Hydrogeochemical characteristics and geothermometry of hot springs in the Altai region, Mongolia

Bolormaa Chimeddorj1,2, Dolgormaa Munkhbat3, Battushig Anitaabatar4, Oyunsetseg Dolgorjav5 and Bolormaa Oyunsetseg1*

1 Department of Chemistry, School of Arts and Sciences, National University of Mongolia, Ulaanbaatar 14201, Mongolia
2 Department of Chemical–Biology, School of Natural Science and Technology, Khovd University, Peace Avenue, 164300 Khovd Province, Mongolia
3 Department of Chemical and Biological Engineering, School of Engineering and Applied Sciences, National University of Mongolia, Ulaanbaatar 14201, Mongolia
4 Mineral Resources and Petroleum Authority of Mongolia, Ulaanbaatar, Mongolia
5 Mongolian Academy of Sciences, Institute of Chemistry and Chemical Technology, Ulaanbaatar, Mongolia

© DM, 0000-0003-4644-7628; BA, 0000-0002-1014-4302; BO, 0000-0002-2861-7841
*Correspondence: bolormaa@num.edu.mn

Abstract: This study determines the properties of hot spring waters and associated rocks, calculates reservoir temperatures and depths in the Mongolian Altai region, and constructs a conceptual model for geothermal water based on these results. The hot springs consist of HCO₃–Na, SO₄–Na, and HCO₃–SO₄–Na mixed waters. They exhibit alkaline pH levels and temperatures in the range 21.3–35°C. X-ray diffraction analyses of outcrop rocks reveal silicate and carbonate-type minerals such as quartz, albite, orthoclase, dolomite, mica and actinolite, while correlation analysis indicates that the chemical composition of the hot spring water is directly related to rock mineral composition, where dissolution of albite, orthoclase and dolomite minerals has played an important role. Reservoir water circulation depths are 2615–3410 m according to quartz and chalcedony geothermometry. The results indicate that the spring water in the Mongolian Altai region comprises a low mineral content with alkaline pH levels and that the reservoir temperature can reach up to 106°C. We also propose a conceptual model for geothermal water in the Chikhertei hot spring. The geothermal water in the Mongolian Altai region has the potential for use in heating systems.

Keywords: hot spring; hydrogeochemistry; Mongolian Altai; geothermal energy; water–rock interaction

Thematic collection: This is part of the Hydrochemistry related to exploration and environmental issues collection available at: https://www.lyellcollection.org/cc/hydrochemistry-related-to-exploration-and-environmental-issues

Received 24 February 2021; revised 29 June 2021; accepted 3 July 2021

Hot springs have been widely used for decades in recreational and medical applications because they impart useful health benefits (Kaven et al. 2020). Nowadays, hot springs are being utilized by heating and power generation systems because the geothermal energy of hot springs is renewable (Lund 2018). Currently, about 82 countries produce 164,635 GWh yr⁻¹ of energy for direct use from geothermal energy resources (Lund, et al. 2011), and this energy is used in agriculture, minor industrial processes, various heating systems, greenhouses and farms (Guðnason et al. 2015). Some countries such as Costa Rica, El Salvador, Kenya and New Zealand produce more than 10% of their total electricity supply using geothermal resources (Van Nguyen et al. 2015).

In Mongolia, there are an estimated 250 hot and cold springs, of which 43 are hot (Oyunsetseg et al. 2015). Numerous authors have classified spring water temperatures to exhibit different ranges of values (Durowoju et al. 2015; Thermal Laboratory et al. 2018; Žunić et al. 2019). We have used the following range, which has been used repeatedly in the literature: cold (<20°C), warm (20–35°C), hot (35–45°C) and very hot (>45°C). Generally, Mongolian springs occur in the following five regions: Mongolian Altai, Khuvsgul, Dormod-Mongol, Khangai and Khentii (Lkhagvadorj and Tseesuren 2005). Most of the hot springs (32) are distributed in the Khangai Mountain region, where they are used for medical treatment and in spa resorts. This area is in the central part of Mongolia, which is located 400 km away from the capital city. For the past 10 years, other hot springs such as the Tsenkher, Khujirt and Shivert have been used for heating systems with constant flow (Bignall et al. 2005; Lkhagvadorj and Tseesuren 2005).

Hot spring research in the Khangai Mountain region has been more intense than in the other four regions (Gendenjams et al. 2003; Ganbat and Demberel 2010; Oyunsetseg et al. 2015). The Mongolian Altai region predominantly has cold springs; however, it also has four hot springs including Chikhertei, Gants Mod, Aksu and Indert. Previous studies in the Mongolian Altai region were focused on hydrochemical analyses of the hot springs (Yousif and El-Aassar 2018); however, a geothermal study in this area has not yet been reported (Tugjamba 2021).

The hydrochemical properties of hot springs are related to the interactions of water with the surrounding rock minerals. Therefore, mineral interactions, mineral components, and geothermal calculations can be used to assess the feasibility of using a hot spring as an energy resource. The purpose of this study was to determine the properties of the hot spring waters and associated rocks and calculate the reservoir temperatures and depths in the Mongolian Altai region; based on these results, we constructed a conceptual model for the geothermal waters.

Study area

Topography and climate

Figure 1 shows a map of Mongolia, which is a landlocked country between Russia and China and whose territory covers an area of...
1.564 116 km². The capital city is Ulaanbaatar, located about 1700 km away from the study area.

The Mongolian Altai range is 900 km long and is located in the NW of the country along the border of Mongolia and China (Fig. 1). The Mongolian Altai range extends through the landscapes and provinces of Bayan-olgii, Khovd, Uvs and Govi-Altai in Mongolia. The total area of the Mongolian Altai region is c. 104 478.5 km² (Blomdin et al. 2016). At the top of the Mongolian Altai range is the Khuiten Mountain, whose summit is 4374 m above sea level.

The Mongolian Altai Mountains have sharp peaks, steep slopes and sharp fault scarps owing to a strong tectonic uplift. In this region, many deep crustal faults are located in the sub-meridional and sub-latitudinal directions. Additionally, the area is a seismically active zone (Lkhagvasuren and Burmaa 2017).

This continental area is located in a harsh climate zone owing to the northern latitudes and high elevations above sea level. The annual average precipitation reaches 50 to 350 mm in the Mongolian Altai Mountains (Lkhagvasuren and Burmaa 2017). Hot springs in the Mongolian Altai region are located 1690–2780 m above sea level (Blomdin et al. 2016; Lehmkuhl et al. 2016).

The distribution of Mongolian hot springs by temperature difference (warm, moderate and high) is shown in Figure 1.

Geological setting

Sedimentary rocks, volcanogenic sediments and granitic intrusions are widespread in the Mongolian Altai region. In the Paleozoic zone, Middle Cambrian–Lower Ordovician sandstone–siltstone was commonly deposited in this area. Gneiss, crystalline shale, granite, syenite, gabbro–syenite, granö–syenite, diorite and granodiorite are the other rocks present in this region.

Furthermore, the region contains valleys, ridges and mountainous areas with saline soils in meadow marshes, saline mountain tundra, and alpine meadow soils in high mountain areas, which were formed by erosion and weathering processes (Byamba et al. 1999; Blomdin et al. 2016). The Mongolian Altai terrain is characterized by overlapping depressions with Middle Ordovician–Lower Silurian volcanogenic–terrestrial compressions, overlapping depressions of the Delum-Sagsai and Yamaat rivers, and Devonian volcanic sediment. It contains a variety of components, including molasse, siliceous sub-granite and the Devonian Altai granite complex (Tomurtogoo 2008, 2012).

The Tawan Bogd terrain range covers the high mountains in the western part of the Mongolian Altai range and includes a 4000–6000 m thick unit of Late Cambrian–Lower Paleozoic sandstone and siltstone. The lithology is composed of granite, leucogranite and granite porphyry that are 385–349 Ma and granite porphyryitic biotite, medium granite and albite granite that are 456 Ma old (Byamba et al. 1999; Bulgatov et al. 2014). The geological setting of the Mongolian Altai region is shown in Figure 2.

Hydrogeology of the study area

The Mongolian Altai terrane is considered to be the remains of oceanic island arcs or a Lake Zone system (Cai et al. 2015 Xiao et al. 2018). It contains various lakes, glaciers and hot springs. The majority of the water resources in the region are located in the Khar-Uus and Khovd river basins, which cover an area of 85 104.8 km² over five provinces in Western Mongolia. The deepest lake, Khoton, is about 26–58 m deep, and the biggest lake, Acht, has an area of c. 296.8 km². In total, 96% of the rivers and lakes are considered to be fed by glaciers, with the longest glacier in the range (Potamine) being 20 km in length. The Khovd River is the largest river in the outflow basin of the Mongolian Altai range, which is situated in Central Asia. The age of the hydrogeology of the region is Quaternary and the aquifer contains alluvial, deluvial and proluvial moraine sediments. The sediments of the Quaternary period represent the layer that is closest to the Earth’s surface; these were formed 10 000 to 80 000 years ago. The Mongolian Altai region exhibits the following spring water types: Na–HCO₃, Na–SO₄, and Na–SO₄–HCO₃ (Lkhagvasuren and Burmaa 2017).

Materials and methods

Hot spring water and rock samples were collected from locations represented by blue dots in Figure 2. Twelve water samples were collected from four hot spring sources: Aksu (AC), Indert (IN), Gants Mod (GM) and Chikheritei (CH). These springs are located in Altai region, Bayan-Olgii Province, Mongolia. Outcrop rock samples were collected from the outflow part of each of the four hot springs. As shown in Figure 2, rock and hot spring water samples were obtained from the same four locations. The bedrock of these hot springs was not exposed.

Each water sample was collected and stored in a polyethylene plastic bottle with 2 ml nitric acid for stabilization. Physicochemical parameters, such as temperature, pH, total dissolved solid (TDS) levels and oxidation reduction potential (ORP), were measured in situ using a Hanna HI-98 (USA) instrument. Bicarbonate ion levels were determined in situ using the 0.01 N hydrochloric acid titration method with methyl orange endpoint. The cations and anions in the hot spring water samples, including Ca²⁺, Mg²⁺, Na⁺ and K⁺, were detected using inductively coupled plasma (ICP) optical emission spectrometry (Shimadzu, Japan). All tests were performed in duplicate and measured in line with ICP standards. Quality assurance and control of the measurements were performed using a certified reference material (SPS–SW2) for the determination of trace elements. Sulfate ion and SiO₂ levels were determined by spectrophotometry using the alkalinity standard and ammonium molybdate methods, respectively, with the 4500-SiO-2.C.3g standard silica sample (UV-MS51, Brazil). Chloride ion concentration was analyzed using the AgNO₃ volumetric method (Ewing and Jordan 1978). The ionic charge balance error of the analyses was less than ±5%.

Rock samples were air dried at room temperature for 48–72 h after collection. The dried samples were sieved using a 0.074 mm mesh and used for elemental and mineral composition analyses. The mineral composition of the rock samples was investigated by X-ray diffraction (XRD) (PANalytical X’Pert PRO, Netherlands) using the Co–Kα line (λ = 1.79 Å) powder method with a 2θ diffraction angle range of 2–50°. Laboratory duplicates and internal standards (Si, kkl113) were used to ensure the precision and accuracy of the XRD analyses. For the X-ray fluorescence (XRF) measurements, a fused borate disk was used for calibration and a reference material (Cu-base-FP) was analyzed to monitor equipment performance and accuracy.

The results of the hydrochemical analyses are presented using a Piper diagram, Figure 3, and chemical geothermometry calculations were performed using Aquachem (PHREEQC Ver 2.0). A Piper diagram describes the chemical composition of groundwater and its different classes (Joshi and Bhanderi 2013). The geological map and geothermometry model were prepared using ArcGIS and CorelDraw. The equations provided in Table 1 were used for geothermometry calculations.

The depth of geothermal waters was calculated using the following equation (Arnórsson et al. 1983; Besser et al. 2018; Shestakova et al. 2018):

\[ D = \left( T - T_s \right)/G \]

where D is the geothermal water circulation depth in m, \( T \) is the reservoir temperature in °C, \( T_s \) is the annual average temperature in °C, and \( G \) is the regional geothermal gradient in °C km⁻¹.

Materials and methods

Hot spring water and rock samples were collected from locations represented by blue dots in Figure 2. Twelve water samples were collected from four hot spring sources: Aksu (AC), Indert (IN), Gants Mod (GM) and Chikheritei (CH). These springs are located in Altai region, Bayan-Olgii Province, Mongolia. Outcrop rock samples were collected from the outflow part of each of the four hot springs. As shown in Figure 2, rock and hot spring water samples were obtained from the same four locations. The bedrock of these hot springs was not exposed.

Each water sample was collected and stored in a polyethylene plastic bottle with 2 ml nitric acid for stabilization. Physicochemical parameters, such as temperature, pH, total dissolved solid (TDS) levels and oxidation reduction potential (ORP), were measured in situ using a Hanna HI-98 (USA) instrument. Bicarbonate ion levels were determined in situ using the 0.01 N hydrochloric acid titration method with methyl orange endpoint. The cations and anions in the hot spring water samples, including Ca²⁺, Mg²⁺, Na⁺ and K⁺, were detected using inductively coupled plasma (ICP) optical emission spectrometry (Shimadzu, Japan). All tests were performed in duplicate and measured in line with ICP standards. Quality assurance and control of the measurements were performed using a certified reference material (SPS–SW2) for the determination of trace elements. Sulfate ion and SiO₂ levels were determined by spectrophotometry using the alkalinity standard and ammonium molybdate methods, respectively, with the 4500-SiO-2.C.3g standard silica sample (UV-MS51, Brazil). Chloride ion concentration was analyzed using the AgNO₃ volumetric method (Ewing and Jordan 1978). The ionic charge balance error of the analyses was less than ±5%.

Rock samples were air dried at room temperature for 48–72 h after collection. The dried samples were sieved using a 0.074 mm mesh and used for elemental and mineral composition analyses. The mineral composition of the rock samples was investigated by X-ray diffraction (XRD) (PANalytical X’Pert PRO, Netherlands) using the Co–Kα line (λ = 1.79 Å) powder method with a 2θ diffraction angle range of 2–50°. Laboratory duplicates and internal standards (Si, kkl113) were used to ensure the precision and accuracy of the XRD analyses. For the X-ray fluorescence (XRF) measurements, a fused borate disk was used for calibration and a reference material (Cu-base-FP) was analyzed to monitor equipment performance and accuracy.

The results of the hydrochemical analyses are presented using a Piper diagram, Figure 3, and chemical geothermometry calculations were performed using Aquachem (PHREEQC Ver 2.0). A Piper diagram describes the chemical composition of groundwater and its different classes (Joshi and Bhanderi 2013). The geological map and geothermometry model were prepared using ArcGIS and CorelDraw. The equations provided in Table 1 were used for geothermometry calculations.

The depth of geothermal waters was calculated using the following equation (Arnórsson et al. 1983; Besser et al. 2018; Shestakova et al. 2018):

\[ D = \left( T - T_s \right)/G \]

where D is the geothermal water circulation depth in m, \( T \) is the reservoir temperature in °C, \( T_s \) is the annual average temperature in °C, and \( G \) is the regional geothermal gradient in °C km⁻¹.
Hydrogeochemical features of Mongolian hot spring

Fig. 1. Hydrogeological regions and locations of hot springs in Mongolia.

Fig. 2. Geological setting of the hot spring area in the Mongolian Altai range.
Results and discussion

Hydrochemical characteristics

Physicochemical parameters and some ion concentrations are shown in Table 2. The accuracy of the analyses was checked through the Standard Method 1030E. The ionic error balance (%), was calculated using cation–anion balance, expressed as milliequivalents per litre. As shown Table 2, ionic error balance for hot spring samples was less than ± 0.2% and it generally should not exceed 5% (Friedman and Erdmann 1982). The results show that the hot spring temperatures ranged between 21.3 and 35°C, and the pH levels were high, between 8.32 and 9.26, indicating alkaline conditions. This temperature range of the hot springs in the Mongolian Altai region corresponds to the warm springs (20–35°C) classification (Thermal Laboratory et al. 2018; Žunic et al. 2019).

Alkaline hot springs might be suitable for heating systems because of the low corrosiveness of water in this pH range (Bleam 2016).

The ORP values are important for determining the state of the elements and their transfer to other forms of groundwater. The ORP values of the GM, CH and AC hot spring samples ranged between 28.8 and 59.1 mV, whereas those of the IN samples were between 124 and 127 mV. The ORP may depend on the mineral salts, as well as on the TDS in the hot spring and the amount of dissolved oxygen (Emelyanov 2005). The SiO2 content (45.3–75.6 mg l⁻¹) can be explained by the weathering of siliceous rocks in the hot spring waters, which caused the transfer of silica from the rocks to the

Table 2. Physical and chemical parameters of hot springs in the Mongolian Altai area

| ID | T°C | pH | ORP, mV | TDS | DO | Na | K | Ca | Mg | HCO₃ | CO₃ | Cl | SO₄ | SiO₂ | F | IEB, % |
|----|-----|----|--------|-----|----|----|---|----|----|-----|-----|---|----|-----|-----|---|-------|
| GM1| 30.4 | 9.21 | 40.0  | 127 | 0.07 | 45.4 | 1.23 | 2.02 | 0.22 | 41.6 | 23.5 | 17.8 | 8.91 | 46.7 | 9.86 | 0.121 |
| GM2| 24.5 | 9.18 | 59.1  | 145 | 0.03 | 45.4 | 1.23 | 2.21 | 0.22 | 47.3 | 22.4 | 13.7 | 10.7 | 48.2 | 7.31 | 0.054 |
| CH1| 22.0 | 9.22 | 43.3  | 221 | 0.53 | 68.6 | 1.84 | 5.33 | 0.42 | 72.6 | 15.1 | 7.70 | 68.3 | 64.5 | 2.97 | 0.133 |
| CH2| 23.8 | 9.26 | 51.9  | 230 | 0.01 | 71.8 | 1.83 | 2.50 | 0.24 | 74.6 | 15.2 | 7.31 | 66.1 | 61.7 | 1.27 | 0.291 |
| CH3| 21.3 | 9.11 | 50.6  | 259 | 0.30 | 71.1 | 1.75 | 2.00 | 0.49 | 68.7 | 17.8 | 9.62 | 61.5 | 64.2 | 3.16 | 0.041 |
| CH4| 24.7 | 9.14 | 44.7  | 225 | 0.40 | 73.3 | 1.86 | 3.22 | 0.41 | 78.0 | 15.9 | 8.11 | 67.4 | 60.9 | 1.99 | 0.016 |
| CH5| 24.7 | 9.13 | 52.6  | 225 | 0.40 | 71.4 | 1.89 | 2.81 | 0.54 | 79.4 | 12.5 | 7.14 | 67.9 | 60.4 | 4.12 | 0.045 |
| AC1| 31.0 | 9.00 | 28.8  | 187 | 0.31 | 38.4 | 1.22 | 2.80 | 0.10 | 50.5 | 18.6 | 3.41 | 14.5 | 51.6 | 4.82 | 0.066 |
| AC2| 35.0 | 8.60 | 30.6  | 188 | 0.28 | 32.9 | 1.02 | 3.20 | 0.16 | 48.7 | 14.6 | 2.62 | 12.9 | 49.5 | 4.05 | 0.148 |
| AC3| 32.0 | 8.65 | 35.8  | 174 | 0.44 | 34.5 | 1.20 | 2.80 | 0.10 | 48.1 | 15.7 | 2.30 | 14.5 | 45.3 | 2.82 | 0.014 |
| IN1| 25.0 | 8.40 | 124   | 725 | 0.45 | 196 | 12.5 | 24.2 | 1.22 | 110  | 8.65 | 47.7 | 322  | 75.6 | 6.2  | 0.080 |
| IN2| 28.0 | 8.32 | 127   | 698 | 0.61 | 186 | 15.2 | 22.2 | 1.22 | 91.5 | 6.12 | 47.7 | 319  | 69.2 | 2.54 | 0.022 |

Fig. 3. Piper diagram of the major chemical compositions of hot springs in the Mongolian Altai region.
water. The fluoride concentrations of the hot spring samples ranged from 1.27 to 9.86 mg l\(^{-1}\). Hot springs contain elements such as aluminum, boron, silicon and iron, which can form stable complexes with fluorine, thereby increasing the solubility of fluorite and the number of fluoride ions in the water. But the solubility of fluorite also depends on hot spring temperature (Edmunds and Smedley 2013). Therefore, another reason for the high fluoride content can be related to the host rocks. Some igneous and sedimentary rocks contain fluoride naturally, which leads to increased fluoride.

The dissolved oxygen (DO) in the hot springs amounted to 0.01 to 0.61 mg l\(^{-1}\); these low levels illustrate that DO decreases as the temperature increases. The TDS values ranged from 127 to 725 mg l\(^{-1}\). Interestingly, the TDS values of the IN1 and IN2 samples ranged between 698 and 725 mg l\(^{-1}\), which were higher than those of the other samples. The IN hot spring is located at a lower elevation area between the mountains. According to these results, we observed that the TDS values increased with a decrease in elevation of the hot springs in the Mongolian Altai region. TDS increase also can be related to evaporation at low altitudes (Oyem et al. 2014).

Moreover, the pH and TDS results support the conclusion that the hot springs of the Mongolian Altai region are advantageous for heating systems because of the low corrosion properties and low amounts of mineral accumulation.

We have illustrated the hydrochemical results for the Mongolian Altai hot springs using a Piper diagram (Fig. 3). According to this diagram, Na\(^+\) and HCO\(_3\)\(^-\) predominated in the CH1, CH2 and CH3 samples, which can be characterized as HCO\(_3\)–Na type waters. The ions Na\(^+\) and SO\(_4\)\(^2-\) predominated in the GM1, IN1 and IN2 samples, which can be characterized as SO\(_4\)–Na type waters. In the AC1, AC2, AC3, CH4 and GM2 samples, which are HCO\(_3\)–SO\(_4\)–Na type waters, Na\(^+\), HCO\(_3\)\(^-\) and SO\(_4\)\(^2-\) predominated.

**Table 3. Correlations of major ions (milliequivalents per litre) in hot spring samples**

|   | Na  | K   | Ca  | Mg  | HCO\(_3\) | SO\(_4\) | Cl  | Si  |
|---|-----|-----|-----|-----|-----------|---------|-----|-----|
| Na | 1.00 |     |     |     |           |         |     |     |
| K  | 0.96 | 1.00 |     |     |           |         |     |     |
| Ca | 0.96 | 0.98 | 1.00 |     |           |         |     |     |
| Mg | 0.93 | 0.90 | 0.89 | 1.00 |           |         |     |     |
| HCO\(_3\) | 0.76 | 0.56 | 0.59 | 0.67 | 1.00 |       |     |     |
| SO\(_4\) | 0.99 | 0.97 | 0.98 | 0.92 | 0.70 | 1.00 |     |     |
| Cl  | 0.98 | 0.95 | 0.94 | 0.94 | 0.93 | 1.00 |     |     |
| Si  | 0.85 | 0.71 | 0.72 | 0.75 | 0.90 | 0.83 | 0.65 | 1.00 |

\(r(x) = 0.96\)

**Table 4. Saturation indices for hot springs of the Mongolian Altai region**

| Samples | GM1  | GM2  | CH1  | CH2  | CH3  | CH4  | CH5  | AC1  | AC2  | AC3  | IN1  | IN2  |
|---------|------|------|------|------|------|------|------|------|------|------|------|------|
| Albite  | 7.59 | 8.13 | 8.31 | 8.11 | 8.07 | 7.99 | 7.81 | 6.65 | 6.56 | 6.07 | 6.81 | 6.56 |
| Alunite | −6.55| −4.77| −2.29| −3.22| −2.71| −3.14| −3.61| −7.51| −7.95| −2.19| −2.64|      |
| Anhydrite| −1.9 | −1.8 | −0.84| −1.12| −1.29| −1.06| −1.09| −1.47| −1.26| −1.32| 0.21 | 0.21 |
| Calcite | 2.34 | 2.37 | 2.62 | 2.39 | 2.18 | 2.4  | 2.32 | 2.41 | 2.28 | 2.23 | 2.6  | 2.44 |
| Fluorite | 2.56 | 2.45 | 1.93 | 0.88 | 1.59 | 1.31 | 1.9  | 2.15 | 2.17 | 1.81 | 2.98 | 2.14 |
| Gibbsite | 1.23 | 1.65 | 1.98 | 1.75 | 1.75 | 1.66 | 1.49 | 0.56 | 0.79 | 0.08 | 0.71 | 0.51 |
| Gypsum  | −1.66| −1.49| −0.51| −0.08| −0.94| −0.75| −0.79| −1.23| −1.07| −1.09| 0.51 | 0.48 |
| Halite  | −4.94| −5.04| −5.18| −5.19| −5.06| −5.14| −5.21| −5.73| −5.89| −5.92| −4.05| −4.08|
| K-feldspar | 8.54 | 9.14 | 9.32 | 9.1  | 9.09 | 8.97 | 8.8  | 7.66 | 7.5  | 7.11 | 8.11 | 7.93 |
| K-mica  | 16.6 | 18.0 | 18.8 | 18.2 | 18.1 | 17.9 | 17.4 | 14.4 | 14.7 | 12.9 | 15.1 | 14.5 |
| Kaolinite | 7.97 | 8.99 | 9.47 | 8.97 | 9.07 | 8.82 | 8.49 | 6.63 | 7.11 | 5.83 | 7.02 | 6.6  |
| Quartz  | 2.33 | 2.43 | 2.34 | 2.32 | 2.38 | 2.34 | 2.34 | 2.33 | 2.34 | 2.42 | 2.38 | 2.38 |

**Fig. 4.** Gibbs diagram for hot springs in the study area.
to silica, HCO$_3^-$ exhibited the highest positive correlation ($r = 0.90$). This correlation may be related to the dissolution of subsurface rock and metamorphic rock containing albite and silicate minerals (Fan et al. 2019). According to the correlation results, gypsum and halite minerals may be dissolved in the hot spring waters.

Saturation indices (SIs) can aid the interpretation of water–rock reactions in hot spring environments. A positive SI value shows oversaturation while a negative value indicates undersaturation (Bouragba et al. 2011). The temperature of hot spring water, major ions, pH and microelements were used as input data in SI calculation. Table 4 shows the SIs of hot spring samples; values for alunite, anhydrite, gypsum and halite indicate undersaturation whereas the other minerals were oversaturated.

### Gibbs diagram

The Gibbs diagram was used to investigate the hydrochemical components of ground water and identify the interactions of water and rock (Fig. 4). The element transfers from rock played an important role in the compositions of the hot springs (Nwankwoala et al. 2015; Nagaraju et al. 2016). Therefore, we investigated the rock sample compositions further as discussed below.

### Rock characterization

#### Elemental composition of the rock samples

Major and minor element compositions of the rock samples determined by XRF are shown in Table 5. The major element oxides were in the following order: SiO$_2$ > Al$_2$O$_3$ > Na$_2$O > K$_2$O > Fe$_2$O$_3$ > CaO > MgO > TiO$_2$ > P$_2$O$_5$ > MnO. Based on these results, we concluded that the hot spring reservoirs included metamorphic layers of granite–gneiss (Joshi and Bhandari 2013).

![X-ray diffraction analyses of rock samples](Q-Quartz Al-Albite O-Orthoclase M-Mica D-Dolomite Ac-Actinolite)

### Table 5. Major and minor element compositions of rock samples

| Sample ID | Major elements, % | Minor elements, mg kg$^{-1}$ |
|-----------|-----------------|-----------------------------|
|           | SiO$_2$ TiO$_2$ Al$_2$O$_3$ Fe$_2$O$_3$ CaO MgO Na$_2$O K$_2$O | Ba Cs Rb Sb Sr Zn Zr V Se Cr |
| CH        | 75.5 0.213 12.5 2.34 1.15 0.290 3.05 3.92 0.033 0.067 | 386 48.0 190 40 176 18.0 64.0 25.1 30.0 43.0 |
| GM        | 75.3 0.176 12.7 2.54 0.74 0.411 4.50 2.57 0.036 0.034 | 289 39.0 80.2 40 116 48.2 104 16.2 30.0 39.0 |
| IN        | 70.2 0.352 13.8 3.04 2.27 0.712 3.42 3.75 0.062 0.119 | 936 42.0 132 40 425 32.3 164 35.0 53.0 39.0 |
| AC        | 69.3 0.355 13.6 4.32 2.69 1.23 2.78 2.41 0.077 0.086 | 235 30.0 199 40 70.2 18.1 36.0 15.3 30.0 33.0 |

#### Table 6. Calculated normative minerals

| Normative minerals | Volume norm, % |
|--------------------|----------------|
|                    | CH  | GM  | IN  | AC  |
| Quartz             | 39.41 | 36.37 | 29.67 | 30.16 |
| Plagioclase        | 31.85 | 42.95 | 41.01 | 38.26 |
| Orthoclase         | 24.53 | 16.14 | 23.93 | 23.94 |
| Corundum           | 0.89  | 0.83  | 0.19  | 0.04  |
| Hypersthenite      | 2.85  | 3.36  | 4.45  | 6.86  |
| Ilmenite           | 0.24  | 0.19  | 0.39  | 0.39  |
| Magnetcite         | 0.09  | 0.10  | 0.12  | 0.17  |
| Apatite            | 0.14  | 0.06  | 0.23  | 0.18  |
Elemental analyses of the rock samples showed the presence of the following 30 elements: Ba, Bi, Ce, Co, Cr, Cs, Cu, Ga, Ge, Hf, La, Mo, Nb, Ni, Pb, Pr, Rb, Sb, Se, Sm, Sn, Sr, Ta, Th, U, V, W, Y, Zn and Zr. Minor element contents in the rock samples were in the order Ba > Sr > Rb > Zr > Nb > Sb > Cs > Cr > Se > Zn > V. High contents of barium and strontium were directly related to the decomposition of dolomite and calcite minerals.

Normative mineral composition (norm) was calculated using the bulk chemical analysis of the rock samples from the four hot springs. Major and minor element compositions of rock samples were used as input data for this calculation. Silicate minerals (quartz, plagioclase, orthoclase, dolomite, mica and actinolite) and oxide minerals (corundum, ilmenite, magnetite) and phosphate minerals (apatite) were present as normative minerals (Table 6).

Mineralogical characterization of rock samples

XRD analyses of the rock samples detected silicate- and carbonate-type minerals such as quartz, albite, orthoclase, dolomite, mica and actinolite. Albite and mica are typically found in the metamorphic rock layer (Joshi and Bhandari 2013; Bleam 2016). The results of the XRD analysis are presented in Figure 5.

Water–rock interactions

Results of the correlation analysis of elemental compositions of hot spring water and rock are presented in Table 7. The correlation analysis for water–rock interaction was calculated using centered log-ratio (CLR) transformation using R software by the R Core Team. The sum of all components was equal to 100% and the f(X) value of correlation analysis was 0.91. As shown in Table 7, the following components showed strong positive correlation with each other: Na(w) and f(CaO(r)); Cl(w) and f(Na2O(r)); Ca(w) and f(CaO(r)) whereas the following components showed strong negative correlation: Na(w) and f(Al2O3(r)) and CaO(r), MnO(r). The positive correlations indicate the interaction between hot spring water and reservoir rock.

Geochemical models of minerals

Rock transformation processes occur over thousands of years and are affected by natural factors. We used the following models to explain the effect of rock weathering processes on the composition of hot springs in the Mongolian Altai region. Water from these hot springs predominantly contain HCO3−-Na, SO42−-Na, and HCO3−-SO42−Na ions, and rocks contain albite, orthoclase, dolomite, calcite, mica and other minerals.

**Geochemical model of the HCO3−-Na-type hot spring.** Hot spring water + albite, orthoclase, and dolomite → HCO3−-Na + (SiO2 + Al2O3 + K2O + Na2O)red

\[
2\text{KAISiO}_8 + 2\text{H}_2\text{CO}_3 + 9\text{H}_2\text{O} \rightarrow 2\text{K}^+ + 2\text{HCO}_3^- + 4\text{H}_2\text{SiO}_4 + \text{Al}_2\text{Si}_2\text{O}_5\text{(OH)}_4 \quad (1)
\]

\[
2\text{NaAlSi}_3\text{O}_8 + 2\text{H}_2\text{CO}_3 + 9\text{H}_2\text{O} \rightarrow 2\text{Na}^+ + 2\text{HCO}_3^- + 4\text{H}_2\text{SiO}_4 + \text{Al}_2\text{Si}_2\text{O}_5\text{(OH)}_4 \quad (2)
\]

Equations (1) and (2) show the dissolution of albite and orthoclase to form KHCO3, NaHCO3 and kaolinite (Oyuntsetseg 2014; Yuan et al. 2017). Moreover, kaolinite can react with hot spring water to form orthosilicic acid and gibbsite.

The interaction of dolomite and calcite minerals with water is shown in Equations (3) and (4), respectively:

\[
\text{MgCa(CO}_3)_2 + 2\text{CO}_2 + 2\text{H}_2\text{O} \rightarrow \text{Ca}^{2+} + \text{Mg}^{2+} + 4\text{HCO}_3^- \quad (3)
\]
CaCO₃(ð) + CO₂ + H₂O → Ca²⁺ + 2HCO₃⁻ (4)

According to Equations (3) and (4), an increase in the calcium and magnesium ion concentrations may lead to cation exchange reactions between sodium and calcium or magnesium. By contrast, minerals containing Na⁺ ions, such as albite, can interact with water and exchange dissolved Ca²⁺, which may cause an increase in Na⁺ ions in the water. The cation exchange reaction is (Kononov 1983; Yuan et al. 2017; Zhang et al. 2020).

Ca²⁺ + 2Na – mineral → 2Na⁺ + Ca – mineral (5)

Geochemical model of SO₄⁻–Na-type hot spring. HCO₃⁻–Na type water shown in the previous model forms SO₄⁻–Na type water and calcite compounds by interacting with gypsum. Equations (6) and (7) show the formation of SO₄⁻–Na type water, which occurs in response to the reactions between HCO₃⁻–Na type water and gypsum:

2NaHCO₃ + CaSO₄ + 2H₂O = Ca(HCO₃)₂ + Na₂SO₄ + 2H₂O (6)

CaSO₄ + 2H₂O → Ca²⁺ + SO₄²⁻ + 2H₂O (7)

According to Equation (8), hydrogen sulfide in groundwater can react with organic carbon and oxygen, and this may lead to the formation of sulfate ions:

H₂S + 2O₂ ↔ 2H⁺ + SO₄²⁻ (8)

Table 8. Reservoir temperature (T, °C)

| Sample ID | Average Quartz(1) | Quartzad(2) | Chalcedony(3) | Chalcedony(4) |
|-----------|-------------------|-------------|---------------|---------------|
| CH        | 96.2              | 111         | 110           | 82.2          |
| GM        | 82.4              | 96.8        | 98.2          | 68.2          |
| IN        | 106.4             | 121         | 119           | 92.6          |
| AC        | 84.4              | 98.8        | 100           | 70.1          |

(1, 2, 4) Fournier (1977); (3) Arnorsson (1983); ad, adiabatic.

Table 9. Temperatures and depths of the hot springs

| Sample ID | Reservoir temperature (°C) | Water circulation depth (m) |
|-----------|----------------------------|----------------------------|
| CH        | 96.2                       | 3070                       |
| GM        | 82.4                       | 2615                       |
| IN        | 106                        | 3410                       |
| AC        | 84.4                       | 2680                       |

Hot springs predominately contain SO₄²⁻, whereas cold springs predominately contain HCO₃⁻. While a temperature increase can lead to a decrease in the dissolution of carbonate ions, a significant rise in temperature can lead to an increase in the dissolution of sulfate ions (Kononov 1983; Joshi and Bhandari 2013).

Geothermal reservoir

Chemical geothermometers

Various types of geothermometers can be used to determine reservoir temperature, including silicate mineral (quartz and chalcedony) and cation (Na–K; Na–K–Ca) thermometers (Klein 2007; Peiffer et al. 2014). These methods are based on the interactions between water and rock (Dolgorjav 2009). We used the methods of Fournier (1977) and Arnorsson (1983) for the geothermometer calculations, and the results are shown in Table 8.

As shown in Table 8, the quartz and chalcedony geothermometry values were ranged between 82.4 and 106.4°C. But the reservoir temperature might be underestimated due to partial equilibration. Moreover, silica geothermometers generally give lower temperature estimation than other cation geothermometers. Hot spring circulation depth is estimated by SiO₂ geothermometers (quartz and
chaledony). The circulation depth ranged from 2615 to 3410 m (Table 8).

**Na–K–Mg ternary diagram.** A ternary diagram (Na–K–Mg) was used to estimate the reservoir temperature and equilibrium state of the water and rock interactions (Sebesan et al. 2015; Fan et al. 2019) (Fig. 6). As shown, hot springs in the Mongolian Altai region included a partially equilibrated water zone. Generally, water–rock interactions are divided into three zones: mature water, partially equilibrated water and immature water. As shown in Figure 6, the hot springs in the Mongolian Altai region include a ‘partially equilibrated water’ zone on the Giggenbach plot. This indicates that the water–rock interactions occurred partially, not fully. On the other hand, mixing with rainwater, cold spring or groundwater in the aquifer may prevent the achievement of complete equilibrium.

**Geothermal water circulation depth**
We calculated the reservoir depth of the hot springs by measuring reservoir temperatures and estimating local geothermal gradients (Shestakova et al. 2018). These are shown in Table 9. The geothermal gradient of the Mongolian Altai region was calculated as 20–30 °C km⁻¹ (Lkhagvadorj and Tseesuren 2005).

According to Mongolian statistics on climate, the annual average temperature of the research area is between −2 and −4°C. Based on this information, we calculated the depth of water circulation as 2615 to 3410 m.

**Conceptual model for the geothermal water of the CH hot spring reservoir**
We propose a conceptual model for the geothermal water for the CH hot spring reservoir in the Mongolian Altai region, based on its geology, the chemical composition of hot springs, and reservoir temperature and depth. The conceptual model of the CH hot spring is shown in Figure 7.

According to the model, precipitation (rain, snow, etc.) from the atmosphere reaches the hot spring system through tectonic faults and other possible geological features at a depth of 2550 m a.s.l., wherein the water interacts with aluminosilicate minerals (Byamba et al. 1999; Bulgatov et al. 2014). The dissolution of albite, dolomite and gypsum minerals generates HCO₃⁻–Na and SO₄⁻–Na types of water in this area. Moreover, rock weathering and precipitation play an important role in the CH hot spring system. According to our calculation, the reservoir temperature of this area is c. 96.2°C at a depth of 3070 m below surface, which corresponds to a geothermal source with a low temperature (Nicholson 1993).

**Conclusions**
In this study, we investigated the chemical composition, mineral components of rocks, geothermal temperature and reservoir depth of hot springs in the Mongolian Altai region.

The chemical compositions of the hot springs revealed that the springs contain HCO₃⁻–Na, SO₄⁻–Na, and HCO₃⁻–SO₄⁻–Na types of water with low TDS values in the alkaline pH range, and their temperatures ranged as 21.3–35°C. According to the Na–K–Mg ternary diagram, hot springs in the Mongolian Altai region have a partially equilibrated state of water–rock interactions. The results suggest that water–rock interactions and the geographical features of the reservoir are responsible for the main hydrochemical properties.

A preliminary study of the reservoir temperature found values ranging from 82.4 to 106.4°C. This result may be estimated lower because of the partial equilibrium. The reservoir depth of the hot springs ranged from 2615 to 3410 m according to the quartz and chaledony geothermometers. We constructed a conceptual model for the geothermal water in the Mongolian Altai region, which can be used in other areas with similar geological settings.

Hot spring water in the Mongolian Altai region has the potential to be used in heating systems. Corrosion and mineral accumulation in heating systems based on these hot springs would be low, based on their alkaline properties and low TDS concentration. This type of hot spring resource could thus be directly used for heating households, greenhouses and swimming pools.

A limitation of this study is the lack of isotopic research on the hot springs and water resource calculations. Further research, including studying the isotopic characteristics of the thermal waters in the Mongolian Altai region, will be carried out in the future. Water source investigations are important for utilizing thermal groundwater in heating systems in remote areas of Western Mongolia.

**Acknowledgments** We thank Dr Eric Steemlter for helping with R software and CLR transformation.

**Author contributions** BC: data curation (lead), formal analysis (lead), investigation (lead), validation (equal), writing – original draft (equal), writing – review & editing (lead); DM: data curation (equal), formal analysis (supporting), investigation (equal), writing – original draft (equal), writing – review & editing (equal); BA: formal analysis (supporting), software (supporting), visualization
Scientific editing by Scott Alan Wood

References

Amorsson, S. 1983. Chemical equilibria in Icelandic geothermal systems—implications for chemical geothermometry investigations. Geothermics, 12, 119–128, https://doi.org/10.1016/0375-6505(83)90022-6

Amorsson, S., Gunnlaugsson, E. and Svarrason, H. 1983. The chemistry of geothermal waters in Iceland. III. Chemical geothermometry in geothermal investigations. Geochimica et Cosmochimica Acta, 47, 567–577, https://doi.org/10.1016/0016-7037(83)90278-8

Besser, H., Mokadem, N., Redhaounia, B., Hadji, R., Hamad, A. and Hamed, Y. 2014. Geothermal system: analysis and geochemical assessment of low-enthalpy resources in the geothermal field of southwestern Tunisia. Euro-Mediterranean Journal for Environmental Integration, 3, 16, https://doi.org/10.4114/2014-018-0055-2

Bignold, J.G., Dori, P., Bakhshig, B. and Tsuchiya, N. 2005. Geothermal resources and development in Mongolia. Proceedings of the World Geothermal Congress, s.n., Antalya, Turkey, 45–51.

Bleam, W. 2016. Soil and Environmental Chemistry. 2nd edn. Academic Press, Amsterdam.

Blomdin, R., Heyman, J. 2011. Geothermometric calculations for geothermal waters in Iceland. III. Chemical geothermometry in geothermal investigations. Geochimica et Cosmochimica Acta, 52, 2749–2765, https://doi.org/10.1016/0016-7037(88)90143-3

Blomdin, R., Heyman, J. 2011. Geothermometric calculations for geothermal waters in Iceland. III. Chemical geothermometry in geothermal investigations. Geochimica et Cosmochimica Acta, 52, 2749–2765, https://doi.org/10.1016/0016-7037(88)90143-3

Bulgatov, A.N., Chen, B.

Funding

This work was funded by the Khovd University (A/51) and the Laboratory of Environmental Analysis, National University of Mongolia and Khovd University, Mongolia.

Data availability

All data generated or analysed during this study are included in this published article.

Scientific editing by Scott Alan Wood

Gendenjants, O.E. 2003. Interpretation of Chemical Composition of Geothermal Fluids from Erskysgurnd, Dalvík, and Hrnsý, N-Iceland and in the Khangai Mountain, Mongolia. United Nations University, Reykjavik, Iceland.

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4

Giggenbach, W.F. 1988. Geothermal soln. equilibria. In: D.D., N.G. McAdam, N.G. McAdam and K. G. McAdam, Coauthors. Geochemistry & Cosmochimica Acta, 52, 777–796, https://doi.org/10.1016/0016-7037(85)90007-4
Yuan, J., Xu, F., Deng, G., Tang, Y. and Li, P. 2017. Hydrogeochemistry of shallow groundwater in a karst aquifer system of Bijie City, Guizhou Province. *Water*, 9, 625, https://doi.org/10.3390/w9080625

Zhang, B., Zhao, D., Zhou, P., Qu, S., Liao, F. and Wang, G. 2020. Hydrochemical characteristics of groundwater and dominant water–rock interactions in the Delingha area, Qaidam Basin, Northwest China. *Water*, 12, 836, https://doi.org/10.3390/w12030836

Žunić, L., Bidžan-Gekić, A. and Gekić, H. 2019. Balneological classification of thermoniminal, thermal and mineral waters at the region of Ilidza-Sarajevo and its impact on tourism. *Ad Alta: Journal of Interdisciplinary Research*, 9, 336–342, www.doi.org/10.33543/0901