

Article

Fluid Chemical and Isotopic Signatures Insighting the Hydrothermal Control of the Wahongshan-Wenquan Fracture Zone (WWFZ), NE Tibetan Plateau

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Abstract: Compared to the southern Tibetan Plateau, the northern part has been regarded as relatively lacking geothermal resources. However, there is no lack of natural hot springs exposed in beads along large-scale fracture systems, and research on them is currently limited to individual hot springs or geothermal systems. This paper focuses on the Wahongshan-Wenquan Fracture Zone (WWFZ), analyzes the formation of five hydrothermal activity zones along the fracture zone in terms of differences in hot water hydrochemical and isotopic composition, and then explores the hot springs' hydrothermal control in the fracture zone. The results show that the main fractures of the WWFZ are the regional heat control structures, and its near-north–south- and near-east–west-oriented fractures form a fracture system that provides favorable channels for deep hydrothermal convection. Ice and snow meltwater from the Elashan Mountains, with an average elevation of more than 4,500 m above sea level, infiltrates along the fractures, and is heated by deep circulation to form deep geothermal reservoirs. There is no detectable mantle contribution source heat to the hot spring gases, and the heat source is mainly natural heat conduction warming, but the “low-velocity body (LVB)” in the middle and lower crust may be the primary heat source of the high geothermal background in the area. The hot springs' hydrochemical components show a certain regularity, and the main ionic components, TDS, and water temperature tend to increase away from the main rupture, reflecting the WWFZ controlling effect on hydrothermal transport. In the future, the geothermal research in this area should focus on the hydrothermal control properties of different levels, the nature of fractures, and the thermal contribution of the LVB in the middle and lower crust.

Keywords: hydrogeochemistry; geothermal gas isotopes; geothermal systems; Wahongshan-Wenquan Fracture Zone (WWFZ)



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1. Introduction

The Tibetan Plateau has the most vigorous geothermal activity in China, with the highest geothermal reserves and various surface geothermal manifestations throughout the region [1,2]. However, it is traditionally believed that the geothermal resources of the Tibetan Plateau are mainly concentrated in tectonically unstable areas such as southern Tibet, western Yunnan, and western Sichuan, while its northern and northeastern parts are in relatively stable blocks with average geothermal potential. Relevant studies have also focused on the north–south Himalayan rift zone [3–5], the Yunnan–Tibet geothermal zone [6,7], and several fault zones where hot springs in western Sichuan are relatively concentrated [8–14]. Only with the exploration of high-temperature hot dry rock resources

in the Gonghe Basin on the northeastern margin of the Tibetan Plateau in recent years [15] has the region gained widespread attention, and a large number of studies have been carried out to address the mechanism of the heat source and the potential for development [16–22].

The Wahongshan-Wenquan Fracture Zone (WWFZ), located in the southwest of the Gonghe Basin, belongs to the southwest boundary fault of this basin, and is the separation zone between the southwest of Gonghe Basin and the northeast of the East Kunlun magmatic belt (Figure 1). Several hot springs were exposed in the northwest–southeast direction of the fault zone, indicating that the fault controlled the distribution of hot springs in this area to a certain extent. Therefore, research on the genesis of hot springs has significant implications for a deeper understanding of the heat control mechanism of the WWFZ, as well as on the deeper analysis of the Gonghe Basin’s high-temperature thermal background. The complexity of fracture systems in orogenic zones and the spatial heterogeneity of the permeability of their fracture–matrix systems are the difficulties in carrying out research on geothermal systems [23], while hydrogeochemical analysis can avoid direct analysis of the medium and channels of geothermal systems; thus, the hydraulic linkages and the evolution process of the chemical components can be analyzed by comparing the hydrochemical characteristics of the different geothermal systems [24], in order to provide a basis for research on the genesis mechanism of the geothermal systems in the fracture zones. Previous authors have conducted relevant hydrogeochemical analyses for some hot springs. Zhang et al. (2016) carried out principal component analysis of the chemical composition of some hot spring water in the area [25]. They proposed the geochemical genesis of the main components and the related hydrogeochemical processes. Ma et al. (2020) carried out a fluid geochemical and water–rock interaction study of the hot water in the southeastern section of the WWFZ [26]. They proposed that aluminosilicate minerals dissolution and recrystallization is the most crucial hydrogeochemical process controlling cations. In contrast, mineral dissolution and filtration during runoff are the main sources of anions. Wang et al. (2023a) conducted geochemical analysis and studied the hot springs genesis mechanism on the southwest side fracture [27]. They proposed that the hot springs belong to a deep-circulating convective geothermal system controlled by the fracture, with a geothermal reservoir temperature as high as 245–261 °C, and a deep circulation depth of 7700–8200 m. The previous studies focused on individual hot springs. They lacked comparative analyses on the elements of different geothermal systems and their genesis mechanisms.

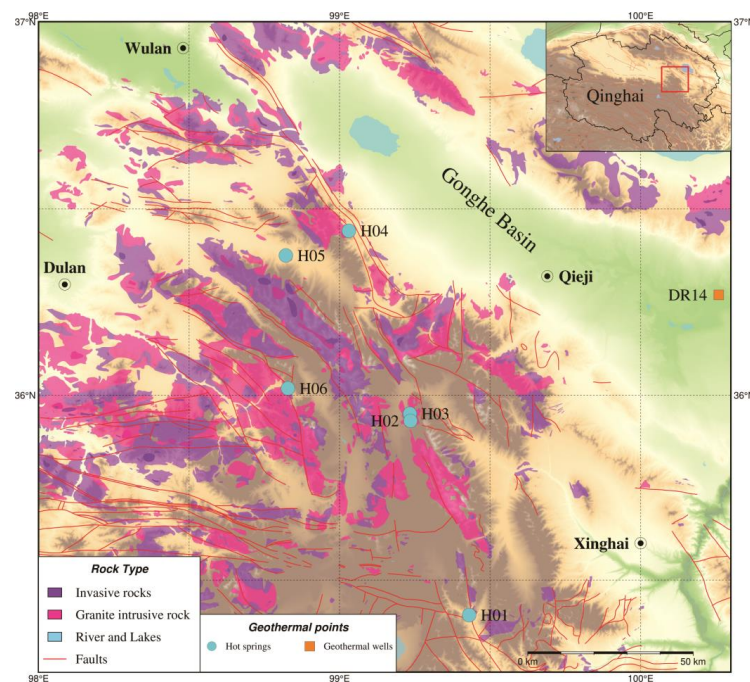


Figure 1. Wahongshan-Wenquan Fractures Zone (WWFZ) geologic map and hot springs distribution.

The tectonic setting is the result of the long-term evolution of the Earth, which determines the temperature, fluid composition, reservoir characteristics, and mode of heat transfer (convection or conduction) in geothermal systems [28]. Geothermal fields around the world are invariably tectonically controlled [29,30]. Fractures are the most prevalent heat-controlling tectonics [31,32]. Under favorable geological and tectonic conditions, meteoric water seeps down along the fracture, and is warmed by deep circulation; then, it gathers and surges upward to be exposed on the surface along different high-angle fracture zones or steeply inclined permeable strata [33]. Thus, bead-like localized thermal anomaly zones can be formed near the fracture zones. In this paper, geothermal fluid and gas geochemical samples of different geothermal anomalies in the WWFZ were systematically collected, and the fluid geochemical characteristics and geothermal reservoir characteristics of the area were comparatively analyzed in combination with the regional geological and tectonic conditions. Research on the heat source mechanism of different geothermal systems was systematically carried out, and the controlling role of the WWFZ on the geothermal system was explored, which will provide a reference for further geothermal exploration in the area in the future.

2. Study Area

The WWFZ starts from Agushar in the south, passes through the Harihad Mountains, and ends in Changshui in the south, with a length of about 275 km, and in the study area about 200 km. The main tectonic trace is NW–NNW oriented, which consists of several high-angle reverse faults trending in the north–northwestern direction, and it is an extrusive reverse strike-cum-rightward strike-slip fracture [34]. A few small-scale near-north–south and near-east–west-oriented fractures cut the study area into diamond-shaped fracture blocks. Along the fracture zone, there are banded bedrock, intermittently distributed ultramafic rock, and localized moderately acidic intrusive rock [35]. The WWFZ has a considerable depth of influence and is a crustal fracture. It was formed at the end of the early Paleozoic Era, and its activity increased from the Hualixi period to the middle Indo-Chinese period. In the late Indo-Chinese-Himalayan period, the region entered the stage of intra-land orogeny, and the WWFZ and a series of basically synchronized fractures formed the north–northwestern Erasas slip tectonic zone with right-lateral slip, which controlled the intrusion of the middle-acidic magmatism in the Indo-Chinese period, and the distribution of the regional late Triassic volcano-sedimentary basins [36]. There is a cluster of distributed small earthquakes, which is an active structure [37].

Along the WWFZ, there are hot springs such as H01 (31–64 °C, Wenquan Township, Xinghai County), H02, H03 (65–78 °C, Sangchigou, Xinghai County), H04 (35–46 °C, Bayinggeli), H05 (67 °C, Angutan, Dulan County), and H06 (87 °C, Reshuigou, Dulan County). These hot springs are located at the intersection of different fractures (Figure 1). Most of these hot springs are in a primitive and undeveloped state, except for the H01 hot spring, which is currently under construction with related tourism and sanatorium facilities for tourists to bathe.

3. Materials and Methods

Sample Collection and Testing

The distribution of sampling sites is shown in Figure 1. The water chemistry and gas samples were collected from May to July 2022. A total of 6 sets of hot water samples and five gas samples were collected, and the testing items included complete analysis, trace elements, stable isotopes, conventional gas fractions, noble gas fractions, and isotopes. The water samples were first filtered through a 0.45 µm microporous filter membrane, then stored in polyethylene bottles, rinsed twice with deionized water, and dried. Hot water chemical composition and stable isotope analyses were carried out at the Institute of Hydrogeology and Environmental Geology, Chinese Academy of Geological Sciences. Testing items included complete analysis and trace elements. The water samples were tested according to the National Standard of the People's Republic of China, "Methods

for examination of drinking natural mineral water" (GB/T8538-2008) [38]. A FP640 flame photometer and PHS-25 digital acidimeter were utilized for chemical composition testing, with a 3% anion/cation balance error. The isotope testing instrument was a Model L2130i Water Isotope Analyzer, and measurements were expressed as thousandths of a percent relative to the VSMOW standard. The drainage method was used to collect hot spring gas samples. They were analyzed and tested at the Oil and Gas Resources Research Center of the Northwest Institute of Ecology and Environmental Resources, Chinese Academy of Sciences (NWIER, CAS). Concentrations and isotopes of He, Ne, and Ar gases were analyzed using the Noblesse rare gas isotope mass spectrometer. Helium isotopes were tested with an error of $\pm 10\%$ for R-values above 1×10^{-7} and $\pm 15\%$ for R-values between 1×10^{-8} and 1×10^{-7} .

4. Results

4.1. Chemical Characterization of Hot Springs

The water chemistry of the different hot springs is shown in Table 1. Due to various geochemical processes occurring along the groundwater path circulation, in general, the chemical composition is relatively homogeneous. The cations are dominated by Na^+ , ranging between 150.0 and 861.0 mg/L. The Na^+ near the primary fracture is around 300 mg/L, and the lowest value occurs at H04, where the spring water temperature is also the lowest (46 °C). Away from the primary fracture, the Na^+ increased significantly, with the highest value (861 mg/L) occurring at H06 and with the highest spring temperature (87 °C). Cl^- and SO_4^{2-} dominate the anions, with ranges of 120.0–1461.0 and 36.3–329.0 mg/L, respectively. The trend of Cl^- was the same as that of Na^+ , and away from the primary fracture, the content of Cl^- was elevated, with the highest value (1461 mg/L) occurring at H06. There is no obvious pattern for SO_4^{2-} , which should be related to the strata through which the hot water of different hot springs flowed on the way up.

Table 1. Chemical compositions and stable isotopes of hot springs in WWFZ.

ID	H/masl	pH	T/°C	Chemical Composition/(mg/L)								Stable Isotopes/‰			Recharge Elevation /masl
				K^+	Na^+	Ca^{2+}	Mg^{2+}	Cl^-	SO_4^{2-}	HCO_3^-	TDS	SiO_2	$\delta^2\text{H}$	$\delta^{18}\text{O}$	
H01	3945	7.91	64.0	21.9	267.0	41.8	5.3	169.0	329.0	140.0	1054	77.7	−90	−11.6	5155
H02	4282	8.29	78.0	13.7	303.0	24.3	3.9	296.0	186.0	116.0	1052	76.6	−92	−11.7	5525
H03	4282	8.29	60	11.66	301.9	25.42	0.51	232.5	301.7	110.4	1065	127.27	/	/	/
H04	3794	8.82	46.0	2.6	150.0	16.3	2.2	120.0	156.0	29.7	552	43.2	/	/	/
H05	3790	8.08	67.0	23.8	482.0	78.5	1.7	610.0	282.0	56.5	1704	108.5	−88	−12.2	5200
H06	4001	7.52	87.0	125.0	861.0	77.7	8.0	1461.0	36.3	134.0	2866	113.9	−85	−11.5	5178

The hydrochemical types can roughly be categorized into the following two types: hot springs near the primary fracture are $\text{Cl}^- \text{SO}_4^- \text{Na}^-$ or $\text{SO}_4^- \text{Cl}^- \text{Na}^-$ -type waters, including H01, H02, H03, and H04; those far away from the fracture show $\text{Cl}^- \text{Na}^-$ -type waters, including H05 and H06. The hydrochemical type of geothermal water is intricately associated with its circulation conditions; if the circulation is faster, then the degree of interaction between it and the surrounding environment is smaller, and the hydrochemical components are relatively low. If the circulation cycle is longer, the runoff pathway is farther, the interaction of the hot water with its surroundings is enhanced, and the hydrochemical components are relatively high. Geothermal reservoirs in the study area, which are located in the tectonic magma belt of the Elashan Mountain, are characterized by granitic rocks. Dissolution and recrystallization of primary aluminosilicate minerals in the geothermal system are the most important hydrogeochemical processes for the control of cations in the hot water, whereas under-seepage of recharge water coupled with dissolution and filtration from salt minerals during the hot-water runoff and upwelling processes is the main source of anions. The hot water in this study area contains high concentrations of Na^+ , SO_4^{2-} , and Cl^- , of which SO_4^{2-} may originate from sulfate mineral dissolution, and may be affected by geochemical processes, for example, pyrite oxidation. In general, however, the water chemistry is mostly the same. All of the hot springs are clustered in the rightmost part of

the diamond on the Piper diagram (Figure 2). However, the difference lies in the size of the TDS, which ranges from 552 to 2866 mg/L, with the smallest value being H04 near the main fracture zone; the most considerable value occurs at H06, which is far away from the primary fracture, reflecting the proximity of different circulating runoff paths.

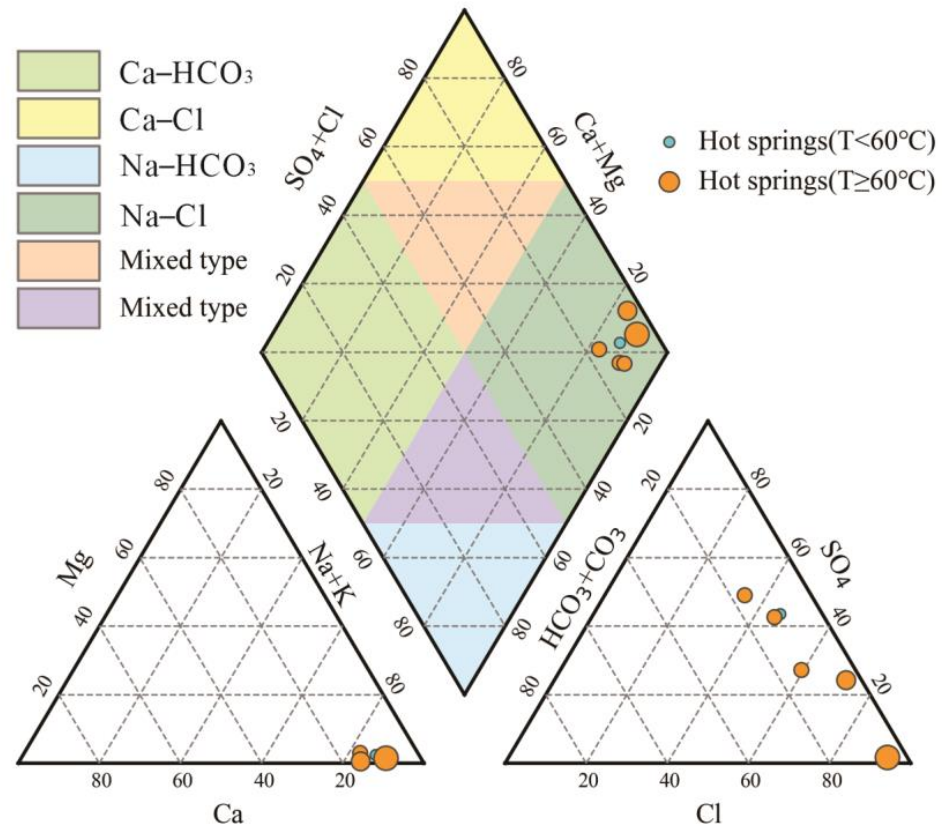


Figure 2. Piper diagram [39] of hot springs in WWFZ.

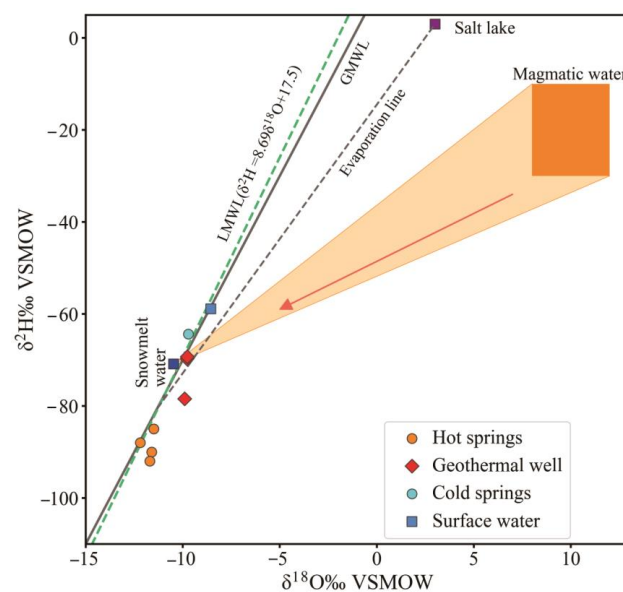
The ratio between some ions in groundwater can help to analyze the various states and groundwater geological effects, which can be used to analyze the hot water transport and hosting environment [40]. Different ionic characterization coefficients were calculated from the hot water chemistry data, as shown in Table 2. $\gamma\text{Na}^+/\gamma\text{Cl}^-$ characterizes the groundwater metamorphism coefficients, which can respond to the strata confinement, the degree of metamorphism of the groundwater, and its activity. The smaller the $\gamma\text{Na}^+/\gamma\text{Cl}^-$, the higher the metamorphism and the better environmental closure of the groundwater; the larger the $\gamma\text{Na}^+/\gamma\text{Cl}^-$, the more the groundwater is affected by infiltrating water. It can be seen that far away from the main fracture of the WWFZ, its $\gamma\text{Na}^+/\gamma\text{Cl}^-$ gradually becomes smaller, reflecting that the environment tends to be closed, and the influence of surface and shallow groundwater recharge is gradually reduced. $\gamma\text{Cl}^-/\gamma\text{Ca}^{2+}$ characterizes the hydrodynamic characteristics of groundwater; the larger $\gamma\text{Cl}^-/\gamma\text{Ca}^{2+}$ is, the condition of groundwater flow is poorer, and the groundwater flow is slower. As can be seen in Table 2, except for H02 and H03, in general, the flow conditions gradually become worse from the primary fracture to the two sides, and H02 and H03 are exposed in the open valley between the mountains. The flow conditions may have received the influence of the thicker quaternary cover on the surface. The salinization coefficient $\gamma\text{Cl}^-/(\gamma\text{HCO}_3^-)$ mainly reflects the concentration degree of stratum water, and the big value reflects the longer groundwater flow path and the slower water circulation. The $\text{Cl}^-/(\gamma\text{HCO}_3^-)$ values near the main fracture of the WWFZ are less than 7. In contrast, H05 and H06 are greater than 18, which reflects that the groundwater cycle is faster in the vicinity of the primary fracture, and gradually slows down far away from the primary fracture.

Table 2. Ionic characterization coefficients of hot water in WWFZ.

ID	Ionic Characterization Coefficients		
	$\gamma_{\text{Na}^+/\gamma_{\text{Cl}^-}$	$\gamma_{\text{Cl}^-}/\gamma_{\text{Ca}^{2+}}$	$\gamma_{\text{Cl}^-}/\gamma_{\text{HCO}_3^-}$
H01	2.44	9.11	2.07
H02	1.58	27.45	4.38
H03	2.00	20.61	3.62
H04	1.93	16.59	6.94
H05	1.22	17.51	18.55
H06	0.91	42.37	18.73

4.2. Hydrogen–Oxygen Stable Isotope Characterization of Hot Water

Hot springs are generally derived from meteoric water, and there is a depth effect in the infiltration process of meteoric water in the fracture zone, i.e., the $\delta^2\text{H}$ and $\delta^{18}\text{O}$ contents in hot water decrease with increasing depth. In the deep high-temperature environment, the hot water interacts with the surrounding rocks, resulting in “oxygen drift”, which increases the $\delta^{18}\text{O}$ content. Therefore, according to the hot springs’ isotopic composition, we can analyze underground hot water’s circulation process, recharge rate, and geological events (water–rock interactions) encountered in the fracture zones, as well as support the analysis of the fracture zones’ controlling role on the geothermal system. The stable isotope $\delta^2\text{H}$ – $\delta^{18}\text{O}$ relationships for geothermal wells, cold springs, salt lakes, rainwater, and snow water in the Gonghe Basin neighboring the WWFZ were analyzed for comparison (Figure 3). All of the collected water samples showed hydrogen and oxygen stable isotopes falling around GMWL and LMWL [41], suggesting that the recharge source is meteoric water. The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values of the hot springs in the WWFZ are significantly smaller than those of other water bodies within the Gonghe Basin, suggesting that the hot water in the study area is derived from meteoric water at higher elevations, or from meteoritic water in colder climates. The WWFZ is located in the Elashan Mountains with an average elevation 4500 m, in combination with the topography. The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values gradually decrease with rising elevation, making the $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values significantly smaller than those of other water bodies neighboring the Gonghe Basin. A map of the possible extent of magma water genesis (Figure 3), based on stable isotope intervals between ice-melt water and magma water in the area, reveals that there is no magma-influenced hot water in the area. The weak oxygen drifts in H01 and H02 suggest that water–rock interactions occur at hot water depths, which is in accordance with the Gibbs diagram results.

**Figure 3.** Stable isotope compositions of different hot springs in WWFZ.

Hot springs' recharge elevations can be determined based on meteoric water isotope elevation effects, calculated as follows [42]:

$$H = \frac{\delta s - \delta p}{k} + h$$

where H is the altitude of groundwater recharge; δs is the isotope value of hot springs at the sampling point; δp is the isotope value of meteoric water; k is the value of the isotope height gradient; and h is the altitude of the sampling point. The average value of the oxygen isotope of local rainwater was taken to be -7.97‰ , and the value of the isotope height gradient was 3‰ ; Table 1 shows the results of the calculations. The altitude of the hot spring recharge area is between 5155 and 5525 m, indicating that the recharge areas of the hot springs are all located in the central main ridge of the Ngola Shan Mountains where they are located.

4.3. Chemical Characterization of Hot Spring Gases

4.3.1. Hot Spring Gas Composition

The gases in hydrothermal areas (Table 3) can be broadly categorized into two types in the Tibetan Plateau according to their composition and genesis: CO_2 -type gases, which have both magmatic hot water and deep circulation genesis, and N_2 -type gases, which have only deep circulation genesis [43]. The gas concentration data (Figure 4) show that the major composition of all hot water is N_2 . The mean N_2 concentration value is $>80\%$, belonging to the N_2 type of gas, indicating that the genesis type of hydrothermal activity in the WWFZ belongs to the deep circulation category. In the samples, H_2 , H_2S , and SO_2 were undetectable due to their low concentrations. Higher levels of O_2 were found in hot spring water samples at higher temperatures, likely because atmospheric oxygen dissolves in the hot water as it rises. However, the contamination of the samples due to unregulated operation during sampling cannot be excluded. Atmospheric composition was not adjusted in this study to accurately characterize the area of hydrothermal activity. Most of the samples had relatively low CO_2 concentrations, averaging only 0.7% , except for sample H06, which had a relatively high CO_2 concentration of 7.83% .

Table 3. Gas compositions and isotopes of hot springs in WWFZ.

ID	Gas Components ϕ b (%)							Gas Isotopes		
	He	N_2	O_2	Ar	CO_2	CH_4	N_2/Ar	$^3\text{He}/^4\text{He}$	$^4\text{He}/^{20}\text{Ne}$	R/Ra
H01	0.1753	83.63	13.87	1.16	1.15	0.01	72.1	4.80×10^{-8}	101.03	0.03
H02	0.0142	79.29	18.53	0.96	1.2	-	82.6	1.00×10^{-7}	3.987	--
H04	0.0052	82.64	15.91	1.2	0.25	-	68.9	6.20×10^{-8}	7.22	--
H05	0.0375	73.67	17.3	1.13	7.83	0.04	65.2	3.90×10^{-8}	45.542	0.02
H06	0.0118	81.42	17.37	0.98	0.19	0.02	83.1	1.80×10^{-8}	11.921	--

Nitrogen (N_2) consistently appears as a major component in low- to medium-temperature geothermal systems. Nitrogen exhibits limited geochemical activity and possesses chemical stability akin to noble gases. Prior to the replenishment of geothermal water by meteoric water, N_2 in geothermal gases, along with Ar, is thought to originate from dissolved air. Following prolonged runoff in a reducing environment, the depletion of oxidizing gas components, such as oxygen, leads to the accumulation of rare gas components like nitrogen and argon, which become the dominant constituents [44]. Giggenbach and Glover He–Ar– N_2 diagrams show (Figure 5) that the N_2/Ar values of all samples fall within the interval between the meteoric water value (38) and the atmospheric value (84) and are closer to the atmospheric value (84), indicating that the N_2 component in the hot springs is dominated by atmospheric genesis [45]. The presence of helium (He) primarily stems from the radioactive decay of uranium and thorium within the Earth's crust [46]. However, it cannot be ruled out that certain samples (such as H01) are in proximity to the helium end, typically associated with mantle origins, hinting at the potential presence of a minor mantle component in these samples. Seismic

profiles along the Gonghe Basin–Yushu highway indicate vertical stacking of high- and low-velocity crustal bodies in Wenquan Township, with the Chabcha area and the LVB buried at depths of 15–43 km and with S-wave velocities below 3.4 km/s [47]. Recently, scholars studying the seismicity of the Gonghe Basin in light of deep hot dry rock exploration generally attribute the phenomenon to a young (Quaternary) intrusion of mantle-derived melt into the crust [48–51]. Therefore, the release of gases from the melt offers the potential for the occurrence of mantle-derived helium in surface hot springs.

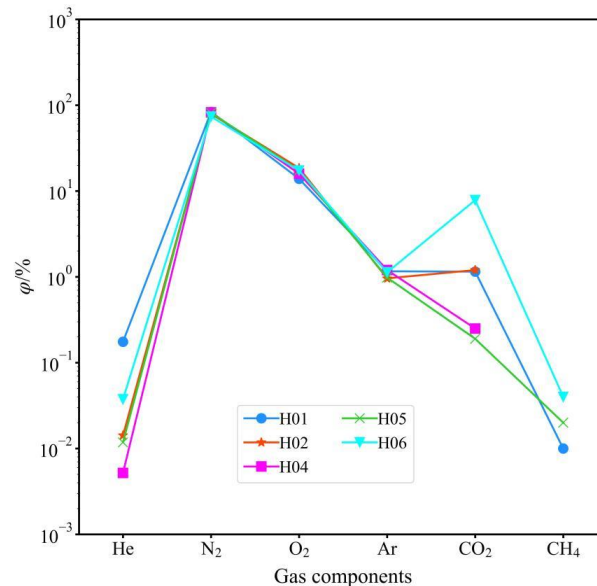


Figure 4. Chemical compositions of hot spring gases in WWFZ.

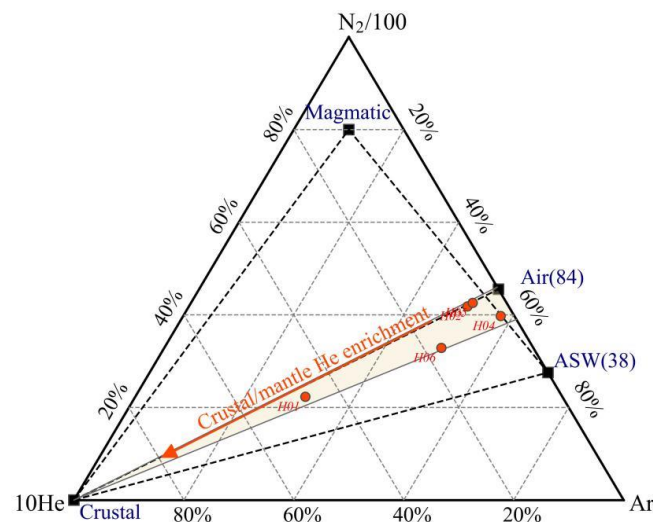


Figure 5. Relative He, Ar and N₂ contents of hot spring gases samples in WWFZ.

4.3.2. Isotopic Characterization of Hot Spring Gases

Helium (He) exists in two natural isotopes, ³He and ⁴He, with the predominant release during mantle degassing being geogenic ³He. This process also involves the transfer of mantle heat to shallow fluids as they ascend along tectonic channels. Conversely, the main origin of helium within the crust stems from the α -decay production of ⁴He through the natural radioactive decay of ²³⁸U, ²³⁵U, and ²³²Th [46]. Variations in helium isotope ratios among different source gases prompt the use of atmospheric helium ($R_a = 1.43 \times 10^{-6}$) as a reference; mid-oceanic ridge basalt (MORB) usually exhibits ratios around $(8 \pm 1) R_a$, representing the upper mantle, while crustal helium ratios range from 0.005 to 0.02 R_a [52].

Consequently, examining the helium isotope ratios in hot spring gases offers valuable insights into the regional geothermal context, fracture propagation patterns, activity, and fluid circulation dynamics.

As atmospheric components may inadvertently mix into samples during gas collection procedures, Duchkov (2010) suggested adjusting the R-value of geothermal gas samples using the $^4\text{He}/^{20}\text{Ne}$ ratio to mitigate the impact of atmospheric contamination during sampling [53]. The atmospheric ratio of $^4\text{He}/^{20}\text{Ne}$ is approximately 0.318. Therefore, when the $^4\text{He}/^{20}\text{Ne}$ ratio of a sample closely approaches 0.318, it indicates that the helium in the sample is predominantly of atmospheric origin. The correction equation is as follows:

$$R_c = \left[R_m \times \left(^4\text{He}/^{20}\text{Ne} \right)_m - R_a \times \left(^4\text{He}/^{20}\text{Ne} \right)_a \right] / \left[\left(^4\text{He}/^{20}\text{Ne} \right)_m - \left(^4\text{He}/^{20}\text{Ne} \right)_a \right]$$

In the equation, the subscripts c, m, and a correspond to the calibration value, test value, and characteristic value of the atmospheric source, respectively. Sano (1982) suggested the presence of three helium sources in natural gases: atmospheric, mantle, and crustal [54]. They devised a formula to determine the mixing ratio of different helium sources in gas samples using the ratios of $^3\text{He}/^4\text{He}$ and $^4\text{He}/^{20}\text{Ne}$, along with a graphical representation of the relationship between $^3\text{He}/^4\text{He}$ and $^4\text{He}/^{20}\text{Ne}$. This method streamlines the process and enhances the clarity of helium source identification. The formulas for helium source determination are as follows:

$$\left(\frac{^3\text{He}}{^4\text{He}} \right)_{\text{sample}} = \left(\frac{^3\text{He}}{^4\text{He}} \right)_A \times f_A + \left(\frac{^3\text{He}}{^4\text{He}} \right)_M \times f_M + \left(\frac{^3\text{He}}{^4\text{He}} \right)_C \times f_C$$

$$\frac{1}{\left(^4\text{He}/^{20}\text{Ne} \right)_{\text{sample}}} = \frac{f_A}{\left(^4\text{He}/^{20}\text{Ne} \right)_A} + \frac{f_M}{\left(^4\text{He}/^{20}\text{Ne} \right)_M} + \frac{f_C}{\left(^4\text{He}/^{20}\text{Ne} \right)_C}$$

$$f_A + f_M + f_C = 1$$

The subscripts A, M, and C represent atmospheric, mantle, and crustal sources, respectively, and f_A , f_M , and f_C represent the proportions of the three sources in the gas samples, of which $(^3\text{He}/^4\text{He})_A = 1.4 \times 10^{-6}$, $(^3\text{He}/^4\text{He})_M = 1.1 \times 10^{-5}$, $(^3\text{He}/^4\text{He})_C = 1.5 \times 10^{-8}$, $(^4\text{He}/^{20}\text{Ne})_A = 0.318$, $(^4\text{He}/^{20}\text{Ne})_M = 1000$, and $(^4\text{He}/^{20}\text{Ne})_C = 0.1000$. The results are shown in Table 2. The helium content in samples H01 and H05 primarily originates from the α -decay of ^{238}U , ^{235}U , and ^{232}Th in the crust's natural radioactive systems, accounting for over 97% on average. A plot of the $^3\text{He}/^4\text{He}$ - $^4\text{He}/^{20}\text{Ne}$ relationship of the geothermal gases in the study area is shown in Figure 6, and the He isotope R/Ra values (ratio relative to the atmosphere) of the hot-water gases in these two locations belong to the crustal helium domain (<0.05), indicating that they are predominantly crustal in origin.

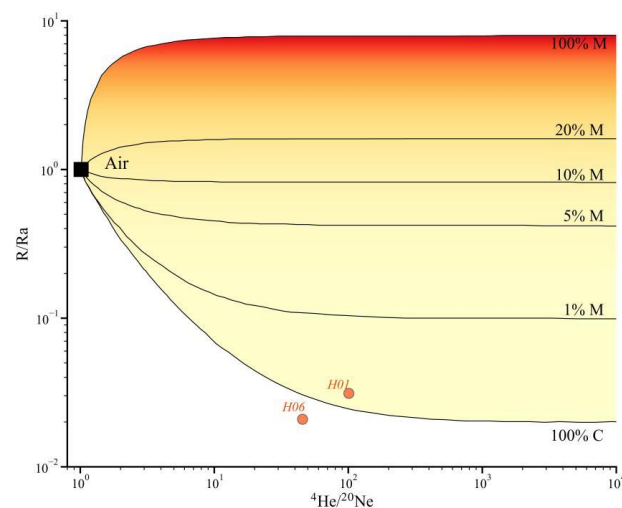


Figure 6. $^3\text{He}/^4\text{He}$ - $^4\text{He}/^{20}\text{Ne}$ relations of hot spring gases in WWFZ.

4.4. Geothermal Reservoir Temperatures and Circulation Depths

4.4.1. Water–Rock Equilibrium Analysis

The Na–K–Mg triangulation diagram, proposed by Giggenbach in 1988, categorizes water samples into fully equilibrated, partially equilibrated, and immature water regions, serving as a common tool to evaluate water–rock equilibrium and differentiate between different types of water samples [55]. The Na, K, and Mg concentrations of various hot springs in the study area were plotted on the Na–K–Mg equilibrium triangulation diagram (Figure 7). Four hot springs were identified in the immature water region at the apex of the lower-right corner, suggesting they are in the initial stage of water–rock interaction. Two hot springs, H03 and H05, were observed in the partially equilibrated or mixed water region, signifying ongoing adjustments in the ionic equilibrium between water and rock. Dissolution continues to occur, or the hot water may have undergone mixing with shallow cold water during its ascent. Moreover, examination of the Na–K–Mg equilibrium triangulation diagram indicates that the geothermal reservoir temperatures of all hot spring sites fall within the range of 120 °C to 270 °C. Notably, H06 exhibits the highest potential geothermal reservoir temperature, aligning with the highest temperature recorded in its spring water.

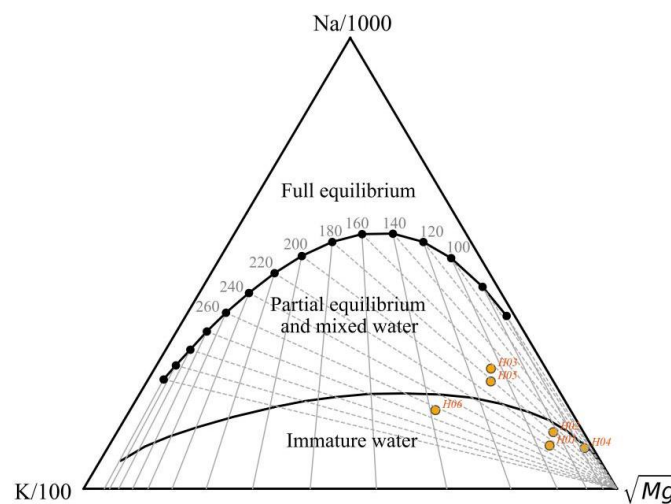


Figure 7. Na–K–Mg diagrams for various hot springs in WWFZ.

4.4.2. Geothermal Reservoir Temperatures by Multi-Mineral Saturation Indices

Reed and Spycher (1983) proposed a multi-mineral equilibrium mapping method for assessing the whole chemical equilibrium between fluids and minerals in geothermal systems [56]. The underlying principle is that the solubility of numerous minerals in water is temperature dependent. If a set of minerals at a specific temperature is nearly in equilibrium simultaneously, it suggests that the hot water and the mineral group have attained equilibrium, with the equilibrium temperature representing the temperature of the deep geothermal reservoir.

Aluminum-bearing silicate minerals could not be produced because the aluminum concentration in the hot water in the study area was not detected below the detection limit. In many geothermal systems, one or two aluminum-containing silicate minerals have typically achieved equilibrium. Consequently, Pang and Reed (1998) suggested employing aluminum fixation to restore equilibrium to aluminum-containing silicate minerals in geothermal systems [57]. Under multi-component, non-homogeneous chemical equilibrium conditions in geothermal systems, the balance of alumino-silicate minerals is interdependent because each of these minerals has a specific chemical composition. When the concentration of aluminum is unknown, it can be estimated at various temperatures by assuming that the activity of aluminum is fixed by an aluminum-bearing mineral such as microplagioclase feldspar. A value for the aluminum concentration is obtained, and

then used to calculate saturation indices for other aluminosilicate minerals. This method of estimating aluminum concentration values applies to specific geothermal waters because most geothermal waters require at least two aluminosilicate minerals to be in equilibrium.

Based on the hydrochemical data, we calculated each mineral's saturation index at different temperatures. We plotted the $\log(Q/K)$ – T curve (Figure 8) using the temperature and $\log(Q/K)$ values as horizontal and vertical coordinates, respectively. Based on the data from the petromineralogical investigations in the study area, we made a choice of 11 common minerals: quartz (Quartz), chalcedony (Chalcedony), potassium feldspar (K-feldspar), calcite (Calcite), fluorite (Fluorite), aragonite (Aragonite), gypsum (Gypsum), and magnesite (Magnesite), and plotted their $\log(Q/K)$ – T curves (H02 and H03 are located in the same geothermal field, and only the higher-temperature H02 hot spring was calculated). Among them, the curves of H01, H02, H05, and H06 converged a certain temperature, and their geothermal reservoir temperatures were 135 °C, 139 °C, 152 °C, and 159 °C, respectively. The curve of H04 is more dispersed, and its geothermal reservoir temperature was about 80 °C. H06 has the highest geothermal reservoir temperature among all of the hot springs, which is consistent with the highest temperature of its spring water.

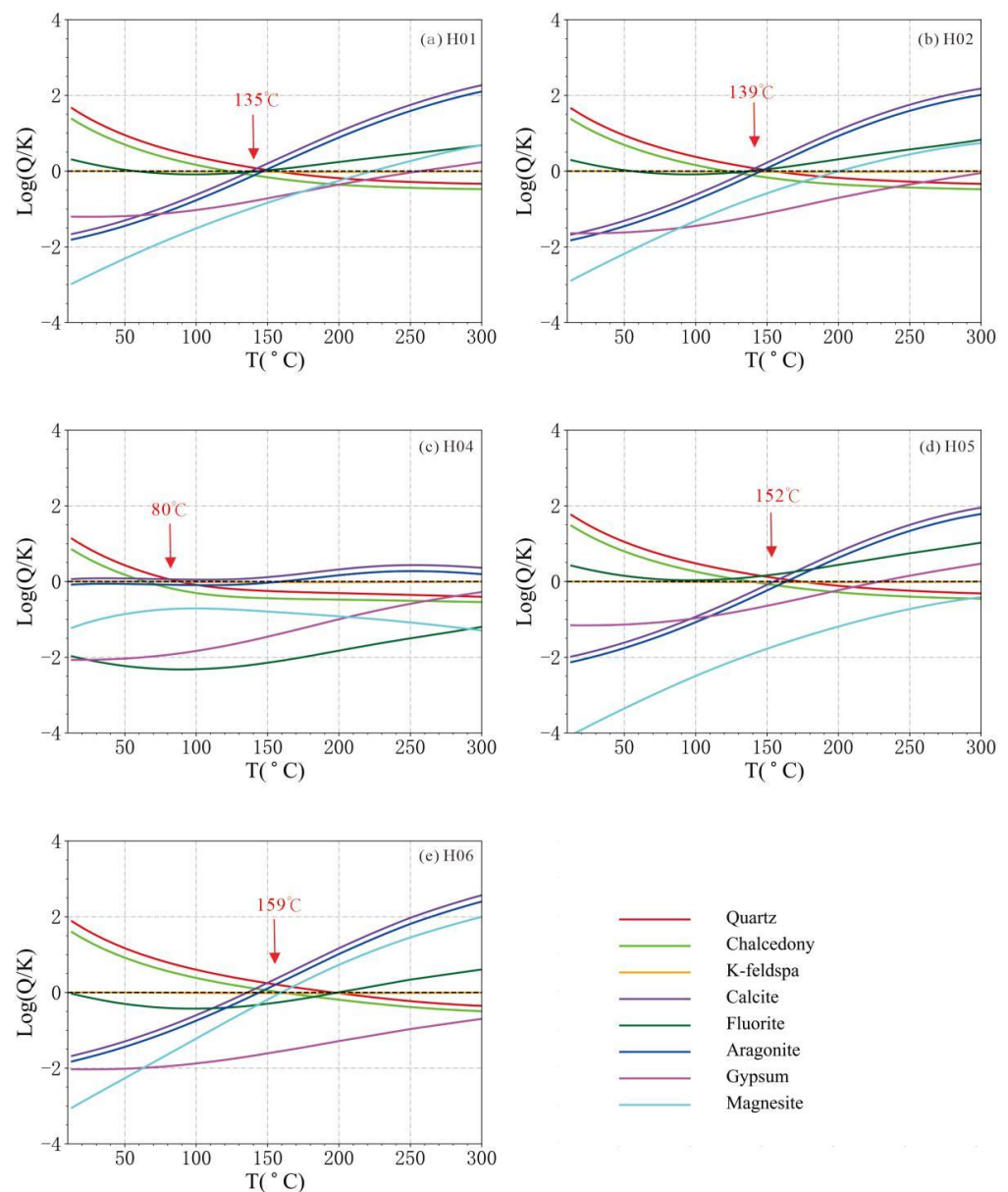


Figure 8. Mineral saturation diagrams for a typical hot spring in WWFZ.

4.4.3. Geothermal Reservoir Temperature Estimation by Geothermometers

According to the previous analysis, the hot spring waters are all immature or partially equilibrated (mixed); this did not meet the conditions for cationic geothermometers, so we adopted the silica geothermometer to estimate geothermal reservoir temperatures. The total concentration of silicon is a function of temperature. When there is no vapor loss from hot water during its ascent to the surface or when only conduction cooling is present, the quartz silica geothermometer uses the no vapor loss expression. When hot water has a maximum steam loss at a given temperature during ascent to the surface or when adiabatic cooling is predominant, the maximum steam loss geothermometer expression is used. Calculations are shown in Table 4.

Table 4. Geothermal reservoir temperatures estimated by different methods and circulation depths of hot springs.

ID	T (°C)	SiO ₂ (Maximum Steam Loss)	SiO ₂ (No Vapor Loss)	Chalcedony	Silicon Enthalpy Diagram	Multi-Mineral Saturation Indices	COLD Water Mixing Ratios/%	Circulation Depths/m
H01	64	121.2	123.6	95.5	193	135	71.0	2732–2985
H02	78	120.6	122.8	94.7	166	139	56.9	2714–3074
H04	44	96.8	95.1	64.7	168	80	79.0	1763–2098
H05	67	136.6	141.8	115.6	230	152	74.5	3136–3363
H06	85	139.0	144.6	118.7	205	159	62.4	3198–3518

4.4.4. Silicon-Enthalpy Mixing Model for Estimating Cold Water Mixing Ratio

The silicon-enthalpy mixing model (SEMM) estimates the mixing proportion of shallow cold water during hot spring ascent. This method obtains the mixing proportion of cold water and estimates the hot spring's geothermal reservoir temperature. However, according to Dong (2000), the result of the SEMM is too high, which may be due to the loss of steam from the expansion of the underground hot water before mixing with the cold water; thus, the measured SiO₂ content of the spring water is too high [58]. The overall geothermal reservoir temperatures calculated using the SEMM are relatively high (Table 4) and are for reference only. The results show a significant amount of cold water mixing in the springs throughout the study area, with the percentage of cold water mixing ranging from 56.9 to 79.0%.

4.4.5. Hot Water Circulation Depth Estimates

Combining the equilibrium states of each hot spring's water, the multi-mineral equilibrium analysis, and the results of different geothermometers (Table 4), the quartz no vapor loss geothermometer is closest to the multi-mineral equilibrium analysis results, so we chose the interval of the results of two methods as the interval of the possible geothermal reservoir temperatures (Figure 9).

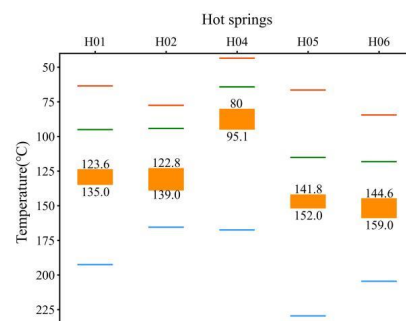


Figure 9. Geothermal reservoir temperatures of different hot springs obtained by different methods. Orange bars show geothermal reservoir temperature ranges as determined by quartz geothermometers (no vapor loss) and the multi-mineral saturation index method; blue lines illustrate geothermal reservoir temperatures derived from mixed silicon enthalpy models; the green lines indicate the geothermal reservoir temperature obtained from the chalcedony geothermometer; and the red line shows the temperature of the hot springs at the surface.

According to the monitoring data, the average annual temperature of the Xinghai, Wulan, and Dulan areas is 2.7 °C, and the depth of the normal temperature zone is about 45 m. The geothermal temperature gradient is about 4.5 °C/100 m, and the ranges of circulating depths of different hot water springs obtained from the calculations are shown in Table 4. Among them, the circulating depth of H06 is the deepest, about 3198–3518 m, and the hot water circulating in H04 is the shallowest, which is only about 1763–2098 m.

Ma et al. (2020) used quartz geothermal geothermometers and calculated the six springs with thermal storage temperatures of 109.3–137.0 °C in the H01 hot spring group [26], which are basically consistent with the results of 123.6–135.0 °C estimated in this paper. Wang et al. (2023a) used quartz geothermometers, the silica enthalpy model, and the Na-Li cation geothermometer to calculate the thermal storage temperatures of the H06 hot spring [27]. The results were as high as 245–261 °C, much higher than the results of 144.6–159 °C in this study. Based on the preceding analysis, it is evident that H06 falls into the category of immature water, rendering the cationic geothermometer inapplicable. Additionally, while the Na-Li geothermometer is suitable for carbonate rocks, its accuracy diminishes for other lithologies. The silicon-enthalpy mixing model assumes that the dissolved SiO₂ in the deep hot water is saturated and subject to the secondary equilibrium of the water–rock reaction that does not occur after mixing with cold water, while the actual situation is often far from the ideal model. Therefore, the authors consider the results to be questionable.

5. Discussion

Fractures are often favorable channels for water (hot or cold), which may affect the geothermal conditions in a particular range to a certain extent, thus forming geothermal anomaly zones. The WWFZ is a crustal fracture that cuts through the entire crust and contacts the upper mantle, which is favorable for mantle heat flow to be transported upward along the fracture, forming a regional heat-controlling fracture. At the same time, the fracture is cut by some near-north–south and near-east–west secondary fractures, and these fracture systems with different depths and crisscrosses provide favorable channels for deep hot water convection circulation in the region. By comparing and analyzing the fluid geochemical characteristics, geothermal reservoir depths, recharge heights, and heat source mechanisms of different hot springs within the WWFZ, a model of the water–heat cycle has been established (Figure 10).

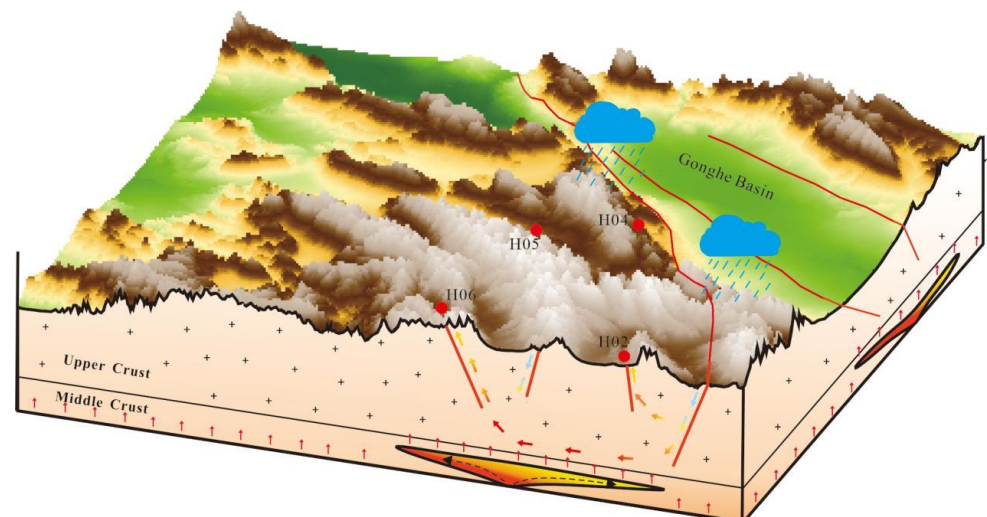


Figure 10. Hot spring formation pattern in the WWFZ.

5.1. Water and Heat Sources

The $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values of all hot springs in the WWFZ are significantly lower than those of other water bodies in the Gonghe Basin, suggesting that hot springs in the study area are recharged from meteoric water at higher elevations or from meteoric water in

colder climates. According to the hydrogen and oxygen isotopes of the hot spring water, it is estimated that the recharge altitude is between 5155 and 5525 m, which means that all the hot spring water is derived from the snow and ice-melt water of the Ngola Shan Mountains, which is 5100 m above sea level. The gas test results show that all of the hot springs in the area are N_2 -type gases, i.e., the hot water is caused by deep-cycle heating. The He isotopes of the gases fall into the helium domain of the crust, i.e., the gases are mainly generated by α -decay of the natural radioactive systems of ^{238}U , ^{235}U , and ^{232}Th in the crust, with no additional heat contribution from the mantle source. Low-velocity bodies of unknown nature in the lower and middle crust of the region may have provided an additional source of heat to the high-temperature thermal background of the study area [16,19,47–51], where hot springs were formed by deep circulation heating of snow and ice meltwater.

5.2. Hydrothermal Transport Channel

The northwesterly Wahongshan-Wenquan fracture and the transverse near-north–south or near-east–west fracture form a fracture system that provides a favorable channel for deep hot water convection circulation in the study area. According to the fluid geochemical characteristics of different hot springs in the area, the hot springs can be categorized into two runoff channels: long and short. Hot springs near the primary fracture of the WWFZ have relatively short runoff pathways, better deep hydrodynamic conditions, relatively open underground environments, and lower TDS of hot water. Away from the primary fracture, the content of Na^+ and Cl^- ions in the hot springs increased significantly, and the groundwater circulation rate slowed down, showing the characteristics of deep circulation and long runoff type.

5.3. Geothermal Reservoir

The study area is a convective geothermal reservoir system, and the fracture system formed by the north–west-trending WWFZ and a small number of fractures that transect the near-north–south or near-east–west direction constitute the primary geothermal reservoir system. Except for H04, which has a low geothermal reservoir temperature (about 80–95.1 °C) due to its shallow hot-water runoff depth of about 1763–2098 m, the rest of the hot springs have a geothermal reservoir temperature higher than 120 degrees; the highest one is located in H06, with a high geothermal reservoir temperature of 144.9–159.0 °C, and accordingly, its hot-water runoff is also the deepest (3198–3518 m).

5.4. Insulation Cover

For convective geothermal systems, the role of the cover is generally not noticeable. The poor hydrodynamic conditions (large $\gamma Cl^- / \gamma Ca^{2+}$ values) of the H02 and H03 hot springs near the primary fracture may be related to the thicker quaternary cover of the intermountain valley in which they are located.

To summarize, under normal or high regional thermal background, Elashan Mountain ice-melt water or meteoric water seeps down through the fracture network or fracture zone of the main fracture of the WWFZ, and runs off to the depths driven by the difference in the hydraulic head, with a gradual increase in temperature at depth; the groundwater continuously draws heat from the surrounding rocks to form geothermal water, and then rises up to the surface due to the difference in density, and is exposed in the appropriate tectonic part (generally at the intersection of two sets of fractures) to form hot springs. Some of the hot springs have a shallow circulation depth and short runoff path, and are exposed in the secondary fractures near the main fracture of the WWFZ, with a lower temperature (H04); meanwhile, the hot springs farther away from the main fracture (H06) have the highest spring water temperature and geothermal reservoir temperature, and their hydrochemistry type is Cl^-Na , and the TDS is nearly twice as high as that of the hot springs near the main fracture zone, reflecting their longer runoff path, deeper circulation depth, and the controlling effect of the main fracture and its secondary fractures of the WWFZ on the geothermal system in the area.

6. Conclusions

The Wahongshan-Wenquan Fracture Zone is located in the northeastern Tibetan Plateau, and several hot springs are exposed in the zone. In contrast to previous studies focusing on individual hot springs, this paper examined the hydrothermal control of the area's hot springs collectively, beginning with variations in hydrochemistry and isotopic composition across all hot springs. The results show that the ice and snow meltwater from the Elashan Mountains is the water source for all the hot springs in the area, and its heat source is dominated by the normal heating of the crust, and no contribution of mantle-source heat is detected. Ice and snow meltwater infiltrated along the fractures and formed a deep geothermal reservoir by deep circulation heating. The chemical components of the hot springs' water display a certain regularity in the plane, and away from the main fault, their main ionic components, TDS, water temperature, etc. have a tendency to increase by degrees, reflecting the controlling effect of the Wahongshan-Wenquan Fracture on the hydrothermal conditions in the area. However, most of the hot springs are located in areas that are rarely visited by people, and it is difficult to develop and utilize them. Based on the understanding of this research, it is important to systematically carry out the distribution and characterization of different levels of fractures in the region, as well as determine the target area of geothermal development potential in the concentrated population zones, which is the focus and difficulty of future research.

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