The alkaline intraplate volcanism of the Antalya nappes (Turkey): a Late Triassic remnant of the Neotethys

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Key-words. – Alkali basalt, Intraplate volcanism, Triassic (Upper), Neotethys, Turkey, Geochemistry.

Abstract. – Late Triassic submarine alkali basalts and hawaiites were collected from two superimposed tectonic slices belonging to the Kara Dere – Sayrun unit of the Middle Antalya nappes, southwestern Turkey. New determinations on conodont faunas allow to date this sequence to the Lower Carnian (Julian). The volcanic rocks show rather homogeneous compositions, with high TiO₂ and relatively low MgO and Ni contents which suggest olivine fractionation. Their primitive mantle-normalised multi-elements plots show Nb and Ta enrichments relative to La, Pb negative anomalies and heavy rare earth element and Y depletions typical of intraplate ocean island basalts. These characteristics are consistent with the major and trace element compositions of their primary clinopyroxene phenocrysts, which do not show any feature ascribable to crustal contamination. The studied lavas display a restricted range of εNd (+4.6 to +5.2) which falls within the range of ocean island basalts. Their initial (143Nd/144Nd)i ratios are too low to be explained by a simple mixing line between depleted MORB mantle (DMM) and HIMU components. Their Pb and Nd isotopic compositions plot along a mixing line between HIMU component and an enriched mantle, the composition of which could be the result of the addition of about 5 to 8% of an EM2 component (recycled marine sediments) to DMM. The lack of evidence for any continental crustal component in their genesis could be consistent with their emplacement in an intra-oceanic setting.

Le volcanisme alcalin intraplaque des nappes d’Antalya (Turquie), témoin de l’histoire triasique de la Néotéthys

Mots-clés. – Basalte alcalin, Intraplaque (volcanisme), Trias (supérieur), Néotéthys, Turquie, Géochimie.

Résumé. – Des basaltes alcalins et hawaiites sous-marins d’âge trias supérieur ont été collectés dans deux écailles superposées de l’unité de Kara Dere – Sayrun des nappes d’Antalya. Cette séquence est datée du Carnien inférieur (Julien) par les faunes de conodontes des sédiments associés. Les laves alcalines sont relativement homogènes, riches en TiO₂ et pauvres en MgO et Ni, ce qui suggère qu’elles ont évolué par fractionnement d’olivine. Leurs spectres multi-élémentaires présentent un enrichissement maximal au niveau du Nb et du Ta, des anomalies négatives en Pb et un relatif appauvrissement en terres rares lourdes et en Y. L’ensemble de ces caractéristiques est typique des magmas basaltiques intraplaque des îles océaniques. Cette conclusion est en accord avec la composition en éléments majeurs et en traces de leurs phénocristaux de clinopyroxènes, qui ne traduit aucune contamination crustale. Les laves étudiées présentent une gamme restreinte d’εNd (+4.6 à +5.2), également compatible avec celle des basaltes des îles intra-oceaniques. Leurs rapports initiaux (143Nd/144Nd)i sont trop faibles pour être expliqués par un mélange entre manteau appauvri source des MORB (DMM) et un composant HIMU. Leurs compositions isotopiques du Nd et du Pb sont compatibles avec un mélange de sources impliquant le composant HIMU et un manteau enrichi. Ce dernier correspondrait à l’addition de 5 à 8 % d’un composant de type EM2 (sédiments marins recyclés) au manteau appauvri DMM. L’absence d’indications de l’intervention de croûte continentale au cours de la genèse des laves d’Antalya est compatible avec leur mise en place en contexte intraplaque océanique.

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Bull. Soc. géol. Fr., 2008, t. 179, no 4, pp. 397-410
INTRODUCTION

The Neotethys opening started during either the Lower Carboniferous [Garzanti et al., 1999] or the Middle Permian [Ricou, 1994; Baud et al., 1996]. It developed north of the Indian plate during the Upper Carboniferous [Bassoullet et al., 1980a,b; Garzanti et al., 1996, 1999] or the Upper Permian [Baud et al., 1996] and north of the Arabian plate during the Lower Permian [Angiolini et al., 2003] or the Middle to Upper Permian [Stampfli et al., 1991; Marcoux and Baud, 1996; Baud et al., 2001].

The Neotethyan ophiolitic suites which occur from the eastern Mediterranean (Antalya, Troodos, Hatay, Baer-Bas; fig. 1) to the Persian Gulf (Samail nappes in Oman) were emplaced during the Upper Cretaceous according to the ages of their metamorphic soles [Celik et al., 2006], and are often tectonically connected with Permian and Triassic submarine volcanic rocks. In the eastern Mediterranean, the latter are stratigraphically associated with pelagic and/or carbonate-platform limestones dated to the Upper Triassic [Marcoux, 1970; Dumont et al., 1972; Robertson and Waldron, 1990]. The Carnian age of the opening of the Izmir-Ankara branch of the Neotethys according to radiolarian faunas [Tekin et al., 2002] sets the problem of the geodynamic setting of the Late Triassic submarine volcanism in the eastern Mediterranean, which might be rift-related, ridge-related or finally oceanic island-type as recently shown for the Mamonia Complex in Cyprus [Lapierre et al., 2007]. In this paper, we present a new set of petrologic and geochemical data on the Antalya nappes (southwestern Turkey) submarine volcanic rocks. Major, trace element and Nd, Sr, Pb isotopic data are used to characterise their magmatic affinities.

GEOLOGICAL NOTES

Belonging to the Taurus orogenic belt, the Antalya nappes are exposed west of the Antalya Gulf (fig. 1). They overlie tectonically the parautochthonous Bey Daglari limestone units, which range in age from Mesozoic to Miocene. Three major sets of nappes can be distinguished (fig. 2). The Lower Antalya nappe (Dereköy Unit [Marcoux and Poisson, 1972]), mostly made up of radiolarites and subordinate calciturbidites, is tectonically overlain by the Middle Antalya nappes, which includes ophiolitic and basaltic units, and is in turn overlain by the calcareous Upper Antalya nappes including the Kemer Gorge, Bakirli Dag and Tahtali Dag units [Marcoux, 1987].

The Middle Antalya nappes comprise three major tectonically superimposed units (fig. 2). The lowest one, the Ala- kir Çay Unit, is composed of a Late Triassic to Late Cretaceous [Marcoux 1987; Tekin et al., 2003] thick pelagic sedimentary sequence tectonically associated with slices of ophiolitic plutonic rocks and Late Triassic volcanics. This unit is tectonically overlain by the Ophiolitic Unit, which includes peridotites, serpentinites, layered gabbros and norites [Juteau, 1975, 1980]. This latter unit displays many petrographic and geochemical similarities with the Troodos (Cyprus) ultramafic and gabbroic sequences. The two units are overlain by the Kara Dere-Sayrun Unit, made up of submarine (pillowed) basalts with intercalated pelagic Late Triassic (Carnian) limestones bearing Halobia insignis (Gemmellaro) and Trachyceras sp. ammonites [Marcoux, 1970] and a thin red pelagic limestone (Hallstatt facies) depositional cover of Upper Carnian age.

We have studied the Kara Dere-Sayrun Unit near its type locality, the Kara Dere gully. The cross-section along the dirt road from Dereköy to Saklikent allows to sample two superimposed basaltic tectonic slices (fig. 2), each of them ca. 400 m thick and showing a normal polarity together with a general 50°-70° dip towards East. The upper slice outcrops at altitudes ranging from 1920 m (top) to 1765 m (bottom), and is overlain by a klippe of Triassic Hallstatt-type limestones. It is made up of basaltic pillows, 50 cm to 3 m in diameter, with occasional pillow breccia levels and minor intercalations of pelagic Halobia-bearing limestones. The lower tectonic slice, which outcrops between altitudes of 1870 m (top) and 1590 m (bottom), is characterised by much larger pillows, some of which reach unusual diameters (5 to 20 m), and display well-preserved radial columnar jointing and thick glassy margins. Up to 15 m thick columnar-jointed lava flows are common in the lower part of this slice. This unit, which also contains intercalated pillow breccias usually associated with pelagic limestone beds, tectonically overlies a serpentinite body.

New determinations on conodont faunas allow us to precise the age of the Antalya submarine volcanism. Sediment sample 87/75 collected near the base of the lower tectonic slice (altitude: 1680 m) in the type Kara Dere section yielded the following Lower Carnian (Julian) fauna: Gladigondolella tethydis (Huckriede), Metapolygnathus polygonathiformis (Burudov and Stefanov). The top of the section (sample 86/79 overlying the pillow basalts from the lower tectonic slice, altitude: 1920 m) is of Upper Carnian age (Tuvanian 3/I) according to the fauna found: Metapolygnathus polygonathiformis (Burudov and Stefanov), Metapolygnathus nodosus.

FIG. 1. – Schematic map of the eastern Mediterranean, showing the location of Tethyan ophiolites and associated Late Triassic volcanics.

FIG. 1. – Carte schématique de la Méditerranée orientale montrant la position des ophiolites téthysiennes et des volcanites triasiques associées.
FIG. 2. – Structural sketch map of the sampled area (Kara Dere unit, Antalya Nappes, SW Turkey), and position of the studied samples.

Fig. 2. – Carte structurale schématique de la zone étudiée (unité de Kara Dere, nappes d'Antalya, Turquie du SW), et position des échantillons analysés.
A third sediment sample (86/81) collected at an altitude of 2000 m, i.e. 80 m below the top of the upper tectonic slice, yielded Gladigondolella tethydis (HUCKRIEDE), Metapolygnathus foliatus (BURUDOV and STEFANOV), also indicative of a Lower Carnian (Julian) age. The mafic volcanism exposed in the Kara Dere section is thus unequivocally of Upper Triassic (Carnian) age.

ANALYTICAL METHODS

Major elements in minerals were analysed with a Cameca SX50 microprobe using a 15kV acceleration voltage and different regulated beam currents, according to the mineral type (10 nA for plagioclase, and 20 nA for olivine, pyroxenes and oxides). The calibration was made on specific natural mineral standards for each type of minerals. Trace element measurements on minerals were made at Lausanne by laser-ablation ICP-MS mass spectrometry using an Ar-F 193 nm Lambda Physics® Excimer laser coupled with a Perkin-Elmer 6100DRC ICPMS. NIST 610 and 612 glasses were used as external standards, Ca and Si as internal standards after microprobe measurements on the pit sites. Ablation pit size varied from 40 to 60 µm. BCR2 basaltic glass was regularly used as a monitor to check the system for reproducibility and accuracy. Results were within ±10% of the certified values.

All the samples were pulverised in an agate mill. Major and trace element analyses were done on whole rocks by ICP – optical emission spectroscopy (ICP-AES) at Brest, using the procedures of Cotten et al. [1995]. Incompatible trace elements, including rare earth elements (REE), of selected samples were also analysed by ICP-MS at Grenoble, after acid dissolution of 100 mg sample, using the procedures of Barrat et al. [1996]. Standards used for the analyses were JB2, WSE, Bir-1 and JR1. Analytical errors are 1-3% for major elements and less than 3% for trace elements.

Sr (static acquisition) and Nd (dynamic acquisition) isotope ratios of selected samples were measured at Toulouse on a Finnigan MAT261 multicollector mass spectrometer using the analytical procedures described by Lapierre et al. [1997]. Results on La Jolla Nd standard yielded 143Nd/144Nd = 0.511850 ± 0.000017 (2σ external reproducibility on 12 measurements). Results on NBS 987 Sr standard yielded 87Sr/86Sr = 0.710250 ± 0.000030 (2σ external reproducibility on 11 determinations). 87Sr/86Sr and 143Nd/144Nd were normalised for mass fractionation relative to 86Sr/88Sr = 0.1194 and 146Nd/144Nd = 0.7219 respectively.

PETROLOGIC, MINERALOGICAL AND GEOCHEMICAL DATA

Rock selection and alteration problems

Twenty three samples were selected for the petrological and geochemical investigations. In situ (laser ablation) plagioclase and clinopyroxene trace element analyses were performed on four petrographically fresh rocks (AY04, AY08, AY20, AY21, tables I and II). ICP-MS analyses were performed on 9 samples, and isotopic analyses on 8 (Nd), 7 (Pb) and 6 (Sr) samples, respectively (table III).

The high levels of the loss on ignition (LOI) are related to the abundant calcite– or zeolite-filled vesicles. Their contents in large ion lithophile elements (LILE) known to be sensitive to alteration and metamorphism may have been modified during the low-grade metamorphism which affected the Antalya volcanic rocks [Juteau, 1975]. Especially, their concentrations in Rb, Ba, K and Sr may not be representative of the primary compositions of these lavas, although most of them do not depart from the “fresh basalt” range (table III). In addition, incompatible elements less soluble or insoluble in hydroxide fluids [McCulloch and Gamble, 1991; Kogiso et al., 1997], such as REE, Th and high field strength elements (HFSE) exhibit crude positive correlation trends in rectangular diagrams (not shown), which seem linked predominantly to crystal fractionation effects.

TABLE I. – Major and trace element analyses of plagioclases. Total iron as FeO.

<table>
<thead>
<tr>
<th>Plagioclase Analysis</th>
<th>AY02-08</th>
<th>AY02-08</th>
<th>AY02-20</th>
<th>AY02-20</th>
<th>AY02-20</th>
<th>AY02-20</th>
</tr>
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<tbody>
<tr>
<td>SrO</td>
<td>46.63</td>
<td>47.39</td>
<td>48.63</td>
<td>47.59</td>
<td>47.38</td>
<td>46.45</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>32.14</td>
<td>33.12</td>
<td>31.99</td>
<td>32.3</td>
<td>32.11</td>
<td>32.00</td>
</tr>
<tr>
<td>FeO</td>
<td>0.52</td>
<td>0.55</td>
<td>0.73</td>
<td>0.76</td>
<td>0.76</td>
<td>0.77</td>
</tr>
<tr>
<td>CaO</td>
<td>16.27</td>
<td>17.02</td>
<td>15.63</td>
<td>15.71</td>
<td>15.67</td>
<td>15.82</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.76</td>
<td>1.84</td>
<td>2.47</td>
<td>2.37</td>
<td>2.45</td>
<td>2.38</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.17</td>
<td>0.11</td>
<td>0.13</td>
<td>0.11</td>
<td>0.12</td>
<td>0.10</td>
</tr>
</tbody>
</table>

La Jolla Sr standard was used as a monitor to check the system for reproducibility and accuracy. Results were within ±10% of the certified values.

For lead separation, sample weights were calculated to obtain approximately 200 ng of lead. The samples were leached with 6N HCl for 30 minutes at 65°C before acid digestion. Pb blanks were less than 40 pg and were negligible for the present analyses. Lead isotope were analysed on a VG Plasma 54 multi-collector inductively coupled plasma-mass spectrometer (MC-ICP-MS) at Lyon. Lead isotope compositions were measured using the Tl normalisation method described by White et al. [2000]. During Pb isotope analyses, samples were bracketed between NIST 981 standards and calculated with respect to the value reported for this standard by Todt et al. [1996]. This technique yielded internal precision of ca. 50 ppm (2σ) and an external reproducibility of ca. 150 ppm (2σ) for 206Pb/204Pb ratios determined on 20 NIST standards.
Petrographic types

Three main types can be recognised based on the texture and mineralogy: (i) intersertal basalts (AY02-04, AY02-21), (ii) clinopyroxene-phyric basalts (AY02-08), and (iii) highly porphyritic plagioclase-phyric basalts (AY02-20). Intersertal basalts (AY02-21) may also display abundant calcite + zeolites-filled vesicles.

**Table II.** – Major and trace element analyses of clinopyroxenes. Abbreviations: sm : small grain; lg : large grain; clg : core of a large grain; rlg : rim of a large grain; ms : mineral separate. Fe₂O₃ and FeO contents calculated by stoichiometry to 4 cations and 6 oxygens.

**FIG. 3.** – Microphotographs of intersertal (AY02-04) and clinopyroxene-phyric (AY02-08) Kara Dere basalts.
### Table III. Major, trace element and isotopic (Nd, Sr, Pb) analyses of the Late Triassic alkaline volcanic rocks from Antalya Nappes. Total iron as Fe₂O₃. Sums of major element oxides recalculated to

<table>
<thead>
<tr>
<th>Sample</th>
<th>Rock Type</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
<th>MgO</th>
<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>TiO₂</th>
<th>Cr</th>
<th>Ni</th>
<th>Sr</th>
<th>Nd</th>
<th>Sm</th>
<th>Eu</th>
<th>Gd</th>
<th>Tb</th>
<th>Dy</th>
<th>Ho</th>
<th>Er</th>
<th>Tm</th>
<th>Yb</th>
<th>Lu</th>
<th>Age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A0501-1</td>
<td>Basalt</td>
<td>46.39</td>
<td>16.50</td>
<td>2.10</td>
<td>1.12</td>
<td>0.13</td>
<td>3.79</td>
<td>1.67</td>
<td>0.12</td>
<td>11.71</td>
<td>0.17</td>
<td>0.97</td>
<td>0.704044 ± 13</td>
<td>0.703888 ± 8</td>
<td>0.704270 ± 10</td>
<td>0.70312</td>
<td>0.109047</td>
<td>0.116234</td>
<td>0.512612</td>
<td>4.80</td>
<td>20.81</td>
<td>15.67</td>
<td>15.66</td>
<td>15.65</td>
</tr>
<tr>
<td>A0501-2</td>
<td>Basalt</td>
<td>46.66</td>
<td>15.29</td>
<td>2.05</td>
<td>1.01</td>
<td>0.14</td>
<td>3.76</td>
<td>1.63</td>
<td>0.13</td>
<td>11.76</td>
<td>0.18</td>
<td>0.99</td>
<td>0.704044 ± 13</td>
<td>0.703888 ± 8</td>
<td>0.704270 ± 10</td>
<td>0.70312</td>
<td>0.109047</td>
<td>0.116234</td>
<td>0.512612</td>
<td>4.80</td>
<td>20.81</td>
<td>15.67</td>
<td>15.66</td>
<td>15.65</td>
</tr>
<tr>
<td>A0501-3</td>
<td>Basalt</td>
<td>48.59</td>
<td>16.33</td>
<td>1.94</td>
<td>1.00</td>
<td>0.15</td>
<td>3.77</td>
<td>1.64</td>
<td>0.14</td>
<td>11.79</td>
<td>0.19</td>
<td>1.00</td>
<td>0.704044 ± 13</td>
<td>0.703888 ± 8</td>
<td>0.704270 ± 10</td>
<td>0.70312</td>
<td>0.109047</td>
<td>0.116234</td>
<td>0.512612</td>
<td>4.80</td>
<td>20.81</td>
<td>15.67</td>
<td>15.66</td>
<td>15.65</td>
</tr>
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<td>A0501-4</td>
<td>Basalt</td>
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<td>1.94</td>
<td>0.99</td>
<td>0.14</td>
<td>3.80</td>
<td>1.61</td>
<td>0.14</td>
<td>11.80</td>
<td>0.17</td>
<td>1.01</td>
<td>0.704044 ± 13</td>
<td>0.703888 ± 8</td>
<td>0.704270 ± 10</td>
<td>0.70312</td>
<td>0.109047</td>
<td>0.116234</td>
<td>0.512612</td>
<td>4.80</td>
<td>20.81</td>
<td>15.67</td>
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<td>A0501-5</td>
<td>Basalt</td>
<td>47.05</td>
<td>14.04</td>
<td>1.93</td>
<td>0.99</td>
<td>0.16</td>
<td>3.79</td>
<td>1.62</td>
<td>0.15</td>
<td>11.82</td>
<td>0.17</td>
<td>1.01</td>
<td>0.704044 ± 13</td>
<td>0.703888 ± 8</td>
<td>0.704270 ± 10</td>
<td>0.70312</td>
<td>0.109047</td>
<td>0.116234</td>
<td>0.512612</td>
<td>4.80</td>
<td>20.81</td>
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<td>A0501-6</td>
<td>Basalt</td>
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<td>14.41</td>
<td>1.94</td>
<td>0.99</td>
<td>0.15</td>
<td>3.79</td>
<td>1.64</td>
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<td>20.81</td>
<td>15.67</td>
<td>15.66</td>
<td>15.65</td>
</tr>
</tbody>
</table>

*Notes:* Fe₂O₃ total iron as Fe₂O₃; Sums of major element oxides recalculated to 100%; Nd, Sr, Pb isotopic compositions for Antalya Nappes given here were provided by the ICP-AES technique. The age of the rocks is 210 Ma.
and plagioclase phenocrysts (10 modal%), and calcite-filled vesicles (20%) set into an abundant groundmass (70%). The latter is formed of quenched plagioclase laths, tiny clinopyroxene grains, dendritic Fe-Ti oxides and interstitial secondary quartz. Olivine is replaced by calcite and includes octahedral brown Cr-spinels, and plagioclase is altered to sericite + albite while clinopyroxene remains fresh.

Highly phryic basalts (e.g. AY02-20) are rich in centimetre-sized plagioclase phenocrysts partially or totally altered into albite + calcite + smectite. Less common clinopyroxene phenocrysts are zoned with pink cores and reddish rims. Their groundmass show an interstitial texture with plagioclase laths, pink clinopyroxene grains, and octahedral Fe-Ti oxides surrounded by zeolites and smectites.

Mineral chemistry

Plagioclase and clinopyroxene major and trace element analyses are given in tables I and II, respectively. Plagioclases (AY02-08, AY02-02, table I) and clinopyroxenes (AY02-04, AY02-08, AY02-20, AY02-21, table II) show homogeneous bytownite and diopside compositions, respectively. Chondrite-normalised REE patterns of plagioclases (AY02-08, AY02-20, fig. 4) are characterised by La and Ce enrichments relative to Nd, Sm and heavier REE (Gd, Tb, Dy), and by marked Eu positive anomalies (4.9 < Eu/Eu* < 9.6). In Leterrier et al.’s [1982] diagrams, clinopyroxenes plot within the alkalic field whatever the sample (fig. 5). They are Cr-poor and Ti-rich, and generally display an enrichment in Fe, Ti and Al from core to rim, correlated with a decrease of Si and Mg (table II and fig. 5). Moreover, differences can be observed between individual samples. AY02-20 clinopyroxene cores show the highest MgO and the lowest Al2O3 and TiO2 contents, while AY02-04 cores display the lowest MgO and comparatively the highest TiO2 contents.

Figure 6 illustrates the clinopyroxene trace element compositions and their variations between individual samples from core to rim within a single grain. The clinopyroxene cores exhibit hump-shaped REE patterns, characterised by relative depletions in light (L) and heavy (H) REE with respect to middle (M) REE. In addition, some rim patterns (AY02-08, AY02-20) display marked Eu negative anomalies (0.76 < Eu/Eu* < 0.92). Geochemical variations are observed between cores and rims and are characterised by an increase of the REE levels (AY02-8, AY02-20) or by a LREE enrichment with respect to the HREE (AY02-04, AY02-21). The multi-element plots (fig. 7) are characterised by a relative Nb and Ta enrichment relative to La, the occurrence of Pb and Sr negative anomalies, and relative depletion in HREE and Y. The Sr negative anomalies indicate plagioclase removal, while the Pb negative anomalies, the enrichment in incompatible elements culminating at the level of Nb and Ta and the depletion in HREE are typical features of OIB. Ti negative or positive anomalies might indicate removal or accumulation of Ti-rich oxides.

Whole rock geochemistry

The Late Triassic volcanic rocks from the Antalya Nappes show rather homogeneous major and compatible element compositions, characterised by high TiO2 (2.7 to 4.2 wt%, table III) and relatively low MgO and Ni contents which suggest olivine fractionation. SiO2 does not exceed 50% while Fe2O3 is not lower than 10%. In usual classification and discrimination diagrams (not shown), they range from alkali basalts to trachybasalts or hawaiites (table III). Their high Cr and V contents are linked to the abundance of clinopyroxene and Fe-Ti oxides.

All the analysed samples are LREE-enriched (12.3 < (La/Yb)CN < 18.3) and yield enriched REE patterns (fig. 7) typical of ocean island basalts (OIB). Some of these patterns show slight Eu negative anomalies (0.92 < Eu/Eu* < 0.98, with the exception of sample AY02-07) suggesting plagioclase early removal. The corresponding multi-elements plots (fig. 7) are characterised by a relative Nb and Ta enrichment relative to La, the occurrence of Pb and Sr negative anomalies, and relative depletion in HREE and Y. The Sr negative anomalies indicate plagioclase removal, while the Pb negative anomalies, the enrichment in incompatible elements culminating at the level of Nb and Ta and the depletion in HREE are typical features of OIB. Ti negative or positive anomalies might indicate removal or accumulation of Ti-rich oxides.

The Late Triassic alkaline volcanics display a restricted range of εNd (+4.6 to +5.2; table III) within the range of OIB. Their 87Sr/86Sr values are lower than 0.704 and range from 0.70312 to 0.70376 (table III). The corresponding initial Pb isotopic ratios range from 19.07-20.03, 15.59-15.72 and 38.69-39.43 for the 206Pb/204Pb, 207Pb/204Pb and 208Pb/204Pb ratios, respectively (table III).
In the \((^{207}\text{Pb}/^{204}\text{Pb})_i\) versus \((^{206}\text{Pb}/^{204}\text{Pb})_i\) correlation diagrams (fig. 8A), most of the analysed rocks are located on or close to the Northern Hemisphere Reference Line (NHRL) [Zindler and Hart, 1986]. Sample AY02-20 differs from the others by its significantly higher \((^{207}\text{Pb}/^{204}\text{Pb})_i\) ratio (15.72). Reported in the \((^{143}\text{Nd}/^{144}\text{Nd})_i\)-(\(^{87}\text{Sr}/^{86}\text{Sr})_i\) diagram (fig. 8C), the samples define a horizontal narrow domain located slightly to the left of the OIB field. In the \((^{143}\text{Nd}/^{144}\text{Nd})_i\)-(\(^{206}\text{Pb}/^{204}\text{Pb})_i\) diagram (fig. 8D), the Antalya samples cluster between the MORB, HIMU and EM2 fields.

**Fig. 5.** – Composition en éléments majeurs des clinopyroxènes : (a) position dans le tétraèdre de classification [Morimoto et al., 1988], (b) affinités magmatiques selon Leterrier et al. [1982], et (c) variations de certains éléments en fonction de MgO.
DISCUSSION

Identification of magmatic processes

Rectangular diagrams using relatively immobile incompatible trace elements such as REE and HFSE (not shown) show positive correlations suggesting the occurrence of crystal fractionation and/or variations of partial melting degrees of the mantle source of the studied volcanics. The almost parallel character of REE and incompatible multi-element patterns of whole rocks (fig. 7) is also consistent with fractional crystallisation processes. In the Th, La, Nb, Y versus La/Yb correlation diagrams (fig. 9), the samples display loose positive trends, a feature suggesting the occurrence of small but significant variations of partial melting degrees, assuming a homogeneous mantle source. In short, the Late Triassic magmas, although rather homogeneous in composition (alkali basalts and hawaiites), probably derived from somewhat variable partial melting degrees of an enriched mantle source, and experienced slight to moderate crystal fractionation during their ascent, as commonly observed in intraplate alkali basalt series.

Interpretation of core-to-rim variations in clinopyroxene phenocrysts

Clinopyroxenes show important differences in their major and trace element chemistry between phenocryst cores and rims or between crystals of various sizes or from different samples (table II and fig. 6). The main differences consist of an enrichment in TiO$_2$ and Al$_2$O$_3$ (and sometimes in FeO and Na$_2$O) correlated with a depletion in MgO, from phenocryst cores to rims and from phenocrysts to microphenocrysts, or even from AY02-21 to AY02-04 clinopyroxenes. The decrease of MgO from AY02-21 to AY02-04 clinopyroxene cores probably reflects crystal fractionation because MgO and Zr contents of AY02-21 clinopyroxenes are higher than those of AY02-04 (table II). Similarly, the increase of the REE and other incompatible trace elements from cores to corresponding rims in AY02-08 and AY02-20 clinopyroxenes (fig. 6) can be explained by crystal fractionation. The composition of the interstitial liquid in which AY02-08 and AY02-20 clinopyroxenes grew derived from the parental magma after crystal fractionation of olivine followed by plagioclase. Indeed, AY02-08 and AY02-20 clinopyroxene rim patterns show negative Eu anomalies that are weaker in the cores, and which likely reflect the concomitant crystallisation of plagioclase characterised by positive Eu spikes (fig. 4).

The important enrichment in the most incompatible elements (Rb, La, Ce, Th, Nb and Ta) of AY02-04 and AY02-21 clinopyroxenes from core to rim (fig. 6 and table II) suggests that their growth took place in a liquid highly enriched in incompatible elements, compared to the parental magma or even to the AY02-08 and AY02-20 interstitial liquid. The AY02-04 and AY02-21 clinopyroxene cores crystallised while plagioclase fractionation had begun or even ended because both core and rim REE patterns are characterised by more or less marked Eu negative anomalies. This important enrichment in incompatible elements of the AY02-04 and AY02-21 clinopyroxene rims can be explained by quench processes. Indeed, when the alkali basalt magmas erupted, olivine, Fe-Ti oxides, clinopyroxene, and plagioclase had already crystallised and the residual melt in which these phenocrysts or microcrysts were included was enriched in incompatible elements. When the hot pillow basalts poured out in the water, clinopyroxene rims, microcrysts and microlites crystallised very quickly at the
expense of the residual liquid enriched in incompatible elements. The core-to-rim enrichment in Na$_2$O of AY02-04 and AY02-21 clinopyroxenes (table II) is consistent with this interpretation. Such differences in incompatible elements contents between clinopyroxene cores and rims have been described in lunar quenched basalts and ascribed to the interplay among the efficiency of the crystallisation process, the kinetics at the crystal melt interface, the kinetics of plagioclase nucleation and the characteristics of the crystal chemical substitutions within both the pyroxene and the

Fig. 7. – Chondrite-normalised rare earth element patterns and primitive mantle normalised incompatible multi-element patterns of selected mafic volcanic rocks (ICP-MS data). The dotted field shows the field of REE patterns for all the samples (ICP-AES data). Normalisation values from Sun and McDonough [1989].

Fig. 8. – Selected isotopic plots for the Late Triassic volcanic rocks. A: (207Pb/204Pb)$_i$ versus (206Pb/204Pb)$_i$; B: (208Pb/204Pb)$_i$ versus (206Pb/204Pb)$_i$; C: (143Nd/144Nd)$_i$ versus (87Sr/86Sr)$_i$; D: (143Nd/144Nd)$_i$ versus (206Pb/204Pb)$_i$. OIB and MORB data are from GEOROC database [http://georoc.mpch-mainz.gwdg.de/georoc/]. DMM, HIMU, EM1, EM2 are reported after Zindler and Hart [1986].
Nd, Sr and Pb isotopic compositions: evidence for an enriched OIB mantle source

The εNd values of the Late Triassic basalts range from +4.6 to +5.2 corresponding to typical intra-oceanic island basalt (OIB) values. No clear correlation is observed between the (143Nd/144Nd)i values and the (87Sr/86Sr)i ratios (fig. 8C). In the (143Nd/144Nd)i versus (206Pb/204Pb)i correlation diagram (fig. 8D), all the Triassic basalts plot within a triangle having as sums the DMM, HIMU and EM2 end-members. This feature suggests that the enriched mantle source of the Triassic basalts could derive from the mixing of these three end-members.

Figures 10A and B illustrate a model accounting for the variations of the Nd and Pb initial isotopic ratios of the studied basalts. The (143Nd/144Nd)i ratios of the Antalya alkaline suite are too low to be explained by a simple mixing line between the DMM and HIMU components (mixing line A, fig. 10A). The Pb and Nd compositions of these volcanics plot closer to the mixing lines C and C’ drawn between HIMU component and an enriched mantle. The composition of this enriched mantle could be the result of the addition of 5 to 8% of an EM2 component to a DMM source (mixing line B). This enriched mantle is indicated in figure 10A by two distinct open boxes corresponding, respectively, to a mixing of DMM + 5% EM2 and DMM + 8% EM2. In the (207Pb/204Pb)i versus (206Pb/204Pb)i diagram, the Antalya alkaline volcanics plot along the mixing line (A) between DMM and HIMU fields, with the exception of sample AY02-20. The latter plots along the mixing line C, involving a DMM source contaminated by an EM2 component. Consequently, we may assume that three components contributed to the genesis of the Antalya volcanic suite: (i) a depleted source mantle (DMM) contaminated by (ii) small amounts of recycled EM2 sediments, and (iii) a HIMU enriched source. The isotope geochemistry of the studied intraplate volcanics can therefore be accounted for by their derivation from mantle components (with a local contribution of pelagic sediments for AY02-20), and there is no need to envision the contribution of continental crust to their genesis. Similar conclusions have been reached by Lapierre et al. [2007] for the Late Triassic Mamonia Complex in SW Cyprus, which differs mostly from the Antalya sequence by the additional occurrence of alkali basalt-related trachytes, and of tholeiites deriving from larger degrees of partial melting of OIB-type mantle.

Incidences on the geodynamic significance of the Antalya Nappes intraplate volcanism

According to tectonic models [Ricou, 1994; Stampfli and Borel, 2002], the Neotethys opening took mostly place during the Middle to Upper Permian, north of the Arabian and Indian plates. The Neotethys opening was accompanied by a northeast drift of several continental blocks, referred to as the Cimmerian Continent [Sengör, 1984], and by the emplacement of intraplate volcanics [Maury et al., 2003; Lapierre et al., 2004] in the basins developed along the southern Neotethyan margin. Seafloor spreading was probably active concomitantly, because of the quick motion of the Cimmerian blocks [Besse et al., 1998] and the continuous subduction of the Palaeotethys. However, remnants of the Late Triassic Neotethyan oceanic crust have never been described until now, possibly because they were entirely subducted beneath the European and Asian plates.

The petrologic, mineralogical and geochemical features of the Late Triassic intraplate alkali basalts exposed in the Antalya Nappes do not suggest that the basement of the basin in which these basalts emplaced was floored by continental crust. The absence of any continental crustal component in the genesis of the Antalya basalts, unlike in those from younger rift-related environments [Chazot and Bertrand, 1993; Baker et al., 1996] or from some of the Middle Permian basins in which the Hawasina Nappes materials were deposited [Lapierre et al., 2004], combined with the pelagic and carbonate-platform types of sedimentation associated with the Antalya basalts, allow to envision their emplacement in an intra-oceanic (plume-related) setting. This OIB volcanism was widespread during the Upper Triassic all along the southern Tethyan margin, from the Himalayas [Bassoullet et al., 1980a, b; Corfield et al., 1999] to Oman [Béchennec et al., 1989, 1991; Pillevuit et al., 1997] and the Eastern Mediterranean, in Cyprus [Malpas et al., 1993; Lapierre et al., 2007], the Baër-Bassit in Syria [Delaune-Mayère, 1984; Al-Riyami and Robertson, 2002] and Antalya.

CONCLUSIONS

The studied submarine alkali basalts and hawaiites, unequivocally dated to Carnian (Julian) by the conodont faunas of the associated sediments, display petrographic, mineralogical and geochemical features typical of intraplate alkaline ocean island basalts (OIB).

**Fig. 9.** – Diagrammes de variation de certains éléments en traces incompatibles en fonction du rapport La/Yb des roches totales.
Their primary Nd, Pb isotopic composition is consistent with their derivation from a mantle source resulting from a mixing between mantle components: a HIMU end-member and another enriched mantle component, resulting from the addition of 5% to 8% of an EM2 end-member (recycled marine sediments) to depleted MORB mantle (DMM).

These isotopic data, as well as the incompatible element compositions of whole rocks and of their preserved clinopyroxene phenocrysts, do not evidence any contribution of continental crust to the genesis of Antalya alkali basalts and hawaiites.

Therefore, the emplacement of these lavas may have occurred in an intra-oceanic (plume-related) setting, such as in the neighbouring island of Cyprus [Lapierre et al., 2007].

Acknowledgements. – This work was funded by the CNRS-INSU program Intérieur de la Terre, UMR 6538, Université de Bretagne Occidentale (Brest) and UMR 5025, Université Joseph Fourier (Grenoble). LK was financially supported by Austrian Academy of Sciences within IGCP Proj. 467 (Triassic Time). We are grateful to the Süleyman Demirel, University of Isparta, and particularly to Prof. F. Yagmurli for their help with logistics. The reviews provided by Mireille Polvé and Georges Mascle led to substantial improvements of the initial manuscript. We dedicate this paper to the memory of our friend Henriette, its second author, who died suddenly on 14 January 2006 during a field study of Late Triassic Baër-Bassit volcanics in Syria. RCM is indebted to Georges Mascle for recovering geochemical data and figures from Henriette’s computers.


