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Lithium enrichment processes in sedimentary formation waters

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Abstract

Reservoirs exploited for oil, gas, or geothermal energy in sedimentary basins may represent an alternative lithium resource to salars and hard rocks. Similarities in the geochemical characteristics of the basins were searched using a database of 2,400 water analyses available from more than 40 sedimentary basins worldwide, notably a selection of representative geochemical trends on 13 basins. The objective was to understand better the factors controlling Li concentrations and the potentiality of such basins for the Li-recovery. Considering Li-Cl and Mg-Cl relationships, major cation (Na/K, Na/Ca) and halogen (Cl/Br) ratios, Li concentrations are monitored in several basins by mixing processes between low salinity recharge waters, seawater and brines. Lithium concentrations span three orders of magnitude in brines between evaporated seawater that has passed saturation with halite and a particularly Li-enriched pole. In the latter, significant Li-enrichment is obtained when brines are primary, in equilibrium with halite, and then enriched in Li through fluid-rock interaction processes. Li-rich lithologies such as volcanic rocks, Li-enriched clay formations, or Li-phyllsilicates dispersed in siliciclastic rocks or the underlying crystalline basement probably act as the primary Li sources. Finally, Li extraction is promoted by the rise in temperature, which explains the high Li concentrations in the waters of

geothermal systems. Therefore, favourable reservoirs for Li-rich brines are characterised by a brine collection from evaporites during tectonic events, such as compressive events or salt tectonics, at a relatively short distance from the present reservoir. They must be preserved from convection and surface water ingress via recharge zones to prevent dilution and Li depletion.

1. Introduction

Lithium is an economically important element in transitioning to low-carbon energy, especially for Li-ion rechargeable batteries (Bibienne et al., 2020, Nate et al., 2021). Most primary lithium resources are currently exploited in pegmatites (Greenbushes, Australia, Yaksic and Tilton, 2009; Mohr et al., 2010; Labbe and Daw, 2012), closed-basin brines (salar), primarily in South America (Salar de Atacama, Chile; Salar de Hombre Muerto and Puna, Argentina; Salar de Uyuni, Bolivia (Lopez Steinmetz et al., 2018, 2020; Lopez Steinmetz and Salvi, 2021) and in Lake Qaidam, China; Ericken and Salas, 1987; Bradley et al., 2017, USGS, 2021). Li-enrichment in closed surficial basins is well-known and has been reviewed by Munk et al., (2016).

Groundwater in sedimentary basins, particularly oilfield brines and saline formation waters, can also contain significant Li concentrations, reaching values above 100 mg.l⁻¹ (e.g. Smackover Formation (Gulf Coast, USA; Collins, 1975). These metal-rich sedimentary formation waters are considered potential "liquid ores" (Alexeev et al., 2020) and represent a good Li resource that can contribute significantly to meeting growing global demand (Dugamin et al., 2021).

The bulk chemical evolution of formation waters in sedimentary basins, such as the role of water-rock interactions (diagenetic reactions), the location of enhanced flow and the timing of fluid migration, have been studied for decades (e.g. White, 1965, Carpenter et al., 1974; Carpenter, 1978; Collins, 1975; Land and Prezbondowski, 1981; Hanor, 1987; Kharaka et al., 1987; McCaffrey et al., 1987; Morton and Land, 1987; Moldovanyi and Walter, 1992; Land and Macpherson, 1992; Stueber et al., 1993; Land, 1995; Davisson and Criss, 1996; Kharaka and Hanor, 2003; Lowenstein et al., 2003; Boiron et al., 2010). However, Li behaviour is much less documented in these deep reservoirs than in closed surficial basins. The sources and mechanisms of Li enrichment remain debated and poorly understood. Fontes and Matray (1993) and, more recently, Coffey et al. (2021) have provided hypotheses on the origin of Li excess in sedimentary basin formation waters: (1) Li-enrichment through diagenetic reactions, including water uptake and Li gains from the host sediments; (2) increased Li in the water source, for instance, paleo- seawaters which may be significantly different from modern seawater and (3) Li enrichment

through evaporation, the evaporated waters being trapped in evaporites or sediments. Understanding such processes is of great interest for the current prospection of reservoirs such as those from the Rhine graben in France and Germany (Sanjuan et al., 2010, 2022) and for recycling water from oil fields (Kumar et al., 2019).

This paper discusses the distribution of Li concentrations in waters from sedimentary basins and the processes that modify these concentrations (mixing, dilution, nature of the different water end-members). The study is based on a compilation of chemical analyses of present-day formation waters from oil and gas reservoirs for which data are available but still not examined from the point of view of Li enrichment processes. Besides, an inventory of Li concentration ranges in rock-forming minerals and rocks has been set up to document the Li cycle. The ultimate objective was to define the standard features of fluids highly enriched in Li and the main processes controlling such enrichment in sedimentary reservoirs. Finally, we examined if such fluids share similar characteristics with brines from salars.

2. Data compilation and processing

All geochemical and physical data for each sedimentary basin used in this study are derived from a compilation of literature data. A database of water analyses from 40 sedimentary basins worldwide, representing approximately 2,400 individual analyses, was first compiled to analyse the factors controlling Li concentrations (Dugamin 2022, Dugamin et al., submitted). All the selected sedimentary waters correspond to reservoir waters (i.e. "formation waters"). Similarities in the geochemical features of Li-rich brines were defined by comparing the relationships between Li concentrations, major cation (Na, K, Ca, Mg) and halogens (Cl, Br) ratios. These observations let us choose 13 out of the 40 basins, which waters share similar features (Fig. 1), and represent the main geochemical trends worldwide. Table 1 provides the location of the selected basins and the chemical composition of their formation waters (Li, Cl, Mg concentrations, Cl/Br, Na/Ca and Na/K ratios).

Data quality and representativeness from the published sources have been scrutinized. The solute concentration data are reported in various units (e.g. ppm, mg.l-1, mmol.l-1, mol.kg-1, meq.l-1). All data were converted to mg.l-1 before being added to the database. The density used for some conversions was either provided in the published sources or set at 1 g.cm-3 (density of pure water at 25°C at 1 atmosphere) by default. The salinity, taken as total dissolved solids (TDS), was calculated as the sum of the available concentration of major ions (Na^+ , K^+ , Ca^{2+} , Mg^{2+} , Cl^- , SO_4^{2-} , NO_3^- , HCO_3^-) compiled

from published sources and expressed in mg.l-1. Cl, Na, Ca, K and Mg concentrations in formation waters are reported in more than 90% of published sources, but other anion concentration data are not always available. The database did not include the solute concentrations for suspected contamination in the published sources. The analysis of the electrical balance values and the limits of detection (in particular, the proportion of data above limits of detection) provide a first-order control on data quality.

The reservoirs considered in this compilation were subjected to oil and gas extraction and one for geothermal energy (Salton Trough), as presented in Tables 1 and 2. The stratigraphy data (age of sedimentary basins, description of the lithology, presence of salt beds and diapirs, presence of volcanic and crystalline rocks as potential source rocks) are presented in Table 2. Reservoirs are mostly siliciclastic formations, except those located on both sides of the Mexico Gulf (Gulf Coast, Salinas-Sureste) and Zagros, where limestones are reservoirs. The selected reservoirs are hosted by sedimentary formations covering an extensive range of ages from Late Proterozoic to Cenozoic. The basins share the presence of evaporite layers affected in several basins by diapirism and salt tectonics.

Salars from South America were considered for comparison. The same chemical parameters were thus compiled for 18 closed basins and 4 salt lake waters (Table 3).

In addition, to assess potential sources of Li, a compilation of rocks and minerals and their associated Li concentrations from published sources was carried out (Supplementary table A).

3; Geochemical data processing

The following sections compare the relationships between Li, Cl, halogen ratio (Cl/Br), and major cation (Na, K, Ca and Mg) ratios for the 13 selected sedimentary basins and salars.

3.1 Water types

The general characteristics of formation waters are shown in the Piper diagram (Fig. 2A), where they are compared to salar waters (Fig. 2B). The prominent feature of the studied waters is the predominance of Na and Cl as most data points plot at the Na + K and Cl + NO₃ end-members. Exceptions are: i) the Delaware waters characterised by a mixing trend between a sulfate Mg-(Ca) rich end-member and the Na-Cl end member, ii) two basins with bicarbonate waters show a Na to Ca-Mg trend typical of waters equilibrated in limestone formations (Zagros, North German for the Mesozoic cover), and iii) the Michigan basin waters with an extended trend from the Na-(Cl) corner to a Ca-Mg-Cl end member. Salar

waters as sedimentary basin waters plot close to the Na+K and Cl+NO₃ corners. Several trends towards a Ca + Mg end member are specific to some salars such as Atacama, Ascota, Loyoques or Alconcha. The four latter also show a more carbonated feature and a wider dispersion than sedimentary basins. A sulfate contribution is found in salt water lakes such as Minique, Miscanti and La Azufrera, as well as in the waters with a significant carbonate content above cited.

The Li-Cl binary plot discriminates four main types of fluid end-members based on Cl and Li concentrations (Fig. 3 and 4): (i) low salinity waters (dilute water noted DW) close to rainwaters or river waters indicated by light blue domains in figure 3, (ii) seawater (noted SW), (iii) evaporated seawater (noted ESW), and (iv) a brine with Cl concentrations close to those of halite saturation, but with exceptionally high Li concentrations (1 to 2 orders of magnitude higher than those of evaporated seawater saturated to halite (Li-rich brine, noted LiRB). Plotting all Li versus Cl data reveals four types of distributions between these four water end-members, as shown in figures 3 and 4:

i) a mixing trend between DW and SW (or ESW-halite). Such a trend characterises the Delaware Basin and the reservoirs above the Triassic series in the Northern German Basin (Fig. 3A and B).

ii) mixing between DW and LiRB, as in the case of the Zagros, Amadeus, and Salinas basins and the Salton Trough (Fig. 3 C).

iii) mixing of LiRB with SW. These mixtures are typical of the Illinois and Appalachian basins (Fig. 3 D).

iv) a trend between ESW and LiRB. Such a trend was found in the reservoir in the basement below the Triassic series in the North German basin and Michigan, Mississippi Salt, North Louisiana Salt, Nepa Botuoba and Qaidam (deep wells) basins (Fig. 3 E and F).

3.2 *Li-enrichment in water end-members*

In a plot of Li versus Cl (Fig. 3), formation waters cover more than six orders of magnitude in Li concentrations, ranging from a few mg.l⁻¹ to ca. 1,000 mg.l⁻¹, as well as in Cl content which ranges from a few mg.l⁻¹ to a few hundred thousand mg.l⁻¹. Lithium and chlorine are roughly correlated when considering all data available. However, the distribution of Li follows very distinct trends from one basin to another. The variations of Li concentrations evolve then between 10⁻³ and a few mg.l⁻¹ in waters with low Cl content and between 10 and 1,000 mg.l⁻¹ in waters with the highest chlorine content, i.e., close to that of halite (NaCl) saturation (from 25°C to 100°C) for evaporated seawater.

The lowest salinity waters distribute in two sub-poles as shown in figure 3B: i) the most common low salinity pole, characterised by chlorinities between a few mg and 10 mg.l⁻¹ close to the river and rain waters and a Li concentration ranging from 10⁻³ to 10⁻² mg.l⁻¹, a little above the concentration of rain and river waters, for instance in the Mesozoic aquifers from the North German basin, and Zagros and Amadeus basins (Fig. 3C). ii) a pole which is ten times more concentrated in Li than the previous one for the same chlorinity, such as in the Delaware basin with Li concentrations of 10⁻¹ mg.l⁻¹ for a chlorinity of around 10 mg.l⁻¹(Fig. 3A and B). Interestingly, the mixing trend DW-SW is located above the trend of seawater dilution by pure water. Therefore, the most representative low salinity waters have a significant Li enrichment factor of around 10 (100 in the case of the Delaware pole), compared to highly diluted seawater and rainwater (Fig. 3 B). For some basins such as Illinois or Appalachian, waters show a predominant LiRB type brine and a few products of probable mixing between LiRB and less saline end-member closer to seawater (Fig. 3D). The central data cluster is considered in the following sections. Still, mixed waters were not deleted from the data sets.

For the LiRB cluster, the chlorinities are high and close to the values of saturated water to halite. The Li concentrations cover a significant range, with some data well above those of evaporated seawater, mostly between 10 and 800 mg.l⁻¹(Fig. 3E and F). Thus, on both sides of the mixing trends, Li concentrations distribute over one to two orders of magnitude (Fig. 3E and F). Therefore, a mechanism independent of the Cl concentration favours the Li-enrichment.

3.3 Brine origin: halogen and stable isotope data

The halogen data: During the formation of evaporites, seawater evaporation results in the so-called "primary" brine, which precipitate successively gypsum and then halides. Primary brines are stored as fluid inclusions in halides or within rock microporosity. Evaporites may dissolve when in contact with recharge waters or seawater, giving rise to secondary brines. To discriminate between brines, the classical use of halogen ratios (Cl/Br ratio) allows differentiation between primary and secondary brines (Banks and Yardley, 1992; Fontes and Matray, 1993). Primary brines have a Cl/Br ratio below that of seawater due to chlorine depletion in brines during halite crystallisation. In reverse, Cl/Br ratio is very high in secondary brines, as halite dissolution does not release bromine.

The Cl/Br ratio in most basins is lower than Cl/Br seawater ratio suggesting their primary origin (Fig. 5A to D). A few basins are exceptions as their waters have Cl/Br ratio higher than seawater, up to 850, and are typical of secondary brines: i) the Illinois Basin and Delaware (Fig. 5 A and C), and ii) the

upper part of the North German Basin (Mesozoic reservoirs above the evaporites, Fig. 5A). In the latter example, low-salinity water enters the reservoir and dissolves the underlying evaporites and diapirs.

The most Li-enriched waters (LiRB pole) have Cl/Br ratios that are lower than seawater and close to those of primary halite or epsomite saturated brines (Fig. 5D). The high Li concentrations in the North German Basin (Rotliegend and the Upper Carboniferous formation waters located in the centre of the basin below the Triassic salt formations) (3,000 times the value of seawater) correspond well from the point of view of Cl/Br ratios to primary brines saturated with halite (Lüders et al., 2010). The Cl/Br ratios in North Louisiana salt, Mississippi salt and Michigan are distributed in a sub-parallel trend to that of seawater evaporation, with Li concentrations multiplied by a factor of nearly 10. In North Louisiana and Appalachian, the Cl/Br spread down to low values of around 30. The only notable exception is the Qaidam basin, where high Cl/Br values of approximately 1,000 indicate that the brines are most likely secondary. In Salton Trough waters, brines are also secondary as their high chlorinity results predominantly from halite dissolution.

Stable isotope data: they are partially available on the studied basins and help to decipher the fluid origin (Craig and Gordon, 1965). When the diluting end-member DW dominates formation waters (recharge meteoric fluid) and is not affected by intense high-temperature fluid-rock interaction, they retain the negative $\delta^{18}\text{O}$ value of meteoric waters. Such an end-member is illustrated by the Delaware and North German (reservoir above Triassic formations) basin waters. In the latter, data points are distributed along the global meteoric water line (Craig, 1961) between -10 and -5‰ ($\delta^{18}\text{O}$ (V-SMOW)) and -75 and -45‰ (δD (V-SMOW)) (Fig. 6).

Evaporated seawater typically shows $\delta^{18}\text{O}$ and δD values ranging between seawater value (i.e. 0‰) and +10‰ at the gypsum saturation stage and +4 and -14‰, respectively, at halite precipitation stages (Knauth and Beeunas, 1986) (Fig. 6). Most considered LiRB basinal brines plot close to the seawater evaporation trend between -5 and +8‰ for $\delta^{18}\text{O}$ and between -25 and +7‰ for δD , with shifts in $\delta^{18}\text{O}$ and δD values specific to each basin. Illinois basinal brine values are between -4 and +2‰ for $\delta^{18}\text{O}$ and between -35 and -10‰ for δD . Data from Salinas- Sureste and Zagros basins show a distribution of $\delta^{18}\text{O}$ (between 0 and +13‰), and δD (between 5 and -35‰) mostly shifted from the seawater evaporation towards higher $\delta^{18}\text{O}$ values. In the Qaidam basin, the trend evolves between the meteoric line and higher $\delta^{18}\text{O}$ values (between -10 and +15‰) at much lower δD values (between -60/-45 and -26‰) than in the other basins. The atypical features of Qaidam waters have been already deduced from considering major cation relationships (see below).

The shift to positive $\delta^{18}\text{O}$ in most LiRB basinal brines accounts for high-temperature fluid-rock interactions (Williams and McKibben, 1989), confirming the effects of temperature-promoted water-rock interactions, which modify both isotopic and chemical features.

3.4 *Water-rock interactions: the cation data*

Relationships between Li concentrations and other cations are represented using Na/K and Na/Ca ratios, as these cations are sensitive to fluid-rock interactions in basins (Fig. 7 and 8). For LiRB brines, the Na/K ratio covers a wide range of values, from low values of 0.1 and around 200, all below the Na/K ratio of SW or ESW (Fig. 7D). Resulting trends specific to each basin, making complex the definition of the location end-member in terms of Na/K values. A relatively identical trend is found for the Mississippi and Michigan basins. The distribution in the North Louisiana basin is steeper and similar to that in the North German basin (lower part under salt). As in Illinois (Fig. 7C), the significant shift in Na/K values towards high values above 100 is due to Na release after halite dissolution and is symptomatic of secondary brines. In the Zagros basin (Fig. 7B), the Li-concentration range is vast, with lower values below those of seawater due to dilution at a sub-constant Na/K ratio. A similar situation is recorded in the upper part of the North German basin (Fig. 7A).

In most LiRB basinal brines, the Li concentration is anti-correlated to the Na/Ca ratio. The latter covers at least two orders of magnitude between 0.2 and 10, below the Na/Ca of seawater and evaporated seawater (Fig. 8D). For the Delaware and upper North German basins (Fig. 8A), the dilution trends between rainwater and seawater end-members cover a large Na/Ca range distinct from LiRB trends. In Zagros (Fig. 8B), the dilution trend starts from an end-member close to LiRB at sub-constant Na/Ca, which may indicate the buffer role of limestones on the Na/Ca ratio. Salinas-Sureste and Amadeus basin plot close to the other LiRB (Fig. 8A and B).

The behaviour of Mg is remarkable compared to Li, as positive correlations characterise many basins, with a slope of 1 similar to the dilution of evaporated seawater or, conversely, evaporation trends. In a given basin, the Mg/Li ratios span an extensive range of values but are constant (Fig. 9). The Mg/Li ratio is around 1,000 in the North German basin (Mesozoic reservoir) and in Delaware (Fig. 9A), 100 in Mississippi Salt, Michigan (Fig. 9D), and Appalachian (Fig. 9C), in between 10 and 100 in Zagros and Amadeus (Fig. 9B), 10 in Delaware and North German (Carboniferous-Permian reservoir) and close to 1 at Salton Trough. For basins in which the LiRB pole is dominant, Mg concentrations have much tighter ranges, mainly from 1,000 to 10,000 mg.l^{-1} .

3.5 *The Li-temperature relationships*

It is well known that during fluid-rock interaction, the chemistry of the fluid is controlled by a series of mineral (silicate)-fluid equilibria, which in turn are a function of temperature (Arnórsson et al. 1983). Cation ratios such as Na/K (Giggenbach, 1988, Fournier 1979) or Na-K corrected from the influence of divalent cations (Ca-Mg) were used as geothermometers, assuming pseudo-equilibria between waters and main silicates. The positive correlation between temperature and Li concentrations is, in addition, at the origin of several geothermometric relationships, including Li (Na/Li (Fouillac and Michard, 1981, Verma and Santoyo, 1997) and Mg/Li (Kharaka and Mariner, 1989). As such relationships are of interest for interpreting the distribution of Li concentrations, a Na/Li versus Na/K diagram is presented in figure 10 with a report of the thermometric relationships.

In brines from Salinas-Sureste and Zagros basins (Fig. 10 B), where measurements are in the range of 80-150°C, a relatively good agreement is found with temperatures from the Na/Li versus Na/K relationships. Conversely, most Na/Li -Na/K pairs for Li-rich formation waters (LiRB) plot outside the thermometric line (Fig. 10D). In the Michigan basin, the high K and Li concentrations move around the trend towards the left. In contrast, high Na concentrations in secondary brine from North Louisiana Salt displace the trend towards the right-hand side of the diagram. The disagreement between calculated and measured temperatures and the excess in Na, Li or K is probably indicative of a lack of equilibrium between brines and host rocks and of a specific enrichment different from those which depend strictly on temperature. The data are close to the expected values for the Salton Trough reservoirs, characterised by high temperatures ranging from 200 to 350°C.

3.6 *Comparison with salar brines*

Salar waters were plotted in the Li-Cl diagram (Fig. 11A). Data distribute between the Li-enriched end-member (LiRB type) and a dilute end-member, with almost the same trend as that recorded in the Zagros basin (Fig. 3C). Therefore, the evaporation/dilution trend for salars is nearly superimposed onto the LiRB and the diluted water mixing trend. Remarkably, the Uyuni brine evolution curve mimics the seawater evaporation curve but with a shift of about two orders of magnitude in Li concentrations. The salar brines in the Altiplano and Puna regions are the only ones that do not show mixing or evaporation trends but predominant LiRB brine features, particularly in deep horizons. In the Puna plateau, surficial

brines have Li values ranging from 57 to 5790 mg.l⁻¹ (López Steinmetz et al., 2020; Lopez Steinmetz and Salvi, 2021). The brines of the salars have Mg/Li ratios ranging from 1 to 100, specific to each basin (Fig. 11B). Most waters from salars have relatively low Na/Ca ratios. Salars in the Altiplano (Uyuni basin) and Puna (Olaroz, Cauchari and Salinas Grandes basins) regions have higher Na/Ca ratios (Fig. 11 C). In the Li versus Na/K ratio (Fig. 11D), the two most enriched salar waters distribute along a similar trend to that obtained for LiRB from sedimentary basins, characterised by the decreasing Na/Ca ratio at decreasing Li content. Similarly, the trend of decreasing Li concentration at sub-constant Na/K corresponds most probably to mixing effects. Finally, the Cl/Br ratio (Fig. 11E) is relatively high in salar brines, similar to those from the Qaidam basin and the Salton Trough and falls within the secondary brine field.

4. Discussion

In some of the considered basins, Li concentrations are correlated with fluid salinity because waters result from the mixing between a dilute end-member, which includes meteoric waters or their evolved products with brines. Almost all mixing rates are encountered in the Delaware, Salinas-Sureste, Zagros, and Amadeus basins. The high-Cl end-members, however, cover a wide range of Li concentrations distributed between the SW-ESW (halite) line and the LiRB, i.e. over more than two orders of magnitude above.

Most LiRB brines have chlorinities close to halite saturation and Li concentrations between 80 and 2,000 ppm, i.e. above Li concentrations produced by extreme evaporation of seawater (up to bischoffite or saturation). Thus, the Li concentration only increases from 0.18 to 5 mg.l⁻¹ during evaporation, i.e. an enrichment factor close to 10 up to halite saturation stage and from 0.18 to 25 mg.l⁻¹ during evaporation, i.e. an enrichment factor close to 100 up to bischoffite saturation stage (Fontes and Matray, 1993). Therefore, an additional enrichment factor is required to reach concentrations 10 to 100 times that of evaporated seawater. The high Li concentrations cannot result from seawater evaporation alone. Therefore, looking for a process other than evaporation is necessary to explain this Li enrichment.

4.1. *The shared features of the basins dominated by Li-rich brines*

The main characteristics of Li-rich brines are: i) they are evaporitic primary brines; ii) they are still in equilibrium with halite as their Cl concentration is close to halite saturation at 25°C or above.

Salt tectonics is one of the main factors responsible for the expulsion of primary brines, particularly the leakage of fluid inclusions into the salt and the migration of fluid inclusions under the effect of the thermal gradient as the expulsion of fluids from micropores. For example, such a process occurred twice in the Pyrenees: first, during Cretaceous rifting, characterised by evaporite stretching under anomalous heat fluxes associated with mantle uplift, and second, during thrusting induced by compression. During these two events, primary brines were expelled from Triassic evaporites, as shown by paleofluids trapped as fluid inclusions (Cathelineau et al., 2021).

The presence of massive evaporite layers may be one of the most favourable factors in obtaining large volumes of primary brines for most of the basins examined. Diapirs are, in some cases, very abundant such as in the Gulf Coast (Mississippi and North Louisiana salt formations and Salinas -Sureste basin) but are also present in the Siberian platform (Nepa-Botuoba basin) or the North German basin. In Gulf Coast, compressive deformation and diapirism cause the salt of the Louann Formation to move for kilometres (up to ten kilometres), intersecting all sedimentary formations, as shown in seismic sections (Hudec et al., 2013 a and b). In the Zagros, the Gasharan formation is partly the result of the dissolution of the thick Hormuz salt formation, and salt tectonics leads to the sealing of the reservoirs (Liu et al., 2018). The amplitude of the movements recorded by the salt formations is significant and explains the widespread expulsion of primary brines into these basins. Brines, generated along deformation zones, then invade the proximal reservoirs.

In addition, brines must be preserved from dilution. This condition is fulfilled when the convection is limited, and the renewal of allochthonous dilute fluids is minimised. Under such conditions, the primary brines may be predominant. Therefore, the presence of primary brines unaffected by dilution constitutes the first positive factor. The selected Li-rich brine basins share similar lithologies (Table 2), particularly the intercalation in siliciclastic and carbonate formations of salt-bearing horizons, sometimes affected by diapirism (salt domes). The favourable factors, compared to other basins hosting evaporitic sequences, thus are: i) a process yielding to brine expulsion (compression, tectonics), and ii) the lack of significant dilute fluid inputs in the reservoirs found in the relatively stable platform or cratons.

4.2. Temperature-dependent processes

The temperature range of the processes that governed the Li enrichment, in the past or more recently, may be considered the third factor of Li enrichment. It is well known that Li concentration

increases with increasing depth and temperature (Chan and Kastner, 2000; James et al., 2003; Coffey et al., 2021; Dugamin et al., 2021). The relationships between Li and other cations (Na, K, Mg) broadly follow the general trends predicted by empirical geothermometers, implying that temperature is essential in controlling Li concentrations. However, the discrepancies observed between reliable thermometric data (in situ measurements and silica in solution) confirm the importance of additional sources of Li, such as the presence of evaporated seawater and the contribution of Li from enriched surrounding rocks.

4.3 *The Li source factor*

Li- enriched source rocks containing leachable source minerals represent a necessary condition for high Li concentrations above the SW evaporation trend. The Li concentration in formation waters could result from significant water-rock interaction with the host rock, either within the immediate sedimentary environments, notably siliciclastic reservoirs or the basement lithologies, as the basin fluids may enter and then exit the crystalline basement.

The Na/K ratio is sensitive to cation exchange and changes when Na silicates (plagioclase, Na amphibole in the basement) or K silicates (K feldspar, muscovite, illite, biotite) are dissolved. During fluid-rock interactions such as plagioclase albitisation (anorthite end-member loss), cation exchange, or calcite dissolution results in a significant Ca release which modifies the Na/Ca mass ratio. The two cation ratios indicate that LiRB waters underwent critical water-rock processes. The primary brines have a Na/K ratio close to SW or ESW between 10 and 100, and the Na/Ca between 0.1 and 10 is lower than SW due to large Ca gains. Water-rock interactions monitor both ratios. Therefore, a process that enhances a long-lived interaction of the primary brines with minerals acting as Li-source during brine migration and storage is required to obtain a substantial Li enrichment.

Several contexts may allow the release of Li from source minerals and its temporary incorporation into newly-formed phases such as clays or its direct release into solution (e.g. Seyfried et al., 1984; Chan et al., 1993; Zhang et al., 1998; Chan and Kastner, 2000; James and Palmer, 2000; Williams and Hervig, 2005; Scholz et al., 2010; Hoyer et al., 2015). Several sources may be considered (see Figure 12 and Table A in supplementary material) :

Differentiated magmatic series: Figure 12 shows that Li is increasingly abundant due to magmatic differentiation from a few ppm in the ultrabasic series to 10 to 75 ppm in granites. Among the plutonic rocks, differentiated peraluminous granites can contain several hundred ppm of Li, primarily in the micas

(200 to 2500 ppm) (Bea et al., 1994; Dostal and Chatterjee, 1995; Raimbault et al., 1995; Förster et al., 1999; Müller et al., 2006; Sokolova et al., 2011; Canosa et al., 2012; Chicharro et al., 2016; Wu et al., 2017, Monnier et al., 2022). Extreme enrichments are found in highly differentiated aplites (Raimbault et al., 1995; Frindt et al., 2004; Moreno et al., 2014) and pegmatites (Teng et al., 2006; Liu et al., 2010; Canosa et al., 2012; Maneta and Baker, 2019) with the highest Li contents in Li-rich micas (polyolithionite, lepidolite) up to 2-3 % Li, petalite, spodumene and Li-phosphates (amblygonite series). Highly differentiated surficial volcanic emissions are the second series of Li-rich magmatic rocks. Ignimbrites, obsidian glass, including atmospheric emissions (tuff, glassy tephra), display an extensive range of Li-concentrations from 10 and 560 ppm (Breiter, 2012; Benson et al., 2017; Castor and Henry, 2020).

Sediments: Supergene degradation of basement lithologies provides the detrital elements of siliciclastic formations in basins (sands and conglomerates), which may be enriched in Li, notably muscovites and biotites from two mica granites. Pre-existing Mg-phyllosilicates, such as phlogopite, can lead to the crystallisation of F- and Li-rich hectorites.

Lake deposits under a climate with high hygrometric variations allowing evaporation and the formation of brackish water up to alkaline brines is the second surface environment particularly suitable for crystallising Li-rich trioctahedral smectites. Although fibrous clays are not as common as smectites and kaolinites, significant amounts of Li (20 - 90 ppm Li in palygorskite and 370 - 1,000 ppm Li in sepiolites) are reported by Tardy et al. (1972). Sepiolites are formed in lagoon environments, often saline during dry periods, such as the large Spanish basins (Madrid basin). The Vilcavaro sepiolites contain between 100 and 800 ppm Li (Torrez-Ruiz et al., 1994; Cuevas et al., 2003).

Marine carbonate environments: They have the lowest Li concentrations, with median values between 1 and 50 ppm (Hoefs and Sywall, 1997; Chan et al., 1997; Huh et al., 1998; Chan and Kastner, 2000; Chan et al., 2006).

Li-clays produced by volcanic glass alteration: Volcanic aerial products may form specific layers during eruptions intercalated within sedimentary series. Alteration of volcanic tuffs by hydrothermal solutions leads to mixtures of montmorillonite and hectorite (Starkey and Mountjoy, 1973). The glass matrix hydrates when in contact with low salinity waters, releasing Li, which can be incorporated in clays (Friedman and Smith, 1958; Ellis et al., 2022). Some siliciclastic sedimentary formations are thus rich in Li (median values over 100 ppm) as hectorite, illite, saponite, stevensite, and sepiolite (Hortsman, 1957 and references therein, Tardy et al., 1972; Vigier et al., 2008). Di-octahedral smectites are generally poor in Li, although Li is a monovalent cation easily exchangeable in the interlayer site of dioctahedral smectites. However, its concentration in the interlayer remains low due to competition with other

monovalent cations (Na, K) at one or two orders of magnitude higher in water. By contrast, trioctahedral smectites contain the highest amounts of Li, notably hectorite (Brenner-Tourtelot and Glanzman, 1978; Starkey, 1982). Hectorite accounts for about 8% of global Li reserves (Evans, 2008).

Pieces of evidence of potential source rocks in the studied basins: The functioning of source rocks and their traceability are challenging to verify. Li extraction from proximal sources, such as volcanic rocks, Li-enriched clay formations (bentonite) or Li-phyllosilicates dispersed in siliciclastic rocks or the underlying crystalline basement, is necessary. But, the information about source rocks may lack in the local geological descriptions, and Li rock and mineral contents are generally unavailable. There are, however, some cases where potential source rocks may be invoked: i) in Appalachian, K-bentonites are described below the Marcellus shale formation. They could be witnesses of ancient volcanic glass, which are good candidates as Li-source; ii) volcanic mafic and felsic intrusions are described in the North German, Gulf Coast, and Salton Trough. Brines, in addition, may have interacted with pre-salt basement rocks, such as in the lower part of the North German reservoirs. Biotite and muscovite in granitic basement may act as potential source minerals, as already invoked in the Soultz-sous-Forêts geothermal field (Rhine graben) by Dugamin (2022).

Exploitability: The Mg/Li ratios between 10 and 100 in the LiRB brines are close to those encountered in salar brines. Such ratios are more favourable than those reaching 1000, relative to seawater, for which the separation of the two cations may be more complicated.

4.4 *The Li- cycle in the crust*

Figure 13 summarises the cycle of Li through the crust. It shows: i) the mechanisms of primary enrichment by evaporation from marine or continental waters yielding marine evaporites on the one hand and salars on the other hand, ii) the formation of secondary brines by salt dissolution, iii) the release of primary brines occurring mainly by crushing of evaporites related to compressive tectonics and salt tectonics (diapir formation), iv) the Li enrichment of primary brines by interaction with lithologies containing Li-rich minerals. The early Li enrichment is generally acquired in the proximity of salt layers (halite saturation) through water-rock interactions preserved from the entrance of low salinity recharge waters. Water-rock interactions may occur at medium to high temperatures for a basin in the case of geothermal activity, mainly when abnormal heat flows are linked to lithospheric thinning during rifting.

The favourable factors for secondary Li enrichment are the presence of: i) highly differentiated rocks (granites and pegmatites within the crystalline basement), ii) late magmatic dykes and intrusive sills in the basement and sedimentary basins, iii) detrital Li-rich minerals (muscovite in particular) in the siliciclastic formation, iv) sedimentary layers dominated by products of hydrothermal alteration or weathering of volcanic emission products from differentiated magmas (bentonite, hectorite rich layers). Enhanced enrichment may then occur thanks to temperature-dependent dissolution reactions involving the Li-bearing phases cited above. In the case of the Clayton Valley, Coffey et al. (2021) show that hydrothermal fluids interacting with volcanic tuffs may produce Li-rich brines, validating water-rock interactions as the primary driver for Li-enrichment.

5. Conclusion

Thirteen basins documented in the literature were selected for their representativeness of basinal brines. The main mechanisms controlling the geochemistry of the Li- enriched brines are as follows:

1- Lithium is correlated with Cl in a series of basins where mixing processes between a dilute end-member close to low-salinity water and a saline end-member relative to seawater or evaporated seawater (the Cl content being fixed by halite saturation) predominate. The dilute end-member has chemical and isotopic features close to meteoric waters, confirming that it corresponds to recharge waters. In Li-rich brines, at a sub-constant chlorinity fixed by the halite saturation, Li spans two orders of magnitude from 80 to 2000 mg.l⁻¹, thus much above the Li concentrations resulting from seawater evaporation. The Li-rich brines were preserved from dilution.

2- Li-rich brines are primary evaporite brines close to halite saturation, with low Cl/Br ratios and $\delta^{18}\text{O}$ similar to or higher than seawater. Favourable geological factors required to obtain such brines are: (i) brine expulsion from evaporites during tectonic events such as compression or salt tectonics; (ii) a relatively short migration distance to the reservoir; (iii) the absence of convection stimulated by thermal anomalies and of connection of reservoirs to recharge zones and surface-related faults to limit dilution processes. Increased Li concentrations can be achieved if primary brines interact strongly with leachable Li source rocks or minerals such as volcanic rocks, Li-rich clay beds or detrital Li-rich phyllosilicates dispersed in siliciclastic rocks. Further enrichment could come from the inflow of Li-rich geothermal waters related to magmatic activities. iv) temperature is a final favourable factor as Li is preferably extracted from source rocks with increasing temperature, a phenomenon well known and expressed by

increased Li/Na ratio with temperature. Symmetrically, the unfavourable factors include: i) large meteoric inflows through boundary faults, ii) inflows into deep connected porous reservoirs, iii) low-temperature gradients, iv) absence of source rocks (series dominated by carbonate platform sediments) such as preserved evaporites, and v) absence of geological events stimulating primary brine expulsion.

3- In some instances, brines are of secondary origin and convect as they are subjected to extreme geothermal gradients. At Salton Trough, high temperatures stimulated the water-rock interactions with evaporites, while a potential magmatic Li contribution cannot be precluded.

4- The exploitability of Li depends on its concentrations and bulk water chemistry. The Li-Mg plots show that the Mg/Li ratios cover various values. The separation of Li from Mg is easier for waters having Mg/Li ranging in between 10 and 100 ratios similar to those of solar brines than for waters with higher Mg/Li ratios.

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Competing interests

The authors declare no competing interests.

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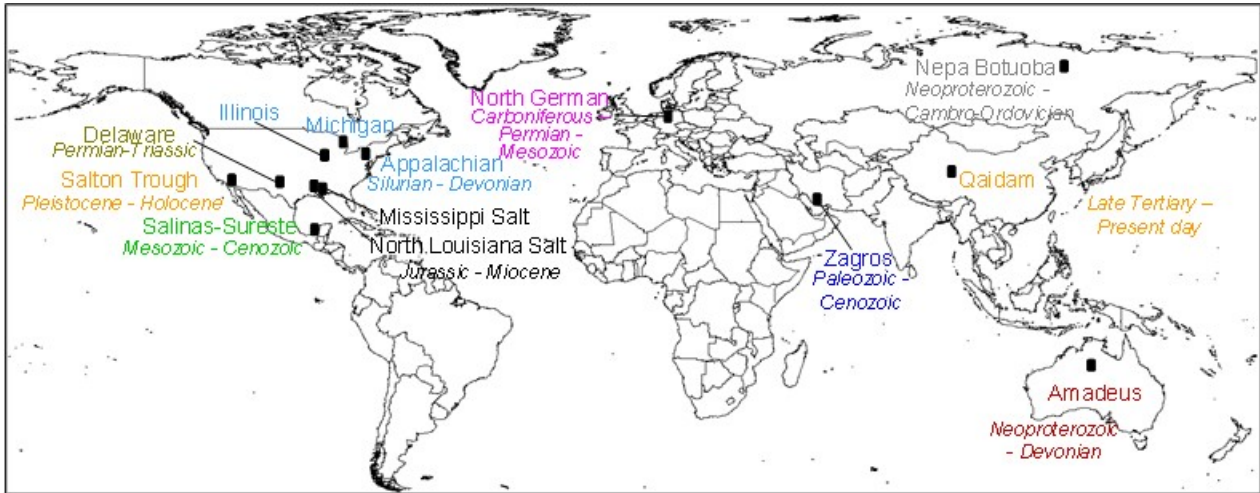


Fig. 1. World map showing the 13 sedimentary basins selected for this study. The age of each sedimentary basin is specified in italics and is indicated by different colours.

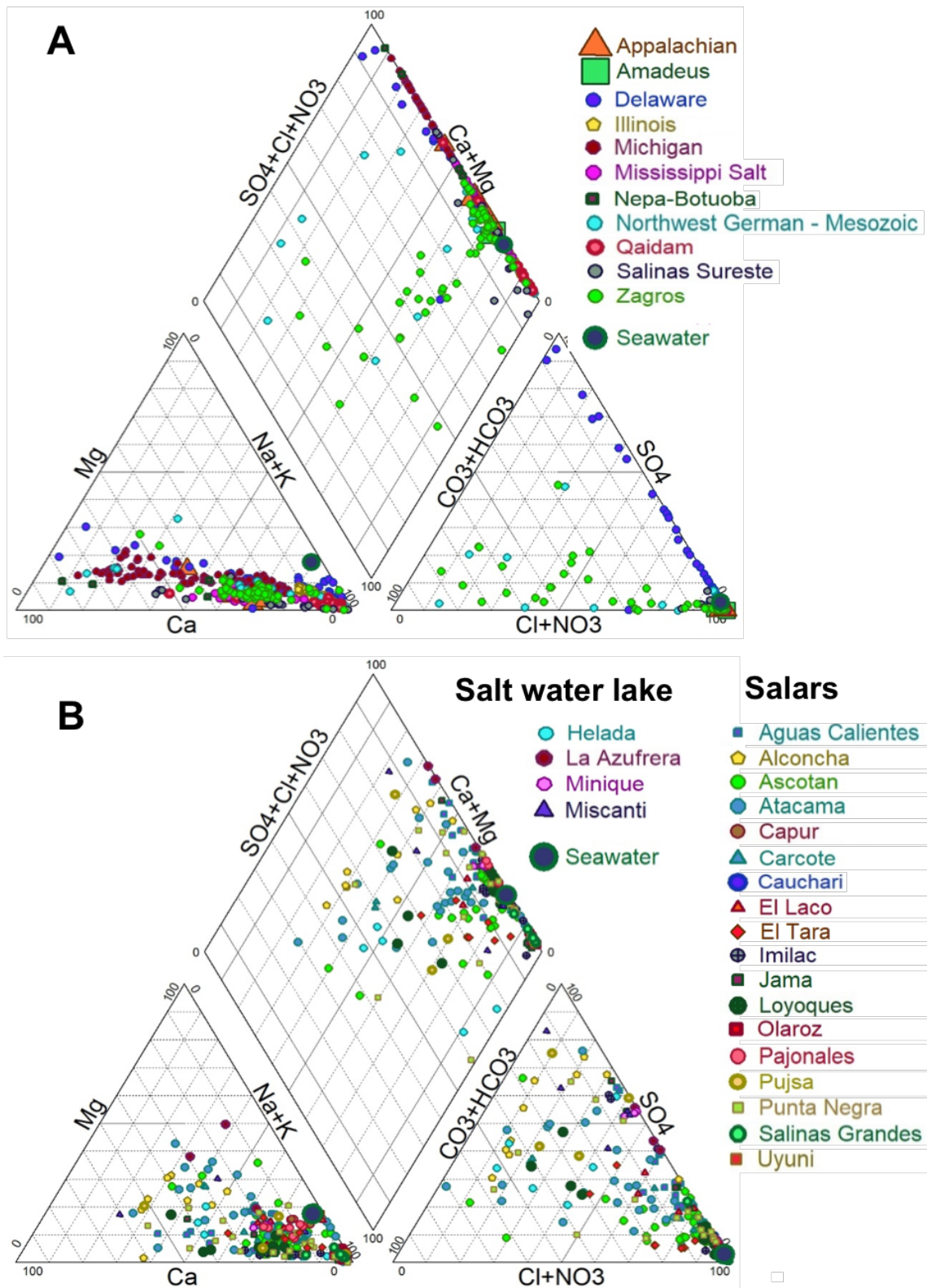


Fig. 2. Piper diagram applied to waters from sedimentary reservoirs (A) and closed basins (B). Data were compiled from literature (see references in table 1 and 3).

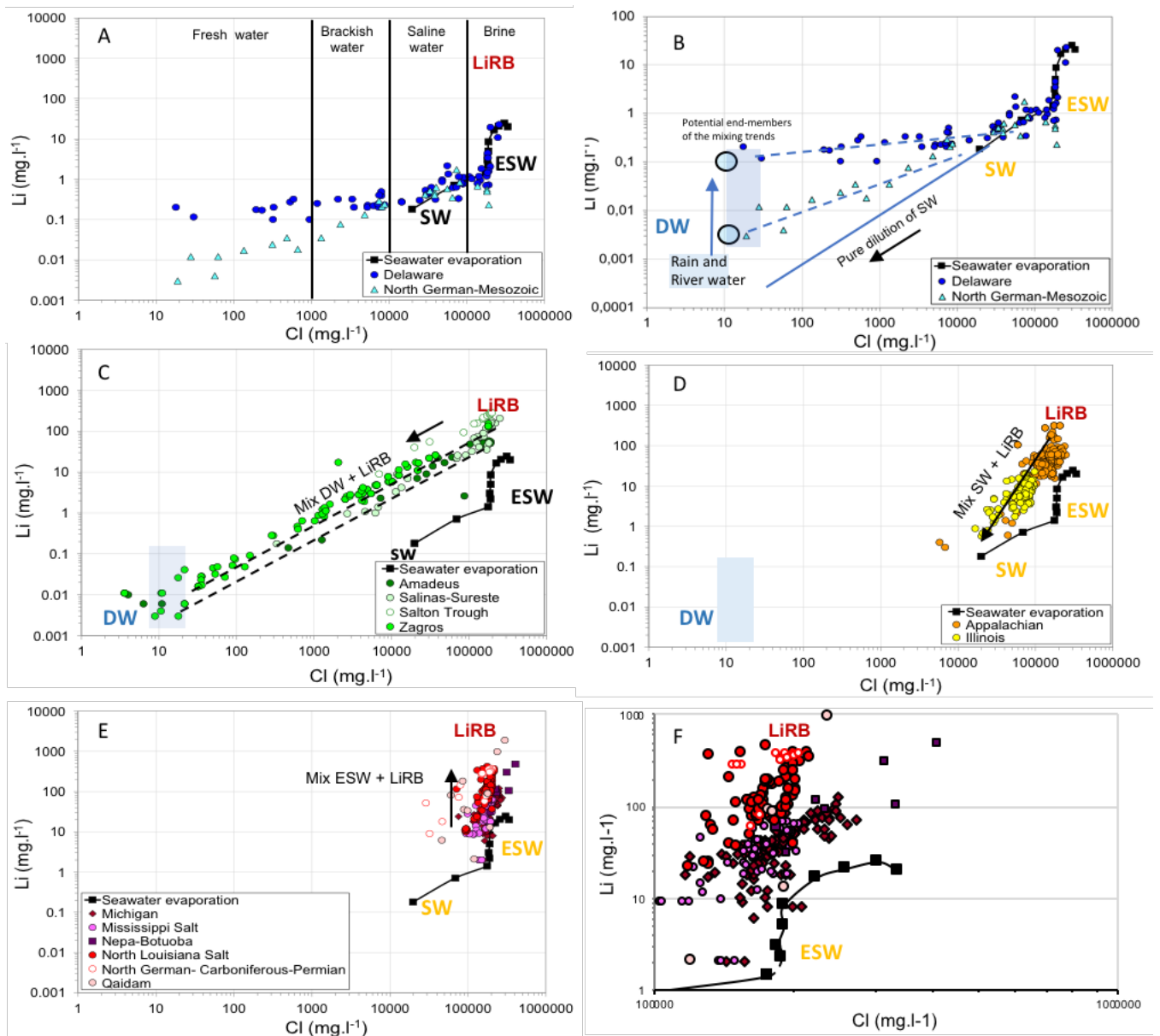


Fig. 3. Lithium concentration in sedimentary formation waters as a function of chloride concentration. DW: dilute water, SW: Seawater, ESW: evaporated sea water, LiRB: Li rich brine. Data for basins with extended dilution trends of seawater are in A, B and C, with B an enlarged and commented version of A. Boundaries between fresh, saline, brackish waters and brines are from Carpenter (1978). D corresponds to sets of brines enriched in Li with a mixing trend towards seawater, and E shows the brines with the highest Li enrichment in between 10 and 100 times the Li content of evaporated seawater. F is an enlarged version of E to show details of data distribution.

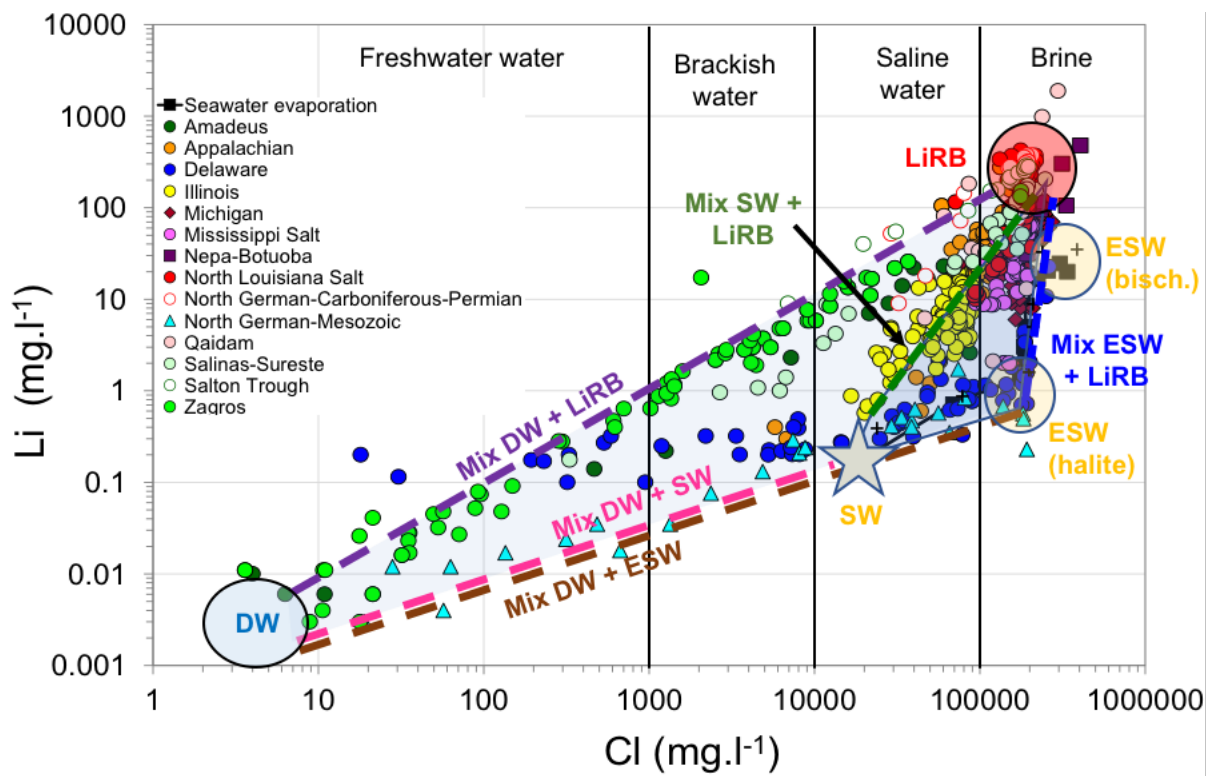


Fig. 4. Main trends of mixing of sedimentary formation waters. LiRB: Lithium rich brines, SW: seawater, DW: dilute water, ESW: evaporated seawater having passed halite saturation (halite) or bischofite saturation (bisch.) (after Fontes and Matray, 1993).

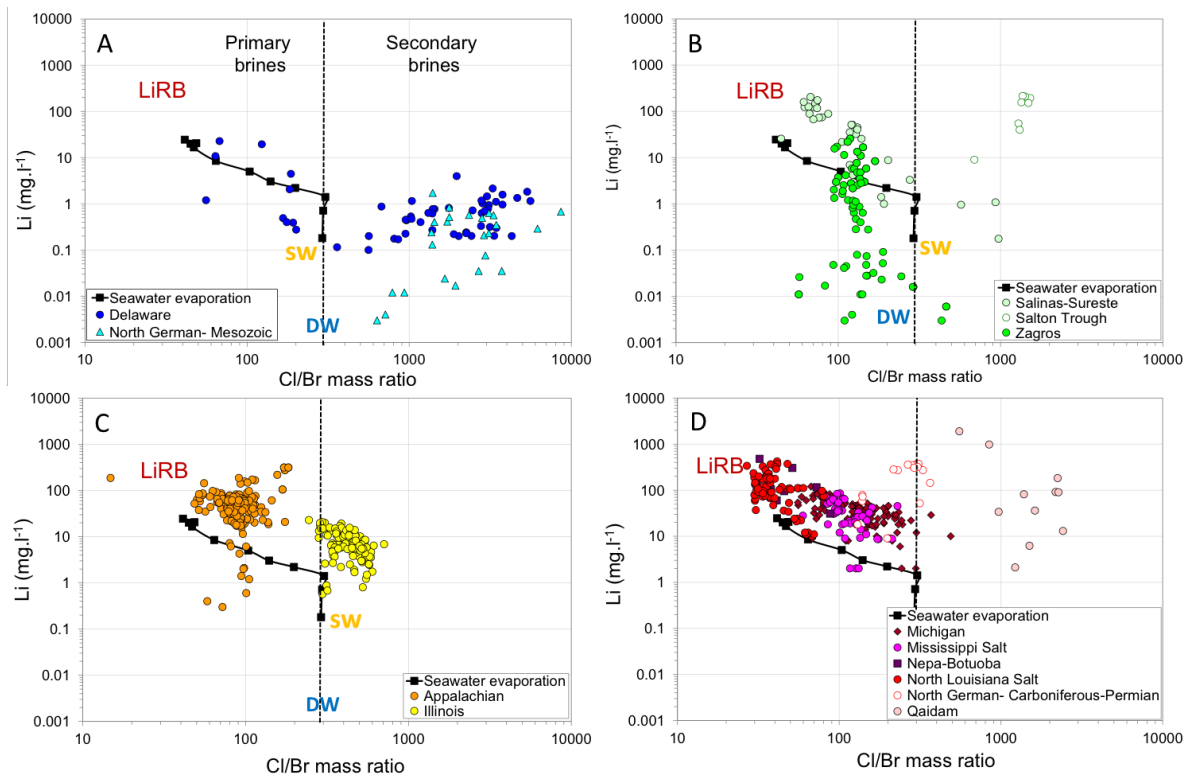


Fig. 5. Lithium concentration as a function of Cl/Br mass ratio in sedimentary formation waters. Data for basins with extended dilution trends are presented in A and B. The basins with high chlorinity close to the values of water saturated with respect to halite are in C and D. LiRB: Lithium rich brines, SW: seawater, DW: dilute water, seawater evaporation trend from Fontes and Matray (1993). Locations of DW and LiRB are indicative.

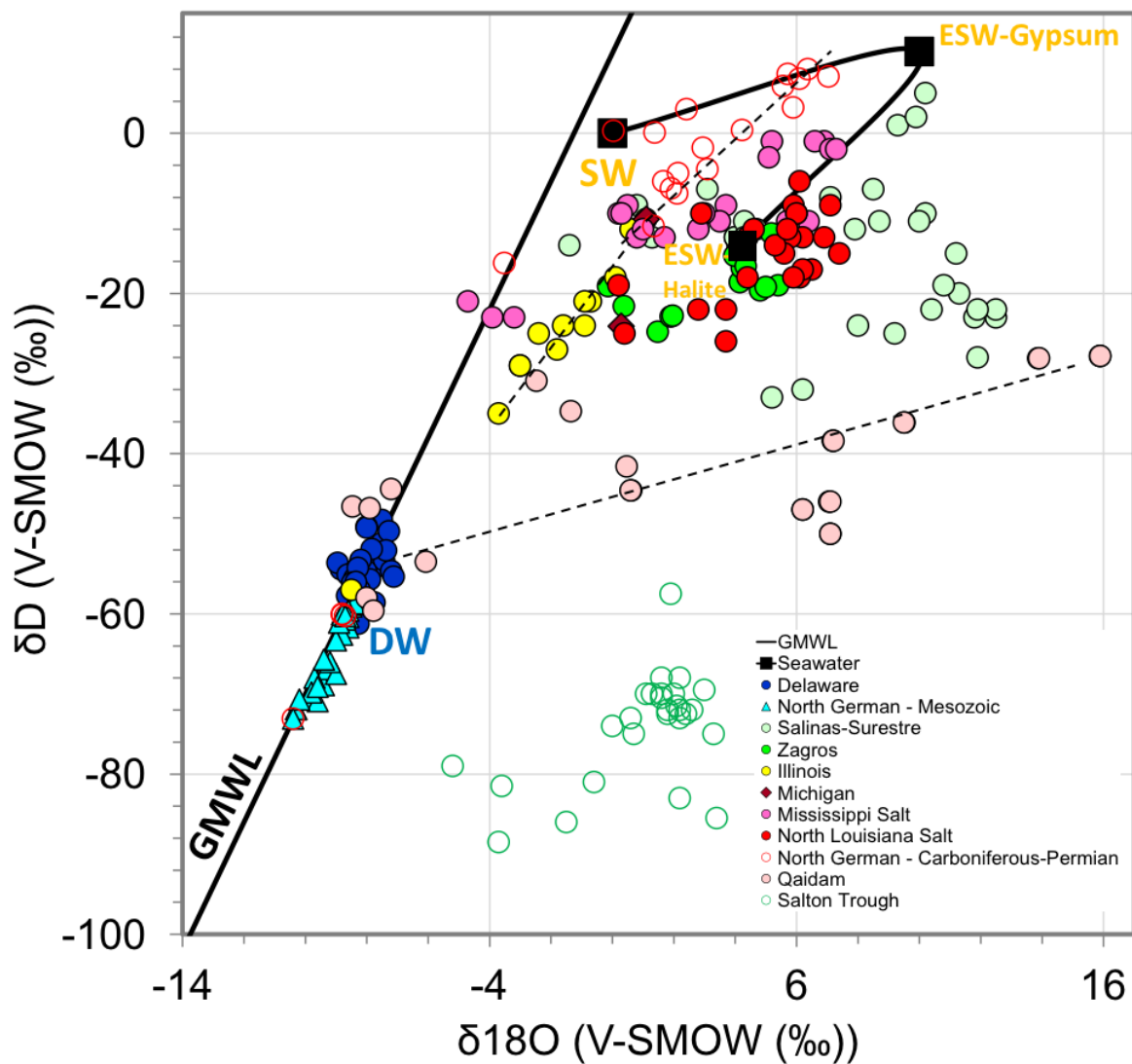


Fig. 6. $\delta^{18}\text{O}$ as a function of δD (V-SMOW ‰). DW: dilute water, SW: seawater, ESW: evaporated seawater up to gypsum and halite, seawater evaporation trend (Knauth and Beeunas, 1986) and global meteoric water line (GMWL) (Craig, 1961). The blackdotted lines show the primary trend of mixing. References data: Delaware (Barnaby et al., 2004), Qaidam (Qichun et al., 2010. Tan et al., 2011), North German (Kloppman et al., 2001; Luders et al., 2010), Zagros (Bagheri et al., 2014a), Illinois (Stueber and Walter, 1991), Michigan (Wilson and Long, 1993a-b), Mississippi Salt (Kharaka et al., 1987), North Louisiana Salt (Moldovanyi and Walter, 1992), Salinas-Sureste (Birkle et al., 2009b) and Salton Trough (Williams and McKibben, 1989).

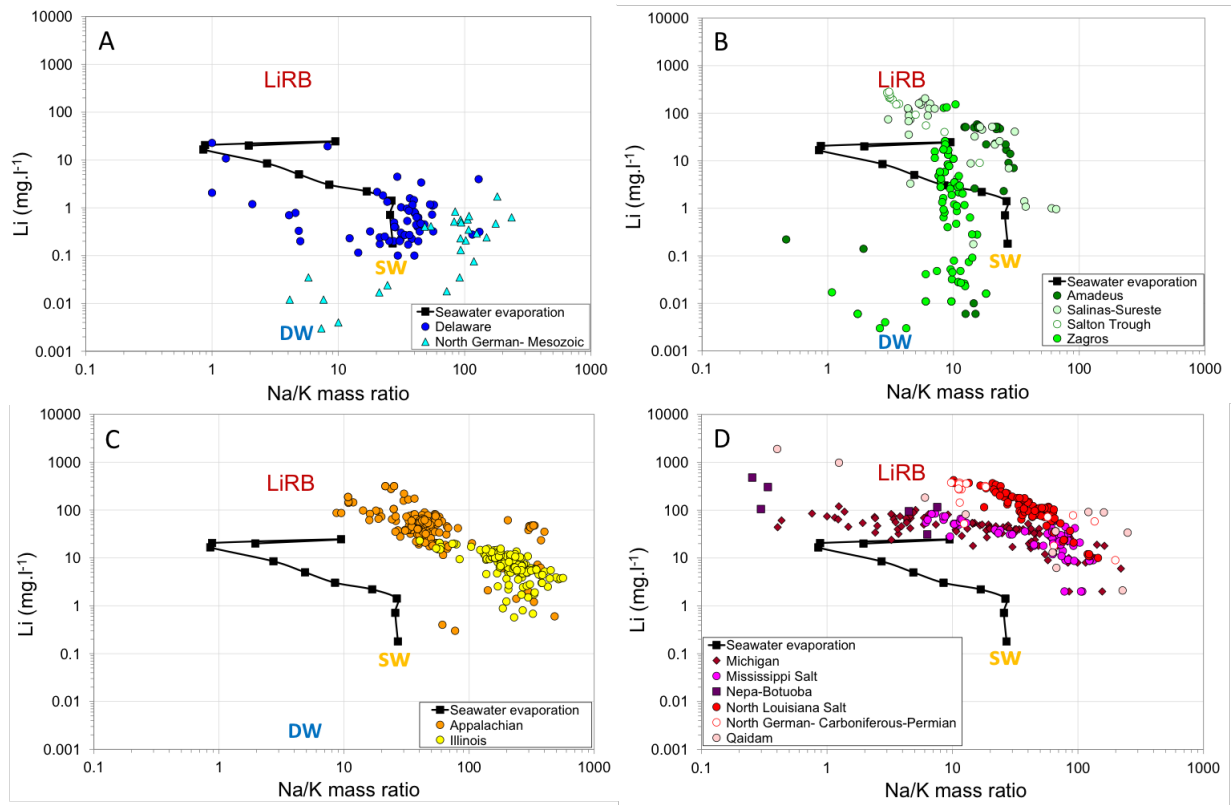


Fig. 7. Lithium concentration in sedimentary formation waters as a function of Na/K mass ratio. Data for basins with extended dilution trends of seawater are in A and B. The basins with high chlorinity close to the values of water saturated with respect to halite are in C and D. LiRB: Lithium rich brines, SW: seawater, DW: dilute water, seawater evaporation trend from Fontes and Matray (1993). Locations of DW and LiRB are indicative.

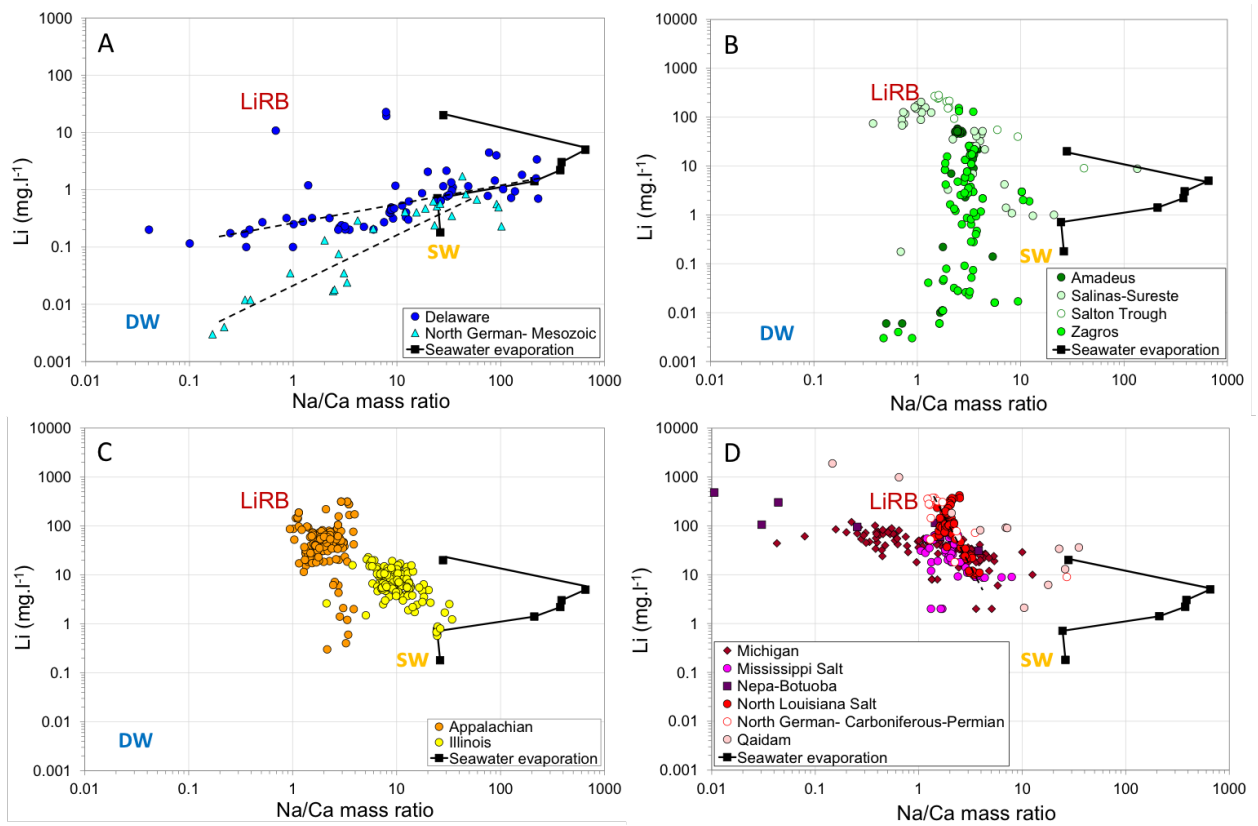


Fig. 8. Lithium concentration in sedimentary formation waters as a function of Na/Ca mass ratio. Data for basins with extended dilution trends are presented in A and B. The basins with high chlorinity close to the values of water saturated with respect to halite are in C and D. LiRB: Lithium rich brines, SW: seawater, DW: dilute water, seawater evaporation trend from Fontes and Matray (1993). Locations of DW and LiRB are indicative.

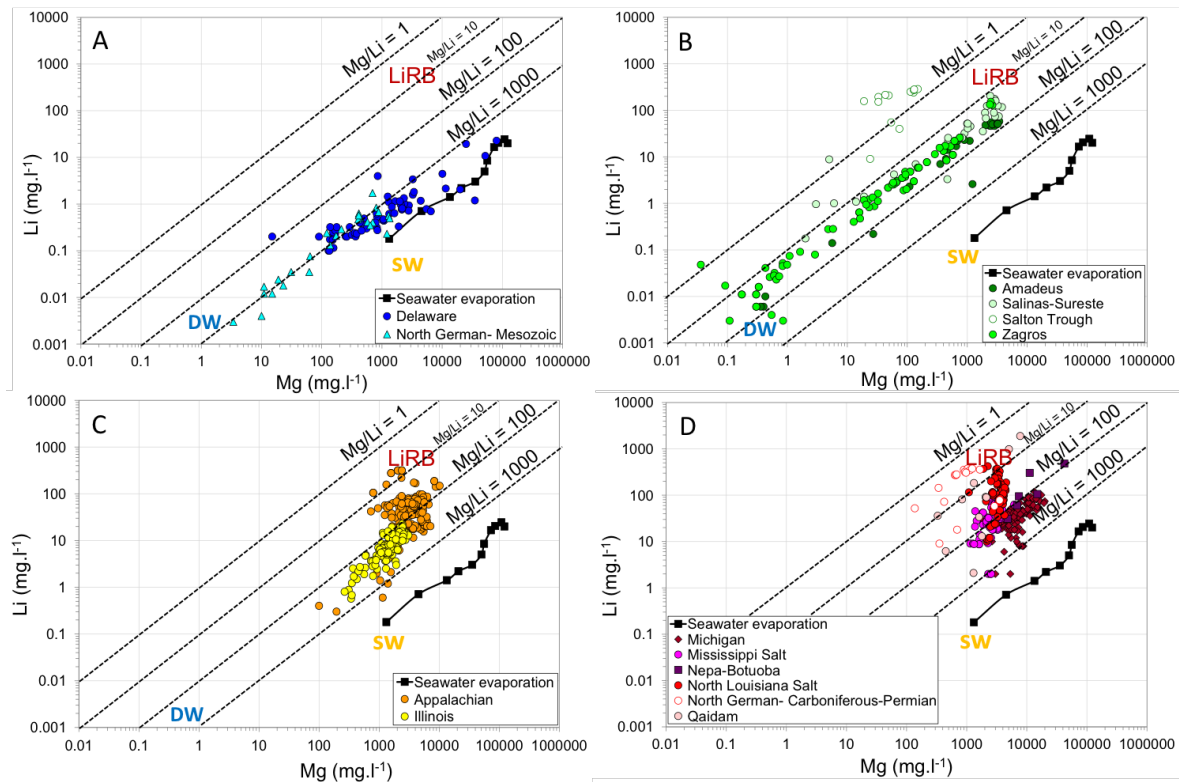


Fig. 9. Lithium concentration as a function of Mg concentration in sedimentary formation waters. Data for basins with extended dilution trends are presented in A and B. The basins with high chlorinity close to the values of water saturated with respect to halite are in C and D. LiRB: Lithium rich brines, SW: seawater, DW: dilute water, seawater evaporation trend from Fontes and Matray (1993). Locations of DW and LiRB are indicative.

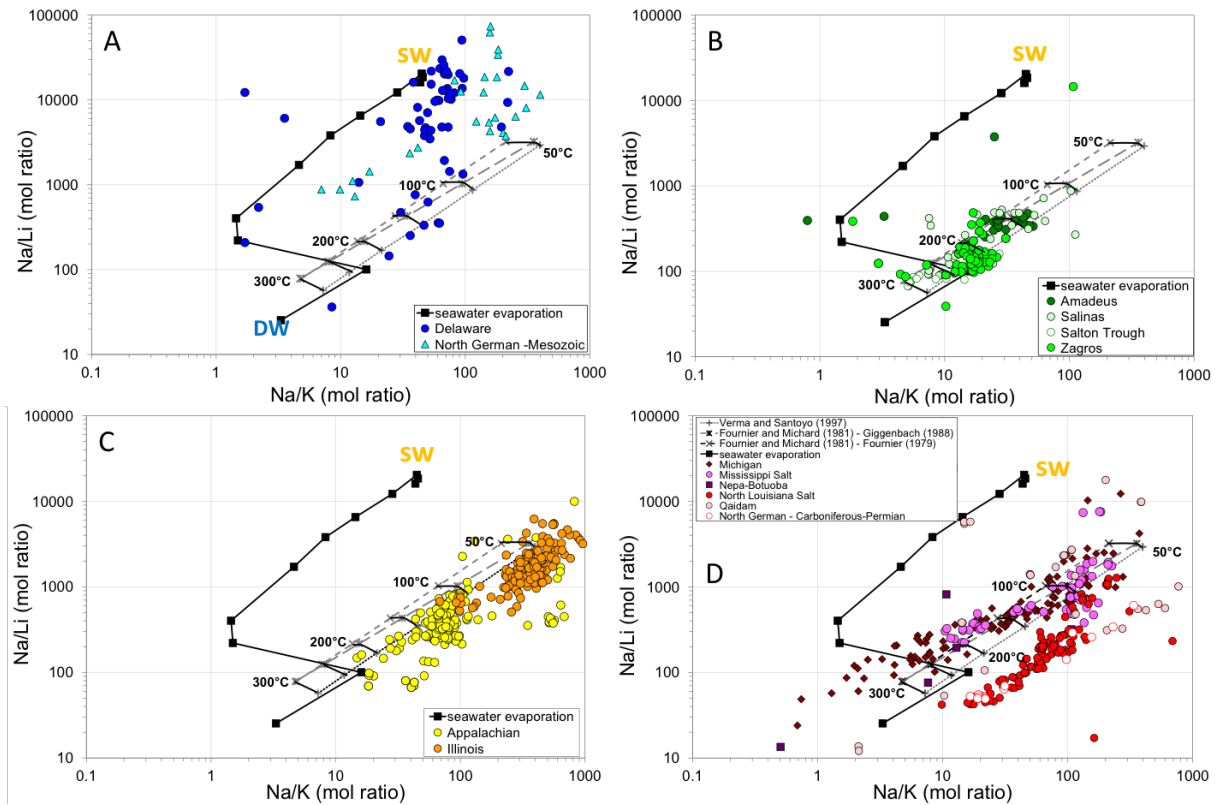


Fig. 10. Na/Li molar ratio in sedimentary formation waters as a function of Na/K molar ratio. Data for basins with extended dilution trends of seawater are in A and B. The basins with high chlorinities close to the values of water saturated with respect to halite are presented in C and D. SW: seawater, seawater evaporation trend from Fontes and Matray (1993). Dotted line: temperature estimation deduced from the geothermometric cation relationships (Fournier, 1979; Giggenbach, 1988; Fouillac and Michard, 1981; Verma and Santoyo, 1997).

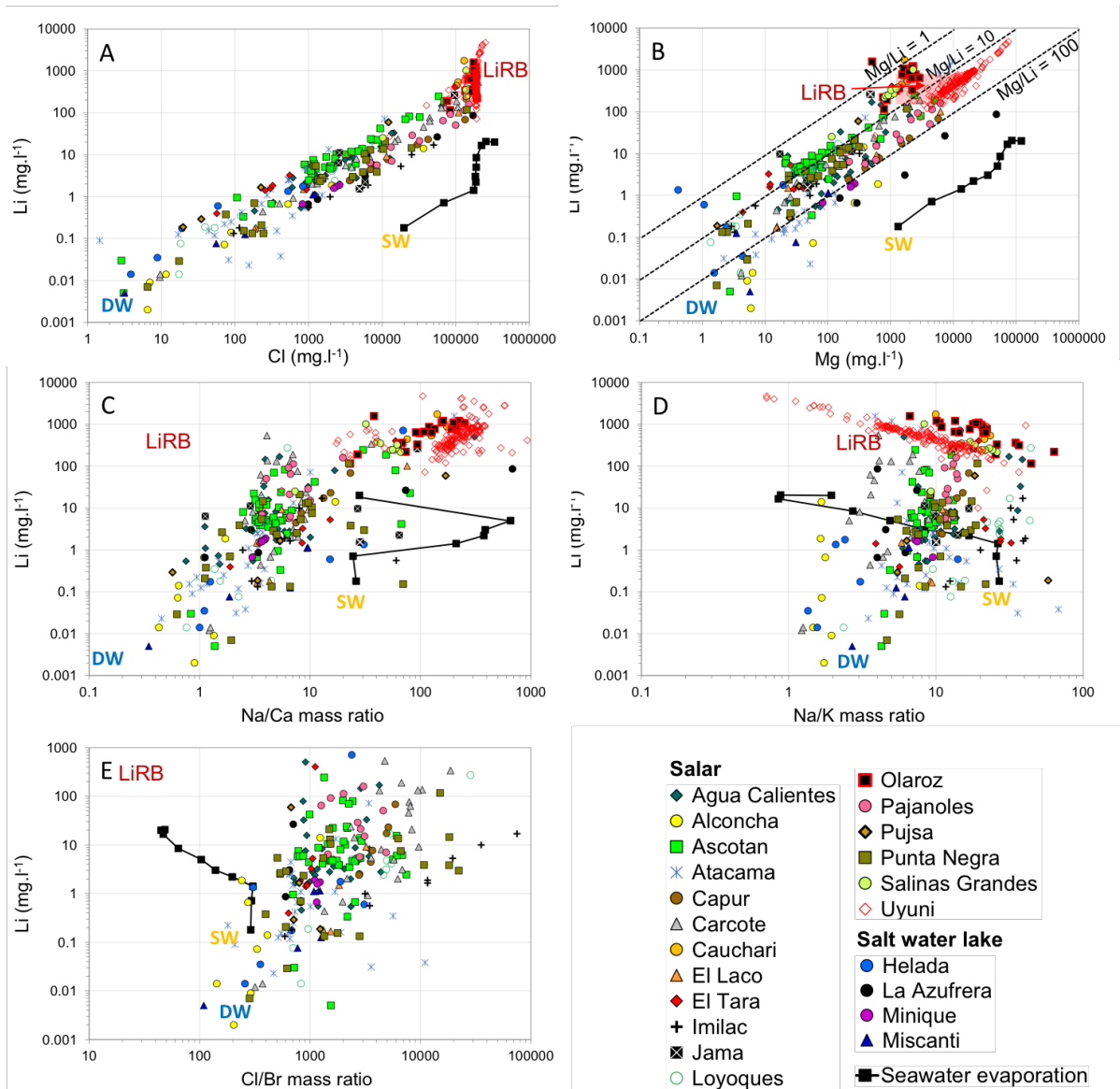


Fig. 11. Data from salars plotted in the chemical diagrams used for the sedimentary basin waters. Lithium concentration as a function of Cl concentration (A), Mg concentration with location of the LiRB brines in the pink inset indicated by the red arrow (B), Na/Ca (C) and Na/K (D) and Cl/Br (E) mass ratios. Main end-members SW: seawater, and end-members defined for sedimentary waters: LiRB: Li rich brines, and DW: dilute water are plotted for comparison. The seawater evaporation trend is reported from Fontes and Matray (1993). References data: Ericksen and Salas (1987), Risacher and Fritz (1991), Risacher et al. (1999), Lopez Steinmetz et al. (2018), Franco et al. (2020).

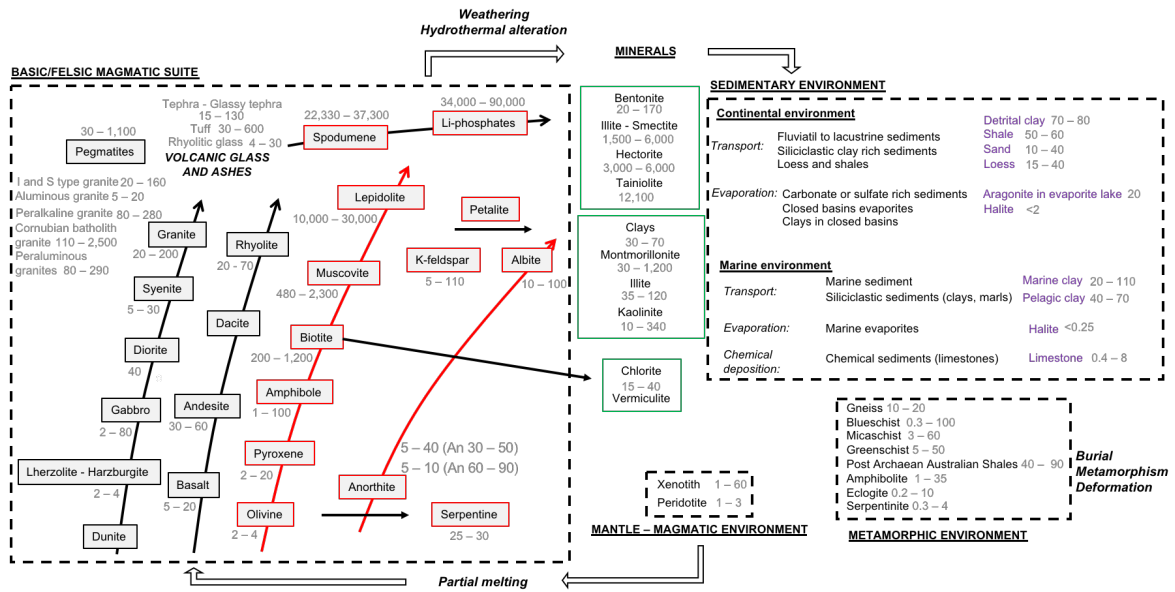
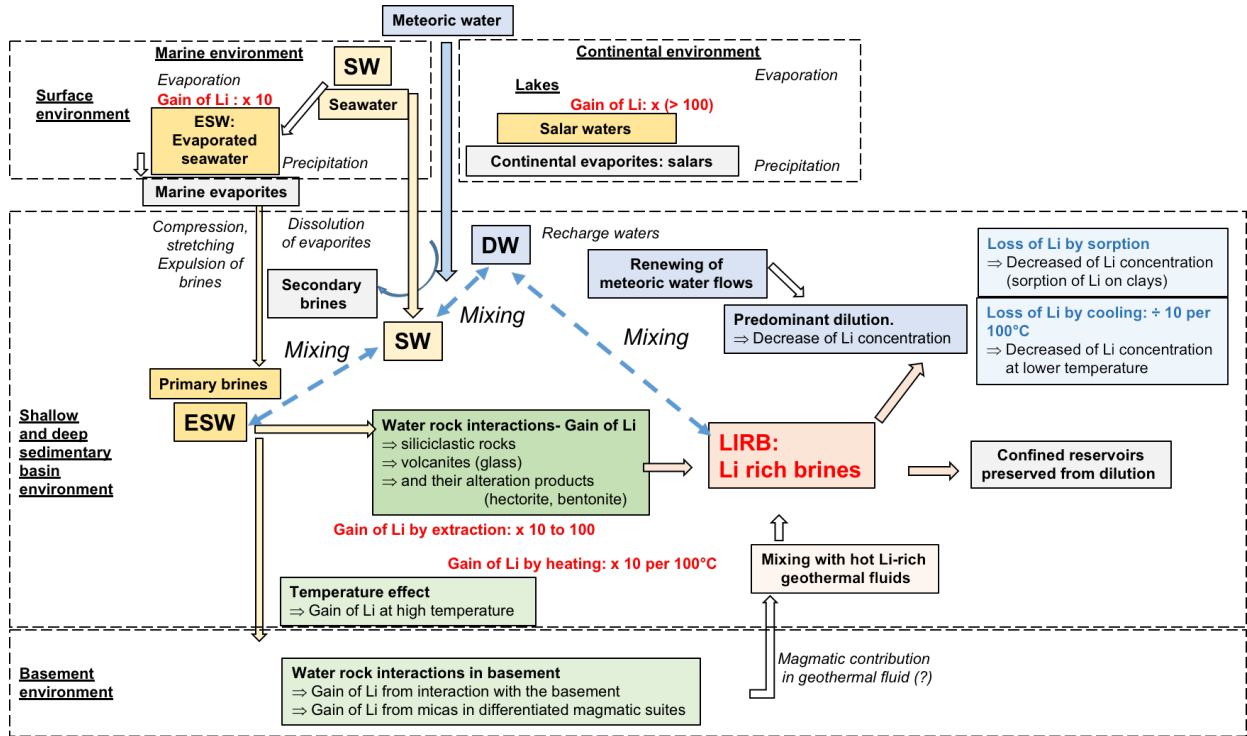


Fig. 12. Geochemical cycle and Li enrichment processes through rocks, minerals and sediments. In the left-hand side inset, the arrows indicate the main magma differentiation trends, with Li concentration range for main rock forming minerals. On the right-hand side, newly formed minerals during weathering and diagenesis, as well as the main sedimentary formations are listed. When possible, the main Li concentrations in rocks, minerals and sediments are indicated according to the 25th and 75th percentile of the data set of the lithium concentration values found in published sources. In a few cases, a single value appears when only one value is found in the literature. References are listed in the Supplementary references section (Supplementary references).

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5 Fig. 13. Lithium through the crustal cycle. At the top of the diagram, the mechanisms of primary Li-
 6 enrichment by evaporation from marine or continental waters yielding to marine evaporites on the
 7 one hand and salars on the other hand; in the middle section, primary brines are expelled from
 8 evaporites by squeezing of evaporites linked to compressional tectonics and salt tectonics, and
 9 secondary brines formed by salt dissolution. They interact with sedimentary reservoir minerals or
 10 layers bearing Li, and become Li-enriched. Interactions with basement rocks (bottom section) may
 11 occur in some instances as well as Li inputs through hydrothermal waters. In blue the effect of recharge
 12 waters in the dilution, cooling of the brines and the decrease of Li concentrations.

13

Table 1. Location of the selected sedimentary basins and chemical compositions (Li, Cl and Mg concentrations, Cl/Br, Na/Ca and Na/K ratios) of the formation waters. The minimum, maximum and mean values (in bracket) are given for set of waters.

Basin	Country	Age of sedimentary basin	Reservoir type	Li (mg.l ⁻¹)	Cl (mg.l ⁻¹)	Mg (mg.l ⁻¹)	Cl/Br mass ratio	Na/Ca mass ratio	Na/K mass ratio	References
Amadeus	Australia	Neoproterozoic - Devonian	Gas	0.006-60 (30)	4-189,960 (88,790)	0.4-3,430 (1,520)		0.5-5 (3)	0.5-30 (20)	Andrew et al., 2005
Nepa Botuoba (Siberian platform)	Russia	Neoproterozoic - Cambro-Ordovician	Oil and gas	31-480 (170)	184,900-405,700 (269,740)	3,700-42,200 (13,030)	30-90 (50)	0.01-4 (1)	0.3-60 (10)	Alexeev et al., 2020
Appalachian	USA	Silurian – Devonian	Oil and gas	0.3-320 (60)	5,760-247,000 (146,480)	100-10,200 (3,490)	15-180 (90)	1-4 (2)	9-480 (70)	Sanders, 1991 ; Skeen, 2010 ; Dresel and Rose, 2010
Illinois	USA	Silurian - Devonian	Oil and gas	0.6-20 (8)	16,630-97,220 (68,635)	270-3,080 (1,490)	240-710 (440)	2-30 (10)	40-560 (230)	Stueber and Walter, 1991 ; Stueber et al., 1993 ; Demir and Seyler, 1999
Michigan	USA	Silurian - Devonian	Oil	2-120 (40)	75,800-275,000 (186,950)	1,820-20,000 (7,960)	60-490 (140)	0.04-10 (2)	0.4-220 (40)	Wilson and Long 1993a - b
North German	Germany	Carboniferous - Permian	Gas	9-375 (220)	29,000-206,000 (145,130)	140-3,530 (1,070)	130-365 (255)	1-30 (3)	10-200 (40)	Lüders et al., 2010
Zagros	Iran	Paleo - Cenozoic	Oil and gas	0.003-30 (9)	4-177,500 (9,990)	0.04-2,630 (190)	60-460 (140)	0.2-50 (4)	1-55 (10)	Bagheri et al., 2014a-b
Delaware	USA	Permian - Triassic	Oil and gas	0.1-55 (2)	20-261,440 (68,800)	15-79,950 (4,870)	60-5,600 (1,890)	0.04-230 (30)	1-130 (35)	Bodine and Jones, 1990
North German	Germany	Mesozoic	-	0.003-2 (0.3)	630-8,625 (2,510)	3-1,340 (410)	20-192,000 (45,625)	0,20-100 (25)	4-235 (85)	Kloppmann et al., 2001
Mississippi Salt	USA	Jurassic - Miocene	Oil and gas	2-90 (35)	94,900-207,400 (164,050)	1,150-6,180 (2,640)	60-230 (120)	1-8 (2)	5-132 (50)	Carpenter et al., 1974; Kharaka et al., 1987
North Louisiana Salt	USA	Jurassic - Miocene	Oil	10-420 (130)	71,255-217,910 (168,340)	1,090-4,540 (3,340)	30-80 (40)	1-4 (2)	10-145 (50)	Moldovanyi and Walter, 1992
Salinas-Sureste	Mexico	Mesozoic - Cenozoic	Oil	0.2-205 (70)	330-248,000 (114,480)	2-3,810 (1,505)	45-970 (175)	0.4-135 (8)	3-70 (20)	Birkle et al., 2009a

Salton Trough	USA	Pleistocene - Holocene	Geothermal	9-280 (170)	6,900-196,800 (122,000)	19-130 (70)	690-1,525 (1,305)	1-40 (6)	3-20 (5)	Thompson and Fournier, 1988; Williams and McKibben, 1989; Lippman et al., 1999
Qaidam	China	Late Tertiary - present-day	Oil	2-1,890 (260)	46,380-297,000 (145,740)	330-7,780 (2,135)	550-2,420 (1,600)	0.1-35 (10)	0.4-250 (100)	Qishun et al., 2010; Tan et al., 2011

Basin names	Country	Age of sedimentary series	Presence of salt beds - diapirs	Description of the sedimentary series	Presence of volcanic and crystalline series	Reference
Amadeus	Australia	Late Proterozoic to Devonian	- Evaporites in Neoproterozoic and Cambrian formations, acting as seal	- Shallow marine to continental deposits - Shales and sandstones		Andrew et al., 2005 Jarrett et al., 2016 Mempes et al., 2020
Siberian platform- Nepa Botuoba	Russia	Neoproterozoic (Vendian) to Cambro-Ordovician	- Widespread Vendian (Lower Cambrian) salt bed - Salt-bearing formation deposited in a giant lagoon that covered the south half of the Siberian craton - Diapirism and salt anticline	- Salts are interbedded with strata of carbonaceous rocks (dolomites, limestones), and siliciclastic formations (sandstone, argillite)	- Archaean-Proterozoic crystalline basement, which is represented by quartzites, gneisses, schists and amphibolites - Devonian rift with volcanic rocks	Ulmishek, 2001
Appalachian Michigan Illinois	USA	Silurian to Devonian	Multiple beds, salt (halite, sylvite and carnallite with polyhalite, potash bearing unit, gypsum, anhydrite)	- Siltstone interbedded with silty shale and clayey shale - Salina group below the Marcellus shale (Mid Devonian oil source) and Oriskany sandstone reservoir - K-bentonite lenses below the Marcellus shale. - Brines in the Utica shale and limestones (Mid Ordovician) coming from the Beemantown group (anhydrite-halite)	Probable volcanic glass (?) transformed in K-bentonite (possible source rocks)	Blondes et al., 2020 Mount, 2014 Rine, 2015 Smith, 2013
Zagros	Iran	Four significant phases: - early Paleozoic intra-cratonic pull-apart basin and platform margin basin, - late Paleozoic platform margin basin, - Mesozoic passive continental margin basin, - Cenozoic foreland basin (Eocene to Pleistocene)	- Proterozoic to Early Cambrian Hormuz salt layer with a thick and continuous salt layer - Evaporite and mudstone of the Gasharan formation (early Miocene): main seals in the foredeep zone - Gasharan evaporite (Fars salt) resulting in part of the dissolution-redeposition of the Hormuz salt - Upper Eocene salt layers with anhydrite, halite and sylvite	- Deep marine hemipelagic – pelagic calcareous shale, marl and limestone with subordinate argillaceous limestone. - Reservoirs in the foredeep zone. Paleozoic reservoir with Silurian, source rock and Meso-Cenozoic reservoirs with Lower Cretaceous source rock.		Liu et al., 2018 Jahani et al., 2009
Delaware	USA	Permian to Triassic	Multiple beds (halite, anhydrite)	- Sand and sandstone - Siltstone and sandstone - Mudstone and siltstone		Engle et al., 2016
North German	Germany	Three significant phases: - rifting initiated during the late Carboniferous, followed by 20 Ma subsidence period (Rotliegend sediments).	- Top of Rotliegend-Zechstein is made of carbonates and evaporate deposits. - Keuper includes dolomite, shale and evaporites. - Several reservoir rocks are	- Siliciclastic series shale, lacustrine evaporites, carbonate, sulphate, salts. - Petroleum system includes Carboniferous and Permian Rotliegend formations	- Abundant volcanic successions of Permian age above a carboniferous basin - Igneous intrusions (diabase dikes)	Luders et al., 2010

		<ul style="list-style-type: none"> - rifting during Triassic followed by subsidence and sediment deposition. - rifting during late Jurassic in response to the North Sea coming event 	<p>observed below and above the massive Trias salt formation.</p> <ul style="list-style-type: none"> - Salt tectonic as diapirism 			
Gulf Coast: Mississippi Salt North Louisiana Salt	USA	<ul style="list-style-type: none"> - Jurassic to Miocene 	<ul style="list-style-type: none"> - Salt tectonics (diapirs) linked to the behaviour of the Mid Jurassic Louann evaporite formation - Louann formation includes halite salt layers (coarsely recrystallised) with streak of anhydrite - Buckner anhydrite formation 	<ul style="list-style-type: none"> - Deltaic, non-marine and continental sandstone and shale. - Carbonate series (laminated lime and dolomitic limestone) - Norplet fluvial and aeolian siliciclastic formation above the Louann salt formation and the Smackover formation 	<ul style="list-style-type: none"> - Local igneous intrusions (diabase dikes) but not acting as evident Li- source 	<p>Carpenter et al., 1974 Land et al., 1988 Rogers et al., 2006 Zhang et al., 2013 Hudec et al., 2013a-b</p>
Salinas- Sureste basin	Mexico	<ul style="list-style-type: none"> - Mesozoic series (from Kimmeridgian to late Cretaceous) to Cenozoic series. 	<ul style="list-style-type: none"> - Salt layers linked to the opening of the Mexico Gulf (Callovia) - Salt tectonics and diapirs forming isolated shallow salt bodies - Strong Miocene diapirism - Anhydrite layer 	<ul style="list-style-type: none"> - Carbonate formations (dolomite, limestone) and mudstones - Shallow oil reservoirs linked to salt tectonics 		<p>Birkle et al., 2009 a-b Comision de hidrocarburos, 2015</p>
Qaidam	China	<ul style="list-style-type: none"> - Late Cenozoic (Pliocene) to present-day 	<ul style="list-style-type: none"> - Evaporite layers corresponding to lacustrine water evaporation - Two periods of salt formation in the Qaidam basin: Late Tertiary Pliocene and Late Quaternary Pleistocene 	<ul style="list-style-type: none"> - Lacustrine sediments: limestone, sandstone, shale, mudstone with anhydrite, halite beds sandstones - Three lakes evolved into salt lakes in the second salt-forming period 		<p>Zhang, 1987 Ma et al., 2021 Cheng et al., 2018</p>
Salton Trough	USA	<ul style="list-style-type: none"> - Pleistocene to Holocene 	<ul style="list-style-type: none"> - Evaporite 	<ul style="list-style-type: none"> - Fluvial, deltaic, lacustrine, evaporite - Low-density siliciclastic sediments and sedimentary rocks from Holocene to Pleistocene - Overlying pre-Salton Trough continental crust - Hydrothermal metamorphism producing greenschist and amphibolite facies hornfels 	<ul style="list-style-type: none"> - Locally intruded by both mafic and felsic igneous rocks - Pre-Salton Trough continental crust consists of Mesozoic granitoids, intrusive in Precambrian gneiss and schists and Miocene volcanic rocks 	<p>Hulen et al., 2002</p>

Table 2. Geological characteristics of the sedimentary basins used in this study, listed as a function of their age from the Proterozoic to the Quaternary.

Basin name	Basin type	Region	Country	Li (mg.l ⁻¹)	Cl (mg.l ⁻¹)	Mg (mg.l ⁻¹)	Cl/Br mass ratio	Na/Ca mass ratio	Na/K mass ratio	References
Agua Calientes (1,2,3,4)	Salar	Puna de Atacama	Chile	0.5-500	190-175000	20-12300	720-4400	1-300	6-40	Risacher et al., 1999
Alconcha		Puna de Atacama	Chile	0.002-15	6-36600	2-630	130-1250	0.5-20	1-10	Risacher et al., 1999
Ascotan		Puna de Atacama	Chile	0.005-250	3-70000	3-5130	700-7500	1-80	4-10	Risacher et al., 1999
Atacama		Puna de Atacama	Chile	0.02-1570	2-189500	2-9650	180-11030	0.5-330	3-70	Ericksen and Salas 1987 Risacher et al., 1999
Capur		Puna de Atacama	Chile	2-70	3620-134000	210-5800	2780-5930	7-20	10-20	Risacher et al., 1999
Carcote		Puna de Atacama	Chile	0.01-540	10-202000	4-9000	320-18750	1-40	2-20	Risacher et al., 1999
Cauchari		Puna	Argentina	440-1740	102200-163900	1510-2100		80-210	10-20	Steinmetz et al., 2018
El Laco		Puna	Chile	0.2-100	190-109630	15-6250	1320-3220	3-80	6-15	Ericksen and Salas 1987 Risacher et al., 1999
El Tara		Puna	Chile	0.5-400	60-98000	4-1050	640-1110	2-60	4-30	Risacher et al., 1999
Imilac		Puna de Atacama	Chile	0.1-20	100-50100	3-600	590-74900	2-60	10-40	Risacher et al., 1999
Jama		Puna	Argentina	2-260	2560-97020	20-500		1-90	8-20	Steinmetz et al., 2018
Loyoques		Puna	Chile	0.01-270	17-148000	1-2070	700-28400	1-6	2-40	Risacher et al., 1999
Olaroz		Puna	Argentina	115-1570	76000-196000	510-2900		20-250	7-60	Steinmetz et al., 2018 Franco et al., 2020
Pajonales		Puna de Atacama	Chile	6-160	6850-151000	220-6200	1260-4890	4-20	10-15	Risacher et al., 1999
Pujsa		Puna de Atacama	Chile	0.2-60	20-12200	2-310	670-1240	1-170	5-60	Risacher et al., 1999
Punta Negra		Puna de Atacama	Chile	0.01-120	7-142000	0.1-1830	280-22230	0.6-70	5-20	Risacher et al., 1999
Salinas Grandes		Puna	Argentina	25-1020	10200-141350	115-2300		7-70	8-30	Steinmetz et al., 2018
Uyuni	Altiplano	Bolivia	10-4720	3400-253000	200-75300		4-930	1-40	Risacher and Fritz 1991	
Helada	Salt water lake	Puna de Atacama	Chile	0.01-710	4-179000	0.4-2450	260-3100	1-70	1-15	Risacher et al., 1999
La Azufrera		Puna de Atacama	Chile	1-90	1000-172130	160-48640	600-820	1-680	4-7	Ericksen and Salas 1987 Risacher et al., 1999
Minique		Puna de Atacama	Chile	1-2	830-2650	80-270	1100-1230	3-4	7-9	Risacher et al., 1999
Miscanti		Puna de Atacama	Chile	0.005-1	3-1210	4-100	110-1260	0.3-10	3-7	Risacher et al., 1999

Table 3. Location of the selected closed basins from South America and related water chemical compositions (Li, Cl and Mg concentrations, Cl/Br, Na/Ca and Na/K ratios). The minimum and maximum values are given for each set of analysed waters.

