.

4. Results

4.1. Petrography–Mineralogy

Distinct petrographic features were noted for two Groups. The Guelb Berrezouk Group-1 volcanic rocks outcrop as basaltic dikes. They have a mainly porphyritic texture, and contain various amounts of olivine and clinopyroxene phenocrysts (Fig. 2a and b). Plagioclase phenocrysts are generally absent. The groundmass consists of plagioclase, clinopyroxene, olivine and oxides. In contrast, the periclinal closure event (axial dikes) Group 2 is comprised of subvolcanic rocks. They are massive and are medium-grained (in the core) to fine grained (at the rim). They display inter-granular and sub-ophitic textures typical of dolerites with euhedral plagioclase, interstitial clinopyroxene, and Fe–Ti oxides (Fig. 2c and d). Accessory quartz and feldspar occur in core-facies.

Representative electron microprobe analyses of minerals are listed as supplementary data in Table A2. In Group 1, olivine compositions vary from phenocrysts (Fo_{90-83}) to microphenocrysts (Fo_{80-77}). Olivine phenocrysts display normal compositional zoning with Fo_{90} core to Fo_{83-78} rim. In the core of the olivine phenocrysts, the CaO contents range between 0.05 and 0.08 wt%. In microphenocrystal olivine, the CaO contents are high (0.25 to 0.45 wt%). Clinopyroxene displays a diopsidic composition ($Wo_{47.6-51.6}En_{40.8-30.9}Fs_{11.6-17.4}$) (Fig. 3a). They are rich in Al_2O_3 (up to 10 wt%) and TiO_2 (2.2 to 5.6 wt%). The amounts of Ca + Na (0.93 to 0.96) are typical of alkali basalts according to Leterrier et al. (1982) (Fig. 3b). Plagioclase phenocrysts were absent, but in the groundmass they display labradorite to andesine ($An_{68.3-46.5}$ $Ab_{30-46.7}$ $Or_{1.7-6.9}$) compositions. Oxides were ulvospinel ($Ti_{6.5-7.3}$ $Al_{1.28-0.79}$ $Fe^{2+}{}_{14.4-14.18}$ $Fe^{3+}{}_{1.72-0.52}Mn_{0.02-1.1}O_{32}$).

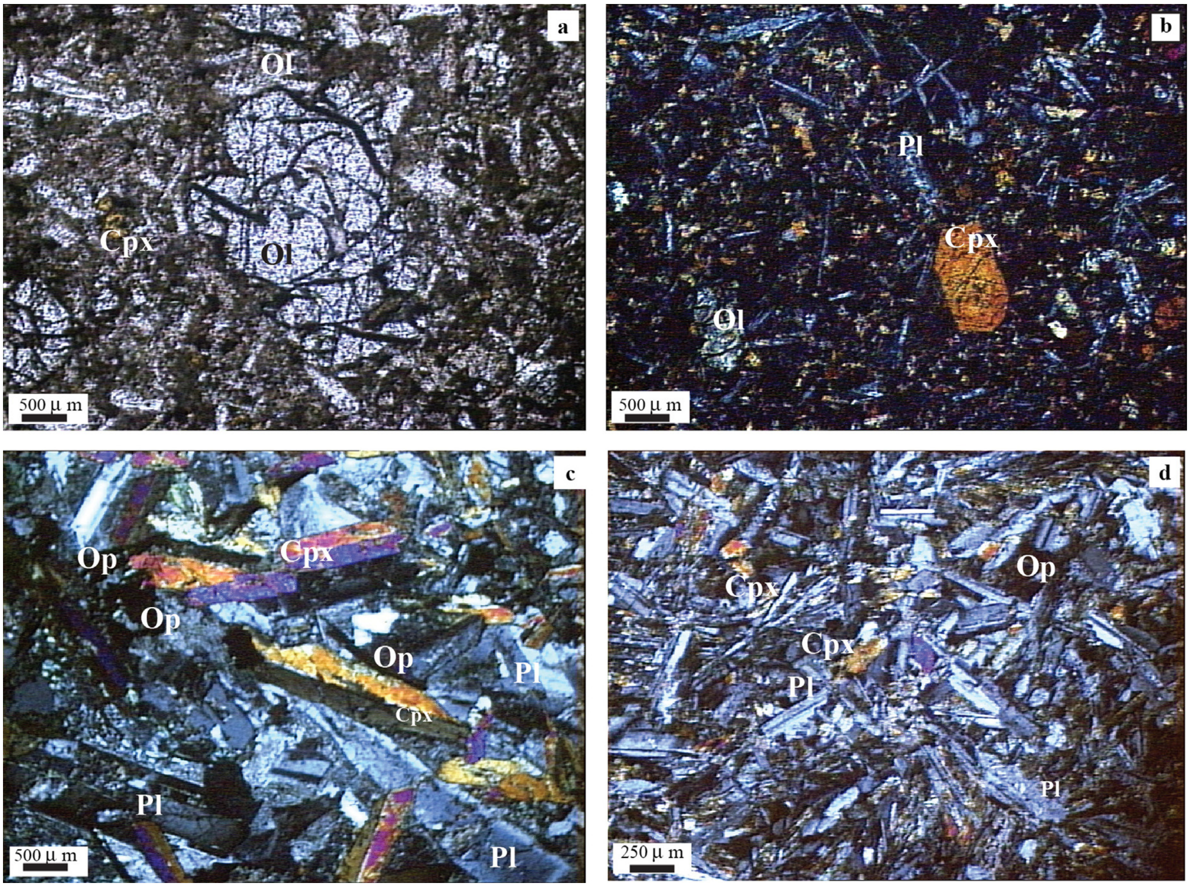

Fig. 2. Representative photomicrographs of petrographic features of KT mafic dikes: a and b: the Guelb Berrezouk Group 1 with porphyritic texture and olivine and clinopyroxene phenocrysts; c and d: the periclinal closure event (axial dike) Group 2 with doleritic composition with medium-grained (core) to fine-grained (rim) respectively. Ol: olivine, Cpx: clinopyroxene, Pl: plagioclase, Op: opaque.

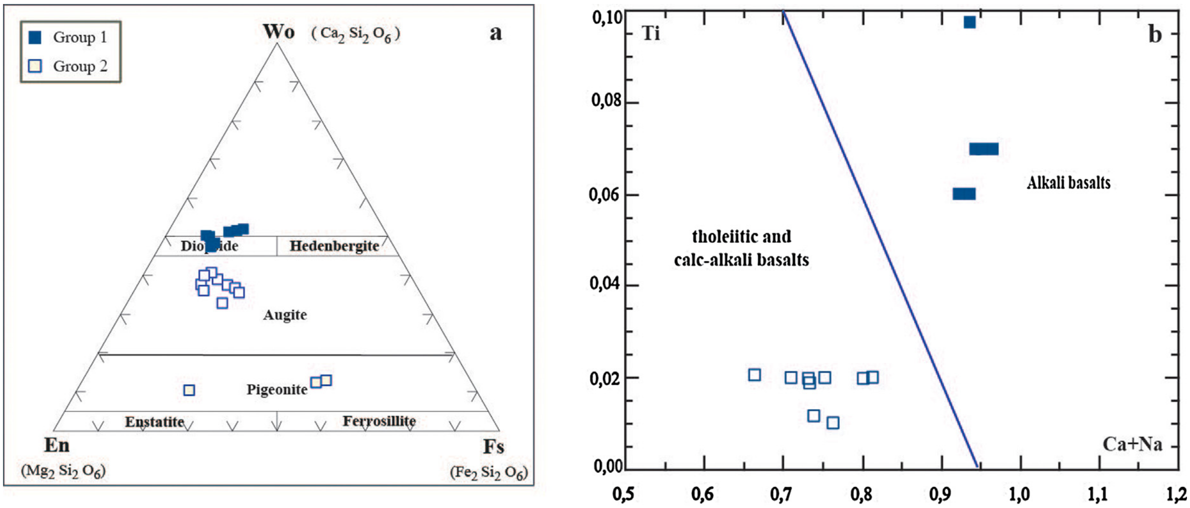

Fig. 3. Classification diagrams of minerals of KT dikes; a: nomenclature diagram of Morimoto (1988); b: distribution of clinopyroxenes in the diagram of Leterrier et al. (1982) (Table A2).

In Group 2, olivine was absent, but plagioclase and clinopyroxene were the predominant minerals; the chemical compositions differ from the core to the rim axial dike. At the rim, plagioclase displayed bytownite–andesine compositions ($An_{71–46.5}Ab_{28.7–51.2.}Or_{0.4–1.3}$) and clinopyroxene crystals are augite ($Wo_{33–41.2}En_{45.5–42.1}Fs_{22.5–16.6}$). In the core of the dike, plagioclase is less calcic (labradorite–oligoclase: $An_{59.2–27.3}$ $Ab_{40–69.5.}Or_{0.7–3.2}$) with augite and Ca-poor clinopyroxene ($Wo_{10.2–13.8}En_{65.4–33.8}Fs_{24.4–52.4}$) (Fig. 3a). Oxides are ilmenite ($Ti_{0.91–0.98}Fe^{2+}{}_{0.83–0.92}Fe^{3+}{}_{0.16–0.04}Mn_{0.08–0.06}O_3$).

4.2. Geochemistry: Major elements, trace elements and Sr–Nd isotopes

Major, trace element, Rb, Sr, Sm and Nd concentrations, measured and age-corrected $^{143}Nd/^{144}Nd$, $^{87}Sr/^{86}Sr$ ratios of KT dikes are listed in electronic annex respectively as Tables A3, A4 and A5.

Most of the analyzed samples were petrographically fresh and show little alteration. The loss on ignition (LOI) values range from 1.41 to 3.77 wt%. The samples exhibit significant variations in major and trace elements and Sr–Nd isotopic compositions.

The Group-1 dikes show basaltic compositions with a restricted range of SiO_2 (43.89–44.72wt%) and potassic–alkaline affinity with $K_2O/Na_2O > 1$ and total alkalis ($K_2O + Na_2O$) contents of 3.56–4.26wt%. In addition, they display relatively high MgO (8.58–10.27wt%), Mg# (100Mg/Mg $+Fe^{2+} = 59–63$), TiO_2 (2.48–2.85 wt%), CaO (9.62–10.80 wt%), Cr (290–396.9 ppm), and Ni (209.4-283.8 ppm). In terms of CIPW normative compositions, these rocks are undersaturated, with up to 1.63 wt% of nepheline and 22.58 wt% of olivine. Positive correlations of MgO versus Fe_2O_3t, CaO, Ni and Cr (Fig. 4b, c, g and h) and negative correlations of MgO versus Al_2O_3,TiO_2,K_2O and Sr are observed (Fig. 4a, d, e and f). They are light rare earth element (LREE) enriched with La/Yb_N ranging from 10.2 to 15.4 and LREE contents up to 122 times chondrite, with no obvious Eu anomaly ($Eu/Eu^* = 0.91–1.06$) (Fig. 5a). In primitive-mantle (PM) normalized multi-element patterns, they are enriched in large ion lithophilous elements (LILE), with positive Ba and Sr anomalies and high-field strength elements (HFSE) with very discrete positive Nb and Zr anomalies (Fig. 5c).

The initial isotopic ratios were all corrected to 204 ± 8 Ma (Group 1) and 183 ± 5 Ma (Group 2) based on Rb–Sr whole-rocks isochrons (Mekkaoui, 2015; Mekkaoui and Remaci-Bénaouda, 2014).

Group-1 mafic dikes display significant variations in $^{87}Sr/^{86}Sr_i$ (0, 7037 to 0, 7069) and relatively homogeneous $(\varepsilon_{Nd})_{204\ Ma}$ values (+2.99 to +2.60); they are consistent with a source mainly composed of depleted mantle.

In contrast, the Group-2 dikes are mafic to intermediate in composition (SiO_2 50.91–56.17 wt%) and exhibit a continental tholeiitic affinity with low TiO_2 (0.81–1.14 wt%), total Fe_2O_3 (8.20–10.04 wt%), P_2O_5 (0.09–0.16wt%), Zr (up to 73.4 ppm), and Nb (up to 15.4 ppm) similar to the case of low-Ti tholeiites (Bellieni et al., 1990; Cox, 1983; De Min et al., 2003; Fodor, 1987; Fodor et al., 1990; Merle et al., 2011), and high abundances in Al_2O_3 (14.26–17.07 wt%) and CaO (7.80–10.65 wt%). They also contain low MgO (4.86–6.66 wt%), Mg# (50–62), Cr (22–149 ppm) and Ni (61.6–105.1 ppm). In term of CIPW normative compositions, the most basic rocks have amounts of quartz (from 1.55 to 3.48 wt%), and hypersthene (20.12 to 23.04 wt%). Positive correlations of MgO versus Al_2O_3, CaO, K_2O, Sr, Ni and Cr (Fig. 4a, c, e, f, g and h), and negative correlations between MgO versus Fe_2O_3, and TiO_2 are observed (Fig. 4b and d). All the analyzed samples show a moderate enrichment in LREE with La/Yb_N values of 4.9–5.5, and LREE contents up to 40 times those of chondrite with no significant Eu anomalies ($Eu/Eu^* = 0.99–1.12$) (Fig. 5b). In the PM-normalized trace element diagram (Fig. 5d), these dolerites also display a moderate enrichment in LILE with distinctive positive Ba, Pb and Sr anomalies and the absence of a negative Nb anomaly. They have relatively homogeneous $(\varepsilon_{Nd})_{183\ Ma}$ values (+0.19 to–0.08) compared with the significant variations in $^{87}Sr/^{86}Sr_i$ (0.7044–0.7083), often observed in continental tholeiites (Bertrand, 1991; Lightfoot and Hawkesworth, 1988; Philpotts and Asher, 1993; Verati et al., 2007).

5. Discussion

The chemical variations of basaltic lavas in continental settings are generally controlled by mantle temperature, source composition, degree of partial melting and processes such as crystal fractionation and crustal contamination (McKenzie and Bickle, 1988). While Cox (1983), Hergt et al. (1991), Turner and Hawkesworth (1995) favor a model with different mantle sources for the two magma types, Fodor (1987) and Arndt et al. (1993) support the consequence of distinct melting conditions for a homogeneous mantle source coupled with crustal contamination.

5.1. Fractional crystallization

According to Desmurs et al. (2002) and Révillon et al. (2002), the criteria of primary mantle melts with $MgO > 8\%$ compositions, and olivine with $Fo_{91–88}$, the most primitive Group-1 rocks seem likely to have resulted from a low degree of magma fractionation, as witnessed by high MgO levels (10.27–9.40 ppm), Mg# (63–61), and compatible elements like Ni and Cr, with contents up 284 ppm and 397 ppm, respectively. Likewise, olivine phenocrysts from the studied rocks are confined to the range $Fo_{90–83}$ (Table A2). In the core of the olivine phenocrysts, the CaO contents range between 0.05 and 0.08 wt%, in a fashion similar to those of mantle olivine, which generally display < 0.1 wt% CaO (Ma et al., 2011; Thompson and Gibson, 2000). In microphenocrystal olivine, the CaO contents are high (0.25 to 0.45 wt%). These olivine phenocryst compositions suggest that they were probably intratelluric minerals that crystallized earlier from a more primitive magma and were subsequently incorporated into a more evolved magma, rather than being xenocrysts of lithospheric mantle origin. These Group-1 dikes may have experienced fractionation of olivine and clinopyroxene from their parental magma. This is supported by the positive correlation between MgO and Fe_2O_3, CaO, Ni and Cr (Fig. 4b, c, g and h) and is also consistent with the

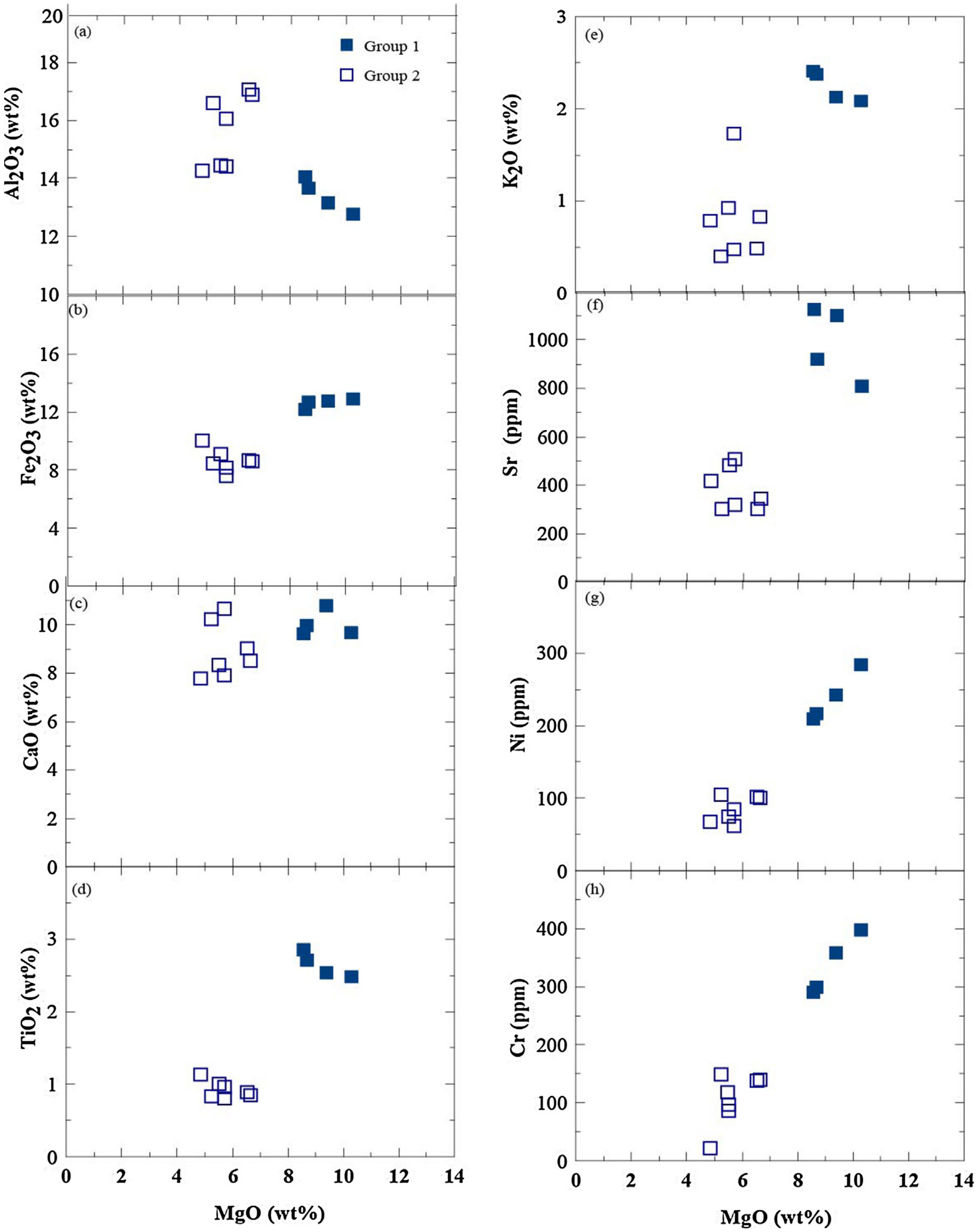

Fig. 4. Variation diagrams for major oxides and trace elements versus MgO contents for mafic dikes from KT.

presence of olivine and clinopyroxene as the dominant phenocrysts in these basaltic rocks. No significant fractionation of plagioclase in the magma as witnessed by the negative correlation between MgO, Al_2O_3 and Sr (Fig. 4a and f) and the absence of plagioclase phenocrysts and of any negative Eu anomaly. The negative correlation between MgO and TiO_2 (Fig. 4d) contents indicates insignificant fractionation of Fe–Ti oxides.

The main features of Group-2 dikes indicate that the most mafic dolerites do not represent primary mantle

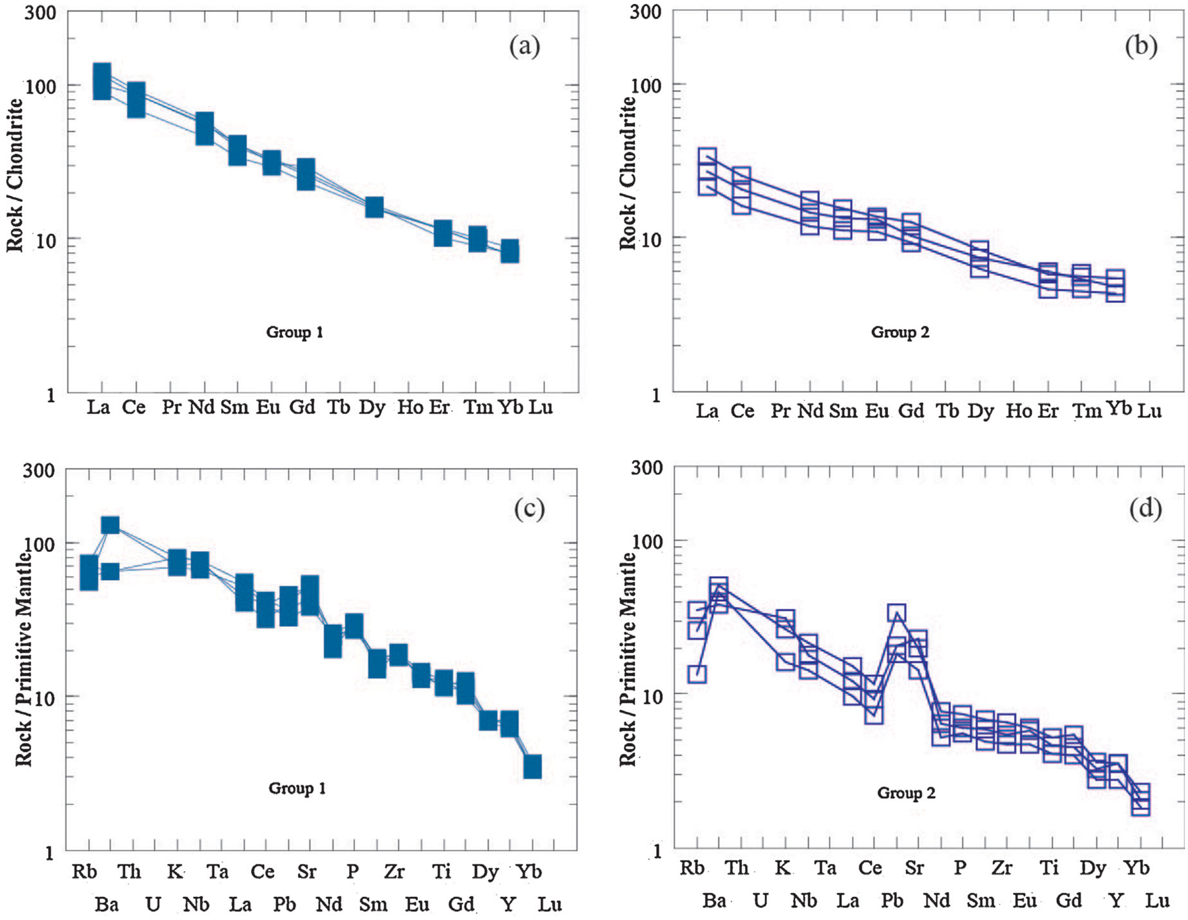

Fig. 5. Chondrite-normalized rare earth element patterns (a and b) and primitive-mantle-normalized diagrams (c and d) for the KT mafic dikes. The normalized values are from Boynton (1985) and Sun and McDonough (1989), respectively.

melts, as judged from their low MgO (<7 wt%), Mg# (≤ 62), Ni and Cr contents (≤ 105 and ≤ 149 ppm respectively). This suggests that they may have undergone significant amounts of fractional crystallization from their parental basaltic magma. The sample features are consistent with extensive plagioclase and clinopyroxene crystals supported by positive correlations between MgO and Al_2O_3, CaO, K_2O, Sr and Cr (Fig. 4a, c, e, f and h). Likewise, a positive Sr anomaly and a distinct lack of a negative Eu anomaly argue against significant plagioclase fractionation (Bogaard and Wörner, 2003).

5.2. Crustal contamination

Since mafic dikes are emplaced into a continental environment, there is general agreement that these mafic mantle-derived magmas experienced some degree of crustal contamination during ascent and/or residence within their high-level crustal magma chambers (Mohr, 1987). Furthermore, the ratios of trace elements in the contrasting mantle and crustal rocks may help to evaluate the possible influence of contamination. The correlation between Sr–Nd isotopic data and some trace element ratios provide evidence of the involvement of various mantle and crustal components in the petrogenesis of the KT rocks. For example, the incompatible trace element ratios, for Ce/Pb and Ba/Nb in the continental crust, are approximately 3.9 and 57 respectively (Rudnick and Gao, 2003).

In the-Group 1 dikes, the geochemical characteristics, including no depletion in Nb, low to moderate Sr isotopic composition (0,7037 to 0,7069) and relatively high ε_{Ndt} (+2.99 to +2.60) do not support significant crustal contamination during magma transport and emplacement. In addition, the same incompatible trace element ratios in the most mafic rocks (Ce/Pb $\sim$ 12 and Ba/Nb $\sim$ 9) are different from those for continental crust, approaching the trace element concentrations of typical mantle-derived rocks like OIB (Ce/Pb $\sim$ 25 and Ba/Nb $\sim$ 7.3) (Sun and McDonough, 1989).

In contrast, Group-2 dikes with large positive Pb anomalies, their elongate trend displaced from the mantle array at relatively constant ε_{Nd} (0.19 to–0.08) in the direction of increasing Sr_i (0.7044 to 0.7083), and their reduced Ce/Pb ratios (4.37–3.32) and elevated Ba/Nb ratios (20.98–118.23) indicate accordingly the involvement of a crustal component in the petrogenesis of the basaltic magmas.

5.3. Mantle source signatures

Finally, the major element trends, the REE patterns and most conclusively the initial Sr and Nd isotope compositions of the Group 1 and Group 2 mafic dike suite in the KT, preclude derivation by crystallization from a common parental magma. The evidence above suggests that these two dike groups could represent melts of a distinct mantle sourced by various petrogenetic processes.

In general, low La/Yb ratios reflect a melting regime dominated by relatively large melt fractions, and/or with spinel as the predominant residual phase (they indicate a thin lithosphere), whereas high La/Yb ratios correspond to smaller melt fractions and/or garnet control (they suggest a thick lithosphere).

In the alkaline Group-1 basalts, the relatively high La/Yb_N ratios (10.24–15.42) suggest that they may have formed by relatively low degrees of melting of a mantle source in the garnet stability field.

In contrast, in tholeiitic Group-2 dolerites, the REE chondrite normalized values are homogeneous and poorly fractionated ($La/Yb_N = 4.86–5.48$); hence the parental magmas may have been derived from partial mantle melts within the spinel stability field.

Furthermore, the heterogeneous composition of these Late Trias and Early Jurassic mafic dikes are confirmed by the most radiogenic compositions. This heterogeneity allows one to support the existence of two Groups: (i) an isotopically depleted reservoir (originated most likely from an OIB-like mantle) with $^{87}Sr/^{86}Sr_i$ (0, 7037) and $(\varepsilon_{Nd})_{204\ Ma}$ values (~ 3); and (ii) a group, close to the BSE isotopic values, with $^{87}Sr/^{86}Sr_i$ (0.7044–0.7083) and $(\varepsilon_{Nd})_{183\ Ma}$ values (~ 0), then contaminated. The isotopic signature of Group-2 dikes demonstrates the close compositional similarity with the low-Ti tholeiitic basalts of the Central Atlantic Magmatic Province (CAMP) (Bertrand, 1991; Chabou et al., 2010; Deckart et al., 2005; De Min et al., 2003; Marzoli et al., 1999; Mekkaoui, 2015; Merle et al., 2011; Sebai et al., 1991; Verati et al., 2005, 2007).

In KT, lithospheric rifting was probably involved during two specific mafic-igneous episodes. Could this continental rifting have been initiated by a plume head impacting beneath the lithosphere? Magmatism may have been induced either by the impact of the plume with the base of the continental lithosphere (Courtillot et al., 1999; Wilson, 1997) or by heat incubation under a thick continental lithosphere and/or edge-driven convection generated by thickness contrasts of different lithospheric domains (Coltice et al., 2007; De Min et al., 2003; Merle et al., 2011).

More mantle source characteristics, as mantle components can be inferred from systematic element ratios. Using basaltic compositions from the Iceland plume, Fitton et al. (1997) suggest that a deep depleted mantle source can be distinguished from the shallow normal MORB (N-MORB) source in plume-related basalts using Zr/Y and Nb/Y ratios. In Fig. 6, the ΔNb line separating plume from non-plume basaltic sources seems to provide good discrimination. The KT groups define a population above the ΔNb line in the mantle plume field, for the oceanic island (Alkaline Group 1) and the oceanic plateau basalt (Tholeiitic Group 2) fields described by Condie (2005). This is consistent with their geochemical composition, and the isotopic radiogenic isotopic diversity of the two Groups presented in this paper. It confirms that their basaltic magmas could have resulted from a heterogeneous mantle-plume source. Therefore, we cannot exclude an evolution of mantle-plume source and various partial melting rates from garnet-bearing peridotite stability field to depleted superficial spinel bearing peridotite stability field and contamination in the continental crust. Such a mantle plume could have supplied source components to the melting zone beneath the KT. Moreover, it is widely accepted that plume-related melts (*i.e.* OIB) do not show any negative Nb or positive Pb anomalies (Weaver, 1991); this is well demonstrated, especially by the chemical characteristics of the Group-1 dikes.

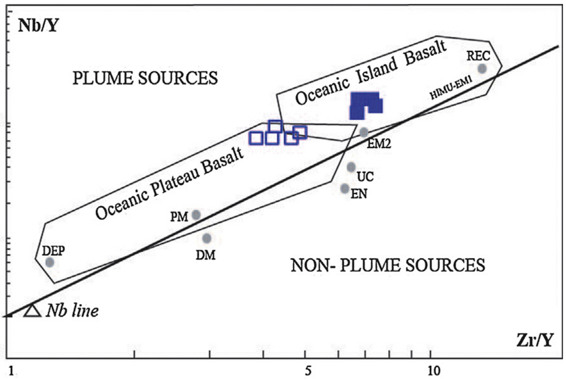

Fig. 6. Samples of KT mafic dikes plotted in a log–log diagram of Nb/Y versus Zr/Y after Fitton et al. (1997). DEP: deep depleted mantle, PM: primitive mantle, DM: depleted mantle, EN: enriched component, UC: upper continental crust, EM1 and EM2: enriched mantle sources, HIMU: high μ (U/Pb).

6. Conclusions

During Mesozoic times, such as the Late Triassic and Early Jurassic, two episodes of mafic magmatism occurred sporadically in the KT, Ougarta Range. The petrographic evidence, as well as chemistry mineral, geochemical and Sr–Nd isotopic compositions, have determined clearly two groups of mafic dikes that cannot be simply explained by crystallization from a common parental magma. They may have been derived from two different mantle sources, by various melting conditions, and were subjected to distinct differentiation and contamination processes.

An alkaline potassic Group 1 of basaltic dikes with olivine ($Fo_{90–83}$), diopside, plagioclase and ulvospinel displays relatively high MgO, Mg#, TiO_2, Cr, and Ni contents indicative of a derivation from mantle-derived melts with minor olivine and clinopyroxene fractionation. The geochemical and isotopic data on these rocks, including relatively high LREE with La/Yb_N ~ 15, suggest no obvious Eu anomaly, but discrete positive Nb and Zr anomalies, with low $^{87}Sr/^{86}Sri$ ratios (~ 0.7037) and relatively high ε_{Ndi} (~ +3), indicating the existence of a depleted reservoir (OIB-like), and preclude a significant contribution of the crustal component. We propose that the parental magmas were derived by a low degree of partial melting of an asthenospheric mantle source within the garnet stability field.

Tholeiitic Group-2 dikes contain plagioclase, augite, pigeonite, and ilmenite-bearing dolerites, and their geochemistry demonstrates low MgO, Mg#, Cr and Ni contents, suggesting that they may have undergone significant crystal fractionation of plagioclase and clinopyroxene. The geochemical and isotopic data in the most primitive compositions of Group-2 samples display a low enrichment in LREE, with $La/Yb_N \sim 5$ and no obvious Eu anomalies, but display distinctly positive Ba, Sr and Pb contents, and demonstrate the absence of any negative Nb anomaly. The large positive Pb anomalies, moderate $^{87}Sr/^{86}Sr_i$ ($\sim$ 0.7044), relatively low ε_{Ndt} ($\sim$ 0), and their reduced Ce/Pb, and elevated Ba/Nb ratios, support an evolved mantle-derived magma with significant crustal contamination during transport and emplacement. The parental magma may have been produced by the large degree of partial melting of a peridotite within the spinel stability field. The isotopic signature would suggest also a mantle source (BSE-like), then contaminated for this group, with close compositional similarity within the large igneous province of low-Ti tholeiitic basalts in the Central Atlantic Magmatic Province (CAMP).

It is therefore pertinent to note that lithospheric rifting was involved during the period of KT basic magmatism. It was most likely initiated by a rising heterogeneous mantle plume impacting the fractured lithosphere below. This plume-head could have supplied source components to the melting zone beneath Daoura, part of the Ougarta Range.

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Appendix A. Supplementary data

Supplementary figure and data associated with this article can be found in the online version at http://dx.doi.org/10.1016/j.crte.2017.06.003.

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