### **ORIGINAL PAPER**



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# From freshwater inflows to salt lakes and salt deposits in the Qaidam Basin, W China



### Abstract

The Qaidam basin in W China is an immense hyperarid intramontane basin with flat vast playas and salt lakes on the Qinghai-Tibet Plateau. The central basin is about 2800–2900 m a.s.l. elevation and enclosed by mountain ranges reaching > 5800 m in the Qilian Mountains and > 6200 m in the eastern Kunlun Mountains. The extensive playas of the basin are covered by gypsum or halite with very subordinate additional solids. In this contribution we report on the chemical composition of salt lakes and inflows to the Qaidam basin (analysis of 30 water samples collected in the summer of 2008 and 2009) together with the composition of 22 salt samples. Salt lakes and small salt ponds formed at topographic depressions. Some of the lakes cover  $> 300 \text{ km}^2$  surface but are very shallow (1–2 m deep). Most salt lakes and salt ponds are NaCl dominated and contain typically 250–300 g kg<sup>-1</sup> total dissolved solids (TDS). Some lakes are industrially used and produce KCI fertilizer, LiCI, and boron or are strongly modified by deep water produced in oil fields. Lakes along the borders to the high mountains are typically not fully saturated with halite. However, also these lakes lost most Ca and are drastically enriched in Mg and some lakes also in B and Li. The chemical development of the most natural salt lakes follows a path producing Ca-deficient water that ultimately precipitate Mg-bearing carbonates and chlorites in addition to halite upon evaporation. The salt lakes form by continuous and drastic evaporation of the waters supplied by the inflows to the lakes in the basin. All inflows carry considerable amounts of CI and are characterized by very high Cl/Br ratios. These chemical characteristics suggest that the salt load of the inflows originates mostly from re-dissolved windblown halite deposited together with sand up to high altitudes in the bordering mountain ranges. Also, thermal waters ascending along deep faults along the Qilian Mountains carry considerable amounts of chloride. Their low Cl/Br ratio however suggests that most of the dissolved Na is derived from minerals of the basement rocks by fluid-rock interaction at T > 130 °C. The thermal fluids also carry considerable amounts of boron, indicating that co-precipitated borax in the salt lakes ultimately also derives from minerals in the basement rocks (tourmaline). Consequently, the presented data improve the understanding how the brines and salt lake waters develop from a wide range of chemically distinct low-TDS inflows and how the sequences of minerals precipitated upon evaporation in the Qaidam basin formed.

Keywords Salt lakes, Evaporites, Qaidam Basin, Brines, Water-rock interaction

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### 1 Introduction

Evaporating seawater produces a predictable set of primary salt minerals crystallizing from increasingly concentrated hypersaline waters. The initial stages of brine concentration tend to form primary evaporitic carbonates such as aragonite (CaCO<sub>3</sub>), followed by gypsum (CaSO<sub>4</sub>·2H<sub>2</sub>O) and later halite (NaCl) precipitates from hypersaline marine waters. Finally, bittern salts, i.e. potassium or magnesium sulphates and -chlorides precipitate



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from the residual brine. Dominant bittern salts are carnallite (MgCl<sub>2</sub>KCl·6H<sub>2</sub>O) and epsomite (MgSO<sub>4</sub>·7H<sub>2</sub>O) (McCaffrey et al., 1987; Warren, 2006).

Unlike marine brines, a more diverse and difficult to predict set of evaporate minerals precipitates during evaporation of continental waters. Evaporation of continental water occurs typically in hydrologically closed basins surrounded by mountains such as the Qaidam Basin in W China. The composition of non-marine brines depends on the rock types exposed in the mountains surrounding the closed basin. The rocks are leached by precipitation water and then evaporate in the drainage basin (Eugster & Hardie, 1978; Warren, 2006; Ye et al., 2018). The consistent general trend is an increase of sodium relative to the other cations. Once halite (Hl) precipitates from the basin brine cations including Mg, Li, and K may increase in salt lakes (all abbreviations of mineral names after Whitney & Evans, 2010). The inflowing dilute waters are the precursors of brines. In nonmarine settings such as the studied Qaidam Basin, rivers and groundwater are the major source of many solutes that are ultimately forming the evaporate salts. Additionally, upwelling waters from deeper strata or springs in the basin can be relevant in the formation of evaporate minerals (Lowenstein & Risacher, 2008; Ye et al., 2015, 2018).

Eugster and Hardie (1978) distinguished five major water types in so-called 'hydrologically closed' continental evaporite basins and developed characteristic suites of evaporite minerals for each water type. Alkaline earth carbonates are the first precipitates forming in evaporating waters due to their low solubility. Thus, in the early stages of the evaporative evolution of saline waters the initial  $(Ca + Mg)/CO_3$  mol ratio is the most important parameter (Eugster & Hardie, 1978). Precipitation of e.g. calcite causes loss of Ca and CO<sub>3</sub> from solution in equal proportion. Unequal proportions in the initial solution will determine, which solute dominates on further concentration, i.e. if the remaining solution becomes carbonate-rich or carbonate-poor. Eventually, evaporation will produce further new minerals and thus change the composition of the brine again (Deocampo & Jones, 2014; Drever, 1997; Spencer et al., 1990).

However, saline lakes often occur in basins that contain evaporites from previous saline lakes. Recycling of these older salt deposits can be a major control on water chemistry, typically resulting in a chloride-rich brine. Generally, saline waters show a wide range of chemical composition, as result of many different chemical processes including mixing, degassing (particularly of  $CO_2$ ), and temperature changes (e.g. Drever, 1997; Warren, 2006). In basins with very low temperature during wintertime, like in the Qaidam Basin, cryogenic effects can be important. This may potentially result in the deposition of mirabilite (NaSO<sub>4</sub> $\cdot$ 10H<sub>2</sub>O) instead of thenardite (NaSO<sub>4</sub>) (Steiger & Asmussen 2008).

Previous hydrochemical research on salt lakes in the Qaidam Basin concentrated mainly on the individual salt lakes (e.g. Zhang et al., 2019; Ye et al., 2015, 2018; Lowenstein & Risacher, 2009; Spencer et al., 1990; Yu, 1986), often with a focus on distinct brine components such as lithium or potassium (e.g. Yu et al., 2012; Zhang et al., 2019) or stable isotopes (e.g. Vengosh et al., 1994; Lowenstein et al., 1994; Fan et al., 2015; Du et al., 2019). Few studies considered the catchment hydrology of particular salt lakes, i.e. inflow of rivers or brooks (e.g. Ye et al. 2018; Spencer et al., 1990; Vengosh et al., 1995). Some studies investigated (particular) evaporative minerals in certain areas of the basin (e.g. Kong et al., 2014; Wang et al., 2016; Duan & Hu, 2001; Kezao & Bowler, 1986; Garret, 1996).

In this article hydrochemical data from 14 salt lakes in the Qaidam Basin are presented and water composition data from 16 rivers, brooks, and springs flowing into the basin from the surrounding mountains. Additionally, mineral data on 30 collected evaporate mineral samples, within or nearby the salt lakes, are given and compared with the chemical composition and saturation state of the salt lakes. The extensive data set offers a snapshot overview on the salt lakes and the inflows into the Qaidam Basin, and on the nature of evaporative minerals in or in close vicinity of the lakes. The further objective of this investigation was to provide insight into the chemical connection between inflows and salt lakes, to understand the brine development from chemically distinct inflows, and to deduce the origin of the different evaporate mineral deposits from the hydrochemical properties of the salt lakes.

### 2 Geology and climate

The Qaidam Basin is a large closed basin located on the northeastern margin of the Tibetan plateau, surrounded by mainly Precambrian basement rocks of the Qilian Mountains to the north, the Kunlun Mountains to the south, and the Altun Mountains in the west (Fig. 1, Xiao et al., 2018). The mountains surrounding the Qaidam Basin reach an elevation of over 5000 m a.s.l. in the northern and southern part and more than 4000 m a.s.l. in the northwest (Zhang, 1987), which results in a basin-and-range type topography of the study area. The Basin covers an area of about 120,000 km<sup>2</sup>. The present day central basin area, located at a minimum altitude of 2800 m a.s.l., is predominantly covered by dry salt flats and many shallow saline lakes. 'Qaidam' means salt flat, referring to the widespread occurrence of salt deposits (Chen & Bowler, 1986; Stober et al., 2016; Ye et al., 2018). The salt flats with densely packed white halite deposits



Fig. 1 Location of the investigation area in northwestern China. *TB* Tarim Basin, *NCC* North China Craton, *YC* Yangze Craton, *CC* South China Craton (after Xiong et al. 2014)

are underlain by stratified salt sequences. Vast desert areas are also covered with gypsum deposits, with gypsum crystals up to several cm in size.

Perennial and ephemeral rivers and brooks flow into the Qaidam Basin (Fig. 2). The larger rivers include in southern direction Togrisay He (south of Mangnai), Narin Gol (Fig. 3A), Golmud He (at Golmud) and from northern direction Tatalin Gol (E of Da Qaidam city), Ige He (NW of Da Qaidam city).

The Qaidam Basin is the largest intramontane endorheic, nonmarine basin on the northern Tibet Plateau with no outlet to the sea. There are several playas in the basin, shown in Fig. 2. These playas and saline lakes occupy over one-fourth of the basin (Miao et al., 2016). The basin contains about 27 salt lakes (Fig. 3B), many of which were linked via thick salt beds and brines (Fig. 3C), and a very thick (> 3000 m) sequence of Quaternary sediments deposited in the central Qaidam Basin (Chen & Bowler, 1986; Zhang, 1987; Liu et al., 1998; Fan et al., 2010a; Miao et al., 2016; Guo et al., 2018). The young sediments of the Qaidam Basin are underlain by further Cenozoic sediments, which are overlying unconformably a variety of rock units including Paleozoic granites, Proterozoic-Paleozoic metamorphic rocks, and Jurassic and Cretaceous sedimentary strata (Yin et al., 2002).

The Qaidam basin is located within a complex system of compressive structures in northeast Tibet and is the largest topographic depression inside the Tibetan plateau. Specifically, four thrust belts control the evolution of the basin squeezing it to a narrow irregular shaped structure: the Kunlun fault to the south, the Altyn Tagh fault to the northwest, the Kunlun Mountains thrust belt on the southern margin, and the Oilian Mountain thrust belt to the northeast of the basin (Fig. 2). The Altyn Tagh fault is the major boundary fault, leading to stronger deformation in the north (Guo et al. 2017; Zheng et al., 2013; Yin et al., 2008; Wang et al., 2006; Yang et al., 2001; Zhang et al., 2001; Cowgill, 2001). The left slip on the Altyn Tagh fault zone is related to and absorbed by crustal shortening within the Qilian Mountains, the Qaidam basin, and other convergent structures south of the fault zone. The deformation history of the Qaidam basin shows that the basin experienced continuous compression with a total shortening of over 25 km since the beginning of Cenozoic (Guo et al., 2017; Meng & Fang, 2008; Zhou et al., 2006; Wang et al., 2006). In later Eocene and Quaternary, the basin had two relatively fast shortening phases. The Cenozoic sedimentary center is located in the central basin with about 15 km deposition (Guo, et al. 2017). Generally, the salt lakes lie within the main areas of late Pleistocene subsidence (Chen & Bowler, 1986).

The Qaidam Basin is one of the largest hyper-arid intramontane basins in the northern hemisphere. It is located at a triple junction of influences from the Southeast Asian monsoon, the westerlies, and the Southwest Indian monsoon (Bryson, 1986) and made it sensitive to past climatic variability (Fan et al., 2010a). The ongoing uplift of the high mountains around Qaidam Basin

![](_page_3_Figure_2.jpeg)

![](_page_3_Figure_3.jpeg)

created conditions for the formation of vast halite deposits combined with clastic sediments (Yuan et al., 1983). Under extremely dry climatic conditions during late Neogene, expansive and thick salt beds were accumulated in the basin (Zhang, 1987). However, there is also evidence for recent climatically controlled water-level fluctuations in shallow-water environments during evaporite deposition (Chen & Bowler, 1986). Today, an arid continental climate with long cold winters, short warm summers and very low precipitation prevails in the Qaidam Basin. The region receives only 55 mm of rainfall on average while evaporation rates are approximately 3000 mm year<sup>-1</sup> (Fan et al., 2014; Miao et al., 2016; Yang et al., 1993). The mean annual air temperature is 3 °C. Temperature may exceed 30 °C in summer (July) and fall below -20 °C in winter (January). In winter there are often strong storms and with wind

![](_page_4_Picture_2.jpeg)

Fig. 3 Typical landscapes in the Qaidam Basin. A Meandering river Nasin Gol on its way into the salt lake Dong Taijnar Hu. Width of the main channel 25 m. B Lake Dabusun in the background; in front solid rugged crust of halite, in places shattered. Lake Dabusun and the salt flats are part of the central Qarhan playa. Distance to the small island c. 500 m. C Salt lake Da Qaidam Hu with rugged salt structures in the foreground. Width of the polygonal shapes 20–30 cm. D Aeolian salt-rich sand deposits in the Qilian mountains, which originally came from the Qaidam Basin. High of the mountains 5827 m a.s.l. and c. 2600 m above earth surface

speeds of up to 40 m s<sup>-1</sup>, transporting sand and salts from the basin high up into the mountains (Fig. 3D). Due to the dry climate and lack of outlets, lake waters gradually evaporate driving an increase in salinity that leads to salt formation (Ye et al., 2018). During the cold winter season, borax and mirabilite (sodium sulfate hexahydrate; hydrated thenardite) precipitate directly from certain salt lake water (Sun Bryson, 1986, Li, 1993, cited in Garrett, 1998).

Even though some of the salt lakes have an extensive surface area, they are generally very shallow, making their hydrochemical composition very sensitive to climatic, temperature, and seasonal changes and their surface areas extremely variable (Zhou, et al. 2016; Gao et al., 2019). Since the evaporitic environments and controlling geochemistry have always differed between local sub-basins of the Qaidam Basin, mineral assemblages in the different playas or hydrochemistry of salt lakes of the Qaidam Basin are highly variable. Borates and halite predominate along the northern margin (Chen & Bowler, 1986), while sulfates together with halite prevail in the western region. The Dalangtan playa in the northwestern Qaidam Basin contains halite (NaCl), carnallite (KCl·MgCl<sub>2</sub>·6H<sub>2</sub>O), gypsum (CaSO<sub>4</sub>·2H<sub>2</sub>O), anhydrite (CaSO<sub>4</sub>), mirabilite (Na<sub>2</sub>SO<sub>4</sub>·10H<sub>2</sub>O), glauberite (Na<sub>2</sub>SO<sub>4</sub>·CaSO<sub>4</sub>), and large quantities of MgSO<sub>4</sub> salts (meridianiite, epsomite, hexahydrite, pentahydrate, starkeyite, sanderite, kieserite) have been reported (Kong et al., 2014; Ye et al., 2018). In contrast, the Qarhan salt lake in the eastern Qaidam Basin (Fig. 2) does not contain  $MgSO_4$  salts. However, it is the largest potassium brine occurrence and the lake also contains a large amount of deposited calcite (with smaller amounts of dolomite, aragonite and magnesite) together with some gypsum, anhydrite, halite, carnallite, and some sylvite (KCl) (Fan et al., 2015; Wang et al., 2014; Warren, 2016; Ye et al., 2018).

As one of the largest basins in China, the Qaidam Basin accumulated commercially interesting amounts of different salt deposits and brine resources. These reserves have been exploited and utilized. Potassium, borax, lithium, strontium, and other commodities are extracted from the salt lakes. Oilfield brines with a great quantity of oil reserves exist in the Cenozoic strata in the western Qaidam Basin originating from Paleogene saline lacustrine carbonate source rocks (Fan et al., 2010b; Xing et al., 2022).

### 3 Sampling and analytical methods

An extensive and broad collection of waters (fresh and slightly brackish waters, brines) and a variety of evaporitic minerals from the Qaidam Basin was gathered during fieldwork in the summers of 2008 and 2009. The collection includes 30 water- and 22 evaporite samples. The sample locations are given in Fig. 2. Description of water samples are presented in Table 1 and of evaporite mineral samples in Table 2. The water samples include fresh water and slightly brackish water from rivers, brooks,

Table 1 Location, coordinates, altitude, and description of the water samples taken in the Qaidam Basin (see Fig. 2)

Sample	Northing	Easting	Altitude (m a.s.l.)	Name of location	Description
QD 1	385,301.72	935,444.75	2801	Lake: Sulhai Hu	Salt lake, sample taken from boat
QD 2	375,705.35	914,737.84	2920	Huanggua Liang	Small salt 'puddle'
QD 3	375,441.87	914,951.97	2892	Spring nearby QD 2	Spring, slight bituminous smell, gas bubbles
QD 4	375,430.52	914,955.72	2925	Spring nearby QD 3	Spring, with white border
QD 5	383,624.10	911,416.76	2713	Pool in Dalangtan playa	Small pool in 'gypsum desert'
QD 6	383,007.90	913,414.85	2746	Lake in Dalangtan playa	Salt lake in oilfield
QD 9	381,425.22	901,953.08	2997	Brook from Karawatag Shan Mountains	Small brook N of Tomorlog
QD 10	381,454.19	902,057.50	2986	Small lake near QD 9	
QD 11	382,413.79	903,039.84	3379	Brook: Seven Springs	Brook with reed in steep valley
QD 12	380,756.05	905,603.46	2859	Small lake near lake Gas Hure Hu	Very small lake with greenish water in white salt pool
QD 13	381,314.82	904,943.53	2901	Small pool near QD 12	Hole at the rim of seismic blasting with mud-dom
QD 15	365,624.99	930,008.10	2951	River: Nasin Gol	Very large, red-brown river, meander, from Kunlun Moun- tains
QD 17	373,207.23	935,624.89	2689	Lake: Dong Taijnar Hu	
QD 18	372,727.49	940,331.84	2694	Deep, broad trench near lake Dong Taijnar Hu	Water light-greenish
QD 19	372,024.45	944,412.16	2707	Trench, NW of lake Dabusun Hu	Water dark-green-black, salt production
QD 20	360,023.64	944,827.08	3326	Spring south of Golmud, near QD 21	
QD 21	360,048.66	944,848.76	3276	River: Golmud He, S of Golmud	Water turbid
QD 22	365,842.86	951,528.26	2685	Lake: Dabusun Hu Dabusan Hu	Potassium firm, water light-green, salt crystals at rim and below water
QD 23	372,936.61	952,745.12	3177	Lake: Xiao Qaidam Hu	At shore some vegetation and fly larvae
QD 24	373,215.49	960,049.56	3513	Brook in Jun Ui Mountains	Canyon in red sanstone (Paleocene)
QD 25	372,744.88	953,633.52	3186	Brook flowing in Xiao Qaidam Hu lake	
QD 26	375,605.13	952,250.14	3653	Thermal spring in Qaidam Shan mountains	Nearby white sinter
QD 27	375,604.06	952,249.67	3644	Thermal spring in Qaidam Shan Mountains	
QD 28	375,605.65	952,251.18	3662	Brook uphill of QD 27	Sign. lower temperature
QD 29	375,154.65	951,531.68	3153	Lake: Da Qaidam Hu	Bizarre salt structures 'vulcanos'
QD 30	374,742.15	952,329.42	3282	River into lake Da Qaidam Hu	Much vegetation
QD 31	380,017.28	950,245.26	3243	River: Ige He	
QD 32	371,540.86	953,923.97	3003	River: Quanji	Xitieshan Mountains, Quanji terrane, old water catch- ment
QD 33	400,515.23	944,009.96	1148	Dunhuang, crescent lake	Water-posture artificial
QD 34	371,545.72	953,921.19	3015	Spring in Xitieshan Mountains, near QD 32	

Sample	Northing	Easting	Altitude (m a.s.l.)	Name of location (see Table 1)	Description
QD 2a	375,705.35	914,737.84	2920	Within QD 2	Fine-grained, white crystals
QD 4a	375,430.52	914,955.72	2925	Within QD 4	Coarse-grained, white, cubic crystals
QD 5a	383,624.10	911,416.76	2713	Within QD 5	Very fine-grained, clumpy, white material
QD 5b	383,449.02	911,749.74	2711	2nd pool, close to QD 5	Coarse-grained, white, transparent, cubic crystals
QD 5c	383,449.02	911,749.74	2711	2nd pool, close to QD 5	Rigid, firm, dense, fine-grained material with white and light-brown layers
QD 5d	383,624.10	911,416.76	2713	Within QD 5	Coarse-grained, clumpy, white material
QD 6a	383,007.90	913,414.85	2746	Within QD 6	Fine-grained, light-orange crystals with larger, transparent, cubic crystals
QD 6b	383,007.90	913,414.85	2746	Within QD 6	Fine-grained, light-orange crystals with larger, transparent, cubic crystals
QD 7a	382,845.26	913,719.17	2761	At edge of QD 6	Very fine-grained, white, transparent material
QD 7b	382,900.54	913,718.72	2764	Near QD 6, in Dalangtan playa	Rigid, firm, dense, coarse-grained material. Crystals: white, transparent
QD 8	375,135.32	914,422.60	2920	Near mine in Mangnai	Crystals light-yellow, transparent, thick-flaky gypsum crystals
QD 12a	380,756.05	905,603.46	2859	Within QD 12	Fine-grained, white cubic crystals
QD 13a	381,314.82	904,943.53	2901	Rim of QD 13	Ochre material at mud-dome, on top small transparent, ochre flakes
QD 14	381,408.62	905,145.47	2949	Near lake Gas Hure Hu	Heaped up 'salt', clumpy, hard, white partly yellowish coarse-grained
QD 16	373,210.39	935,824.85	2696	Near lake Dong Taijnar Hu	Lithiumcarbonat production firm, sample stalky, clear
QD 17a	373,207.23	935,624.89	2689	Rim of QD 17	White, clumpy, hard porous material
QD 17b	373,207.23	935,624.89	2689	Rim of QD 17	Yellow-brownish hard material
QD 18a	372,727.49	940,331.84	2694	Within QD 18	Coarse-grained, white, cubic crystals
QD 19a	372,024.45	944,412.16	2707	Within QD 19	Large, cubic, clear crystals
QD 26a	375,606.41	952,249.91	3661	Near QD 26	White powder sinter
QD 29a	375,154.65	951,531.68	3153	Within lake QD 29	At bottom of QD 29 coarse-grained, white crystals, few yellowish crystals
QD 29b	375,210.99	951,553.93	3157	Near QD 29, 'salt for sale'	Agglutinated, former coarse-grained, white material, porous like sponge

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springs, or small lakes (16 samples) and salt lake brines (14 samples). Care was taken, to collect the samples preferably in the center of the rivers or brooks, while brine samples had to be taken in most cases near the shore of the salt lakes. The evaporate samples were mainly taken within (below water table) or directly at the shore of the salt lakes; however, some had to be taken in close vicinity of the salt lakes, when subsoil was not passable. At some localities several evaporite samples were collected if they differed in color or in macroscopic crystal shape.

Water samples for chemical analyses were collected in 500 ml acid-cleaned polyethylene bottles. Care was taken to fill the bottles completely to avoid gas exchange. Temperature, pH-value, EH-value, and electrical conductivity (EC) were measured in the field with the following devices: HANNA HI 98150, WTW Multi 340i, WTW TetraCon 325. Alkalinity was determined by titration with 0.1 N HCl.

The chemical composition of the water was determined in the laboratory of the IMPG (Institute of Mineralogy, Petrology, and Geochemistry, University of Freiburg) after filtration through cellulose acetate membrane filters with a 0.45  $\mu$ m pore size, using ion chromatography for the anions (Dionex 120), atomic absorption spectroscopy for cations (AJ Vario 6), and photometry for Si and B. Detection limits for Ca, Sr, F, Br, NO<sub>3</sub> are 50  $\mu$ g/l, for Na, K, Li, Mg 20  $\mu$ g/l, for Cl 500  $\mu$ g/l, and for SO<sub>4</sub> 800  $\mu$ g/l. The total of dissolved solids (TDS) has been calculated from the analyzed concentrations.

Evaporite mineral samples were ground carefully. To obtain a homogeneous powder. 1 g of the resulting powder was mixed with 100 ml bi-distilled water and stirred until entire dissolution. The chemical composition of the solution was analytically determined using the same methods as for the water samples and is described above.

The evaporite minerals have been identified by powder X-ray diffraction (XRD) measurements with a Bruker D8 Discover at IMPG. System parameters were a Cu anode at 40 kV and 40 mA and a silicon strip detector (lynxeye XE-T). The scanning range was  $2-82^{\circ}$ , with a step size of 0.01° 2  $\mu$ .

The saturation state (SI) of the waters with respect to selected minerals was calculated with the computer program PHREEQC (Parkhurst & Appelo, 1999) using the Pitzer thermodynamic data set at 25 °C for brines (Pitzer, 1991) and for low TDS waters the LLNL thermodynamic data set (Lawrence Livermore National Laboratory, 2020) was used. The PHREEQC code was as well used to determine activities and to carry out theoretical calculations of SI of gypsum and anhydrite as a function of NaCl concentration. Additionally, the online version of SUPCRTBL (Zimmer et al., 2016) of the software package SUPCRT (Johnson et al., 1992) combined with data of Holland and Powell (2011) was used to calculate thermodynamic properties of reactions, like equilibrium constants, in dependence of T, p. The computer program SNORM (Bodine & Jones, 1986) was used to determine an equilibrium assemblage of salts persisting with evaporation to dryness from waters.

### 4 Results

### 4.1 Water composition data

Water composition data can be found in Table 3 for brines and in Table 4 for fresh and slightly brackish water

(termed low TDS waters throughout the text). The concentration data refer to amount of solute per kg solution. The overall charge error of the analyses expressed as electroneutrality (EN) was calculated from (meq L<sup>-1</sup>): EN = 100 \* (sum of cations - sum of anions)/(sum of cations + sum of anions), concentrations in meq L<sup>-1</sup>. EN forall 30 samples was significantly < 4%. In one sample theEN was 5%. This sample had a very low TDS and thus,small analytical errors may lead to high EN values.

The brine samples show TDS between 33 g/kg and 351 g/kg and the low TDS water samples between 0.3 g/kg and 5.2 g/kg. The pH varies from 5.6 to 9.3, with the exception of QD 6 (pool in western oilfields of Dalangtan playa) with a pH=2.7. The corresponding sample of this brine pool showed also the highest TDS (351 g/kg) of all samples. Generally, pH of the brines (median 7.20) is lower than that of the low TDS waters (median 8.25). Temperature of the water samples was generally surprisingly high (10–30 °C), especially in shallow lakes and was probably caused by the high air temperature of 30 °C and higher, around noon time. Very high water temperatures of C. 70 °C were measured in the thermal springs of Da Qaidam (Stober et al., 2016).

Table 3 Composition of salt lake waters (brind	es)
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	QD1	QD2	QD3	QD4	QD5	QD6	QD12	QD13	QD17	QD18	QD19	QD22	QD23	QD29
Temp (°C)	16.1	26.4		21.3	15.5	31.5	21.1	27.2	29	21		21.5	18.4	17.5
рН	8.96	7.08	6.32	6.42	7.41	2.7	6.8	7.2	8.28	7.31	6.81	5.63	8.36	7.88
Density (g/ml)	1.0272	1.1086	1.2108	1.2068	1.2480	1.3167		1.2147	1.1098	1.2176	1.2025	1.2912	1.1030	1.1255
Ca	72.5	3638	4179	6211	172	66,580	145	211	298	412	1577	151	354	177
Mg	1958	1617	696	2170	11,570	16,070	29,090	25,130	2173	9912	36,380	73,320	2249	8109
Na	8313	95,810	96,210	93,760	83,980	9792	58,880	51,680	51,020	84,940	36,840	4686	41,700	57,920
К	461	690	969	626	2929	19,170	4003	3460	1419	3976	3571	5532	623	2753
Sr	20.84	82.4	115.3	132.3	15.97	4611	3.47	3.06	8.34	21.5	36.4	8.21	6.8	9.63
Li	1.07	10.05	19.22	13.42	11.8	1055	22.03	23.9	54.3	80.58	71.79	178.3	5.57	67.33
HCO <sub>3</sub>	6.102	1220.4	9153	9153	9153	6.102	12,204	12,204	6102	4271.4	3051	6.102	6.102	6.102
SO <sub>4</sub>	11,895	1414	1224	1053	37,237	73	40,465	47,440	6076	13,276	2710	5787	22,559	15,621
Cl	10,398	157,086	154,144	155,407	132,375	220,944	141,302	112,846	78,535	152,221	168,879	225,620	55,356	103,954
F	9.79				9.19									13.1
Br	4.39	69.2	65.4	145	18.3	244	30.7	25.7	4.96	8.36	17	24.6	16.1	44.2
J		23.8	23.7	40.7		33.7								
NO <sub>3</sub>		11.6			103				6.76	22.7	36.9	50.2		
SiO <sub>2</sub>					10.3				6.6				1.3	8.7
HBO <sub>2</sub>	101	1125	1388	2139	851	12,788	432	327	312	519	396	692	781	1500
TDS	33,241	262,797	268,187	270,850	278,436	351,367	286,577	253,351	146,016	269,661	253,566	316,055	123,658	190,183
Cl/Br (mass)	2369	2270	2357	1072	7234	906	4603	4391	15,834	18,208	9934	9172	3438	2352
Na/Cl (mol)	1.23	0.94	0.96	0.93	0.98	0.07	0.64	0.71	1.00	0.86	0.34	0.03	1.16	0.86
Ca/SO <sub>4</sub> (mol)	0.0146	6.1658	8.1822	14.1355	0.0111	2186	0.0086	0.0107	0.1175	0.0744	1.3946	0.0625	0.0376	0.0272
log m <sub>SiO2</sub>					- 3.77				- 3.96				- 4.66	- 3.84

Sampling localities given on Table 1 and Fig. 2

Blank values indicate concentration below detection limit after dilution of the sample for analysis. Solute concentrations in mg/kg

	QD9	QD10	QD11	QD15	QD20	QD21	QD24	QD25	QD26	QD27	QD28	QD30	QD31	QD32	QD33	QD34
Temp (°C)	12.5	17.2	12					12.4	71.8	67	22	18	9.5			
РН	8.34	9.34	8.12	8.22	8.24	8.25	8.08	8.26	7.79	7.91	8.35	8.43	8.55	8.36	8.51	7.79
Ca	46.70	14.70	36.80	39.30	46.70	44.40	47.50	37.60	25.80	26.00	49.70	118.00	36.50	62.40	36.80	231.00
Mg	44.00	66.50	498.00	18.67	27.90	18.95	19.04	11.41	0.15	0.29	8.68	314.90	8.37	45.40	48.90	189.00
Na	117.00	231.00	1007.00	112.00	48.00	37.00	750.00	31.30	337.00	330.00	158.00	776.00	23.20	646.00	72.30	921.00
$\mathbf{x}$	5.47	7.97	17.40	11.80	3.07	4.67	4.51	1.64	13.50	13.70	7.43	50.70	1.25	8.64	5.97	15.40
Sr	0.69	0.45	3.61	0.71	0.61	0.49	2.34	0.29	0.34	0.33	0.46	2.16	0.24	4.29	1.00	7.44
Li	0.05	0.06	0.44	0.41	0.04	0.03	0.06	0.02	3.23	3.12	1.21	3.47	0.04	0.33	0.31	0.35
HCO <sub>3</sub>	336.2	297.8	945.8	186.1	246.5	201.4	256.3	116.6	206.9	207.5	170.9	396.6	123.9	215.2	213.4	211.7
$SO_4$	119.0	209.0	639.0	81.0	54.3	45.6	191.0	32.6	191.0	190.0	117.0	448.0	28.1	435.0	152.0	1191.0
C	129.0	269.0	2055.0	163.0	65.5	52.1	1003.0	61.9	348.0	341.0	175.0	1813.0	44.8	815.0	88.4	1517.0
ц	1.47	1.69	0.78	0.56	0.36	0.26	1.20	0.31	8.88	8.93	4.51	8.69	0.32	2.69	0.77	3.30
Br	0.05	0.05	0.68	0.05			0.46	0.04	0.59	0.61	0.26	2.33		0.31	0.03	0.14
NO <sub>3</sub>	0.56	0.10		3.99	3.70	3.10	6.56	2.61					3.68			
SiO <sub>2</sub>									83.76	85.01	39.76			8.19	11.57	10.57
HBO <sub>2</sub>	1.34	2.55		5.69	1.08	0.77	5.99	2.41	165	165	68.8	127	2.63	10.4	2.24	3.06
TDS	802	1101	5205	623	498	409	2288	299	1540	1519	871	4061	273	2254	634	4301
Cl/Br (mass)	2580	5380	3022	3260			2180	1548	590	559	673	778		2629	2947	10,836
Na/Cl (mol)	1.32	0.76	1.06	1.13	1.09	1.15	0.78	1.49	1.49	1.39	0.66	0.80	1.22	1.26	0.94	
Ca/SO <sub>4</sub> (mol)	0.94	0.17	0.14	1.16	2.06	2.33	0.60	2.76	0.32	0.39	1.02	0.63	3.11	0.34	0.58	0.47
log m <sub>sio2</sub>									- 2.947	- 2.952	- 3.279					
Sampling localiti Blank values: Bel	ies given on f ow detectior	ig. 2 and Tab I limit. Solute	le 1 concentratior	ns in mg/kg												

Table 4 Composition of low TDS waters, inflows to the Qaidam Basin

![](_page_9_Figure_2.jpeg)

Fig. 4 Main composition of the salt lakes (brines) (A) as well as fresh waters and slightly brackish waters (low TDS waters) (B), presented as pie charts in eq.%. TDS of each sample is given at the bottom of the corresponding pie chart. Location of the samples is presented in Fig. 2

### 4.1.1 Main water components

Both groups, the brines and the low TDS waters have a remarkable variability in hydrochemical composition (Fig. 4, Tables 3, 4).

Most brines contain more alkalis than alkaline earths. Also, the majority of the brines show higher Mg than Ca concentrations (mg/kg). Generally, Cl is the main anion in all brines (Table 3).  $SO_4$  showed second highest values (up to 47,440 mg/kg). HCO<sub>3</sub> was usually extremely low and in several salt lakes at detection limit.

The normalized main brine components (eq.%) of the salt lakes (Fig. 4A) are arranged roughly from W to E. The distinct composition patterns show that characteristic water types are present in the lakes. The key pattern is that of Na-Cl brine. Some lakes contain a significant portion of Mg, whilst other lakes contain a considerable quantity of  $SO_4$ . In more detail, the data (Fig. 4A) show that salt lakes in the west QD 13, QD 12, and QD 1 (Gas-Hure Hu, Suhai Hu) are dominated by sodium (Na), magnesium (Mg), and chloride (Cl), whereas QD 1 has a significantly higher sulfate content (SO<sub>4</sub>) and generally the highest SO<sub>4</sub> content of all investigated salt lakes. Further to the east and in the central areas of the Qaidam Basin water composition in the salt lakes QD 2, QD 3, QD 4, QD 17, and QD 18 (lakes between Shulzhan and Mangnai, Dong Taijnar Hu) changes to Na-Cl brines. In contrast, the eastern lakes QD 19 and QD 22 (Dabusun Hu area) are dominated by Mg and Cl. Farther to the north the lakes QD 23 and QD 29 (Xiao Qaidam Hu, Da Qaidam Hu) show a Na-(Mg)-Cl water-signature, guite similar to that of the salt lake QD 5 (NE Mangnai Zhen). Remarkably different is the chemical composition of the salt lake QD 6 (western oilfields), dominated by Ca, Mg, and Cl with extraordinarily low Na and enhanced potassium (K) and lithium (Li).

The low TDS water samples from rivers, brooks, cold and hot springs, and from lakes with outflows to the Qaidam Basin differ drastically from the patterns of the brines (Table 4). Figure 4B shows the main water components (eq.%) of the inflows into the basin, approximately arranged from SW to NE. The inflows in the west contain relatively high Na and Mg with varying amounts of Cl, HCO<sub>3</sub>, and SO<sub>4</sub>. Inflows from the southern Kunlun Mountains QD 15, QD 20, and QD 21 (Narin Gol, spring into Golmud He, Golmud He) are typically Ca-Na-Mg-HCO<sub>3</sub>-Cl rich waters. In contrast, the water composition of the influx from eastern and northeastern direction QD 32, QD 34, and QD 24 (near Xitieshan) shows, that Na and Cl dominate with variable amounts of SO<sub>4</sub>, but generally low HCO<sub>3</sub>. The major components of the brook water (QD 25) flowing into Xiao Qaidam Hu and of the river water Ige He (QD 31) are similar. Both waters are characterized by high Ca and HCO<sub>3</sub> contents. The thermal springs in the NW of Da Qaidam (QD 26, QD 27) are rich in Na and Cl with significant amounts of  $SO_4$  and  $HCO_3$  and extremely low Mg contents. Chemical composition of the nearby thermal brook (QD 28) shows additionally enhanced Ca and Mg contents. However, all three thermal waters contain a significant amount of fluoride (F) and lithium (Li) in contrast to the other low TDS waters.

The general composition patterns (Fig. 4) of the two groups of water (brines and low TDS water) are distinctly group specific despite the large group internal variability. Brines are generally dominated by chloride with a few localities, where also substantial sulphate is present. Low TDS water has typically comparable amounts of the anions Cl, SO<sub>4</sub> and HCO<sub>3</sub>. Brines contain very little Ca (with one exception) and the absolutely leading cation is Na. However, in a few brines the Mg concentration is significant or even dominant. In contrast, many low TDS waters contain a significant amount of Ca, but Na is also the prime cation and several contain also a high proportion of Mg.

### 4.1.2 Trace components in water

Strontium (Sr), Rubidium (Rb), Lithium (Li), Fluoride (F), Bromide (Br), Iodine (J), Nitrate (NO<sub>3</sub>), and Boron (B) were analyzed in all water samples (Tables 3, 4). However, not all elements could always be measured in the samples, either because TDS in the salt lakes was too high with respect to the relatively low concentration of trace elements or because TDS was too low (rivers, brooks, thermal springs) and thus the trace elements were below detection limit. Generally, Rb was below detection limit in all samples.

Sr in brines varies between 3.1 and 132.3 mg/kg. Exceptionally high Sr was found in QD 6 (4,611 mg/kg). Li ranges from 1.1 mg/kg in QD 1 to 178.3 mg/kg in QD 22. However, the highest Li concentration was observed in QD 6 with 1,055 mg/kg. F could be determined only in three salt lakes (QD 1, QD 5, QD 29) with concentrations around 10 mg/kg and J only in four salt lakes (QD 2, QD 3, QD 4, QD 6) varying between 23.7 and 40.7 mg/kg. NO<sub>3</sub> could be analyzed in 6 salt lakes with concentrations up to 103 mg/kg. Br and B was detected in all salt lakes with Br-values between 4.4 mg/kg (QD 1) and 244.0 mg/kg (QD 6), resp. B-values between 25 mg/kg (QD 1) and 3,155 mg/kg (QD 6) (Table 3).

In low TDS waters Sr varies generally between 0.24 mg/kg (QD 24) and 7.44 mg/kg (QD 34) and Li between 0.02 mg/kg (QD 25) and 3.47 mg/kg (QD 30). In most samples Li is far below 0.50 mg/kg, only the thermal waters (QD 26, QD 27, QD 30) showed significantly higher Li concentrations (3.1–3.5 mg/kg). F could be analyzed in all waters reaching concentrations between

0.26 mg/kg (QD 21) and 8.93 mg/kg (QD 27). However, J was below detection limit in all samples. NO<sub>3</sub> was found in 7 waters with 0.1 mg/kg (QD 10) up to 6.56 mg/kg (QD 24). Nearly all water samples contained Br and B. Br varied between 0.03 mg/kg (QD 33) and 2.33 mg/kg (QD 30) and B between 0.19 mg/kg (QD 21) and 40.7 mg/kg in the two thermal springs QD 26 and QD 27 (Table 4).

### 4.2 Evaporate minerals

Evaporite mineral samples differ in shape, grain size, and color. First, they were investigated macroscopically and the main findings are summarized in Table 2. The first inspection showed that the samples are very different in composition (Fig. 5) and motivated further examinations.

Water compositions gained from dissolved powders of the evaporite samples have been converted into molar equivalents of the components in each solution. These data were used to determine the percentage of pure evaporative minerals in each sample, ignoring carbonates. Only simple pure evaporite minerals, e.g. halite (NaCl), bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O), or sylvite (KCl), were considered and the variation of hydration H<sub>2</sub>O ignored. Thus, the possible presence of more complex salts, e.g. carnallite (KMgCl<sub>3</sub>· $6H_2O$ ) or kainite (KMgClSO<sub>4</sub>· $3H_2O$ ), is disregarded. This effect of the computational procedure should be kept in mind when examining the data in Table 5. However, the data in Table 5 can easily be converted to more complex salts if there is evidence for their presence. The data show a widely varying range of salt mineral composition. The X-ray analysis of the salt samples confirmed these results and often showed a dominating halite pattern. Dolomite, gypsum and celestine have been identified in QD 8 (Fig. 5D), gypsum is present in QD 13 together with halite. QD 2, QD 4, and QD 6b contain a borate mineral, probably ulexite.

The halite salt crusts in the Qaidam Basin typically contain admixed clay minerals, iron oxides, gypsum, and other evaporite minerals. These impurities occur as separate minerals between halite crystals or as inclusions in the crystals together with fluid inclusions. Detailed analyses of halite often show small amounts of Ca, K, Mg,  $SO_4$ , or Br (Chang et al., 1998; Golding & Phaup, 1936; Johnson, 1970; Johnstone, 1954; Junner & James, 1947;

![](_page_11_Figure_8.jpeg)

Fig. 5 Examples of evaporite mineral samples taken within, at the shore, or nearby salt lakes in the Qaidam Basin. A QD 13a small ochre gypsum flakes covering the total sample, B QD 18a large halite crystal cubes, C QD 16 fibrous form of halite in soil environment, D QD 8 large gypsum crystals partly covered with a white powder of dolomite (and celestite)

Salt minerals	Formula <sup>a</sup>	QD2a	QD4a	QD5a	QD5b	QD5c	QD5d	QD6a <sup>2</sup>	QD6b <sup>2</sup>	QD7a	QD7b
Halite	NaCl	98.8	99.4	79.2	98.1	61.1	94.9	88.9	90.5	79.7	95.0
Anhydrite, Gypsum	CaSO <sub>4</sub>	0.2	-	0.7	0.5	34.7	0.2	-	-	0.1	2.4
Kieserite, Epsomite	MgSO <sub>4</sub>	0.2	-	4.1	0.8	0.1	3.5	-	-	0.1	0.2
Bischofite	MgCl <sub>2</sub>	-	0.2	-	-	1.7	-	2.0	1.5	-	-
Sylvite	KCI	-	0.0	-	-	2.5	-	4.4	3.8	-	-
Thenardite	Na <sub>2</sub> SO <sub>4</sub>	-	-	15.8	0.5	-	1.0	-	-	20.0	2.3
Arcantite	K <sub>2</sub> SO <sub>4</sub>	-	-	0.2	0.1	-	0.4	-	-	0.1	0.1
Sellaite	$MgF_2$	-	-	-	-	-	-	-	-	-	-
Antarcticite	CaCl <sub>2</sub>	0.9	0.4	-	-	-	-	4.6	4.2	-	-
Nitromagnesite	Mg(NO <sub>3</sub> ) <sub>2</sub>	-	-	-	-	-	-	-	-	-	-
Total		100.1	100.0	100.0	100.0	100.1	100.0	99.9	100.0	100.0	100.0
Salt minerals	Formula <sup>a</sup>	QD8 <sup>3</sup>	QD12a	QD13a	QD14	QD16	QD17a	QD18a	QD19a	QD29a	QD29b
Halite	NaCl	6.8	95.0	31.5	36.1	99.2	97.7	98.8	99.4	98.1	99.5
Anhydrite, Gypsum	CaSO <sub>4</sub>	66.7	0.5	41.1	0.1	-	1.8	-	0.1	0.8	0.2
Kieserite, Epsomite	MgSO <sub>4</sub>	-	0.8	6.6	19.6	-	-	-	-	-	-
Bischofite	MgCl <sub>2</sub>	2.1	3.6	19.3	30.0	0.3	0.1	0.9	0.3	0.9	0.1
Sylvite	KCI	0.2	0.2	1.1	14.2	0.1	0.1	0.2	0.1	0.1	0.1
Thenardite	Na <sub>2</sub> SO <sub>4</sub>	-	-	-	-	-	-	-	-	-	-
Arcantite	K <sub>2</sub> SO <sub>4</sub>	-	-	-	-	-	-	-	-	-	-
Sellaite	MgF <sub>2</sub>	-	-	0.1	-	-	-	-	-	-	-
Antarcticite	CaCl <sub>2</sub>	24.2	-	-	-	0.5	0.3	0.1	0.1	0.1	0.1
Nitromagnesite	Mg(NO <sub>3</sub> ) <sub>2</sub>	-	-	0.3	-	-	-	-	-	-	-
Total		100.0	100.1	100.0	100.0	100.1	100.0	100.0	100.0	100.0	100.0

Table 5 Minerals in salt samples recalculated from leached salts

Carbonate minerals not considered

<sup>a</sup> Anhydrous composition

<sup>b</sup> Small amounts of insoluble residues present

<sup>c</sup> QD8 contains dolomite and calcite. Sample lable in italics: analyzed samples were close to pure halite with very low impurities

Wells, 1937). Fluid inclusions in halite contain brines often enriched in K,  $SO_4$ , Ca, and Mg (Galamay et al., 2020). In celestite, Ba is the principal substituent. Gypsum shows generally only minor compositional variations. Small amounts of Sr and Ba may occur replacing Ca (Chang et al., 1998). Consequently, minor, pure evaporite mineral components of samples in Table 5 may be caused by impurities in the main minerals, mainly from fluid and solid inclusions in halite.

### **5** Discussion

### 5.1 Saturation states of the collected water samples

Mineral precipitation also depends on the degree of saturation of the water. If water is supersaturated with respect to a certain mineral, it has the potential that this mineral precipitates. To better understand the possible origin and genesis of the different evaporite minerals in the Qaidam Basin, the degree of saturation (SI=saturation index) of all waters was determined with the code PHREEQC (Parkhurst & Appelo, 1999).

### 5.1.1 Salt lakes

All salt lakes are nearly saturated or supersaturated with respect to celestine  $(SrSO_4)$  (Fig. 6). Generally, the saturation state with respect to minerals increases with rising TDS of the brines. However, supersaturation (SI>0) in high TDS salt lakes is only achieved with respect to celestine (SrSO<sub>4</sub>), anhydrite (CaSO<sub>4</sub>), and gypsum (CaSO<sub>4</sub> $\cdot$ 2H<sub>2</sub>O) (Fig. 6). In contrast, when saturation (SI=0) with respect to halite (NaCl) is reached, the brines will not develop halite supersaturation, even though TDS or Cl concentration continues to increase. In a pure NaCl-solution saturation with respect to halite is achieved with 263 g/kg NaCl (Fig. 6). In salt lakes with higher Cl-concentrations the saturation indices of sylvite (KCl), kainite (KMgClSO<sub>4</sub>·3H<sub>2</sub>O), bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O), and/or carnallite (KMgCl<sub>3</sub>·6H<sub>2</sub>O) continue to increase if the relevant components are available, but saturation (SI=0) with respect to these minerals is not reached, solely in QD 22 for carnallite. In the two salt lakes QD 22 and QD 6 the saturation state with respect

![](_page_13_Figure_2.jpeg)

Fig. 6 Saturation state with respect to selected minerals for the different salt lakes in dependence of TDS. Saturation condition (SI = 0) is shown by the 'light red' line

to glauberite  $[Na_2Ca(SO_4)_2]$  is very low in contrast to saturation in the brines of QD 5 and QD 12. This circumstance is caused by comparatively low SO<sub>4</sub> and Na or Ca concentrations in these salt lakes (Fig. 6).

The salt lakes in the TDS range 250-285 g/kg are saturated with respect to both anhydrite and gypsum (Fig. 6). For brines with TDS < 250 g/kg the saturation index of anhydrite is lower than that of gypsum and for brines with TDS > 285 g/kg it is reversed.

A PHREEQC model (Fig. 7A) shows the SI of gypsum  $(\mathrm{SI}_{\mathrm{Gp}})$  and anhydrite  $(\mathrm{SI}_{\mathrm{Anh}})$  as a function of the NaCl concentration expressed as  $\mathrm{SI}_{\mathrm{Hl}}$ . The simulation begins with a solution saturated with respect to gypsum (red curve on Fig. 7A). If NaCl in the brine is low  $(SI_{Hl} < -1.0)$  $SI_{Gp}$  exceeds  $SI_{Anh}$ . For high NaCl in the brine ( $SI_{Hl} > 0.0$ ),  $SI_{Anh}$  (blue curve Fig. 7A) exceeds  $SI_{Gp}$ . At  $SI_{Hl} \sim -0.25$ the solution is approximately saturated with respect to both, anhydrite and gypsum (Fig. 7A). The 'black dashed' curve (Fig. 7A) shows the activity of water  $(a_{H2O})$ as a function of  $SI_{Hl}$ . In diluted, low TDS water  $a_{H2O}$ approaches 1. In concentrated brines  $a_{H2O}$  is significantly below 1. At about  $a_{H2O} \sim 0.8$  and lower the saturation index of gypsum  $(SI_{Gp})$  decreases drastically. Even if the activity of water is only slightly below  $a_{H2O} = 1.0$  the saturation index of anhydrite (SI<sub>Anh</sub>) increases ('blue curve' in Fig. 7A). Thus, at very high salinities not gypsum but anhydrite will be the stable phase precipitating from the brine. In more complex systems, common in nature,

further solutes may also influence the solubility of gypsum/anhydrite considerably.

Most salt lakes in the Qaidam Basin are saturated or supersaturated with respect to both gypsum and anhydrite (Fig. 6). The salt lakes with  $SI_{Gp}$  and  $SI_{Anh} \ge 1$ (Fig. 7B) follow a trend line with a negative slope (m = -1). Along this trend line brines with low Ca concentration are rich in SO4 and brines rich in Ca have a low SO<sub>4</sub> content, whereas the supersaturated brine QD 6 (Fig. 7B) has the lowest  $SO_4$  and highest Ca content. The intersection of the 'red' trend line with the 'blue' gypsum/anhydrite dissolution line (Fig. 7B) corresponds to a saline solution containing about 5.1 g/ kg calcium sulfate ( $\sim 75 \text{ meg/kg}$ ). This is close to the computed model composition of 4.9 g/kg CaSO<sub>4</sub> for a brine with  $SI_{Anh} = 0$  and  $SI_{Hl} = -0.13$ . Higher gypsum/ anhydrite solubility (5.9-6.3 g/kg) has been reported in brines containing also high concentrations of MgSO<sub>4</sub> (Mel'nikova & Moshkina, 1973, in: Klimchouk, 2015).

In Fig. 7C the activity coefficients of calcium  $(a_{Ca})$  are plotted versus the activity coefficients of sulphate  $(a_{SO4})$ , which are calculated for all salt lake brines with the code PHREEQC (Parkhurst & Appelo, 1999). Brines saturated (or slightly under- or supersaturated) with respect to anhydrite follow a trend line with a negative slope (m = -1) ('red line' in Fig. 7C). The related solubility reaction of anhydrite is given in Eq. (1), where K is the equilibrium constant of this chemical reaction at

![](_page_14_Figure_2.jpeg)

**Fig. 7** Calcium and sulfate content in waters and relationship to saturation with respect to anhydrite and gypsum. Saturation states and activities are calculated with PHREEQE (Parkhurst & Appelo, 1999). **A** Saturation indices (SI) of anhydrite and gypsum for a Ca-SO<sub>4</sub>-saturated solution versus halite saturation at 22 °C and 1 bar. With increasing Na-Cl-content the activity of water decreases (standard state: pure H<sub>2</sub>O at P and T). **B** Calcium and sulfate content (meq/kg) of all waters. 'Red numbers' refer to the individual salt lakes. All brines saturated or slightly supersaturated with respect to gypsum and/or anhydrite (legend see **C**) follow the 'red trend line'. **C** Activity of sulphate (log a<sub>SO4</sub>) versus activity of calcium (log a<sub>Ca</sub>). The 'blue dotted line' shows the gypsum/anhydrite dilution line. The symbols indicate SI<sub>Anh</sub> of the brines (CaSO<sub>4</sub>). The 'red line' is the trend line for brines saturated with respect to anhydrite. The salt lake brines define the equilibrium constant K<sub>Anh</sub> of anhydrite dissolution at 22 °C and 1 bar (Eq. 2). The standard state refers to unit activity for a hypothetical one molal solution at infinite dilution

constant p-T conditions, i.e. K is the value of its reaction quotient at chemical equilibrium (e.g. Drever, 1997).

$$\log K = \log a_{Ca} + \log a_{SO4} - \log a_{Anh}.$$
 (1)

Equation 2 translates to the straight line in Fig. 7C:

$$\log a_{SO4} = -1 \log a_{Ca} + (\log K + \log a_{Anh}). \quad (2)$$

For pure anhydrite log  $a_{Anh}=0$  (using an appropriate standard state). The intersection with the trend line at log  $a_{Ca}=0$  results from Eq. (2) in log K = - 4.3 (Fig. 7C). Impure anhydrite would result in lower K values, whereas supersaturation with respect to anhydrite indicates higher K values. The computed field derived anhydrite saturation conditions were verified by using the code SUPCRTBL (Johnson et al., 1992; Zimmer et al., 2016) and the anhydrite data of Holland and Powell (2011). At atmospheric pressure and T=22 °C the log equilibrium constant is equal to log K=-4.30, confirming the field derived value.

### 5.1.2 Low TDS waters

Inflows into the Qaidam Basin are supersaturated with respect to calcite  $(CaCO_3)$ , aragonite  $(CaCO_3)$ , and dolomite  $(CaMg(CO_3)_2)$ . The saturation index of dolomite  $(SI_{Dol})$  is generally higher than that of calcite  $(SI_{Ca})$ , with the exception of the thermal spring QD 26. All low TDS waters are saturated or supersaturated

with respect to strontianite (SrCO<sub>3</sub>). The western inflows (QD 9, QD 10, QD 11), the NE inflow QD 30, and QD 33 (Dunhuang crescent lake) are supersaturated with respect to magnesite (MgCO<sub>3</sub>), as well as QD 20 (Golmud spring) in the south and QD 32, QD 34 (Xitieshan river and spring). Some of these inflows into the Qaidam Basin (QD 9, QD 10, QD 11, QD 30, QD 33) are also saturated or supersaturated with respect to huntite [CaMg<sub>3</sub>(CO<sub>3</sub>)<sub>4</sub>].

Despite  $SI_{Dol} > SI_{Cal}$ , dolomite may not precipitate from the inflow waters due to kinetic causes, including very high activation energy for nucleating crystals of dolomite (Blatt & Tracy, 1996). Although, some of the low TDS waters are saturated or supersaturated with respect to magnesite, magnesite also may not precipitate, because the 'dehydration barrier' of aqueous Mg hampers the low-temperature formation of anhydrous magnesium carbonate (Deelman, 1999; Lippmann, 1973). However, low-temperature formation of magnesite, dolomite, and huntite may occur, but it requires fluctuations in pH and in other parameters (Deelman, 1999). Such fluctuations can be related to seasonal changes in water composition, daily variations caused by the photosynthesis of e.g. algae. Thus, the internal parts of the Qaidam Basin are potentially appropriate for Mg-carbonate formation from low TDS waters.

Some of the NE inflows, like the thermal waters (QD 26, QD 27, QD 28) and QD 34, are only slightly undersaturated with respect to fluorite (CaF<sub>2</sub>) and one sample (QD 30) is fluorite saturated. All other low TDS waters are considerably undersaturated with respect to fluorite. QD 30 is the only water sample saturated with respect to sellaite (MgF<sub>2</sub>).

In all thermal waters (QD 26, QD 27, QD 28) and some cold water samples (QD 32, QD 33, QD 34) dissolved silica concentration was measured (Table 3). These waters are all supersaturated with respect to quartz (SiO<sub>2</sub>) and undersaturated with respect to amorphous silica. Thus, these waters have no tendency to form silica sinter and all silica remains in solution.

All inflows are undersaturated with respect to gypsum (CaSO<sub>4</sub>·2H<sub>2</sub>O) and anhydrite (CaSO<sub>4</sub>). Calcium and sulfate in low TDS waters are positively correlated, i.e. with increasing calcium concentration sulfate increases as well (Fig. 7B).

### 5.2 Halogens in water samples

Halogens in waters and distinct halogen ratios can give an important contribution to the origin of the waters and they can help to get closer to a potential connection between the inflows into the Qaidam Basin (low TDS waters) and the salt lakes. The fact that certain elements in the salt lakes showed elevated concentrations, which usually occur only in traces, was also a reason to take a closer look at halogen ratios.

### 5.2.1 Chloride and sodium

Seawater (SMOW) has a Na/Cl mol ratio of about 0.86, whereas halite's mol ratio is 1.0. In most salt lakes the mol ratio of Na/Cl varies between 0.64 and 1.23 (median Na/Cl = 0.89). However, in two brines the Na/Cl ratio is extremely low (QD 22: Na/Cl=0.03, QD 6: Na/Cl=0.07) (Fig. 8). The two salt lakes have also the highest TDS (>300 g/kg). Brines, saturated with respect to halite (QD 2, QD 3, QD 4, QD 12, QD 18), are close to halite saturation in a pure NaCl-solution (Na=Cl=4,497 mmol/ kg solution, or 263 g/kg NaCl), despite additional other ingredients (Fig. 8). Also, the two salt lakes QD 6 and QD 22 with very low Na/Cl ratios are saturated with respect to halite. Their position on the Na vs Cl diagram (Fig. 8) relates to the presence of large amounts of other cations balanced with chloride (Ca in QD 6 and Mg in QD 22: Fig. 4A). None of the salt lakes is supersaturated with respect to halite.

The mol ratio of Na/Cl in low TDS waters is generally slightly higher than in brines and varies between 0.66 and 1.49 (median Na/Cl=1.14). The thermal springs of Da Qaidam (QD 26, QD 27) have with Na/Cl=1.49 the highest Na/Cl ratio of all water samples (Fig. 8). The relatively high Na/Cl ratio in the thermal springs is caused by plagioclase alteration along the deep circulation flow path in the granitic rocks of the Qilian mountains (Stober et al., 2016) leading i.a. to an enrichment of Na over Cl.

### 5.2.2 Chloride and bromide

Both, chlorine and bromine are important trace elements of magmatic and metamorphic rocks. They can be found in the crystal lattice of hydroxyl minerals like apatite, biotite, or amphibole, along grain boundaries of minerals and in fluid inclusions. Magmatic rocks typically have chlorine-bromine-ratio far below Cl/Br = 100 (e.g. Behne 1953, Correns 1956). Cl/Br ratios reported from waters in fractured granitic rocks are typically 50–100 (Frape & Fritz, 1987; Gascoyne et al., 1987; Savoye et al., 1998; Davis et al., 2001; Stober & Bucher, 2005; Drüppel et al., 2020). In contrast, the Cl/Br ratio of water derived from dissolved halite deposits is on the order of some 1000 (Davis et al., 1998; Stober & Bucher, 1999) and Cl/Br of seawater is 288 (Stumm & Morgan, 1975).

The Cl/Br ratio (mass ratio) in all water samples is much higher than that of seawater and water that has solely interacted with crystalline basement rocks (Fig. 9A).

Most brines are saturated with respect to halite (Fig. 9), i.e. halite precipitates from solution upon further evaporation. Since halite does not incorporate Br in its lattice,

![](_page_16_Figure_2.jpeg)

Fig. 8 Sodium and chloride concentrations (mmol/kg) of all water samples. Saturation with respect to halite for a pure NaCI-solution is shown by the black dashed lines. 'Green dotted line' refers to seawater dilution and evaporation.' Orange line' corresponds to Na and CI contents in water, in which halite is dissolved. 'Blue dotted line' corresponds to Na/CI ratio of 1.49, found in the thermal springs

Br concentration in these brines will increase and the Cl/ Br ratio will decrease. Figure 9A shows the 'halite-saturation-line' for a pure NaCl-solution. This boundary will be followed towards lower Cl/Br as a result of expected increasing residual bromide concentration in the brines during prolonged evaporation.

The brines and most low TDS waters show a clearcut 'halite-signature', with the exception of the thermal waters (QD 26, QD 27, QD 28) and the nearby inflow (QD 30) to the salt lake Da Qaidam Hu. The Cl/Br ratio of these inflows into the Qaidam Basin is relatively low (560–780), but higher than waters originating from basement rocks (Cl/Br < 100) (Fig. 9A).

Stober and Bucher (2016) explained the unusual high Cl/Br ratio of waters originating from crystalline basement of the Qilian mountains with the circumstance that these waters contain a significant amount of dissolved halite due to aeolian deposits from the basin up to steep slopes of surrounding mountains (Fig. 3D). These deposits are rich in halite and other evaporate minerals. For this reason, also the Cl/Br ratio of the hot springs (QD 26, QD 27) is lower compared to the thermal brook water (QD 28) and other low TDS waters with origin in the mountains. Figure 9A clearly shows, that all low TDS waters have a much higher Cl/Br ratio than waters typically found in crystalline basement. They are affected by salt-rich deposits in the surrounding area of the Qaidam Basin, which belongs to the catchment of these rivers, brooks, and springs.

### 5.2.3 Lithium in brines and low TDS waters

Most inflows to the Qaidam Basin showed lithium (Li) concentrations considerably below 0.50 mg/kg; only in the thermal springs (QD 26, QD 27) and in the nearby river QD 30 Li was higher (3.1–3.5 mg/kg). In the large river Nasin Gol (QD 15) a Li concentration of 0.41 mg/kg was measured. Yu et al. (2013) associated the source of lithium in the Nasin Gol River with hydrothermal fields and two active faults in the upper catchment area of the river.

Due to Drüppel et al. (2020) the main Li-sources are phyllosilicates (mainly muscovite, biotite) in granitic rocks. The chemical reactivity of Li is generally very low and the sorption of Li is extremely low, much lower than that of all other cations. For this reason, Li remains preferentially in solution once released from the primary minerals.

During hydrothermal alteration of the granitic rocks in the Qaidam Shan and Kunlun Mountains Li will be released from granitic phyllosilicates leading to an increase in the Li concentration of the subsurface fluids over time (hydrothermal origin). Therefore, the Li

![](_page_17_Figure_2.jpeg)

Fig. 9 Chloride versus bromide (A) and chloride versus lithium (B) concentration (mg/kg) in salt lakes (red dots) and low mineralized waters (blue dots). 'Red' numbers indicate salt lake samples (Table 1). The 'red line' shows saturation with respect to halite for a pure NaCl-solution. A The 'green dotted line' is the seawater dilution/evaporation line (Cl/Br = 288). Seawater data from Stumm and Morgan (1975). 'Red dots' with 'violet edges' indicate brines saturated with respect to halite. Brines with extremely high TDS have as well very high Br contents. B Lithium is enriched in evaporating halite-saturated brines. 'Green arrow' marks overall Li-Cl evolution path. Low TDS water from a hydrothermal environment with crystalline basement alteration show high Li/Cl ratios. The 'violet dotted line' is the trend line as observed in deep thermal waters from the crystalline basement of Central Europe (Black Forest and Upper Rhine Graben)

concentrations in the Da Qaidam thermal springs (QD 26, QD 27) and also in the Nasin Gol river (QD 15) are enhanced. Thus, the observed high Li in certain salt lakes is ultimately derived from crystalline basement rocks in the north (Qaidam Shan Mountains) and south (Kunlun Mountains). Long-term input of Li into the basin via rivers and brooks and high evaporation results in the observed accumulation of lithium in some brines (e.g. QD 29, QD 19).

The Li/Cl-ratio (mass ratio) is in most low TDS waters below Li/Cl=0.01. However, there are some exceptions: the thermal waters (QD 26, QD 27, QD 28) with Li/ Cl>0.04, QD 15 (river Nasin Gol), and QD 33 (crescent lake Dunhuang). All inflows, which are not associated with a significant hydrothermal water component from deep crystalline basement rocks, follow a trend line in a Li vs. Cl diagram (Fig. 9B). However, the thermal waters of Da Qaidam with higher Li/Cl ratios follow the same trend line as observed in deep thermal waters from the crystalline basement of Central Europe (Black Forest and Upper Rhine Graben) (Drüppel et al., 2020) ('violet dotted' trend line in Fig. 9B).

During extended water-loss from the inflows during evaporation both Li and Cl increase in concentration (Stober & Bucher, 1999, 2004). In the downstream salt lakes, during ongoing evaporation the increase of the Cl concentration is drastically slowed down, when halite saturation is reached and halite precipitates. In contrast, Li concentration continues to increase resulting in a characteristic overall Li-Cl evolution path ('green arrow' in Fig. 9B).

A similar development can be observed for potassium (K). The concentration of K increases in halite saturated salt lakes with continued evaporation, while total Cl (NaCl-KCl) experiences only small relative increases (Table 3).

### 5.3 Salt norm calculations of the collected water samples

Generally, water composition data can be converted into an equilibrium solid salt assemblage or 'salt norm' using the code SNORM (Bodine & Jones, 1986). The 'salt norm' can be considered as the equilibrium assemblage of salts persisting with evaporation to dryness (all free water removed) under atmospheric conditions at 25 °C. However, the 'salt norm' is not directly comparable with the saturation state of the solutions (Sect. 5.1), because the code SNORM considers evaporation of the solution, i.e. removing all H<sub>2</sub>O from the sample, but certain similarities can be expected.

The water composition data (Tables 3, 4) were used to calculate the 'salt norm' and results are presented for brines in Fig. 10A and for low TDS waters in Fig. 10B. The computed 'salt norm' minerals can be grouped into 5 classes: Chlorites, (bi)carbonates, sulphates, borates, and fluorides.

In all salt lakes chlorides outweigh the other salt classes and halite (NaCl) is by far the dominating equilibrium salt mineral (Fig. 10A). However, there are three exceptions: the salt lakes QD 6 (western oilfields), QD 22 (Dabusun Hu), and QD 19 (E of Dabusun Hu). Due to the high Ca content in salt lake QD 6 antarcticite (CaCl<sub>2</sub>.6H<sub>2</sub>O) predominates in the norm followed by carnalite (KMgCl<sub>2</sub>.6H<sub>2</sub>O), as a result of the high Mg and K content in the brine, and halite (Fig. 10A). QD 22 and QD 19 contain both abundant Mg, resulting in a considerable amount of bischofite (MgCl<sub>2</sub>.6H<sub>2</sub>O) (Fig. 10A). From enhanced Mg and SO<sub>4</sub> contents in the salt lakes QD 13 and QD 12 (Gas-Hure Hu) bischofite and the sulphate kieserite (MgSO<sub>4</sub>.H<sub>2</sub>O) appear in the 'salt norm'. Significant quantities of other sulphates like bloedite  $(Na_2Mg(SO_4)_2.4H_2O)$  appear in the 'salt norm' of the salt lakes QD 23 (Xiao Qaidam Hu) and QD 1 (Suhai Hu), for the latter brine additionally the sulphate epsomite (MgSO<sub>4</sub>·7H<sub>2</sub>O) is a member of the 'salt norm' assemblage. QD 1 is the only salt lake with a resulting considerable amount of epsomite. Due to the comparatively high Li content in salt lake QD 6, almost 6 mol% lithium-chloride-monohydrate (LiCl.H<sub>2</sub>O) were modeled by SNORM (Fig. 10A; Table 3).

'Salt norm' results of the low TDS waters show as well a dominating impact of halite (Fig. 10B). Water from E of Mangnai (QD 9, QD 10, QD 11) is highest in normative magnesite (MgCO<sub>3</sub>). Comparatively high normative dolomite [CaMg(CO<sub>3</sub>)<sub>2</sub>] resulted for the southern inflows into the Qaidam Basin (QD 9, QD 15, QD 20, QD 21) and for two NE inflows (QD 25, QD 31). Enhanced calcite (CaCO<sub>3</sub>) was modeled for the river Ige He (QD 31) and the thermal brook (QD 28). For the NE inflows (QD 25, QD 30, QD 31) anhydrite (CaSO<sub>4</sub>) is the prime salt modeled by SNORM. QD 11 and QD 30 are the only inflows into the Qaidam Basin, which are expected to produce a considerable amount of bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O) (Fig. 10B).

Thenardite  $(Na_2SO_4)$  (8.1–9.7 mol%) has been predicted for QD 9, QD 10, and the thermal waters (QD 26, QD 27, QD 28) (Fig. 10B). The SNORM models are supported by the occurrence of white sinter material, collected close the thermal spring QD 26, which has been identified as halite (NaCl) and thenardite  $(Na_2SO_4)$  using XRD (Stober et al., 2016). The collected and analyzed white sinter material formed by evaporation of thermal spring water fits ideally the results of the SNORM code.

The minerals pirsonite  $[Na_2Ca(CO_3)_2 \cdot 2H_2O]$ (5.5 mol%), borax  $(Na_2B_4O_7.10H_2O)$  (c. 7.0 mol%), and lithium-fluorite (LiF) (c. 3.5 mol%) were predicted to precipitate from the thermal springs of Da Qaidam (QD

![](_page_19_Figure_2.jpeg)

Fig: 10 'Salt norm' model of the waters computed using SNORM (Bodine & Jones, 1986) and presented as pie-charts, in mol%. A Salt lake brines. B Low TDS waters, inflows into the Qaidam Basin.

26, QD 27). These thermal waters contain comparatively high B (40.7 mg/kg), Li (c. 3.2 mg/kg), and F (8.9 mg/kg) concentrations. However, for the thermal book (QD 28), containing significantly lower B, Li, and F concentrations (Table 3), smaller amounts of borax were predicted. Surprisingly, the river flowing into Da Qaidam Hu (QD 30) contains similar B, Li, and F concentrations as the thermal springs (Table 3), but SNORM models for QD 30 indirite ( $Mg_2B_6O_{11}.15H_2O$ ) instead of borax and lithium-sulfate-monohydrate ( $Li_2SO_4.H_2O$ ) instead of lithium-fluorite.

The results gained using the PHREEQC code are consistent with results of the 'salt norm' calculations. Waters more or less saturated or supersaturated with respect to certain evaporite minerals show when converted by the code SNORM into an equilibrium solid salt assemblage these minerals as well. However, there are some few exceptions, which are most probably caused by different minerals in the data bases of the two computer programs. The collected and analyzed salt minerals (Table 5) support the results as well.

### 5.4 Alteration of low TDS waters by evaporative concentration and precipitation

Hydrochemistry of the low TDS waters will alter by evaporative concentration and precipitation of solids from supersaturated waters. Thus, these waters change considerably along their flow path into and in the Qaidam Basin, both in TDS and water type. Therefore, a hydrochemical connection between low TDS waters and brines might be possible and the goal hereby was to find out, if low TDS waters are potential precursors of the salt lakes. The investigation thereby is based exclusively on the main water components.

Generally, the first minerals precipitating from evaporating low TDS waters are alkaline-earth carbonates, specifically calcite (CaCO<sub>3</sub>). The initial molar (Ca + Mg)/HCO<sub>3</sub> ratio is the key value in the early stages of the evolution of waters (Eugster & Hardie, 1978). Most low TDS waters show (Ca + Mg) > HCO<sub>3</sub> and all are supersaturated with respect to calcite (CaCO<sub>3</sub>), aragonite (CaCO<sub>3</sub>), and dolomite [CaMg(CO<sub>3</sub>)<sub>2</sub>] (Sect. 5.1). The effect of evaporation can best be shown on ternary phase diagrams. Sodium and chloride (and if relevant also potassium) are present in excess.

The ternary Ca-SO<sub>4</sub>-(HCO<sub>3</sub> + CO<sub>3</sub>) diagram shows all collected low TDS waters (Fig. 11A). Chemical divides ('red lines' from calcite to SO<sub>4</sub> [calcite divide] and calcite to gypsum/anhydrite [gypsum divide]) separate waters into three types, based on their compositions as equivalents of Ca, HCO<sub>3</sub> + CO<sub>3</sub>, and SO<sub>4</sub> (Eugster & Hardie, 1978; Spencer et al., 1990; Lowenstein & Risacher, 2008; Deocampo & Jones, 2014):

- Na-(HCO<sub>3</sub>+CO<sub>3</sub>)-SO<sub>4</sub> brines form from waters with equivalents of (HCO<sub>3</sub>+CO<sub>3</sub>)+SO<sub>4</sub>>Ca during precipitation of calcite.
- Waters with (HCO<sub>3</sub>+CO<sub>3</sub>)+SO<sub>4</sub><Ca evolve into Ca-Cl brines lacking SO<sub>4</sub> and (HCO<sub>3</sub>+CO<sub>3</sub>) following precipitation of calcite and gypsum and ultimately produce an antarcticite (CaCl<sub>2</sub>·6H<sub>2</sub>O) evaporite assemblage.
- Waters in the Cl-SO<sub>4</sub> field form Ca-poor, Na-Cl-SO<sub>4</sub>rich brines following precipitation of calcite and gypsum.

All low TDS waters plot with one exception (QD 34) into the Na-(HCO<sub>3</sub>-CO<sub>3</sub>)-SO<sub>4</sub> field (Fig. 11A).  $HCO_3 + CO_3 \ge Ca$  of these waters causes precipitation of calcite (CaCO<sub>3</sub>) upon evaporative concentration. Because Ca and CO<sub>3</sub> are removed in equal molar proportions, these waters evolve to a brine lacking Ca ('blue dotted lines' in Fig. 11A). Evaporation and precipitation of calcite produces alkaline, Ca-poor brines enriched in Na–K-(HCO<sub>3</sub>+CO<sub>3</sub>)-SO<sub>4</sub>-Cl. Thus, NaSO<sub>4</sub>, Na-HCO<sub>3</sub>-CO<sub>3</sub>, and NaCl salts can be expected to precipitate from these inflows.

However, some of the low TDS waters (QD 31, QD 25, QD 28, QD 30, QD 32) are close to or on the 'chemical divide: calcite-SO<sub>4</sub>' and will thus evolve into a brine extremely poor or even lacking  $(HCO_3 + CO_3)$  (Fig. 11A). These waters originate from the mountains (Qaidam Shan and Xitieshan Mountains) surrounding the Qaidam Basin in the NE (Fig. 2). From these inflows  $NaSO_4$  and NaCl salts are expected to precipitate during further evaporation, but Na-HCO<sub>3</sub>-CO<sub>3</sub> salts will not be formed. Precipitation of larger amounts of calcite can be expected in waters with relatively equal and high Ca and high HCO<sub>3</sub> contents like in QD 31 or QD 28 (Fig. 11A). These results are in line with the predicted significant precipitation of calcite from QD 31 and QD 28, using the SNORM code (Fig. 10B). Contrarily, the thermal springs (QD 26, QD 27) do not precipitate calcite as a result of evaporation, using both methods (Figs. 10B, 11A). QD 31 and QD 25 are comparably rich in Ca and lie close to or on the 'chemical divide: calcite-SO<sub>4</sub>' (Fig. 11A). Therefore, a significant amount of anhydrite (or gypsum) precipitation is expected from these waters during further evaporation, fitting with the results of the SNORM calculation (Fig. 10B).

The thermal springs (QD 26, QD 27), the brook originating in the Jun Ui Mountains (QD 24), and the river Nasin Gol from the Kunlun Mountains (QD 15) are expected to evolve into brines enriched in Na–K-SO<sub>4</sub>-Cl with low (HCO<sub>3</sub>+CO<sub>3</sub>), while in the evolving waters QD 20, QD 21 (Kunlun Mountains S of Golmud), and in QD 9, QD 10, QD 11(west of Mangnai) the (HCO<sub>3</sub>+CO<sub>3</sub>)

![](_page_21_Figure_2.jpeg)

**Fig. 11** Ternary diagrams showing all low TDS waters, inflows into the Qaidam Basin (in eq.%). 'Blue numbers' refer to water samples. Chemical divides are presented as 'red lines'. **A** Ternary diagram Ca-  $SO_4$ -(HCO<sub>3</sub> + CO<sub>3</sub>). The 'blue dotted lines' show the theoretical development paths for two typical waters in the Na-HCO<sub>3</sub>-CO<sub>3</sub>-SO<sub>4</sub> field. and the 'green dotted line' for water (QD 34) in the Cl-SO<sub>4</sub> field. No samples plot into the Ca-Cl field. **B** Ternary diagram Mg-SO<sub>4</sub>-(HCO<sub>3</sub> + CO<sub>3</sub>). The pathways ('black lines') result from initial precipitation of calcite, then precipitation of dolomite and magnesite (after Deocampo & Jones, 2014)

content in the evolving brines is expected to be higher (Fig. 11A). Thus, these waters should produce more Na-HCO<sub>3</sub>-CO<sub>3</sub> salts than those from the other source regions.

QD 34 (spring SE of Xitieshan) is the only water plotting into the Cl-SO<sub>4</sub> field (Fig. 11A). Most probably this water will form a Ca-poor, Na-Cl-SO<sub>4</sub>-rich brine following precipitation of calcite and gypsum and then will shift toward the sulfate corner ('green dotted line' in Fig. 11A). Thus, NaSO<sub>4</sub> and NaCl salts may be produced from this inflow in addition to calcite (carbonate).

After calcite  $(CaCO_3)$  and gypsum/anhydrite  $(CaSO_4 \cdot H_2O/CaSO_4)$  deposition the solutions become progressively more concentrated and the Mg/Ca ratio increases. Ultimately, Mg-bearing carbonates (e.g. dolomite, magnesite) will be produced. However, as stated above reliable predictions for the precipitation of Mg-rich carbonates are problematic, because precipitation kinetics and metastable conditions are normally not quantifiable.

The interaction of Mg and Ca can be explored using the ternary  $Mg-SO_4$ -( $HCO_3 + CO_3$ ) diagram (Fig. 11B). The 4 arrows on this diagram represent 4 conditions of varying Ca/Mg rations, which produce Ca-Mg- and Mg-carbonates at different stages (Deocampo & Jones, 2014):

- Low-Ca waters will produce Mg-rich carbonates earlier than high-Ca waters.
- The more calcite can be produced, the farther toward the 'upper left' of the triangle the water will evolve.
- Once calcium has been exhausted, if CO<sub>3</sub> still remains present, magnesite precipitation will drive the fluid away from the Mg–CO<sub>3</sub> axis and determine the Mg/SO<sub>4</sub> or SO<sub>4</sub>/CO<sub>3</sub> ratio in the residual fluid.

Generally, most surface waters have relatively high  $HCO_3$  and low Mg contents, i.e. they plot somewhere in the 'lower right area' of the triangle and will develop during prolonged evaporation towards waters with relatively more  $SO_4$  and/or Mg content, i.e. they evolve along a pathway towards the 'left side' of the triangle, moving away from the calcite or Mg-calcite point (Fig. 11B).

The Mg/Ca ratio (in equivalents) of the low TDS waters is in most samples < 1.0. However, Mg/Ca > 2.0 is measured in 4 probes (QD 10, QD 11, QD 30, QD 33), whereas QD 11 has with Mg/Ca = 22.3 by far the highest ratio. All waters (exception QD 34) show lower Ca than (HCO<sub>3</sub> + CO<sub>3</sub>) contents (Fig. 11A). QD 11 and QD 30 plot in the uppermost part of Fig. 11B. Thus, during continued evaporation and precipitation of calcite the remaining fluids are expected to get extremely enriched in Mg with marginal SO<sub>4</sub> content, in addition

to high Na, K, and Cl contents (Fig. 11B) and will thus potentially precipitate magnesite (MgCO<sub>3</sub>), bischofite (MgCl<sub>2</sub>· $6H_2O$ ), and they may produce MgSO<sub>4</sub>-bearing salts such as kieserite (MgSO<sub>4</sub>· $H_2O$ ). These anticipated developments are in agreement with the SNORM calculations (Fig. 10B). However, MgSO<sub>4</sub>-bearing salts precipitate at progressively later stages of evaporation and the evolved brines will finally not form MgSO<sub>4</sub>-bearing salts. These brines produce 'MgSO<sub>4</sub> deficient' mineral sequences and can precipitate CaCl<sub>2</sub>-bearing salts upon extreme evaporation (Deocampo & Jones, 2014). Also, highly soluble potash salts like sylvite (KCl) or carnallite (KMgCl<sub>3</sub>· $6H_2O$ ), lacking MgSO<sub>4</sub>, can be produced.

In contrast, the spring-water QD 34 (SE of Xitieshan) is expected to evolve during prolonged evaporation into a fluid with high SO<sub>4</sub>, low Ca and high Na, K, and Cl contents. Ultimately, additionally to calcite and gypsum precipitation of bloedite (MgSO<sub>4</sub>·H<sub>2</sub>O) and potentially polyhalite (K<sub>2</sub>SO<sub>4</sub>·MgSO<sub>4</sub>·4H<sub>2</sub>O) may occur (Fig. 11B). For the low TDS waters QD 10, QD 32, and QD 33 precipitation of magnesite and probably a mineral like then ardite  $(Na_2SO_4)$  is predictable, due to their position in the ternary phase diagram. After prolonged evaporation, precipitation of dolomite  $[CaMg(CO_3)_2]$  and magnesite could be expected from the waters QD 9 and QD 20 and from QD 31, QD 21, QD 25, and QD 15 rather dolomite (Fig. 11B). Precipitation of carnallite (KMgCl<sub>3</sub> $\cdot$ 6H<sub>2</sub>O) is predicted for the inflows QD 30, QD 31, and QD 25 in the northeastern part of the Qaidam Basin by the SNORM modeling (Fig. 10B).

For all waters precipitation of halite and other highly soluble salt minerals follows in late steps of the evaporation process due to their significantly higher solubility. The precipitation of the chlorides cannot be derived from the composition relations shown on the ternary diagrams (Fig. 11A, B) but follows from SNORM models (Fig. 10B).

### 6 Analysis and conclusions: from evaporating inflow to salt lakes, regional differences

The composition of inflow (low TDS water) has three different components: (1) Composition of precipitation, mainly rainwater. (2) Rock-derived components from rainwater-rock reactions. (3) Redissolution of eolian salt deposits at the mountain slopes (Stober et al., 2016; Zhu, 2016). Mainly during winter time heavy storms transport salts from the Qaidam Basin to the mountains (Fig. 3D). As a consequence, all inflow becomes enriched in TDS along its flow-path into the basin due to re-dissolution of salt minerals. Rivers and brooks rapidly reduce their flow velocity and ultimately become stagnant when entering

the basin. Simultaneously, and due to the desert climate conditions, water ( $H_2O$ ) evaporates and the least soluble minerals start precipitating.

In the basin, processes including mixing, degassing, temperature changes, local dissolution of salts, cyclic wetting and drying, ion exchange and adsorption further modify the inflows. In addition, interaction of presentday water in the generally shallow salt lakes containing oscillating  $H_2O$ -content with older salt deposits further adds to the complexity of lake water composition with time. Reports from upwelling deep(er) waters into salt lakes suggest significant additional effects on salt lake water composition (e.g. Lowenstein & Risacher, 2009; Spencer et al., 1990; Xiao et al., 2018).

In the following, the impact of inflows (low TDS waters) on the chemical composition of the salt lakes (brines) is shown. For this purpose, areas had to be selected, where investigated inflows and salt lakes occur in close proximity.

### 6.1 Gas Hure Hu area: inflows QD 9, QD 10, QD 11 and salt lakes QD 12, QD 13 (Fig. 2)

Gas Hure Hu salt lake is large in terms of its surface area (about 117.5 km<sup>2</sup>) but with a maximum depth of 1.3 m very shallow (Ye et al., 2015) making the hydrochemical composition very sensitive to for instance climatic or temperature variations. The samples QD 12 and QD 13 originate from two small salt lakes in close vicinity, north of Gas Hure Hu, where additionally two evaporate mineral samples (QD 12a, QD 13a) were collected. A third mineral sample was taken a bit farther north (QD 14). The inflow QD 9 comes from the Karawatag Shan Mountains in the west, and QD 10 was taken in a small fresh water lake nearby. QD 11 originates in the northern Altun Shan Mountains (Fig. 2; Tables 1, 2).

The inflows (low TDS waters) are characterized by relatively high Mg (especially QD 11), moderate SO<sub>4</sub>, and high Na and Cl contents. K contents are low and Sr is enhanced, especially in QD 11. TDS of the inflows are increasing from SW to NE, i.e. from QD 9, to QD 10, and to QD 11 (Table 4). All low TDS waters show higher  $(HCO_3 + CO_3)$  than Ca contents (Fig. 4B), i.e. after calcite  $(CaCO_3)$  precipitation there will still be carbonate in solution. Calcite precipitation (Fig. 11A) causes a relative increase of the SO<sub>4</sub> content in the remaining fluid, which will be for the most part still in the remaining solution after anhydrite/gypsum precipitation, controlled by low limiting Ca. Consequently, total anhydrite/gypsum precipitated is predicted to be low. Precipitation of calcite and anhydrite/gypsum in the first evaporation steps gradually develops high Mg residual water forming the salt lakes (QD 12, QD 13) (Table 3; Fig. 4A).

Precipitation of other carbonates, i.e. Mg-bearing carbonates, like dolomite or magnesite, or Mg-sulphates is possible after the first evaporation steps, because of still high Mg, high  $SO_4$  and moderate (HCO<sub>3</sub> + CO<sub>3</sub>) contents in the remaining solution (Fig. 11B). Precipitation of magnesite (QD 9-QD 11) and partly dolomite (QD 9) is also predicted by the SNORM models (Fig. 10B). The next major 'chemical divide' is created by the low-solubility Ca-sulphates: gypsum/anhydrite. The evaporating residual waters are dominated by Cl, Na, Mg, and SO<sub>4</sub>. Figure 11B shows, that prolonged evaporation of the water QD 9 is expected to lead to a solution with very low Mg and high SO<sub>4</sub> contents, favoring precipitation of sodium sulfate, e.g. thenardite (Na<sub>2</sub>SO<sub>4</sub>). QD 10 will probably include precipitation of bloedite  $(Na_2Mg(SO_4)_2 \cdot 4H_2O)$ and QD 11 is expected to precipitate a magnesium chloride, like bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O) (Fig. 11B). The occurrence of magnesium salt deposits, including bloedite  $(Na_2Mg(SO_4)_2 \cdot 4H_2O)$ , and other sulfates such as mirabilite (Na<sub>2</sub>SO<sub>4</sub> $\cdot$ 10H<sub>2</sub>O) were reported by Chen and Bowler (1986) and Ye et al. (2015), confirming the findings in this investigation. Further, SNORM models predict precipitation of magnesite (MgCO<sub>3</sub>), kieserite (MgSO<sub>4</sub> $\cdot$ H<sub>2</sub>O), bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O) during prolonged evaporation of the salt lakes (Fig. 10A). Due to enhanced Sr input (particularly QD 11) and the still high remaining  $SO_4$  contents in the salt lakes saturation with respect to celestite  $(SrSO_4)$  is reached (Fig. 6) and its precipitation can be expected.

During prolonged evaporation, Na and Cl increase until saturation with respect to halite is reached (Fig. 6) and halite (NaCl) precipitates, leading to a relative enrichment of remaining ions in solution, e.g. K or Li (Fig. 9B). The modeling shows, that in the present salt lakes saturation with respect to potassium salts (sylvite, carnallite) or lithium salts is not yet reached (Fig. 6). However, the occurrence of potassium salt deposit is described in Chen and Bowler (1986), but it is not clear if the potassium deposit results from earlier, perhaps dryer conditions, or from salt lake waters artificially concentrated. Sylvite was found during field work exclusively as artificially generated evaporative salt mineral (QD 14).

The evaporative mineral samples (QD 12a, QD 13a, QD 14), taken in or nearby the salt lakes (Fig. 2; Tables 2, 5), are of different chemical composition:

- Sample QD 12a is nearly pure halite (95%), with a small proportion of bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O) and traces of a MgSO<sub>4</sub> mineral, anhydrite (CaSO<sub>4</sub>) resp. gypsum, and sylvite (KCl).
- Sample QD 13a (Fig. 5A) was taken from a salt lake at the rim of a seismic blasting. It mainly consists of anhydrite/gypsum (41%), halite (32%), and bischofite

(19%). Additionally, in the sample a small amount of magnesium sulphates (7%), and traces of sylvite, sell-aite (MgF<sub>2</sub>) and magnesium nitrate [Mg(NO<sub>3</sub>)<sub>2</sub>] are present. Magnesium nitrate is used in pyrotechnics. The sample was taken at the rim of a hole, resulting from former seismic blasting.

• Sample QD 14 was taken from a heaped-up pile of evaporative minerals, an interim step in the production chain from artificially up-concentrated water of the Gas Hure Hu salt lake. The investigation indicates that the main components are halite (36%), bischofite (30%), a magnesium sulphate (20%), and additionally sylvite (14%) as well as traces of anhydrite.

The evaporative minerals in the individual samples confirm the investigation of the waters, when concentrated. The results of this study shows as well that the inflows (low TDS waters) are precursors of the salt lakes. Additionally, these findings are consistent and in line with reported evaporite deposits in the western region of the Qaidam Basin (Chen & Bowler, 1986; Decampo & Jones, 2014; Zhang et al., 2019; Ye et al., 2015, 2018).

### 6.2 Lake Dong Taijnar Hu area with the inflow QD 15 and brines QD 17 and QD 18 (Fig. 2)

The big river Narin Gol (QD 15) flows from the Kunlun Mountains into the Qaidam Basin disemboguing into the salt lake Dong Taijnar Hu (QD 17) (Figs. 2, 3). The river water shows high Na, enhanced Ca and Mg, significant K concentrations and is rich in Cl and (HCO<sub>3</sub> + CO<sub>3</sub>) with lower SO<sub>4</sub> (Table 4; Fig. 4B). When Narin Gol finally reaches the area of the salt lake Dong Taijnar a significant amount of water (H<sub>2</sub>O) must already have been evaporated due to its long flow-path within the basin and first minerals already precipitated, particularly carbonates.

QD 15 contains higher ( $HCO_3 + CO_3$ ) than Ca (Fig. 4B). Thus, after calcite precipitation excess carbonate remains in solution (Fig. 11A) and SO<sub>4</sub> slightly increases passively. Low Ca of the remaining fluid results in very minor anhydrite/gypsum precipitation. After the first evaporation steps in the basin Mg drastically increases and as a result, characterizes the salt lake Dong Taijnar Hu (QD 17) and the nearby trench (QD 18) (Table 3; Fig. 4A). During prolonged evaporation, precipitation of Mg-bearing carbonates, e.g. dolomite, are possible and expected (Fig. 11B) und supported by SNORM models (Fig. 10B).

Prolonged evaporation leads to increasing Na and Cl until finally precipitation of halite occurs. This process causes a relative enrichment of remaining ions in solution, such as K and Li (Fig. 9B). The computations show that in the actual salt lake water Dong Taijnar Hu (QD 17) saturation with respect to halite was not yet reached in contrast to the nearby trench (QD 18), in which water is passively concentrated by the man-made trench design. Li concentration is high in both brines, firstly because river-water of Narin Gol (QD 15) contains considerable amounts of primary Li (Table 4) and the river has a high discharge rate and secondly because extensive evaporation increases Li in the residual fluids, i.e. in the brines and thus especially in QD 18 (Table 3). K is enriched correspondingly.

Both brines (QD 17, QD 18) are not saturated with respect to potassium salts (sylvite, carnallite) or lithium salts (Fig. 6). However, the SNORM model predicts for QD 18 precipitation of kieserite (MgSO<sub>4</sub>·H<sub>2</sub>O), bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O), and carnallite (KMgCl<sub>3</sub>·6H<sub>2</sub>O) (Fig. 10A) in addition to the dominant halite. This is mainly caused by increasing Mg contents during the evaporation and first precipitation processes, additionally enhanced SO<sub>4</sub> contents, the low Na/Cl ratio (0.86) of the brine (Fig. 4A), and because the river water Narin Gol (QD 15) contains a significant amount of K (Table 4; Fig. 4B), which will be passively enriched during the evaporation and precipitation processes.

The salt sample QD 16 collected near salt lake Dong Taijnor Hu consists mainly of halite (99%), likewise the sample QD 18a (Fig. 5B) collected within the nearby trench (QD 18). The fibrous form of sample QD 16 is most probably caused by capillary rise of halite-saturated water in soil environment (Fig. 5C). The salt analysis of QD 16 produced very small amounts of antarcticite (CaCl<sub>2</sub>·6H<sub>2</sub>O, 0.5%), bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O, 0.3%) and traces of sylvite (KCl), similar to sample QD 18a (Table 5). The modeled trace minerals probably represent 'halite-impurities'.

Yu et al. (2012) describe halite-dominated strata with carnallite, being the most common potassium mineral either disseminated in thin-layers on the lowermost locations of the playa surface, occasionally coexisting with bischofite. Halite with some sulfate impurities has been reported by Chen and Bowler (1986). The observations of these authors fit the findings of this study.

### 6.3 Lake Dabusun Hu area (QD 22, QD 19) with southern QD 20, QD 21 (Golmud He) and northern inflows QD 32, QD 34 (Fig. 2).

River water from Golmud He (QD 21) and the nearby spring water (QD 20), that flows downstream into Golmud He, originate from the Kunlun Mountains. Golmud He flows in northern direction into lake Dabusun Hu (QD 22) (Fig. 3B). The river- and spring water QD 32 and QD 34 derive from the Xitieshan Mountains northeast of Dabusun Hu. Most of this water seeps into the soil before eventually reaching the salt lake in the subsurface. The river Golmud He (QD 21) is the main inflow, having by far the highest discharge rate. QD 21 and QD 20 are compositionally quite similar and rich in Ca, Mg and HCO<sub>3</sub>, in contrast to the NE-inflows (Table 4; Fig. 4B).

Lake Dabusun Hu (QD 22) lies within the central Qarhan playa. With a maximum depth of only 1.72 m and a varying surface area of around 250 km<sup>2</sup> (Lowenstein et al., 1994; Casas et al., 1992). Its hydrochemical composition is very sensitive to climatic or temperature changes. The surface of Qarhan playa is predominantly dry and consists of a solid rugged crust of halite (Fig. 3B).

QD 21 and QD 20 are saturated with respect to calcite, aragonite, and dolomite, and  $(HCO_3 + CO_3)$  exceeds Ca in both waters (Fig. 4B). After calcite precipitation there will still be a relatively high amount residual carbonate in solution (Fig. 11A). In the remaining fluid a passive increase of SO<sub>4</sub> can be expected. Anhydrite/ gypsum precipitation is predicted to be very low, due to low Ca contents in the remaining fluid. The results are supported by SNORM models (Fig. 10B). After the first evaporation steps in the basin, high Mg contents in the remaining fluid will develop and are as well found in the salt lake Dabusun Hu (QD 22) and in the trench east of Dabusun Hu (QD 19) (Table 3; Fig. 4A). During further evaporation precipitation of Mg-bearing carbonates, e.g. dolomite (QD 20, QD 21) or even magnesite (QD 20) is anticipated (Fig. 11B) and supported by SNORM models (Fig. 10B). The spring-water QD 20 was already saturated with respect to magnesite before it flows into Golmud He (QD 21). Due to elevated Sr content and residual alkalinity QD 21 and QD 20 are saturated with respect to strontianite (SrCO<sub>3</sub>). The high discharge rate of Golmud He transports considerable amounts of 'trace' elements like K, B, and Li into the basin, which will be enriched during prolonged evaporation in the remaining solution.

In contrast, the northern inflows (QD 32, QD 34) have very low (HCO<sub>3</sub> + CO<sub>3</sub>) in the remaining solution after calcite precipitation (Fig. 11A). Prolonged evaporation of spring-water QD 34 results in predicted precipitation of gypsum/anhydrite and bloedite (Na<sub>2</sub>Mg(SO<sub>4</sub>)<sub>2</sub>·4H<sub>2</sub>O) (Fig. 10A, 11B). However, the influence of QD 32 and QD 34 on the bulk hydrochemistry of lake Dabusun Hu is expected to be small, due to low discharge rate.

The brine of lake Dabusun Hu (QD 22) is a highly concentrated solution (TDS = 316.1 g/kg) and extremely rich in Mg and Cl. The trench northwest of Dabusun Hu (QD 19) has lower TDS with Mg, Na and Cl as main constituents (Fig. 4A). The highly concentrated brine in Dabusun Hu also contains high loads of Li (Fig. 9B) and K (in QD 22: Li = 178.3 mg/kg and K = 5532 mg/kg). Lake water (QD 22) is supersaturated with respect to anhydrite and celestite (SrSO<sub>4</sub>) and saturated with respect to carnallite (KMgCl<sub>3</sub>·6H<sub>2</sub>O) and halite (Fig. 6). The trench-water (QD 19) is also supersaturated with respect to celestite and saturated with respect to anhydrite and gypsum. In contrast, Dabusun Hu (QD 22) is undersaturated with respect to gypsum but supersaturated concerning anhydrite because of reduced H<sub>2</sub>O activity in the brine (Fig. 7A). The snorm model predicts precipitation of bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O), carnallite (KMgCl<sub>3</sub>·6H<sub>2</sub>O) and halite from the salt lake (Fig. 10A) and a similar development for the nearby trench (QD 19).

The evaporite sample (QD 19a) from the trench shows large, clear, cubic crystals, consisting mainly of halite (99.4%), some magnesium chloride (0.3%), and traces of anhydrite, antarcticite, and sylvite (Table 5), probably 'halite-impurities'.

Kong et al. (2018) reported the occurrence of halite, bischofite and carnallite in the Qarhan playa, but give no precise information on the sampling location. However, accumulated halite and carnallite have been identified along restricted margins of the lake Dabusun Hu (Chen & Bowler, 1986; Zhang, 1987; Spencer et al., 1990; Vengosh et al., 1995; Zheng, 1997; Yu et al., 2013). Casas et al. (1992) mapped the precipitated evaporation minerals along the flow path in the delta of the branching river Golmud He and found the following succession: aragonite – carnallite+halite – salt plain (playa). Xiao et al. (2018) suggest that groundwater also evolves from freshwater to brine water along the flow path based on hydrochemical and isotopic data. Zhang et al. (2019) traced the high K concentration in the salt lake to accumulated input of the southern inflows, in agreement with the here reported findings.

## 6.4 Lake Da Qaidam Hu (QD 29) with northeastern thermal water inflows (QD 26, QD 27, QD 28) and southeastern inflow (QD 30) (Fig. 2)

Hot water from two springs (QD 26, QD 27) and warm water from a brook (QD 28), uphill of QD 27, represent northern inflows into the salt lake Da Qaidam Hu (QD 29) (Fig. 3C). The hot water originates from fractured granite at about 5 km depth in the Qilian Mountains (Stober et al., 2016). Also from the Qilian Mountains, but from southeastern direction another river (QD 30) runs to the salt lake.

The thermal spring waters (QD 26, QD 27) are characterized by high Na, low Ca, and lower Cl than Na concentrations (Fig. 4B; Table 4). Chloride equals about  $SO_4$ plus HCO<sub>3</sub>. The waters contain considerable amounts of K (13.7 mg/kg), Li (3.2 mg/kg), B (40.7 mg/kg), and F (8.9 mg/kg). Like lithium, boron (B) also originates from granitic rocks, where B occurs in minerals like tourmaline; furthermore, B can be present in rock-forming minerals, like muscovite, biotite, plagioclase (e.g. Saurer & Troll, 1990; Barth, 2000). Thus, the origin of B in thermal waters might be related to crystalline basement. The composition of the relatively low TDS thermal brook water (QD 28) is similar. In contrast, TDS of the southeastern inflow (QD 30) is significantly higher and the river has a relatively high Mg concentration. QD 30 is also rich in the 'trace' elements (K, Li, B, F) (Fig. 4B; Table 4). However, all inflows into the salt lake are supersaturated with respect to aragonite, calcite, dolomite, and strontianite (SrCO<sub>3</sub>). Inflow QD 30 is additionally supersaturated with respect to magnesite (MgCO<sub>3</sub>), monohydrocalcite (CaCO<sub>3</sub>·H<sub>2</sub>O), and sellaite (MgF<sub>2</sub>).

The salt lake Da Qaidam Hu (QD 29) has a mean surface area of 30 km<sup>2</sup> (Fig. 3C). The lake is very shallow with a maximum depth less than 1 m during the summer flood season (Gao et al., 2019). Thus, its hydrochemical composition is very sensitive to climatic, seasonal, hydrological, and temperature changes, causing a continuous interplay between formation and solution of different minerals. During the study period, TDS of the salt lake Da Qaidam Hu was with 190 g/kg lower than in most other salt lakes of the Qaidam Basin at that time. However, TDS of the salt lake changes dramatically in short time intervals (Gin 2013). The lake water is a Na-Mg-Cl- $(SO_4)$  solution (Fig. 4A; Table 3) with comparably high concentrations of K (2753 mg/kg), Li (67,3 mg/kg), and B (370 mg/kg). The brine is saturated with respect to celestite, but not saturated with respect to gypsum/anhydrite or halite (Fig. 6).

After calcite precipitation from thermal spring waters (QD 26, QD 27) there will still be a relatively high amount of carbonate in solution. In contrast, the inflows QD 28 and QD 30 contain low carbonate after calcite precipitation (Fig. 11A). The higher Ca and Mg in the thermal brook water (QD 28) compared to the thermal spring waters (QD 26, QD 27) (Fig. 10A) may form calcite (or dolomite) in the thermal brook. This is also supported by SNORM models. The thermal spring waters are expected to precipitate the remaining carbonate, e.g. as strontianite  $(SrCO_3)$  or pirsonite  $(Na_2Ca(CO_3)_2 \cdot 2H_2O)$  (Fig. 10B). Figure 11B shows, that the inflow QD 30 with its high Mg concentration favors precipitation of magnesiumcarbonates and -sulphates, such as magnesite  $(MgCO_3)$ and kieserite (MgSO<sub>4</sub>· $H_2O$ ), which is also supported by SNORM modeling (Fig. 10B). After precipitation of carbonates a slight passive increase of  $SO_4$  can be predicted and is as well observed. In contrast to inflow QD 30, anhydrite/gypsum precipitation from the thermal waters is expected to be very low (Fig. 11A) also supported by the SNORM model (Fig. 10B). High Mg and increasing  $SO_4$  in the residual water (QD 30) are finally reflected in the salt lake Da Qaidam Hu (QD 29) (Table 3; Fig. 4A).

Extensive evaporation from the thermal waters may produce thenardite  $(Na_2SO_4)$  and other sulphates

predicted by SNORM modeling (Fig. 10B). The occurrence of thenardite is described in Stober et al. (2016). From the southeastern inflow QD 30 ultimately precipitation of bischofite (MgCl<sub>2</sub>·6H<sub>2</sub>O) and carnallite (KMgCl<sub>3</sub>·6H<sub>2</sub>O) is predicted (SNORM model) (Fig. 10B). Consequently, brine water of Da Qaidam Hu (QD 29) predominantly forms halite after prolonged evaporation and potentially bischofite, carnallite, and kieserite in addition (Fig. 3C, 10A).

Input of K, B, Li and F from the inflows into the salt lake Da Qaidam Hu (QD 29) is comparatively high. Figure 9B shows, that Li/Cl ratios of all inflow-waters and the water of the salt lake are equally high. During prolonged evaporation and precipitation of the first evaporative minerals, these solutes will get enriched in the brine (see Table 3). Gin (2013) reports high fluctuations of brine concentration within short time intervals. Thus, under appropriate conditions saturation with respect to borates or Li-sulphate may develop. This also holds for evaporating smaller basins and pools in close vicinity to the salt lake.

The two salt samples from the bottom of lake Da Qaidam Hu (QD 29a) and the nearby salt flat (QD 29b) (Table 2) consist of nearly pure halite. QD 29a are coarsegrained white cubic crystals with a few yellowish crystals. The salt consists of 98.1% halite, 0.9% bischofite, 0.8% anhydrite, and some trace minerals probably caused by 'halite-impurities'. Sample QD 29b is an agglutinated, white, porous material with halite (99.5%) and some 'halite-impurities' (Table 5).

Previous studies found that halite and borates are the predominate salts in Da Qaidam Hu salt lake area (Chen & Bowler, 1986; Gao et al., 2019). Zhen (1997) reports of Ca-, Mg-, and Mg-Ca-carbonates and borates. In addition, Gao, et al. (2019) describe the occurrence of halite, carbonate minerals (calcite, aragonite, dolomite, hydromagnesite), and gypsum in two wells, drilled within the lake. Furthermore, they reported pinnoite (MgB<sub>2</sub>O<sub>4</sub>·3H<sub>2</sub>O) deposits in the lakebed and borate earth in the northeastern lakesite and farther to the south ulexite (NaCaB<sub>5</sub>O<sub>6</sub>(OH)<sub>6</sub>·5H<sub>2</sub>O) and hydroboracite (CaMgB<sub>6</sub>O<sub>8</sub>(OH)<sub>6</sub>·3H<sub>2</sub>O). The latter possibly caused by the relatively Ca-rich southern inflow QD 30. Thus, the observations of these authors fit the findings of this study.

### 7 Conclusions

The intramontane Qaidam basin is  $a \sim 500$  km long and  $\sim 250$  km wide high altitude drain-off free hydrologic basin located at the northern rim of the Tibet plateau in western China. Playas, vast gypsum and salt flats and salt lakes shape the plane basin. The basin has numerous inflows derived from the surrounding mountain chains. Two major rivers originate in the Kunlun Mountains in the south and one in the Qaidam mountains in the north of the basin. The major rivers contain an unusually large amount of chloride when they enter the basin from the mountains. The Cl/Br ratio of the river water indicates that the carried chloride originates from dissolved windblown halite. The large rivers also contain a high amount of Mg and boron and a high TDS for the same reason. Thus, a large amount of the yielded solutes represents recycled solids from the salt flat playas of the basin. Only about 20–50% of the solutes have been contributed by weathering of basement rocks in the surrounding mountains.

The original relatively pristine deep water contributed by fluid-rock interaction is exemplified by the thermal springs produced by the Qaidam Mountains near the town Da Qaidam. These waters contain very low Mg and the TDS is twice to three times that of the major rivers. The high  $SiO_2$  content suggests that the water reacted with basement rock at temperatures in excess of 130 °C. The thermal water contains high Li and B indicating that these valuable trace elements ultimately originate from the basement rocks. The relatively low Cl/Br ratio of the thermal water suggests that dissolved windblown halite contributes some however small amount to the composition of the water.

All other inflows, small rivers and brooks, show high to very high Cl/Br ratios typical of recycled halite. TDS varies considerably in these waters. The most pristine spring water has been collected in the Kunlun Mountains south of Golmud. The similarity with the main Golmud river suggests that also the main river has little recycled solutes. A special group of smaller inflows is represented by some high-TDS brooks in narrow small valleys rich in vegetation in the rugged topography bordering the flat land. The brooks contain up to 5.2 g/ kg TDS and the Cl/Br ratio is high. Most likely windblown halite causes the observed water composition. The rich vegetation in the small valleys may retain this halite. The observation also shows that recycling of halite is also possible in areas where the exposed basement rocks are free of salt minerals.

All salt lakes and salt pools contain high TDS water with a high to very high Cl/Br ratio. Typically, most salt lakes and small salt ponds are saturated with respect to halite and contain about 263 g NaCl per kg solution. The majority of the salt lakes contain massive halite crystals covering the floor and the rims of the lakes and ponds. The halite lakes dissolve and precipitate the salt deposits depending on the seasonal temperature and precipitation variations. There are several departures from the simple halite lake. Some lakes close to the margins of the basin touching the enclosing mountains are undersaturated with halite and contain less than 200 g/kg TDS. These lakes have very low Ca and Mg > > Ca. The lakes are in a state of evolving towards Na-Mg-Cl-SO<sub>4</sub> dominated lakes.

The central lakes of the basin surrounded by vast salt playas are massively influenced by salt mining activities and/or mineral oil production. The Dabusun Lake contains very high MgCl<sub>2</sub> and KCl in solution. Its salt load is used for fertilizer production. One small salt lake in the west of the basin consists of CaCl<sub>2</sub>, KCl and LiCl and only minor NaCl. Its unusual composition results from mixing with deep fluids produced by the nearby oil field.

Within this article it could be shown, that there is a direct chemical connection between the salt lakes and the inflows into the Qaidam Basin, i.e. the inflows are indeed precursors of the brines.

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#### Author contributions

IS carried out fieldwork, analyzed and interpreted the hydrochemical data, drafted the figures, tables, and the manuscript. JZ carried out fieldwork and contributed to the manuscript. KB analyzed and interpreted geochemical data and worked as well on the manuscript. The main revision of the manuscript was made by IS, supported by KB and JZ. All authors approved the submission. All authors read and approved the final manuscript.

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### Availability of data and materials

All relevant data generated or analyzed during this study are included in this published article. Additional data are available from the corresponding author on reasonable request.

### Declarations

**Ethics approval and consent to participate** Not applicable.

### **Consent for publication**

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### **Competing interests**

The authors declare that they have no competing interests.

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### References

- Barth, S. R. (2000). Geochemical and boron, oxygen and hydrogen isotopic constraints on the origin of salinity in groundwaters from the crystalline basement of the Alpine Foreland. *Applied Geochemistry*, 15, 937–952.
- Blatt, H., & Tracy, R.J. (1996). Petrology; Igneous, sedimentary, and metamorphic. In W. H. Freeman, 2nd edn. ISBN 0-7167-2438-3.

- Bodine, M.W., Jr. & Jones, B.F. (1986). The salt norm: a quantitative chemicalmineralogical characterization of natural waters. U.S. Geological Survey, Water-Resources Investigations.
- Casas, E., Lowenstein, T. K., Spencer, R. J., & Zhang, P. (1992). Carnallite mineralization in the nonmarine, Qaidam Basin, China; evidence for the early diagenetic origin of potash evaporites. *Journal of Sedimentary Research*, 62, 881–898.
- Chang, L.L.Y., Howie, R.A., & Zussman, J. (1998). Non-silicates: Sulphates, carbonates, phosphates, halites. Rock-Forming-Minerals, 5B, 2nd edn. (p. 383), London: The Geological Society.
- Chen, K., & Bowler, J. M. (1986). Late pleistocene evolution of salt lakes in the Qaidam basin, Qinghai province, China. *Palaeogeography, Palaeoclimatol*ogy., 54, 87–104.
- Cowgill, E. (2001). Tectonic evolution of the Altyn Tagh-Western Kunlun fault system, northwestern China. PhD thesis, (p. 311) Los Angeles: University of California.
- Davis, S. N., DeWayne, C. L., Zreda, M., & Moysey, S. (2001). Chlorine-36, bromide, and the origin of spring water. *Chemical Geology, 179*, 3–16.
- Davis, S. N., Whittemore, D. O., & Fabryka-Martin, J. (1998). Uses of chloride/bromide ratios in studies of potable water. *Ground Water*, *36*, 338–350.
- Deelman, J.C. (1999). Low-temperature nucleation of magnesite and dolomite. Neues Jahrbuch für Mineralogie, Monatshefte, 7, 289–302, Stuttgart.
- Deocampo, D.M., & Jones, B.F. (2014). Geochemistry of saline lakes. In J. I. Drever (Ed.), *Treatise on Geochemistry*, Vol. 7: Surface and Groundwater, Weathering, and Soils, 2nd edn., Chap. 7.13.
- Drever, J.I. (1997). The geochemistry of natural waters: Surface and groundwater environments. 3rd edn., (p. 436) *Prentice Hall Upper Saddle River*, New Jersey 07458.
- Drüppel, K., Stober, I., Grimmer, J. C., & Mertz-Kraus, R. (2020). Experimental alteration of granitic rocks: Implications for the evolution of geothermal brines in the Upper Rhine Graben, Germany. *Geothermics*, *88*, 101903.
- Du, Y. S., Fan, Q. S., Gao, D. L., Wei, H. C., Shan, F. S., Li, B. K., Zhang, X. R., Yuan, Q., Qin, Z. J., & Ren, Q. H. (2019). Evaluation of boron isotopes in halite as an indicator of the salinity of Qarhan paleolake water in the eastern Qaidam Basin, western China. *Geosience Frontiers*. https://doi.org/10.1016/j.gsf. 2018.02.016
- Duan, Z., & Hu, W. (2001). The accumulation of potash in a continental basin: The example of the Qarhan Saline Lake, Qaidam Basin, West China. *European Journal of Mineralogy*, *13*, 1223–1233.
- Eugster, H. P., & Hardie, L. A. (1978). Saline lakes. In A. Lerman (Ed.), *Lakes, chemistry, geology, physics* (pp. 237–293). New York: Springer.
- Fan, Q. S., Lai, Z. P., Long, H., Sun, Y. J., & Liu, X. J. (2010a). QSL chronology for lacustrine sediments recording high stands of Gahai Lake in Qaidam Basin, northeastern Qinhai-Tibetan Plateau. *Quarternary Geochronology*, 5, 223–227.
- Fan, Q. S., Ma, H., Lai, Z., Tan, H., & Li, T. (2010b). Origin and evolution of oilfield brines from Tertiary strata in western Qaidam Basin: Constraints from <sup>87</sup>Sr/<sup>86</sup>Sr, δD, δ<sup>18</sup>O, d<sup>34</sup>S and water chemistry. *Chinese Journal of Geochemistry*, 29, 446–454.
- Fan, Q. S., Ma, H. Z., Wei, H. C., Shan, F. S., An, F. Y., Xu, L. M., & Madsen, D. B. (2014). Late Pleistocene paleoclimatic history documented by an oxygen isotope record from carbonate sediments in Qarhan Salt Lake, NE Qinghai-Tibetan Plateau. *Journal of Asian Earth Sciences*, 85, 202–209.
- Fan, Q. S., Ma, Y., Cheng, H., Wei, H. C., Yuan, Q., Qin, Z., & Shan, F. S. (2015). Boron occurrence in halite and boron isotope geochemistry of halite in the Qarhan Salt Lake, western China. *Sedimentary Geology*, *322*, 34–42. https://doi.org/10.1016/j.sedgeo.2015.03.012
- Frape, S. K., & Fritz, P. (1987). Geochemical trends for groundwaters from the Canadian shield. In P. Fritz & S. K. Frape (Eds.), *Saline water and gases in crystalline rocks*, 33 (pp. 19–38). Geological Association of Canada Special Paper.
- Galamay, A. R., Bukowski, K., Sydor, D., & Meng, F. (2020). Ultramicrochemical Analyses (UMCA) of fluid inclusions in halite and experimental research to improve the accuracy of measurement. *Minerals*, 10, 823. https://doi.org/ 10.3390/min10090823
- Gao, C., Yu, J., Min, X., Cheng, A., Hong, R., & Zhang, L. (2019). The sedimentary evolution of Da Qaidam Salt Lake in Qaidam Basin, northern Tibetan Plateau: Implications for hydro-climate change and the formation of pinnoite deposit. *Environmental Earth Siences*, 78, 463. https://doi.org/10. 1007/s12665-019-8480-0

- Garret, D.E. (1996). Potash: deposits, processing, properties, and uses. Chapman & Hall, 2–6 Boundary Row, (p. 733), London SE1 8HN, UK.
- Garrett, D. E. (1998). Borates: Handbook of deposits, processing (p. 483). Academic Press.
- Gascoyne, M., Davison, C. C., Ross, J. D., & Pearson, R. (1987). Saline groundwaters and brines in plutons in the Canadian Shield. In P. Fritz & S. K. Frape (Eds.), Saline Water and Gases in Crystalline Rocks, 33 (pp. 53–68). Geological Association of Canada Special Paper.
- Golding, E., & Phaup, A.E. (1936). Chemical analyses of the rocks, ores and minerals of Southern Rhodesia. *Bulletin Geological Survey Southern Rhodesia*, 29.
- Guo, J., Wei, X., Long, G., Wang, B., Fan, H., & Xu, S. (2017). Three-dimensional structural model of the Qaidam basin: Implications for crustal shortening and growth of the northeast Tibet. *De Gruyter Open Access*, 9, 174–185.

Guo, P., Liu, C., Yu, M., Ma, D., Wang, P., Wang, K., Mao, G., & Zhang, Q. (2018). Paleosalinity evolution of the Paleogene perennial Qaidam lake on the Tibetan Plateau: Climatic vs. tectonic control. *Intern. Journal of Earth Sciences*, 107, 1641–1656. https://doi.org/10.1007/s00531-017-1564-8

Holland, T. J. B., & Powell, R. (2011). An Improved and Extended Internally Consistent Thermodynamic Dataset for Phases of Petrological Interest, Involving a New Equation of State for Solids. *Journal of Metamorphic Geol*ogy, 29, 333–383. https://doi.org/10.1111/j.1525-1314.2010.00923.x

- Johnson, J. W., Oelkers, E. H., & Helgeson, H. C. (1992). SUPCRT92 A software package for calculating the standard molal thermodynamic properties of minerals, gases, aqueous species, and reactions from 1-bar to 5000-bar and 0°C to 1000°C. Computer and Geosciences, 18, 899–947.
- Johnson, K. S. (1970). Salt produced by solar evaporation on Big Salt Plain, Woods County, Oklahoma. *Oklahoma Geology Notes, 30*, 47–54.
- Johnstone, S. (1954). *Minerals for the chemical and allied industries*. Chapman and Hall.
- Junner, N.R. & James, W.T. (1947). Chemical analyses of Gold Coast rocks, ores and minerals. *Gold Coast Geological Survey Bulletin, 15*.
- Kong, F., Zhen, M., Hu, B., Wang, A., Ma, N., & Sobron, P. (2018). Dalangtan saline playa in a hyperarid region on Tibet Plateau: I. Evolution and environments. Astrobiology, 18, 1243–1253. https://doi.org/10.1089/ast.2018.1830
- Kong, W. G., Zheng, M. P., Kong, F. J., & Chen, W. X. (2014). Sulfate-bearing deposits at Dalangtan Playa and their implication for the formation and preservation of martian salts. *American Mineralogist*, 99, 283–290. https:// doi.org/10.2138/am.2014.4594
- Lawrence Livermore National Laboratory (2020). Lawrence Livermore National Laboratory (LLNL) Open Data Initiative. UC San Diego Library Digital Collections. https://doi.org/10.6075/J0HD7T2Q
- Lippmann, F. (1973). Sedimentary carbonate minerals (p. 228). Springer.
- Liu, Z. C., Wang, Y. J., Chen, Y., Li, X. S., & Li, Q. C. (1998). Magnetostratigraphy and sedimentologically derived geochronology of the Quaternary lacustrine deposits of a 3000 m thick sequence in the central Qaidam Basin, Western China. *Palaeogeography, Palaeoclimatology, Palaeoecology,* 140, 459–473.
- Lowenstein, T.K., Spencer, R.J., Casas, E., Zhang, P., Zhang, B., Fan, H., & Krouse, H.R. (1994). Major-element and stable-isotope geochemistry of fluid inclusions in halite, Qaidam Basin, western China: Implications for late Pleistocene/Holocene brine evolution and paleoclimates. *Geological Society of America, Special Paper 289*. https://doi.org/10.1130/SPE289-p19.
- Lowenstein, T. K., & Risacher, F. (2008). Closed Basin Brine Evolution and the Influence of Ca–Cl Inflow Waters: Death Valley and Bristol Dry Lake California, Qaidam Basin, China, and Salar de Atacama, Chile. Aquatic Geochemistry, 15, 71–94. https://doi.org/10.1007/s10498-008-9046-z
- Mel' nikova, Z.M., & Moshkina, LA. (1973). Solubility of anhydrite and gypsum in the system Na-Mg-Ca-Cl-S0<sub>4</sub>-H<sub>2</sub>O. *Jzvestija Sib Otdel AN SSSR Ser Khim Nauk*, 4(21), 176–182. in Russian.
- Meng, Q., & Fang, X. (2008). Cenozoic tectonic development of the Qaidam Basin in the northeastern Tibetan Plateau. *Geological Society of America* Special Papers, 2008(444), 1–24.
- Miao, W., Fan, Q., Wie, H., Zhang, X., & Ma, H. (2016). Clay mineralogical and geochemical constraints on late Pleistocene weathering processes of the Qaidam Basin, northern Tibetan Plateau. *Journal of Asian Earth Sciences*, 127, 267–280.
- Parkhurst, D.L., & Appelo, C.A.J. (1999). PHREEQC (Version2)—a computer program for speciation, batch-reaction, onedimensional transport, and inverse geochemical calculations. US Geological Survey Water-Resources Investigation Report, 99-4259.

- Pitzer, K.S. (1991). Activity coefficients in electrolyte solutions (2nd ed.). Chapter 3, In K.S. Pitzer (Ed.), *Ion interaction approach: theory and data correlation* (pp. 75–153), C.R.C. Press.
- Sauerer, A., & Troll, G. (1990). Abundance and distribution of boron in the Hauzenberg (Bavaria) granite complex. *Geochemica Et Cosmochemica Acta*, 54, 49–55.
- Savoye, S., Aranyossy, J.-F., Beaucaire, C., Cathelineau, M., Louvat, D., & Michelot, J.-L. (1998). Fluid inclusions in granites and their relationships with present-day groundwater chemistry. *European Journal of Mineralogy, 10*, 1215–1226.
- Spencer, R.J., Lowenstein, T.K., Casas, E., & Pengxi, Z. (1990) The potash salts and brines in the Qaidam Basin, China. In Spencer & I-Ming Chou (Eds.) Fluid– mineral interactions: A Tribute to H. P. Eugster, The Geochemical Society, Special Publication No. 2.
- Steiger, M., & Asmussen, S. (2008). Crystallization of sodium sulfate phases in porous materials: The phase diagram Na<sub>2</sub>SO<sub>4</sub>–H<sub>2</sub>O and the generation of stress. *Geochimica Et Cosmochimica Acta*, 72, 4291–4306.
- Stober, I., & Bucher, K. (1999). Origin of salinity of deep groundwater in crystalline rocks. *Terra Nova*, 11, 181–185.
- Stober, I., & Bucher, K. (2004). Fluid sinks within the earth's crust. *Geofluids*, 4, 143–151.
- Stober, I., Zhong, J., Zhang, L., & Bucher, K. (2016). Deep hydrothermal fluid– rock interaction: The thermal springs of Da Qaidam, China. *Geofluids*, 16, 711–728. https://doi.org/10.1111/gfl.12190

Stumm, W., & Morgan, J. J. (1975). Aquatic chemistry (2nd ed.). Wiley.

- Sun, D., & Li, B. (1993). Origins of borates in the saline lakes of China. Seventh Symposium on Salt, 1, 177–194.
- Vengosh, A., Chivas, A. R., Starinsky, A., Kolodny, Y., Zhang, B., & Zhang, P. (1995). Chemical and boron isotope compositions of non-marine brines from the Qaidam Basin, Qinghai, China. *Chemical Geology*, *120*, 135–154.
- Wang, E., Xu, F. Y., Zhou, J. X., Wan, J., & Burchfiel, B. C. (2006). Eastward migration of the Qaidam basin and its implications for Cenozoic evolution of the Altyn Tagh fault and associated river systems. *Geological Society of America Bulletin*, 118, 349–365.
- Wang, X., Miller, J. D., Cheng, F., & Cheng, H. (2014). Potash flotation practice for carnallite resources in the Qinghai Province. *PRC Mining Engineering*, 66–68, 33–39. https://doi.org/10.1016/j.mineng.2014.04.012
- Warren, J. K. (2016). Evaporites a geological compendium (2nd ed.). Springer. https://doi.org/10.1007/978-3-319-13512-0
- Wells, R.C. (1937). Analyses of rocks and minerals from the laboratory of the United States Geological Survey 1914–36. U.S. Geological Survey Bulletin, 876.
- Whitney, D. L., & Evans, B. W. (2010). Abbreviations for names of rock-forming minerals. American Mineralogist, 95, 185–187.
- Xiao, Y., Shao, J., Frape, S. K., Cui, Y., Dang, X., Wang, S., & Ji, Y. (2018). Groundwater origin, flow regime and geochemical evolution in arid endorheic watersheds: A case study from the Qaidam Basin, northwestern China. *Hydrology and Earth System Sciences, 22*, 4381–4400. https://doi.org/10. 5194/hess-22-4381-2018
- Xing, L., Xu, L., Zhang, P., & Wang, P. (2022). Organic Geochemical Characteristics of Saline Lacustrine Source Rocks: A Case Study from the Yingxi Area, Qaidam Basin, China. *Geochemistry International*, 60, 92–108.
- Xiong, Q., Zheng, J.-P., Griffin, W. L., O'Reilly, S. Y., & Pearson, N. J. (2014). Pyroxenite dykes in orogenic peridotite from North Qaidam (NE Tibet, China) track metasomatism and segregation in the mantle wedge. *Journal of Petrology*, 55, 2347–2376.
- Yang, J. S., Xu, Z. Q., Song, S. G., Zhang, J., Wu, C., Shi, R., Li, H., & Brunel, M. (2001). Discovery of coesite in the North Qaidam Early Palaeozoic ultrahigh pressure, UHP metamorphic belt NW China. *Comptes Rendus De L Academie Des Sciences Serie II Fascicule A-Sciences De La Terre Et Des Planetes*, 11, 719–724.
- Yang, Q., Wu, B. H., Wang, S. Z., Cai, K. Q., & Qian, Z. H. (1993). Study on Potash Deposit of the Qarhan Salt Lake (pp. 28–36). Geological Publishing House.
- Ye, C., Mao, J., Ren, Y., Li, Y., Lin, Y., Power, I. M., & Luo, Y. (2018). Salt crystallization sequences of nonmarine brine and their application for the formation of potassium deposits. *Aquatic Geochemistry*, 24, 209–229. https://doi.org/ 10.1007/s10498-018-9340-3
- Ye, C., Zheng, M., Wang, Z., Hao, W., Wang, J., Lin, X., & Han, J. (2015). Hydrochemical characteristics and sources of brines in the Gasikule salt lake, Northwest Qaidam Basin, China. *Geochemical Journal*, 49, 481–494. https://doi.org/10.2343/geochemj.2.0372

- Yin, A., Dang, Y., Wang, L., Jiang, W., Zhou, S., Chen, X., Gehrels, G. E., & McRivette, M. W. (2008). Cenozoic tectonic evolution of Qaidam basin and its surrounding regions (Part 1): The southern Qilian Shan-Nan Shan thrust belt and northern Qaidam basin. *Geological Society of America Bulletin.*, 120, 813–846.
- Yin, A., Rumelhart, P. E., Butler, R., Cowgill, E., Harrison, T. M., Foster, D. A., Ingersoll, R. V., Zhang, Q., Zhou, X.-Q., Wang, X.-F., Hanson, A., & Raza, A. (2002). Tectonic history of the Altyn Tagh fault system in northern Tibet inferred from Cenozoic sedimentation. *Geological Society of America Bulletin*, 114, 1257–1295.
- Yu, J. Q., Gao, C. L., Cheng, A. Y., Liu, Y., Zhang, L., & He, X. H. (2013). Geomorphic, hydroclimatic and hydrothermal controls on the formation of lithium brine deposits in the Qaidam Basin, northern Tibetan Plateau, China. Ore Geology Reviews, 50, 171–183.
- Yu, S. (1986). The hydrochemical features of salt lakes in the Qaidam Basin. Chinese Journal of Oceanology and Limnology, 4, 383–403.
- Yuan, J., Huo, C., & Cai, K. (1983). The high mountain-deep basin saline environment—A new genetic model of salt deposits. *Geological Review*, 29, 159–165.
- Zhang, J. X., Zhang, Z. M., Xu, Z. Q., Yang, J. S., & Cui, J. W. (2001). Petrology and geochronology of eclogites from the western segment of the Altyn Tagh, Northwestern China. *Lithos*, *56*, 187–206.
- Zhang, P. (1987). Salt lakes in Qaidam Basin. Science Press.
- Zhang, X., Fan, Q., Li, Q., Du, Y., Qin, Z., Wie, H., & Shan, F. (2019). The source, distribution, and sedimentary pattern of K-rich brines in the Qaidam Basin, western China. *Minerals*, *9*, 655. https://doi.org/10.3390/min9110655
- Zheng, M. (1997). An Introduction to Saline Lakes on the Qinghai-Tibet Plateau (p. 324). Springer. https://doi.org/10.1007/978-94-011-5458-1
- Zheng, W., Zhang, P., He, W., Yuan, D., Shao, Y., Zheng, D., Ge, W., & Min, W. (2013). Transformation of displacement between strike-slip and crustal shortening in the northern margin of the Tibetan Plateau: Evidence from decadal GPS measurements and late Quaternary slip rates on faults. *Tectonophysics*, 584, 267–280.
- Zhou, S. I., Zhang, W. C., & Wang, F. (2016). Spatial-temporal variations and their dynamics of the saline lakes in the Qaidam Basin over the past 40 years. *IOP Conference Series Earth and Environmental Science, 46*, 012043. https://doi.org/10.1088/1755-1315/46/1/012043
- Zhou, J., Xu, F., Wang, T., Cao, A., & Yin, C. (2006). Cenozoic deformation history of the Qaidam Basin, NW China: Results from cross-section restoration and implications for Qinghai-Tibet Plateau tectonics. *Earth Planetary Science Letters*, *243*, 195–210.
- Zhu, B.-Q. (2016). Atmospheric significance of aeolian salts in the sandy deserts of northwestern China. Solid Earth, 7, 191–203. https://doi.org/10.5194/ se-7-191-2016
- Zimmer, K., Zhang, Y. L., Lu, P., Chen, Y. Y., Zhang, G. R., Dalkilic, M., & Zhu, C. (2016). SUPCRTBL: A revised and extended thermodynamic dataset and software package of SUPCRT92. *Computer and Geosciences*, 90, 97–111.

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