1 The genesis of the Ali Ou Daoud Jurassic carbonate Zn-Pb Mississippi Valley-type deposit, Moroccan
2 Central High Atlas: Constraints from bulk stable C-O-S, in situ radiogenic Pb isotopes, and fluid inclusion
3 studies.

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16 17 18 **Abstract**

19 The proximity of the Ali Ou Daoud Zn-Pb ore deposit (AOD) to the Ikkou Ou Ali diapiroic and
20 magmatic paleohigh raises the question of the role of halokinesis and magmatism in the emplacement of
21 the ore. The AOD ore deposit is hosted in hydrothermally dolomitized Bajocian carbonates. The
22 mineralogical paragenetic sequence is simple and consists of dolomite-pyrite-galena-sphalerite-calcite. The
23 mineralization is tectonically and unconformity controlled. Microthermometry of fluid inclusions in
24 sphalerite indicates that ore precipitated from a NaCl-KCl-CaCl₂-MgCl₂ basin-derived hot (Th= 94 - 170°C)
25 and saline (Salinity = 13.8 - 27.3 wt% NaCl eq.) mixed ore-forming fluid. Post-ore calcite also formed from
26 a similar brine (Th = 176- 206 °C, Salinity = 17.6 - 20.5 wt% NaCl eq.). The $\delta_{34}\text{S}$ values for sulfides range
27 from 7.8 to 23.4‰ V-CDT with most values in the range of 16 to 23.4‰ V-CDT. Sulfur was mainly derived
28 from the Triassic-Jurassic sulfates through thermochemical-sulfate reduction. The carbon-oxygen isotope
29 values of calcite are overall similar to that of the barren Bajocian limestone and dolostone indicating that
30 the calcite-forming fluid is in equilibrium with the carbonate host rock. Galena samples have a
31 homogeneous Pb composition with ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb ratios ranging from 18.358 to
32 18.376, 15.636 to 15.663 and 38.486 to 38.606 respectively. This Pb isotopic composition points to the
33 Paleozoic siliciclastic rocks as the main source of lead and presumably other metals with a possible
34 contribution of Precambrian rocks. The long-distance fluid flow through a thick sequence of Paleozoic-
35 Jurassic rocks ensured the homogeneity of Pb isotopes of the ore-forming fluids before they reached the

loci of deposition. The emplacement of the ore occurred during Late Middle Jurassic-Early Cretaceous extensional tectonic activity in relation to thermal doming when a deep-seated, Paleozoic-derived metal-bearing fluid migrated upward along E-W- and NE-SW- trending faults. At the site of deposition, these hot, metal-carrying fluids mixed with a warm, shallow, metal-poor, and sulfur-rich fluid as revealed by a wide range in salinity and almost similar temperatures. The proposed fluid-mixing led to the formation of hydrothermal metric-sized karsts and to the precipitation of sulfides in the Bajocian carbonates.

1. Introduction

Mississippi Valley-type (MVT) ore deposits worldwide form either in an extensional tectonic setting related to rifting/post-rifting or in a compressional setting related to orogenies. The western part of the Zn-Pb province of the Circum-Mediterranean Sea and Alpine Europe (e.g. Rouvier et al., 1985; Bouhlef, 2005; Bouabdellah and Slack, 2016) host numerous ore deposits. Many of these ore deposits were formed during the Triassic-Jurassic rifting in relation to the opening of the Atlantic Ocean (e.g. Europe: Muchez et al., 2005; North Africa: Margoum et al., 2015). Others, however, formed during the Alpine orogeny (e.g. Morocco: Toussit-Bou Beker, Bouabdellah et al., 2015, Bou-Dahar, Rddad and Bouhlef, 2016). In Western Europe, North Africa, and several other locations worldwide, local factors play an important role in the emplacement of the ore. In the Tunisian-Algerian diapiric zone, for example, Triassic diapirs and the organic matter-rich Cretaceous formations have a determinant role in the genesis of the Pb-Zn-Ba-F ore in this zone (e.g. Orgeval et al., 1986; Charef and Sheppard, 1987; Montacer et al., 1988; Orgeval, 1994; Sheppard et al., 1996; Bouhlef, 2005; Rddad and Belayouni, 2014; Bouhlef et al., 2016). Similar factors were also emphasized in the ore emplacement of Toussit-Bou Beker (Bouabdellah et al., 2015). The Moroccan central High Atlas, which hosts numerous Zn-Pb ore deposits, is part of this western Zn-Pb province. The ore deposits located in the center of the basin and along the southern border are of economic importance whereas those on the northern border are just prospects of no economic value (Rddad and Mouguina, 2018).

In the center of the Moroccan central High Atlas, the juxtaposition of the Jurassic-Early Cretaceous magmatic rocks and the Triassic diapiric evaporites to the Jurassic carbonate-hosted Pb-Zn ore deposits is evident and frequent (e.g. Tazoult, Tirrhist, Ali Ou Daoud). This juxtaposition raises the question of the role played by the Jurassic-Early Cretaceous magmatism and halokinesis in the emplacement of the ore in the Moroccan central High Atlas. Among these ore deposits is the Ali Ou Daoud deposit hosted in the Jurassic formations and is located near the Ikkou Ou Ali paleohigh. The latter was intruded by the Triassic diapirs and the Jurassic igneous intrusions. Despite its small size, the Ali Ou Daoud ore deposit is an ideal case study to determine the role and interplay of the three geological components (Jurassic-Early Cretaceous magmatism, halokinesis, and tectonism) in the emplacement of the Zn-Pb ore. This study

provides an insight into the genesis of the other ore deposits in the central High Atlas, and also shows the importance of local factors in the formation of the ore. Moreover, ridges with magmatic intrusion and evaporitic diapir may constitute a potential target for ore exploration in central High Atlas.

The Bajocian carbonate-hosted Ali Ou Daoud ore deposit, the focus of this study, is located 7 km from Imilchil. The Ali Ou Daoud ore deposit has produced around 130 tonnes of ore grading 45% Zn and 18% Pb (Bouladon, 1953). Afterward, local artisanal miners have been permitted to exploit this mine. Currently, the mine is closed with a research license allocated to Managem, a division of the Société Nationale d'Investissement (SNI). Very few studies were undertaken in the Zn-Pb-Ba High Atlas province. Although Rddad and Bouhlef (2016) constrained the genesis of the Bou-Dahar Zn-Pb-(Ba) ore deposits located in the oriental High Atlas, the genesis of Zn-Pb ore deposits in the central High Atlas have not been fully constrained. Based on geological and geochemical studies, Mouguina (2004) concluded that there is both syngenetic and epigenetic mineralization and that two fluids were involved in the genesis of the Ali Ou Daoud Zn-Pb ore deposit. Researchers who studied this ore deposit (e.g. Mouguina, 2004; Mouguina and Daoudi, 2008) did not reach a comprehensive ore genetic model. Moreover, the nature of the fluids involved in the ore emplacement, the source of sulfur and metals, and modes and time of fluid migration and ore precipitation are not fully constrained. The present study, thus, aims at thoroughly constraining the nature, interrelationships of the ore-forming fluids, the sources, mode and age of migration of metals and sulfur. Therefore, this work combines C-O-S stable isotopes, Pb radiogenic isotopes, and fluid inclusion microthermometry. This work sheds light not only on the genesis of the small AOD ore deposit but also provides an insight into the ore genesis in the central High Atlas metallogenic province.

2. Regional geologic setting

The Atlas Mountains extend from Morocco to Tunisia with significant lateral variations in the style of deformation. The Moroccan segment consists of the NE-SW-trending Middle Atlas and the ENE-WSW-trending High Atlas (Fig. 1A). It consists of Mesozoic to Cenozoic sedimentary rocks covering a Precambrian to Paleozoic basement, which was deformed by the Hercynian orogeny (Piqué and Michard, 1989). This basement is composed of volcano-clastic and metamorphic rocks (Nelter, 1938; Nefly, 1998; El Archi et al., 2004). The Paleozoic succession, mainly composed of siliciclastic rocks, also outcrops in the Mougues and Tamlet inliers.

The central High Atlas region underwent Triassic rifting which was related to the opening of the central Atlantic Ocean and the formation of the Tethys Ocean (Laville, 1988; Saura et al., 2014). The Hercynian NE-SW-trending faults, entrenched in the Paleozoic basement, were reactivated (Laville, 1985) and led to the formation of the Triassic rift basins. These basins received detrital siliciclastic sediments which were deposited under arid conditions (Courel et al., 2003) and are intercalated with tholeiitic basalts with a

doleritic texture (Du Dresnay, 1987). This magmatism represents the first magmatic event, known as the central Atlantic Magmatic event. During the Late Triassic-Early Jurassic arid climate, thick evaporite sequences were deposited mainly in the center of the central High basin (Saura et al, 2014; Moragas et al., 2016). The Lower Liassic carbonate platform, overlapping the Triassic series, was developed and evolved into marls alternating with limestone (Du Dresnay, 1979). This change in sedimentation was related to the Middle Liassic-Middle Jurassic rifting (Warme, 1988; Ellouz, et al., 2003). This second rifting phase resulted in the deepening of the basin and the differentiation of the ridges (Du Dresnay, 1979; Laville, 1988). In the central High Atlas, these ridges were interpreted as the result of transpressive tectonics that occurred during the first rifting phase (Laville and Harmand, 1982). Recently, new data showed that these ridges, mainly of the central High Atlas, were formed as a result of Triassic salt diapirism (Bouchouata, 1994; Ettaki et al., 2007; Ibouh et al., 2011; Michard et al., 2011, Saura et al., 2014; Martín-Martín et al., 2016; Moragas et al., 2016; Vergés et al., 2017). Therefore, diapirism and transpressive tectonics contributed to the formation of these ridges. The compartmentalization of the region into mini-basins and ridges is the result of the combined effect of extensional faulting and halokinesis (Saura et al., 2014; Moragas et al., 2016). These NE-SW-trending ridges were intruded by Upper Triassic evaporite diapirs (Moragas et al., 2016; Vergés et al., 2017) during the Early Jurassic time and later by magmatic intrusions (Laville, 1985). These evaporites and igneous intrusions cross-cut the Middle Jurassic sedimentary series (e.g. Seufert, 1988). The age of the emplacement of mafic, gabbroic igneous intrusions along the NE-SW-trending ridges and the associated basaltic, alkaline lava flows is constrained to the Middle Jurassic - Early Cretaceous (Jenny et al., 1981; Laville and Harmand, 1982; Laville et al., 2003; Lhachmi et al., 2001; Zayane et al., 2002). This Jurassic-Early Cretaceous magmatism represents the second magmatic event and is related to the crustal extensional tectonic activity (Frizon de Lamotte et al., 2009). This extensional event is related to the Late Middle Jurassic-Early Cretaceous thermal doming induced by the upwelling of the asthenosphere (Frizon de Lamotte et al., 2009). At the end of the Middle Jurassic (Late Bajocian-Bathonian), a regional uplift occurred (Saddiqi et al., 2009) and a marine regression led to the deposition of continental red beds (Jenny et al., 1981; Charrière and Haddoumi, 2016).

Tectonic inversion due to the convergence between the African and European plates started toward the end of the Cretaceous period (e.g. Tapponier, 1977). Successive compressive events occurred during the Cenozoic and Quaternary (e.g. Laville, 1985). The first tectonic Atlasic event occurred during the Middle-Late Eocene, followed by a quiescent period during the Oligocene-Miocene (Frizon de Lamotte et al., 2009). The second shortening Atlasic event, responsible for the uplift and the formation of the High Atlas, took place during the Late Miocene (Frizon de Lamotte et al., 2009). It is during the Alpine orogeny that a third alkaline magmatic event took place, i.e. during the Eocene (Tisserant et al., 1976; Bouabdellah et al., 2010). This magmatic event is characterized by the formation of carbonatites and undersaturated rocks of

Tamazert located at the northern border of the central High Atlas. These magmatic rocks were formed from the mantle as a result of the delamination of the subcontinental lithosphere in response to the African-European collision (Duggen et al., 2009; Bouabdellah et al., 2010).

The tectonic model responsible for the formation of the Moroccan High Atlas is predominantly a thick-skinned model which involved the Hercynian faults, the Paleozoic basement, and the Mesozoic-Cretaceous cover (Teixell et al., 2003). Currently, the High Atlas is marked by large-scale folds dominated by the NE-SW-trending ridges, separated by wide synclines.

The Moroccan High Atlas Zn-Pb metallogenic province hosts several prospects and ore districts (Ovtracht, 1978). It constitutes the western part of the Zn-Pb province of the Circum-Mediterranean Sea and Alpine Europe (e.g. Rouvier et al., 1985; Bouhlel, 2005). Several Zn-Pb ore deposits, hosted in Jurassic carbonates, occur in the center of the High Atlas (i.e. Ali Ou Daoud, Tazoult), northern and Southern borders, and also near the Precambrian-Paleozoic paleohigh (i.e. Tadaghaste). The central High Atlas, in particular, hosts numerous polymetallic Zn-Pb±Cu±Co±Ni±W±Au ore deposits and prospects with most of them located at the borders of the central High Atlas. Some ore deposits, including the AOD mine, are located proximal to the paleohighs, which are most often occupied by Triassic salt (diapirs) and Jurassic igneous intrusions. The ore deposits are hosted either in the Middle Jurassic gabbros or in carbonates of Liassic to Dogger age.

According to Mouguina (2004), two main ore-forming events were responsible for the emplacement of the central High Atlas ore deposits. A first Liassic Zn-Pb mineralization event is related to the Triassic-Liassic thermal event that occurred during the opening of the central Atlantic Ocean. The mineralization of this event consists of syngenetic Zn-Pb-Fe ore deposits (Upper Moulouya ore district, Emberger, 1965; Tizi n'Firest, south border of the oriental High Atlas, Bazin, 1968; Tazekka ore deposit, Eastern Morocco, Auajjar and Boulègue, 1999). A second Dogger Zn-Pb mineralizing event is related to the evolution of the Jurassic magmatism well expressed along the paleohighs (ridges) of the central High Atlas. The Pb-Zn ore is mainly disseminated in gabbro, with (i) an early magmatic related Cu-Ni mineralization (Chèvremont, 1975), (ii) a syngenetic Bajocian-hosted Zn-Pb episode, and (iii) an epigenetic Zn-Pb hydrothermal episode (e.g. AOD, Mouguina, 2004). On the basis of structural and mineralogical studies, Mouguina (2004) and Mouguina and Daoudi (2008) proposed that some of the Middle Jurassic Zn-Pb ore deposits (e.g. AOD, Tazoult) of the central High Atlas are Mississippi Valley Type deposits similar to the ones described in the oriental High Atlas (e.g. Bou Dahar, Rddad and Bouhlel, 2016), and the High Plateau zone (e.g., Touissit and Mibladen) (Dagallier, 1977; Wadjinny, 1989; Bouabdellah et al., 2012). A third ore-forming event took place during the Eocene and is characterized mainly by the fluorine rare-earth ore deposits of Tamazert (El Ouazzani, 2005; Bouabdellah et al., 2010; Achmani, 2017). These ore deposits occur about 100 km away from the AOD ore Zn-Pb deposits. This mineralizing event is related to the emplacement of the Eocene

alkaline magmatism ([Achmani, 2017](#)).

3. Geology of the Ali Ou Daoud ore deposit.

The AOD Zn-Pb ore deposit is located at the center of the High Atlas basin, 7 km SW of the town of Imilchil (Fig. 1 A, B). The rocks outcropping in the Imilchil area are represented in Figure 2. The mine, located on NE limb of the Ikkou Ou Ali paleohigh, is hosted in the Bajocian carbonates (Fig. 3). The lithostratigraphic facies outcropping in the AOD area are represented by the Tislit carbonate formation of Middle to Late Bajocian age ([Kadmiri, 1998](#)). A synthetic geologic cross-section at the AOD mine shows two main members, which are from the base upwards ([Mouguina, 2004](#)) (Fig. 4A):

1- Lower member (~ 100 m) — This member has been deposited on the Jurassic carbonate platform and hosts the ore. This member consists of the following units:

- Unit 1 (~ 10-20 m) is composed of metric thick gray to beige layers of limestone and dolostone. It represents the footwall of the ore.
- Unit 2 (10 m) consists of dolomitic limestone and dolostone associated with mud-mounds. This is marked by an erosional surface/unconformity above which the Zn-Pb mineralization is encountered.
- Unit 3 (40 m) represents the ore hanging wall and consists of massive black limestone beds showing slumps and mud-mounds.
- Unit 4 (40 m) is composed of bioclastic limestones alternating with yellowish marls and red sandstones.

The upper limestone bed shows a paleo-erosional surface with bioturbation and debris of wood fossils.

2- Upper member— This member consists of a ~300 m-thick sequence of marls alternating with few limestone beds. This member is marked by an angular unconformity which is overlain by an iron-rich carbonate which has been deposited as a sedimentary bar.

With respect to the tectonic setting, the AOD district is controlled by Triassic-Jurassic rifting and halokinetic activity ([Saura et al., 2014](#)). The rifting reactivated the roughly E-W- and NE-SW-trending faults during the Triassic-Jurassic time. The Ali Ou Daoud area is affected by these major deep-seated faults (Fig. 3). Of particular importance are the two synsedimentary faults (F_1 and F_2) that affect the Bajocian carbonate and host the ore (Fig. 3 and Fig. 5). Considering the occurrence of the Triassic diapirs in the Imilchil area ([Saura et al., 2014](#); [Martín-Martín et al., 2016](#); [Moragas et al., 2016](#); [Vergés et al., 2017](#)), it is likely that the Ikkou Ou Ali ridge is the result of the emplacement of Triassic salt diapirs during Early-Middle Jurassic (Fig. 3) (e.g. [Saura et al., 2014](#); [Vergés et al., 2017](#)). These ridges were later intruded by igneous rocks during Middle to Late Jurassic. It is the combined effect of the Early-Middle Jurassic rifting, magmatic and halokinetic activities that have led to the development of the Bajocian platform carbonates which host the ore (Fig. 3). It also led to the uplift of the carbonate platform, which resulted in the development of the erosional surfaces/unconformities. The post-Bajocian basin evolution is mainly

controlled by the Middle to Late Jurassic magmatic activity (Laville and Harmand, 1982; Laville et al., 2003) and halokinetic salt activity (Moragas et al., 2016; Vergés et al., 2017). The Bajocian carbonate was sealed by Bajocian marls. During the Late Bajocian-Early Bathonian, part of the Ikkou Ou Ali ridge was submerged with concomitant subsidence of neighboring sectors (Ibouh and Chafiki, 2008). It is worth noting that the Ikkou Ou Ali ridge was emplaced during the Bajocian -Bathonian time in an extensional tectonic setting (Frizon de Lamotte et al., 2008, 2009). The Bathonian red beds overly the SE of the Ikkou Ou Ali ridge with an angular discordance (Ibouh and Chafiki, 2008).

During the Alpine Tertiary inversion, the rocks outcropping in the Ali Ou Daoud as well as in the Imilchil area were folded. The Jurassic intrusions and the Triassic salt diapirs became squeezed. The mine is located in one of these Alpine-related structures, i.e. the Ikkou Ou Ali anticline. The ore, hosted in the Bajocian carbonate, is located in the northern limb of the anticline (Fig. 3). This limb forms a monoclinical structure dipping 30° NNW. This structure consists of tilted blocks delineated/bordered by E-W- to ENE-WSW-trending faults.

4. Dolomitization and Mineralization

The mineralized faults F₁ and F₂ are surrounded by a halo of dolomite (Fig. 6). The macroscopic and microscopic observations of the mineralized zones reveal two types of dolomitization: an early pre-ore diagenetic dolomitization (D₀) and a syn-ore dolomitization (D₁). The former is characterized by anhedral to subhedral Fe-rich dolomite which formed within birds-eyes, substitute bioclasts, and cements pellets. This pre-ore dolomitization could have increased the porosity and permeability of the limestone and thus created a suitable reservoir for the emplacement of the ore. The syn-ore dolomitization occurs in mineralized vugs and fractures and is represented by anhedral dolomite and saddle dolomite which shows a sweeping extinction (Figs. 6A, B, C). A layer of porous, friable dolostone of 0.5 m thickness marks the base of karsts (Fig. 6G). Saddle dolomite commonly forms at a temperature range of 50 to 150 °C (Radke and Mathis, 1980; Gregg and Sibley, 1984). These dolomites are cemented by sulfides which consist mainly of sphalerite and galena.

Based on the orebodies' description, two main mineralization styles with a similar and simple paragenesis (pyrite-galena-sphalerite-calcite) are distinguished: open-space fillings and replacement (Figs. 6D, E, F, G, H, I). The ore occurs mainly in open-space fillings (fractures, faults, karsts, vugs). Two sets of mineralized faults can be distinguished. The first set of mineralized faults and associated fractures are oriented NE-SW to E-W. Of great importance are the two local NE-SW-trending mineralized faults called F₁ and F₂ (Fig 5). These faults are parallel to the regional NE-SW-trending-faults (e.g. Major fault, Fig. 5) and are entrenched at the base of the karsts described below. The second set of faults are the NW-SE- to N-S-trending faults that cross-cut the former faults. These faults, delineating the karsts, were formed during Late Bajocian-

Early Bathonian extensional tectonic activity ([Mouguina, 2004](#)).

Sulfide mineralization (sphalerite and galena) occurs as euhedral, coarse-grained crystals (Fig. 6F). Locally, a dense, thin network of fractures forms a mineralized stockwork. Locally, sulfides replace bioclasts in the partially dolomitized limestone (Fig. 6D). Dissolution cavities delimited by NNW-SSE- and ENE-WSE-trending faults/fractures form karsts (Figs. 4B and 7). These ore-hosting karsts are confined to the dolostone strata and are oriented N-S to NNW-SSE and form lens-shaped orebodies. They extend over 20 m with a width of 2 m and a thickness of 3 to 6 m. These karsts are filled with a 1 to 2 m-thick sequence of sulfides. These sulfides display a mineralogical zonation, which starts with a ~0.5m thick layer consisting of massive, euhedral, coarse-grained brecciated pyrite (Fig. 6I). This pyritic layer is overlain by a 20 cm-thick layer of coarse-grained galena. This galena-rich layer is followed by coarse-grained, euhedral sphalerite and galena (Figs. 6E, C). The final 10 cm-thick mineralization is composed of euhedral, medium- to coarse-grained sphalerite. Within the same mineralized layer, the size of sulfide crystals increases upward. A 2-m thick dolomitized limestone hosts numerous mineralized cm-sized vugs (up to 20 cm) (Fig. 6B) occurring along and parallel to the stratification. The number and size of the vugs increase upward where they give the dolostone a zebra texture. Their anastomosing morphology is similar to that of fenestrae. At the top of the first mineralized dolostone occur porous, friable dolostone-bearing vugs, disseminated galena and veins of post-ore-related coarse-grained calcite. The vugs are lined with baroque dolomite and successively filled with sphalerite and galena. Disseminated ore (sulfides) occurs along the dolomitic limestone as swarms each of which is guided by fractures parallel to the NE-SW- to E-W-trending faults. These sulfides occur as replacement of the host rock and locally substitute bioclasts. The sulfide ores represented by pyrite, sphalerite, and galena were oxidized to hematite-goethite, cerussite, and smithsonite-hemimorphite respectively.

The macroscopic and microscopic study of the orebodies led to distinguish four mineral stages (Tab. 1). The first one is the pre-ore dolomite which consists of dolomite crystals which formed during the diagenetic phase, prior to the main ore-related dolomitization of stage 2. The latter is represented by ore-related dolomite (iron-rich subeuhedral to xenotopic dolomite and saddle dolomite) that occurs in the vugs, fissures, and stockworks (Figs. 6A, B, C). The third mineralization stage consists of pyrite, sphalerite, galena (Figs. 6D, E, F, G, H, I) which are often cross-cut by the post-ore calcite (Fig. 6H). These sulfides occur mainly in open-space fillings (faults, fractures, dissolution cavities, vugs, stockwork). The last and fourth stage corresponds to the oxidation of sulfide ores.

5. Sampling and analytical procedures

A total of ten carbonate and sixteen sulfide representative samples were collected from different ore styles within the mineralization. Five sulfide samples and one calcite sample were prepared as standard 200

µm-thick, double-polished wafers. Fluid inclusion assemblages (FIAs) of Goldstein and Reynolds (1994) and the genetic classification as primary, secondary or pseudosecondary of Roedder (1984) were used to classify the fluid inclusions. The microthermometric measurements were carried out at the Department of Geology, Cortland College, State University of New York (SUNY), using a Linkam THM600 freezing-heating stage mounted on a Leitz microscope. The stage was calibrated with synthetic fluid inclusions at -56.6°C (the CO₂ triple point), 0.0°C (the H₂O triple point), and +374.1°C (the H₂O critical point). The precision is calculated at ±0.1°C. The accuracy of our measurements less than -20°C is estimated to be ±0.2°C, between -20°C and +30°C, it is ±1°C, and above +100°C, it is 5–10°C (Xavier and Foster, 1999). Fluid salinity was calculated as weight percent NaCl equivalent (wt% NaCl eq.) using the equations of Bodnar (1993) and Steele-MacInnis et al. (2011). Previous fluid inclusion data from Mouguina (2004) are also used in this study.

Mineral separates for C, O, S, Pb isotope analyses were prepared under a binocular microscope by handpicking or by using a micro-drill. Carbon and oxygen stable isotope measurements were performed at the Stable Isotope Mass Spectrometer facility at the Memorial University of Newfoundland, Canada. Approximately 0.2 mg of powder sample was placed in vials and heated at 50 °C in an inert atmosphere. Samples were then injected with approximately 0.05 ml of phosphoric acid (100%) manually, using a syringe and needle. A minimum of one hour was allowed for samples to react in a Thermo Finnigan Gasbench II, and the produced CO₂ was automatically delivered to the source of a Thermo Finnigan Delta V plus isotope ratio mass spectrometer in a stream of helium, where the gas was ionized and measured for isotope ratios. The carbon isotopic compositions are reported in delta notation (δ₁₃C) relative to V-PDB, and the oxygen isotopic compositions are reported in delta notation (δ₁₈O) relative to V-PDB and V-SMOW. The precision of carbonate analyses was better than ± 0.1‰ for both δ₁₃C and for δ₁₈O.

Sulfur isotopes analyses were carried out on six sphalerite samples and ten galena samples from different ore levels at the Spectrometry facility of the Department of Engineering, University of Nevada. Sulfur isotope analyses were performed using a Eurovector elemental analyzer interfaced to a Micromass isoprime stable isotope ratio mass spectrometer, after the methods of Giesemann et al. (1994) and Grassineau et al. (2001). Vanadium oxide (V₂O₅) was added to sulfate samples as a combustion aid. S-isotopic compositions in sulfide minerals are expressed as δ₃₄S values relative to the V-CDT standard. Reproducibility was ±0.2‰.

Lead isotope analyses were performed on nine galena samples at the Department of Earth Sciences, University of New Brunswick, Fredericton, Canada. The galena samples were analyzed using the Laser Ablation Quadrupole Inductively Coupled Plasma Mass Spectrometry (LA Q-ICP-MS) instrument, which consists of ASI (formerly Resonetics) Resolution™ M-50 193 nm ArF (excimer) laser system synchronized to an Agilent 7700x quadrupole inductively coupled plasma mass spectrometer equipped with dual external rotary pumps. For more details on this technique, the reader is referred to McFarlane et al. (2016). The

precision of Pb isotope analyses was better than ± 0.06 %.

6. Results and interpretation

6.1. Fluid inclusion petrography and microthermometry

Fluid inclusions studies carried out on sphalerite and calcite samples shed light on the nature of the fluids involved in ore precipitation. Sphalerite and calcite samples were collected from different ore styles and were analyzed microthermometrically. The fluid inclusions hosted in sphalerite and post-ore calcite are rounded, elongated, and rarely irregular in shape, and the inclusions measured have an average size of 30 μm . The primary fluid inclusions (L+V) are aqueous, two-phase (L/L+V ratio of 90 to 95%) and occur along growth zones or isolated clusters in sphalerite (Fig. 8A) and calcite (Fig. 8B). The secondary fluid inclusions are also two-phase (L + V) aqueous and are abundant. They are, however, not investigated because of their very small size, which renders their microthermometric study extremely difficult.

The microthermometric data of the primary fluid inclusions are listed in Table 2 and shown in Figures 9A, 9B, and Table 2. The eutectic temperatures (T_e) of the primary fluid inclusions for both mineral populations range from -50 to -27 $^{\circ}\text{C}$, far below the eutectic of the H_2O -NaCl system (-21.2 $^{\circ}\text{C}$). Taking into account the carbonate host-rock and gangue mineral in which some of the fluid inclusions measured occur, the presence of a H_2O -NaCl- CaCl_2 fluid system is inferred (cf. [Roedder, 1984](#)). Ice melting temperature values ($T_{\text{m,ice}}$) vary from -9.9 to -26 $^{\circ}\text{C}$ (avg. = -18.0 $^{\circ}\text{C}$) and from -13.8 to -20.4 $^{\circ}\text{C}$ (avg. = -17.1 $^{\circ}\text{C}$) for fluid inclusions hosted in sphalerite and calcite respectively (Fig 9A). Calculated salinities range between 13.8 and 27.3 wt% NaCl eq. (avg. ~ 20.6 wt% NaCl eq.) and between 17.6 and 20.5 wt% NaCl eq. (avg. ~ 19.1 wt% NaCl eq.) respectively for sphalerite and calcite. Homogenization temperatures (T_h) vary from 94 to 292 $^{\circ}\text{C}$ (avg. = 190 $^{\circ}\text{C}$) and from 176 to 206 $^{\circ}\text{C}$ (avg. ~ 193 $^{\circ}\text{C}$) for sphalerite and calcite respectively. The high T_h values ($\geq 192^{\circ}\text{C}$) in sphalerite from the study of [Mouguina \(2004\)](#) were obtained on large (100-200 μm) inclusions only and are therefore suspected of stretching and/or leaking during heating runs. Large inclusions in sphalerite decrepitate at lower internal pressures than small inclusions ([Ulrich and Bodnar, 1988](#)). Therefore, these fluid inclusions are not further considered in this study. Taking this into consideration, fluid inclusions recorded in sphalerite have T_h values in the range of 94 to 170 $^{\circ}\text{C}$ (avg. = 132 $^{\circ}\text{C}$).

The homogenization-salinity diagram (Fig. 10) shows a wide range of temperatures in sphalerite (T_h = 94 to 170 $^{\circ}\text{C}$) and salinities (13.8 to 27.3 wt% NaCl eq.) The diagram shows at least two fluids with contrasting salinities and almost similar temperatures: a high temperature, very saline fluid and a warm, low saline fluid (Fig. 10). The range of salinities reflects mixing between these two end-member fluids. The first fluid referred hereafter as ore-forming fluid (F_{ore}) had a high temperature (up to 170 $^{\circ}\text{C}$) and high salinity (up to ~ 27 wt% NaCl eq.). The second fluid (F_{sub}) is characterized by a high temperature and a low

salinity and represents the in-situ fluid. The ascending ore-forming fluid (F_{ore}) encountered and mixed with a warm, less saline fluid (F_{sub}) at the site of the deposition.

6. 2. Carbon and oxygen isotopes

In addition to fluid inclusion studies, C-O isotopic analysis was performed on post-ore calcite to obtain further insight into the nature of the fluids that prevailed during the ore formation. Barren limestone and mineralized dolostone were also analyzed for their C-O isotopic compositions.

The C-O isotopic composition for barren Bajocian limestone, dolostone, and calcite are reported in Table 3 and Figure 11. The limestone and dolostone host rocks display a uniform C-O isotopic composition with $\delta_{13}\text{C}$ and $\delta_{18}\text{O}$ values varying between -2.5 and 2.2‰ V-PDB (avg. = - 0.2‰ V-PDB) and between 24.2 and 25.1 ‰ V-SMOW (avg. = 24.7‰ V-SMOW). This C-O composition range falls within that of Jurassic marine limestone (Veizer and Hoefs, 1976; Land, 1980). This C-O isotopic signature is similar to that reported for the Bou Dahar Liassic limestone (Rddad and Bouhlef, 2016). The calcite samples display a relatively wide range of carbon isotopic values ($\delta_{13}\text{C}$ = -9.0 - 1.5‰, avg. ~ - 5.3 ‰) and a narrow range of oxygen isotopic values ($\delta_{18}\text{O}$ = 23.2 - 27.2 ‰; avg. ~ 25.2 ‰) similar to that of the carbonate host rock.

6. 3. Sulfur isotopes

To constrain the source(s) of sulfur for sulfides, a sulfur isotopic analysis was carried out on different generations of sulfides (pyrite, galena, sphalerite) hosted in different ore styles (Fig. 12). The sulfur isotopic composition of sphalerite, galena, and pyrite are listed in Table 4 and Figure 15. The $\delta_{34}\text{S}$ values are all positive, ranging from 7.8 to 23.4 (avg. = 22.1 ‰), 16.5 to 18.8 ‰ (avg. = 17.7 ‰), and 20.6 to 22.1 ‰ (avg. = 21.4‰) respectively for sphalerite, galena, and pyrite. These sulfur isotopic values fall in the range of those of the Triassic and Jurassic seawaters (+13 to +20‰) (Claypool et al., 1980; Kampschulte and Strauss, 2004). The fractionation factors between coexisting sphalerite-galena pairs range from 2.5 ‰ to 5.6 ‰. Using the equation $\alpha_{\text{ZnS-PbS}} = (0.73/T_2) \times 10^6$ (T in Kelvin) (Ohmoto and Rye, 1979), the calculated temperatures vary between 90 °C and 195 °C, excluding an outlier (270 °C) (Table 3). These temperatures are in agreement with those obtained from fluid inclusion data. The $\delta_{34}\text{S}$ values show the following isotopic increasing trend: $\delta_{34}\text{S}_{\text{Galena}} < \delta_{34}\text{S}_{\text{Pyrite}} < \delta_{34}\text{S}_{\text{Sphalerite}}$, regardless of the ore style and the level at which sulfides samples are collected.

6. 4. Lead isotopes

The lead isotopic composition of galena is commonly used to constrain the possible origin of Pb and presumably other metals in ore deposits. Galena samples were collected from different ore styles and were analyzed for their Pb isotopic composition. Lead isotope compositions of galena samples (n = 9) from the

AOD ore deposit are presented in Table 5 and plotted in the $^{208}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ (thorogenic) diagram and the $^{207}\text{Pb}/^{204}\text{Pb}$ vs. $^{206}\text{Pb}/^{204}\text{Pb}$ (uranogenic) diagram (Fig. 12). The Pb isotopic composition of galena from the Bou Dahar MVT Jurassic carbonate-hosted Pb-Zn-(Ba) district located in the oriental High Atlas (Rddad and Bouhlel, 2016), from the Tadaghaste Jurassic carbonate-hosted Pb-Zn ore deposit located in the proximity of the Precambrian/Paleozoic basement (Rddad and Mouguina, 2018), and from Touissit-Bou Beker (Bouabdellah et al., 2012) are also plotted for comparison. The Pb isotopic composition of the AOD ores is homogenous with $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$, and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios ranging respectively from 18.358 to 18.376, 15.636 to 15.663 and 38.486 to 38.606.

7. Discussion

On the basis of the geological and geochemical data (C-O-S-Pb isotopes, fluid inclusions) presented above, important constraints to the ore genesis of the AOD are highlighted here. The nature of the fluids involved in the ore emplacement, the possible source(s) of sulfur and metals, precipitation mechanism of the ore, and the age of the ore are discussed. Ultimately, an ore genesis model for the AOD ore deposit is proposed.

7.1. Nature of the ore-forming fluid and ore deposition

Microthermometric analysis of the fluid inclusions hosted in sphalerite allows the physico-chemical characterization of the fluids involved in the ore formation. The T_h -Salinity diagram reveals the involvement of an ascending, hot, ore-bearing, saline fluid and a subsurface, warm fluid of lower salinity in the precipitation of the ore in AOD deposit.

The salinities and temperatures of the fluids, involved in ore deposition, fall within those of most MVT ore-forming fluids (salinity = 15-30 wt% NaCl eq. and temperature = 75 °C-200 °C) and are typical of basinal brines (Basuki and Spooner, 2004; Leach et al., 2005). The high salinity recorded in sphalerite-hosted fluid inclusions can be attributed to evaporated seawater or the dissolution of evaporites by the ambient fluids. The relatively high temperatures recorded in both fluids (F_{ore} and F_{sub}) and in calcite-hosted fluid inclusions (up to 206 °C) clearly indicate the presence of an abnormal geothermal gradient. Taking into account the estimated maximum burial depth of the Triassic-Jurassic rocks, a normal geothermal gradient cannot explain the very high temperature of the ore-forming fluid. With a maximum depth of 5 km (Malaval, 2016) and a gradient of 30°C/km, a temperature of 150°C can be reached at the Paleozoic-Triassic boundary. This temperature would have further decreased to a lower temperature during the ascension of the ore-forming fluid towards the loci of deposition and subsequent mixing with a subsurface fluid. The resulting temperature would have been far below the maximum temperature of 170 °C and 206 °C respectively recorded in sphalerite and calcite. The high temperatures recorded in the ore and calcite can

be, therefore, explained by the combination of geological processes. The first is an unusually high geothermal gradient present during ore emplacement. This abnormally high geothermal gradient can be explained by the Jurassic-Early Cretaceous magmatic event in relation to thermal doming (see section 7.4). During the Middle Jurassic-Early Cretaceous, an unusually high geothermal gradient of 130 °C /km (in the proximity of the magmatic intrusions) and 60 °C /km (2 km away from the intrusions) was present (Zayane, 1992). The emplacement of ore during the Middle Jurassic-Early Cretaceous renders this scenario most likely. The second geological process is the ascent of deeply circulating fluids from basement rocks beneath the deposits during extensional tectonic activity (e.g., Irish Midlands and Upper Silesia; Wilkinson et al., 2003; Heijlen et al., 2003). Due to the presence of the deep-seated faults, connecting the Paleozoic basement to the Jurassic host rocks, this second proposed process could have also contributed to the high temperatures observed. The last, but not least mechanism is advective heat transport from deeper parts of the basin as observed for many other MVT and sediment-hosted Pb-Zn ore districts (e.g., Touissit-Bou Beker, Bouabdellah et al., 1996; Ozark MVT province, Leach et al., 2010). Based on the geological setting, though this mechanism may have contributed in increasing the geothermal gradient, the two proposed scenarios/mechanisms, mentioned above, are likely the main processes responsible for the abnormally high geothermal gradient. The emplacement of the ore at a depth not exceeding 1 km and the high temperature of the subsurface fluid (F_{sub}) imply and confirm the presence of a high geothermal gradient to account for the high temperature recorded in the fluid inclusions. The relatively high temperature recorded in the post-ore calcite (up to 206 °C) compared to that of sphalerite (up to 170°C) likely indicates that a magmatic pulse was present during the precipitation of this calcite. This post-ore heating could explain the decrepitation/leakage of the large-sized fluid inclusions in the sphalerite.

Using a temperature range of 177 - 206 °C, the calculated $\delta_{18}\text{O}_{\text{SMOW}}$ of the fluid from which calcite precipitated, using the equation of Friedman and O'Neil (1979), is around 8 - 13 ‰ V-SMOW. This isotopic composition falls in the range of that of basinal brines ($\delta_{18}\text{O} = +4$ to $+10$ ‰) (Hoefs, 2009). Fluid inclusions data indicate that the fluids involved in ore emplacement are modified basinal brines and are categorized as a $\text{H}_2\text{O}-\text{NaCl}-\text{CaCl}_2$ fluid as deduced from the first melting temperatures ($T_e = -50$ to -25 °C). On the basis of the $\delta_{18}\text{O}_{\text{SMOW}}$ of the calcite-forming fluid (8 - 13 ‰ V-SMOW) and the high temperature, the calcite-forming fluids and likely also the ore-forming fluids were basinal brines. The negative $\delta_{13}\text{C}$ values recorded in post-ore calcite samples suggest the presence of hydrocarbon-derived organic carbon during the precipitation of calcite and likely also of sulfide ore.

The ore precipitation mechanisms, common in numerous MVT and sediment-hosted Pb-Zn ore deposits, include (i) cooling of the hydrothermal system, (ii) pH decrease, and (iii) mixing between two or more fluids (Anderson, 1973; Helgeson, 1970; Sverjensky, 1981). Fluid inclusion data point to a mixing process as the main process that caused ore precipitation in the AOD ore deposit. The upward migration of a metal-

bearing, sulfur-depleted, hot brine (F_{ore}), originating from the deep Paleozoic basement, most likely mixed with a shallow, warm, metal-depleted, sulfur-rich reservoir (F_{sub}). Such mixing accounts for the relatively wide range of salinity obtained from sphalerite-hosted fluid inclusions.

The mixing of these two fluids resulted in the formation of a hot saline fluid, triggering the precipitation of sulfides and the generation of large amounts of acids. This led to the formation of the hydrothermal metric-sized karsts and dissolution cavities. In fact, [Corbella et al. \(2004\)](#) demonstrated on the basis of reactive transport modeling that fluid mixing concomitantly triggers carbonate dissolution and precipitation of sulfides in MVT/sediment-hosted Pb-Zn deposits. The increase of the pH due to carbonate dissolution and a decrease in temperature most likely contributed to the precipitation of the ore. Such a fluid-mixing model has been proposed for numerous sediment-hosted Zn-Pb ores where the mixing of a hot, metal-bearing fluid with a cooler, diluted fluid triggered ore precipitation ([Anderson, 1973](#); [Beales and Jackson, 1966](#); [Anderson and Macqueen, 1988](#); [Sverjensky, 1986](#); [Plumlee et al., 1994](#); [Leach et al., 2005](#)). Magmatism could have played a role in the circulation of the fluids involved in the ore emplacement. Such likely magmatic involvement places this small Zn-Pb ore deposit apart from other MVT deposits in North Africa in general and the Moroccan High Atlas in particular. Based on the aforementioned discussion, we propose that the AOD ore resulted from the mixing of an ascending, hydrothermal, metal-bearing brine (F_{ore}) with a local warm fluid (F_{sub}) at the site of ore deposition in an extensional Jurassic-Early Cretaceous tectonic setting in relation to the Jurassic-Early Cretaceous magmatism.

7.2. Source of sulfur

The reduced sulfur required for the precipitation of the sulfides could have been derived from different reservoirs, which include sulfate-bearing evaporites, connate seawater, diagenetic sulfides, sulfur-bearing organic material, or H_2S reservoir gas ([Leach et al., 2010](#)). Owing to the absence of sulfates in the ore deposits, the total sulfur isotopic composition ($\delta^{34}\text{S}_{\Sigma\text{H}_2\text{S fluid}}$) of the hydrothermal fluid can be approximated except for galena ([Ohmoto and Goldhaber, 1997](#)): $\delta^{34}\text{S}_{\Sigma\text{H}_2\text{S fluid}} \approx \delta^{34}\text{S}_{\text{sphalerite}} = +7.8$ to $+23.4\%$.

The isotopic composition of sulfides/hydrothermal fluid ($\delta^{34}\text{S}_{\Sigma\text{H}_2\text{S fluid}} \approx \delta^{34}\text{S}_{\text{sphalerite}}$) ranges from $+7.8$ to $+23.4\%$ with most values in the range of $+16.3$ to $+23.4\%$. This range in the sulfur isotopic composition falls within that of the Triassic gypsum in the Toussit-Bou Beker ore deposit ($\sim +13$ - $+17\%$, [Bouabdellah et al., 2015](#)). This sulfur isotopic composition range indicates that the sulfur is likely derived from the Triassic-Jurassic seawater sulfates and/or coeval sulfates/evaporites (Fig. 12). This Triassic-Jurassic sulfur reservoir was also proposed for the Bou Dahar Zn-Pb-Ba-(Sr) ore deposits in the oriental High Atlas ([Rddad and Bouhlel, 2016](#)) and the Touissit-Bou Beker Zn-Pb ore deposits in the Moroccan-Algerian confines ([Bouabdella et al., 2012](#)). The observed decreasing trend $\delta^{34}\text{S}_{\text{galena}} < \delta^{34}\text{S}_{\text{pyrite}} < \delta^{34}\text{S}_{\text{sphalerite}}$ suggests

isotopically overall homogenous H₂S in the ore-forming fluid (Ohmoto, 1986). This conclusion is consistent with the Triassic-Jurassic seawaters and coeval evaporites as the main source of sulfur.

There are two main pathways through which sulfates can be reduced to sulfur: abiotic thermochemical sulfate reduction (TSR) and bacterial sulfate reduction (BSR) processes. These two processes take place at different temperature conditions. BSR occurs at low temperature, usually below 80 °C (e.g. Machel, 1987). In contrast, TSR occurs in the presence of e.g. hydrocarbons at higher temperature conditions (80-100 °C < T < 150-200 °C), often in the range of 80 to 130 °C (Orr, 1974; Krouse et al., 1988; Machel, 1987). In this study, the high temperatures and the heavy isotopic composition of sulfides point to TSR as the main sulfur-producing process. The latter is known to be an effective mechanism that rapidly produces a large amount of reduced sulfur (Ohmoto et al., 1990). The TSR mechanism also explains the heavy isotopic composition (up to 23.4 ‰) recorded in some sulfides.

7. 3. Source of the metals

The Pb isotopic composition of the sulfides in the sediment-hosted, base metal deposits is used to constrain the source(s) of lead and presumably other metals. The brines can leach metals from the reservoirs they passed through. Lead isotope values for galena samples of the AOD ore deposit form a very tight cluster (²⁰⁶Pb/²⁰⁴Pb: 18.358 to 18.376; ²⁰⁷Pb/²⁰⁴Pb: 15.636 to 15.663 and ²⁰⁸Pb/²⁰⁴Pb: 38.486 to 38.606). The plots of the galena Pb isotopic composition in the uranogenic and thorogenic diagrams put the Pb isotope values above the orogen Pb evolution curve and between the upper crust and orogeny curves of Zartman and Doe (1981) (Fig. 12). The Pb isotopic data and specifically the higher values of ²⁰⁷Pb/²⁰⁴Pb suggest that lead and likely other metals originated from the upper continental crustal reservoir (Stacey and Kramers, 1975; Doe and Zartman, 1979). This is consistent with the geodynamic setting of the central High Atlas.

The homogenous Pb isotopic composition of galena samples implies that either (i) the source of Pb is an overall homogenous reservoir, and/or (ii) the lead was derived from different reservoirs and subsequently mixed and thoroughly homogenized. Given the Late Middle Jurassic to Early Cretaceous age for the ore emplacement and the fact that the ore-forming fluids were derived from the deep-seated Paleozoic basement, rock formations post-dating the Early Cretaceous are not a potential metal source. This narrows the source of metals to the Paleozoic rocks and the Triassic-Early Cretaceous rocks. The presence of deep-seated E-W- to NE-SW-trending Hercynian faults, connecting the Paleozoic basement with the Mesozoic cover (Laville, 1985) and the Pb radiogenic signature of galena samples render the Paleozoic rocks the main source of metals. These regional faults were reactivated during the Jurassic-Early Cretaceous extensional tectonic activity. The Pb isotopic composition of the Paleozoic-derived fluids flowing through the Mesozoic cover over long distances, via these faults, was homogenized before reaching the site of deposition. This

mechanism of Pb homogenization was proposed for many MVT ore deposits (Vaasjoki and Gulson, 1986). The Paleozoic basement was also proposed as the main source of lead and presumably other metals for the Upper Moulouya ore district on the basis of Pb isotopes (Jébrak et al., 1988) and Sr and rare earth elements and yttrium (Margoum et al., 2015). In fact, the Upper Moulouya ore district shares a similar Paleozoic basement as the central High Atlas. The Paleozoic rocks were also proposed as the main metal source for several other Moroccan ore deposits (Tighza -Aouam ore district, central Meseta, Marcoux et al., 2015; Rossi et al., 2017; Touissit-Bou Beker, Bouabdellah et al., 2015; Bou Dahar, oriental High Atlas, Rddad and Bouhlel, 2016; Amensif ore deposit, Western High Atlas, Ilmen et al., 2016). The Paleozoic basement was also proposed as a source of base metals for several North African ore deposits (e.g. Tunisian-Algerian diapiric zone, Bouhlel et al., 2005, Jemmali et al., 2011; Bouhlel et al., 2016; Laouar et al., 2016). Worldwide, the compiled lead isotopic composition of several MVT ore deposits (Leach et al., 2005) suggests a strong relationship between the isotopic composition of the ore in a district and that of the corresponding basement.

It is interesting to note that the ore becomes more radiogenic towards the west of the High Atlas. The Jurassic carbonate-hosted Pb-Zn deposit of Tadaghaste is located west of the High Atlas at the proximity of the Precambrian/Paleozoic basement locally known as “le massif ancien”. The galena of this ore deposit has relatively radiogenic Pb isotopic composition ($^{206}\text{Pb}/^{204}\text{Pb}= 18.709$ $^{207}\text{Pb}/^{204}\text{Pb}= 15.658$ and $^{208}\text{Pb}/^{204}\text{Pb} = 38.808$) (Rddad and Mouguina, 2018) compared to those of the AOD and the Bou Dahar ore deposits. This could indicate a contribution of the Precambrian rocks as a metal source for the ores in the central High Atlas ore district. The Precambrian basement is also proposed as a source of lead for the Upper Moulouya ore district (Jébrak et al., 1988).

7.4. Timing and migration of ore-forming fluids.

The ore is hosted in Middle-Upper Bajocian carbonates and is clearly epigenetic. Hence, the ore was emplaced after the lithification of Bajocian carbonates and, therefore, must be post-Bajocian. On the basis of stratigraphic data, Charrière et al. (2009; 2011) concluded that the Jurassic-Early Cretaceous igneous rocks along with the associated mineralization were emplaced before the Eocene. This conclusion places the age of the AOD ore emplacement prior to the major Alpine compressional tectonic events. Moreover, the high temperatures recorded in sphalerite- and calcite-hosted fluid inclusions (respectively up to 170 °C and 206 °C) reflect abnormally high geothermal gradient. The latter can be explained by one the three magmatic events that occurred during the High Atlas basin evolution during the Trias, the Jurassic-Early Cretaceous and the Eocene. The former magmatic event can be excluded owing to the fact that the host rock is Middle Jurassic. The last magmatic event, consisting of the alkaline carbonatite complexes of Tamazert, occurred locally along the northern part of the central High Atlas and is mainly associated with

fluorite mineralization, rather than Zn-Pb mineralization typical of the AOD ore deposit. Furthermore, the mineralized NNE-SSW-trending fractures were formed during Late Bajocian-Early Bathonian. Therefore, the emplacement of the ore post-dates the Early Bathonian. The aforementioned discussion led us to constrain the age of the ore emplacement to the Late Middle Jurassic-Early Cretaceous period during which an abnormal thermal anomaly was present in relation to mantellic doming. The latter induced crustal extension which triggered Jurassic-Early Cretaceous magmatism and the circulation of the ore-forming fluids. In this respect, the high geothermal gradient was caused by several pulses of magma, which started during the Early Jurassic (Laville and Harmand, 1982; Bougadir, 1988; Zayane, 2001) and continued until the Early Cretaceous (Laville, 1985). The proposed Late Middle Jurassic-Early Cretaceous age falls within that proposed by Mouguina (2004).

The magmatic upwelling that started during the Early Jurassic along with the Jurassic diapirism of the Triassic evaporites (e.g. Saura et al., 2014) led to the development of the central High Atlas with depocenters separated by paleohighs that are occupied by Jurassic intrusions and Triassic evaporites. This morphology along with the Late Middle Jurassic-Late Jurassic extensional event (Frizon de Lamotte et al., 2009) facilitated the migration of the fluids from deep parts of the basin to the paleohighs. The Jurassic-Early Cretaceous magmatism formed a potential heat engine that provided a long-lived heat source which facilitated the circulation of the hydrothermal fluids (Bougadir, 1988; Berraouz, 1995; Zayane, 2001). Moreover, the reactivation of the E-W and ENE-WSW- striking faults rooted in the Paleozoic basement have also contributed in triggering large-scale circulation of the deep-seated fluids. At the scale of the AOD deposit, the magmatic intrusion of Ikkou Ou Ali, which is about 4 km away from the ore deposit, may have established a small-scale fluid flow.

All the aforementioned fluid-flow driving mechanisms could have acted together to induce the circulation and upward migration of a large amount of a deep-seated fluid towards paleohighs through regional E-W to ENE-WSW-trending faults. Locally, the NE-SW-trending faults (F₁ and F₂) along with the accompanying veins and fractures, encountered in the mine, were likely local conduits for the flow of the fluids involved in the ore formation. The proposed driving mechanisms, involved in the circulation of a large amount of deep-seated ore-forming fluids in an extensional tectonic setting, are also suggested for numerous ore deposits worldwide. Among them is the Upper Moulouya ore district (Aouli-Zeïda ore deposits) located in central Morocco. For this ore district, Jébrak et al. (1988) and later Margoum et al. (2015) asserted that the ore was emplaced during the early stages of the Permian-Triassic Pangea rifting. Heijlen et al. (2003) and Muchez et al. (2005) proposed expulsion and subsequent ascent of deep-circulating fluids in the basement during the early Cretaceous extensional activity for a major basin-hosted deposit in Europe. Boni et al. (2002) also proposed that a Triassic-Jurassic rifting-related hydrothermal event in relation to Tethys spreading is one of the ore-forming events that formed the ore deposits in Sardinia. The

Zn-Pb-(Ag) Jabali ore deposit in Yemen is another ore deposit which formed in an extensional setting in relation to the Late Jurassic-Early Cretaceous rifting (Ostendorf et al., 2015).

7.5. Ore genesis model

On the basis of the geodynamic setting, the fluid type (basinal brine) involved in ore formation, the source of sulfur (Triassic-Jurassic evaporites and coeval seawater), the source of metals (Paleozoic basement), an ore genesis model for AOD ore deposit is proposed (Fig. 14).

As early as the Liassic, salt halokinesis caused the Triassic evaporites to migrate from the active subsiding area to the paleohigh of Ikkou Ou Ali. Owing to the thick Jurassic sedimentary pile, this salt movement was accentuated during Jurassic time. During the Early Jurassic, magma intruded the paleohigh of Ikkou Ou Ali. The emplacement of the Triassic evaporites and the Jurassic magmatic intrusion along the Ikkou Ou Ali led to the formation of Ikkou Ou Ali paleohigh. This local uplifting created a shallow marine environment (subtidal-intertidal) on which a Bajocian carbonate platform developed (Mouguina, 2004). This halokinesis- and magmatism-induced uplift ultimately led to the emersion of the carbonate platform and the subsequent formation of unconformities. The latter were sealed with dark limestones and later with a marly Bajocian formation. Within this tectonic-lithostratigraphic setting, the ore was emplaced during the Late Jurassic-Early Cretaceous extensional tectonic event. The reactivation of the deep-seated faults and the high heat flow caused the expulsion of a substantial amount of the Paleozoic basement-derived fluids along deeply penetrating faults. Numerous authors highlighted the importance of these fluid flow driving forces in the circulation of large amounts of the fluids and their subsequent ascent to the loci of deposition (e.g. Wilkinson et al., 2003; Heijlen et al., 2003; Muchez et al., 2005; Bouabdellah et al., 2015; Margoum et al., 2015).

In the current study, ascending fluids evolved into Zn-Pb -rich brines (F_{ore}) through interaction with rocks through which they flowed. On the basis of Pb isotopes, the metals were leached and extracted mainly from the Paleozoic basement. The ore-forming fluids migrated upwards into the Jurassic via E-W- and ENE-WEW-trending faults and locally via NE-SW-trending faults (e.g. F_1 and F_2). Considering the presence of a thick impermeable sequence of marls overlaying the host rocks, the metal-bearing fluid was focalized within the Bajocian host carbonate rocks where it mixed with a warm, H_2S -rich fluid (F_{sub}). The reduced sulfur was derived from TRS of the Triassic-Jurassic seawaters and/or coeval evaporites. This fluid-mixing is reflected by fluid inclusions which recorded fluids with contrasting salinities. The metals, carried by the ascending fluids, combined with the reduced sulfur, leading to the precipitation of the ore. Such mixing mechanism is known to be the main cause of ore formation in sediment-hosted Pb-Zn ore deposits worldwide (Anderson, 1973; Beales and Jackson, 1966; Anderson and Macqueen, 1988; Sverjensky, 1986; Plumlee et al., 1994).

8. Conclusion

Geological and mineralogical observations, fluid inclusions data, and C-O-S-Pb isotopic data of the Ali Ou Daoud Zn-Pb ore deposit hosted in Bajocian carbonates led to the formulation of a metallogenic model. The sources of metals and sulfur, the mode of fluid migration and ore precipitation, and the age of ore emplacement are constrained. These constraints are:

- 1) Deposition of the ore occurred during the Late Middle Jurassic-Early Cretaceous extensional tectonic event in relation to thermal doming. Deep-seated fluids migrated from the deep parts of the basin towards the Jurassic carbonates where ore precipitation took place.
- 2) Sulfur in the metal sulfides ($\delta_{34}\text{S} = 7.8$ to 23.4‰) was mainly derived from the Triassic-Jurassic seawaters and/or coeval evaporites through TSR.
- 3) The uniformity of the Pb isotope compositions of galena ($^{206}\text{Pb}/^{204}\text{Pb}$: 18.358 to 18.376; $^{207}\text{Pb}/^{204}\text{Pb}$: 15.663 to 15.486 and $^{208}\text{Pb}/^{204}\text{Pb}$: 38.486 to 38.606) reflects the derivation of metals mainly from the Paleozoic rocks.
- 4) At the site of deposition, the ascending hydrothermal metal-rich, sulfur-poor fluid encountered and mixed with a warm, metal-poor, and sulfur-rich basinal fluid. This fluid-mixing process caused the precipitation of the sulfides within karstic cavities and along faults and fractures.

The interplay of magmatism, halokinesis, and tectonism which occurred during the Jurassic to early Cretaceous distinguishes the small-size AOD Zn-Pb ore deposit from other MVT ore deposits in North Africa and in Western Europe. This trilogy is determinant in the emplacement of the AOD ore deposit and likely several other ore deposits in the central High Atlas. This paper shows that besides regional factors (e.g. regional extension), local factors can be important in the emplacement of ore in specific ore districts. In that regard, diapirism, magmatism, and local tectonic are key factors in the ore genesis of the AOD ore deposit. With regard to ore exploration in central High Atlas and in North Africa, ridges intruded with igneous rocks and Triassic evaporitic rocks can be potential targets for base metals exploration.

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634 **Figures and Tables**
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Fig. 1

Age	Log	Lithology	Formation
Middle Cretaceous		Marl and limestone beds	Tasraft Formation
Callovian		Marls with igneous intrusion	Timzought Formation
Bathonian		Marls and limestone	Isli Formation
Bajocian		Alternating siltstone and marl beds	Imilchil Formation
		Alternating marl and limestone layers	Tislit Formation
		Massive limestone	Bab-n-Ouayad Formation
Aalenian		Alternating marls and limestone layers	Tassent Formation
Toarcian		Alternating marls and sandy limestone layers	

Fig. 2

Fig. 3

Fig. 4

Fig. 5

Fig. 6

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Fig. 8

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Fig. 11

Fig. 12

Fig. 13

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Fig. 14

Table 1

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Mineral	<u>Hydrothermal mineralizing event</u>			
	Pre-ore Stage (S ₀)	Ore-related Dolomitization stage (S ₂)	Sulfide stage (S ₃)	Oxidized stage (S ₄)
Dolomite (D ₀)	-----			
Dolomite (D ₁)		-----		
Pyrite			-----	
Sphalerite			-----	
Galena			-----	
Calcite				-----
Calamine				-----
Cerussite				-----
Covellite				-----
Goethite				-----
Hematite				-----

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Table 2

Sample	Host mineral	Num.	T _{m, ice} (C)		Salinity (Wt.% NaCl eq.)		Th (°C)	
			Range	Mean	Range	Mean	Range	Mean
ALIDA2	Sphalerite	5	-9.9 to -15.9	-12.9	13.8 – 19.4	16.6	101-144	123
ALIDA8	Sphalerite	12	-20.8 to -24.0	-22.4	23.2 - 25.4	24.3	111-144	128
ALIDA10	Sphalerite	2	-23.3 to -24.3	-23.8	24.8 – 25.6	25.2	140-145	143
DEMaA(r)*	Sphalerite	20	-19 to -26.0	-22.5	23.1-27.3	25.2	94-292	193
DEMaA(j)*	Sphalerite	6	-23.2 to -23.9	-23.6	24.8-25.7	25.3	146-290	218
DEMaA(c1)*	Sphalerite	5	-23.3 to -24.0	-23.7	24.8- 25.4	25.1	110-151	131
ALIDA3	Calcite	10	-13.8 to -20.4	-19.1	17.7 - 25.6	21.7	177-206	192

Table 3

Sample	Rock/Mineral	δ ¹³ C (VPDB, ‰)	δ ¹⁸ O (VPDB, ‰)	δ ¹⁸ O (VSMOW, ‰)
ALIDA1A1h1	Limestone	1.3	-6.5	24.2
ALIDA1A1h2	Limestone	0.7	-7.0	23.7
ALIDA1a1d1	Dolostone	1.9	-5.7	25.1
ALIDA1a1d2	Dolostone	1.7	-5.7	25.1
ALIDA2h	Dolostone	-2.5	-6.0	24.7
ALIDA3h	Dolostone	2.1	-6.3	24.5
ALIDA2c	Calcite	1.4	-5.2	25.6

ALIDA3c	Calcite	1.5	-7.5	23.2
ALIDA9c1	Calcite	-9.0	-3.6	27.2
ALIDA9c2	Calcite	-8.9	-3.7	27.1

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Samples	$\delta_{34}\text{S}$ ‰, (VCDT)	Δ Sph-Gn ‰	T °C
ALIDA1 Ga	18.8		
ALIDA2 Sp	7.8		
ALIDA3 Ga	17.9		
ALIDA4 Py	20.6		
ALIDA5 Ga	16.3		
ALIDA6 Ga	16.5		
ALIDA6 Sp	20.9	4.4	135
ALIDA7 Ga	17.8		

ALIDA7 Sp	23.4	5.6	90
ALIDA8 Ga	18.6		
ALIDA8 Sp	22.0	3.4	195
ALIDA9 Ga	16.5		
ALIDA10 Ga	17.0		
ALIDA10 Sp	21.0	4.0	155
Deb-Ga	18.3		
Deb-Sp	20.8	2.5	270

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Table 5

Sample	Pb ₂₀₆ /Pb ₂₀₄	Pb ₂₀₇ /Pb ₂₀₄	Pb ₂₀₈ /Pb ₂₀₄
ALIDA1	18.365	15.658	38.588
ALIDA2	18.364	15.661	38.606
ALIDA3	18.364	15.657	38.541
ALIDA5	18.369	15.650	38.486
ALIDA6	18.376	15.663	38.578
ALIDA7	18.364	15.656	38.552
ALIDA8	18.358	15.636	38.504
ALIDA9	18.359	15.661	38.554
ALIDA10	18.361	15.639	38.559

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