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Hydrological and Chemical Budgets of Okama Crater Lake in Active Zao Volcano, Japan

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Abstract: The Okama Crater Lake is located in the highly active Zao Volcano on the boundary of Miyagi and Yamagata Prefectures, Japan. At present, the lake stays relatively calm with neither bubbling, steaming nor gas smell at a pH of 3.2–3.4, though the lake did change color with steaming from the water surface in 1939 as the result of one of Zao’s volcanic activities. In order to clarify the geothermal effect on Okama and the groundwater flow system below or around Okama, field observations were performed in 2019 and 2020. Groundwater inflow and outflow in Okama were separately evaluated by estimating the hydrological and chemical budgets of the lake, based on the hydrometeorology, water temperature and river inflow measured in the field. The average groundwater inflow and outflow were estimated at 0.012 m$^3$/s and 0.039 m$^3$/s during the non-rainfall periods of 2020, respectively. A surplus of groundwater outflow makes the lake level consistently decrease during non-rainfall periods or the completely ice-covered season. In the completely ice-covered periods, the water temperature consistently increased at 0–15 m above the lake bottom, which is probably due to thermal leakage from a hydrothermal reservoir below the lake bottom. The heat fluxes averaged over December 2019–April 2020 and December 2020–March 2021 were calculated at 2.5 and 2.9 W/m$^2$, respectively. A coupling between the estimated groundwater inflow and the calculated geothermal heat flux was used to evaluate the temperature of inflowing groundwater.

Keywords: Okama Crater Lake; Zao Volcano; groundwater flow system; hydrological budget; chemical budget; geothermal heat flux

1. Introduction

When a volcano has high potential to produce a disaster, its volcanic activity is monitored at present by video cameras, seismometers, gravimeters, tiltmeters and a GNSS (Global Navigation Satellite System) survey (US Geological Survey; URL https://www.usgs.gov/media/images/volcanic-monitoring-types-and-methods-employed-usgs-vhp (accessed on 4 December 2021)). However, the prediction of volcanic eruptions is sometimes very difficult and, consequently, has produced many serious disasters, even in the 21st century (e.g., the 2010 eruption of Merapi Volcano, Indonesia [1,2], and the 2014 eruption of Mt. Ontake, Japan [3]). In order to increase the prediction accuracy and improve understanding, thermal and chemical studies are effective additions to the above geophysical approaches, although continuous thermal and chemical monitoring is rarely used among volcanoes because the sensors suffer heavy damage due to the high temperatures and/or corrosive environments.

If a lake or pond exists in an active volcano, the volcano carries a high risk of producing phreatic/phreatomagmatic explosions and lahars at the time of eruptive activity (e.g., Taal Volcano, the Philippines, with Taal Lake [4], Kusatsu-Shirane Volcano, Japan, with Yugama...
Crater Lake [5] and Poas Volcano, Costa Rica, with an acid crater lake [6,7]). Observation of volcanic earthquakes and ground deformation can catch mechanical variation inside a volcano, but what causes (or triggers) such variations is unclear in many cases. Questions remain whether magma is rising or the hydrothermal system is activated. Volcanic lakes provide a place for a research that can directly address the states of the heat source beneath volcanoes. Hence, it is important to establish methods for directly exploring the interaction between the heat source and the lake.

Lake Kuussharo, Hokkaido, Japan, exists in the Kuusahro Caldera (43°37′35″N, 144°20′08″E) which includes Atosanupuri Volcano and is thermally affected by hot spring water from a geothermal heat source. Chikita et al. [8] evaluated the groundwater inflow and outflow in Lake Kuussharo by estimating the hydrological and chemical budgets of the lake and specified the location of the lake bottom area to outflow the lake water as groundwater. Kuttara Volcano, Hokkaido, Japan (42°30′19″N, 141°08′40″E), has an active volcanic area with hydrothermal ponds and geothermal fields in the westernmost region and Lake Kuttara in the eastern region. Chikita et al. [9] found that, during the development of thermocline in the lake, the lake water temperature consistently increases at 0–45 m above the bottom of the deepest point. Consequently, the geothermal heat flux at the lake bottom was calculated at 4.1–10.9 W/m².

Meanwhile, the groundwater flow system in volcanos has been explored in volcanos of low activity. From the geochemical viewpoint, Ono et al. [10] inferred the groundwater flow system of the stratovolcano Mt. Fuji, Japan. Aizawa et al. [11] proposed a resistivity model that reflects the groundwater flow system in each of five stratovolcanoes by electric self-potential and magnetotellurics, and, thereby, two-dimensionally simulated the groundwater flow accompanied by hydrothermal zone and groundwater zone.

In Zao Volcano, Miyagi Prefecture, Japan, lahars occurred from Okama Crater Lake many times in the 19th century [12]. At present, the lake keeps relatively calm but has the potential for an eruption. The lake has kept high acidity (pH = 2.6–3.3) for ca. 80 years since its last activity, suggesting a volcanic fluid supply from the underground. In this study, the lake water temperature is monitored throughout the year, and the hydrological and chemical regimes in and around Okama Crater Lake were explored, in order to clarify the geothermal effect on the lake and the groundwater flow system in Zao Volcano.

2. Study Area

Zao Volcano, which crosses Miyagi and Yamagata Prefectures, Japan, has contained high volcanic activity for about one million years. Following the Great East Japan Earthquake of 2011, an increase in volcanic activities has been recognized since 2013, such as volcanic earthquakes, volcanic tremors and tilt motion. However, in 2020, the activity level of Zao returned to normal (URL http://www.data.jma.go.jp/svd/vois/data/tokyo/STOCK/monthly_v-act_doc/monthly_vact_vol.php?id=212 (accessed on 14 December 2021). Okama Crater Lake was probably formed in 1625–1694 through a series of volcanic eruptions from Zao that began in 1230, in which the crater was moved from the south of Mt. Goshiki to its present position (Figure 1) [12]. As one of the eruptions centered on Okama, a phreatomagmatic eruption occurred in 1894–1897, which produced pyroclastic surges and lahars containing lake water and the lower hydrothermal water [12,13]. In 1918, Okama’s water boiled, and, in August 1923, its water became milky with floating amorphous sulfur due to H₂S or SO₂ gas spewing from the lake surface [13], although no eruptive activity occurred at or around Okama. Okama remained calm after a weak steam eruption in 1940 at the Maruyamasawa geothermal field, 1.5 km northeast of Okama.
Okama is located at an altitude of ca. 1550 m above sea level (asl) with a depth of ca. 25 m at maximum and water surface area of \(8.208 \times 10^4\) m\(^2\) (Figure 2). The drainage area, including the lake area, is \(6.312 \times 10^5\) m\(^2\) (Figure 1). The drainage basin is adjacent to the headwater region of the Nigori River, where part of Okama’s water is likely to leak [14]. However, the location and altitude of the springing points are not specified in Konno [14].

Okama has three inflowing streams: the Goshiki Stream, the Western Stream and the Southern Stream. Of all the streams, only the Goshiki Stream is perennial with streaming water at any time; the other streams are ephemeral with streaming water only during a rainy period or in the snowmelt season. Referring to Konno [14] and Yoshimura [15], the lake was iron-rich and acidic at pH of 2.6 and 37 m deep at maximum in 1936. In 2020, the lake was still iron-rich, and the pH and maximum depth were 3.1–3.3 and ca. 25 m, respectively. Thus, the lake type of Okama belongs to a siderotrophic lake within inorganic acidotrophic lakes [15]. On the topographic map with a 1/25,000 scale, in 2014, the highest lake level is at an altitude of ca. 1550 m asl, which is the same as the level in 1936 [14]. Hence, the decrease in maximum depth from 37 m to 25 m indicates that sedimentation is in progress, probably due to surrounding slope collapse or sediment runoff from the inflowing streams.
Figure 2. Bathymetric map of Okama Crater Lake with inflowing streams. The thick solid lines, thin grey lines and dotted lines in the map show isopleths of water depth at 5 m, 1 m and 0.2 m intervals, respectively (modified after Yamasaki and Yagi [16] as well as Tsuchiya and Hirano [17]).

3. Methods

3.1. Estimates of Hydrological and Chemical Budgets

When a crater lake is located in an active volcano, it is important to know how the groundwater in or around the lake is connected to the underground hydrothermal flow system and magma. Thus, in order to separately evaluate the groundwater inflow and outflow in Okama, the hydrological and chemical budgets of the lake were estimated. The hydrological budget equation for a closed lake with no outflowing river, such as Okama, is given as follow:

\[
\frac{\Delta V}{\Delta t} = (P - E)A_0 + R_{in} + G_{in} - G_{out} \tag{1}
\]

where \( V \) is the water volume (m\(^3\)), \( A_0 \) is the water surface area (m\(^2\)), \( P \) is the precipitation (m/s) onto the lake surface, \( E \) is the evaporation (m/s) at the lake surface, \( R_{in} \) is the river inflow (m\(^3\)/s), \( G_{in} \) and \( G_{out} \) are the groundwater inflow and outflow (m\(^3\)/s), respectively, and \( t \) is the time. The left side of Equation (1) indicates the water volume change per budget period \( \Delta t \). Evaporation \( E \) was calculated by using the following bulk transfer method:

\[
Q_E = -\lambda \left( \frac{\rho_w \beta}{\rho_a} \right) \cdot (a_E u_z) \cdot (e_z - e_0) \tag{2}
\]

\[
E = \frac{Q_E}{(\lambda \rho_w)} \tag{3}
\]

where \( Q_E \) is the heat flux (W/m\(^2\)) by evaporation, \( \lambda \) is the latent heat (J/kg) for evaporation, \( \beta \) is the ratio of water vapor density to dry air density (=0.622), \( a_E \) is the dimensionless bulk transfer coefficient for latent heat, \( u_z \) is the wind speed (m/s) at \( z \) (m) above the lake surface, \( p \) is the air pressure (Pa) at \( z \), \( e_z \) is the vapor pressure (Pa) at \( z \), \( e_0 \) is the saturated vapor pressure (Pa) at lake surface temperature \( T_s \) (K), and \( \rho_a \) and \( \rho_w \) are the air and water densities (kg/m\(^3\)), respectively. Here, \( z = 4.0 \) m was adopted in Equation (2) as the height of the wind speed sensor as well as of the air temperature and relative humidity logger above the lake water surface. The dimensionless bulk transfer coefficient, \( a_E \), for latent heat
was given at a constant of 0.0013 for z = 4.0 m [18]. Under the condition of non-rainfall, Equation (1) is as follows:

\[ G = G_{\text{in}} - G_{\text{out}} = \Delta V/\Delta t + E - P_{\text{in}} \]  

(4)

Here, the G value, or the net groundwater inflow, is provided via calculation of the water volume change and surface water area change from the measurement of lake level (Figure 2), the measurement of river inflow and the calculation of evaporation E by applying the hydrometeorological data at site M to Equations (2) and (3).

The chemical budget equation for the closed lake is as follows:

\[ \Delta (C_L V)/\Delta t = C_{\text{Rin}} R_{\text{in}} + C_F P A_0 + C_{\text{Gin}} G_{\text{in}} - C_L G_{\text{out}} - S \]

(5)

where \( C_L \) is the mean ionic concentration (g/L) of the lake; \( C_{\text{Rin}} \), \( C_F \) and \( C_{\text{Gin}} \) are the ionic concentrations (g/L) of inflowing rivers, precipitation and inflowing groundwater, respectively; and S is the net depositional flux (kg/s) related to a chemical reaction of the ion. The mean ionic concentration, \( C_L \), of the lake is given as that of groundwater outflow, since the depth at which the lake water leaks out as groundwater is unknown. The magnitude of S is inferred by considering the lake water chemistry based on the ionic analysis.

If a non-rainfall period is adopted as the budget period, the second term on the right side of Equation (5) is zero. Then, the simultaneous equations from Equations (4) and (5) lead to:

\[ G_{\text{out}} = (C_{\text{Gin}} G - B)/(C_L - C_{\text{Gin}}) \]  

(6)

\[ G_{\text{in}} = G_{\text{out}} + G \]  

(7)

where \( B = \Delta (C_L V)/\Delta t - C_{\text{Rin}} R_{\text{in}} + S \). Here, the B values are given by a coupling with the analyzed chemistry of the stream and lake waters. Then, an ion that is not influenced by the chemical reaction should be selected as the ion specified in the ionic concentration in Equation (5), since quantifying the flux S is sometimes difficult. However, the flux S, if any, could be estimated by setting sediment traps on the lake bottom and then by identifying the mineralogy or chemical compounds of the trapped deposit with the X-ray diffraction (XRD) method.

3.2. Field Observations

In order to evaluate the temporal change in the water volume of Okama on the left side of Equation (1), lake level was measured at 1 h intervals by fixing two pressure loggers (type HOBO U20, Onset Computer Inc., Bourne, MA, USA) at site L (for water pressure) and site M (for air pressure) (Figure 1). The difference between the water pressure and the air pressure provided the water depth at site L as the lake level. The pressure logger at site L was fixed throughout the year.

In order to calculate the first term on the right side of Equation (1) and evaporation E in Equation (3), a weather station (URL: https://www.onsetcomp.com/products/kits/u30-nrc-sys-c/ (accessed on 17 November 2021)) was set at site M to record air pressure, air temperature, relative humidity, wind velocity and rainfall at 1 h intervals. The weather station was removed from site M before snowfall began in October, because the snow accumulation becomes more than 1 m deep in winter. Meanwhile, in order to measure the hourly air temperature, relative humidity and air pressure throughout the year, two loggers were fixed at the Daikokuten observation hut (altitude, 1446 m asl), belonging to the Research Center for Prediction of Earthquakes and Volcanic Eruptions, Graduate School of Science, Tohoku University, which is 1.55 km southeast of site MD. Therefore, the daily mean air temperature and air pressure at Okama after the removal of the weather station at site M were acquired using the high correlations (\( R^2 = 0.991, p < 0.01 \)) for the daily mean air temperature and (\( R^2 = 0.982, p < 0.01 \)) for the daily mean air pressure between site MD and
the observation hut. The annual mean air temperature at site MD was also evaluated using annual data from the observation hut.

On the lake shore near site M, the discharge of the three inflowing streams (Figure 2) was measured on non-rainfall days to evaluate the river inflow $R_{in}$ in Equation (1). The stream water was also manually sampled to analyze the inorganic chemistry, related to $C_{in}$ in Equation (5). The EC25 (electric conductivity at 25 °C) and pH of the stream water were then measured by a thermocouple thermometer (Type TX1001, YOKOGAWA, Co., Ltd., Tokyo, Japan) and a portable pH and EC meter (Type D-74, HORIBA, Co., Ltd., Kyoto, Japan). The Nigori River water was sampled at site N (Figure 1), its pH and EC25 were similarly measured, and the chemistry was analyzed. The river water at site N is unlikely to contain water leaked from Okama because its altitude (1557 m asl) is higher than that of Okama’s water surface (1550 m asl).

In order to explore the effect of the geothermal heat flux on the thermal condition of Okama, the water temperature was measured every hour at nine depths of the deepest point (site MD) using a mooring system of temperature loggers (type TidBit v2 with accuracy, ±0.2 °C, Onset Computer Inc., Bourne, MA, USA; URL https://www.onsetcomp.com/products/data-loggers/utbi-001/ (accessed on 17 November 2021)) (Figures 1 and 3). In order to ascertain the temporal change in lake water chemistry related to the left side of Equation (5), an electric conductivity (EC) logger for fresh water (type HOBO U24-001 with accuracies of ±0.5 mS/m and ±0.1 °C for electric conductivity and temperature, respectively, Onset Computer Inc., Bourne, MA, USA) was added in June 2020 to a position of 2 m above the lake bottom. The mooring system was set in the lake throughout the year.

![Figure 3. Mooring system for fixing nine loggers at site MD.](image)

Meanwhile, by lowering a profiler on boat (type ASTD102 with accuracies of ±0.1 mS/m and ±0.01 °C for EC and water temperature, respectively, JFE-Advantech, Inc., Japan; URL https://www.jfe-advantech.co.jp/products/ocean-rinko.html (accessed on 16 November 2021)), the water temperature and EC25 were vertically measured at 0.1 m depth intervals in June–October of 2019 and 2020. The EC and water temperature were recorded only during the downward movement of the profiler. Thus, the profiler’s record is not influenced by the disturbance of suspended matters from its bottom touch. The water temperature and EC25 from the loggers were calibrated by those from the profiler with relatively high accuracies.

Lake water and bottom sediment at site MD were sampled on boat by a Van Dorn sampler (URL https://www.kc-denmark.dk/products/water-sampler/van-dorn-water-sampler.aspx (accessed on 6 September 2021)) and an Ekman-Burge grab sampler [19], respectively. The water temperature, EC25 and pH of the lake water were then measured by a thermocouple thermometer (Type TX1001, YOKOGAWA, Co., Ltd., Tokyo, Japan), a compact EC meter (Type LAQUAtwin EC-11, HORIBA, Co., Ltd., Kyoto, Japan) and a compact pH meter (Type...
In order to obtain the concentration $C_{Gin}$ of inflowing groundwater in Equation (5), pore water in the bottom sediment of site MD was sampled by an extractor kit with 60 mL syringe and a porous cup (URL https://www.soilmoisture.com/1900K2/ (accessed on 24 November 2021)), and its EC25 and pH were measured by a compact EC meter (Type LAQUAtwin EC-11, HORIBA, Co., Ltd., Kyoto, Japan) and a compact pH meter (Type LAQUAtwin pH-11B, HORIBA, Co., Ltd., Kyoto, Japan), respectively.

Concentrations of major ions ($K^+$, $Na^+$, $Mg^{2+}$, $Ca^{2+}$, $Cl^-$, $SO_4^{2-}$) for the sampled river water, lake water and pore water were measured by ion chromatography (type Dionex ICS-1600 (cation) and ICS-2100 (anion), Thermo Fisher Scientific Inc., Tokyo, Japan; URL https://www.thermofisher.com/ (accessed on 24 November 2021)). Only the bicarbonate ion ($HCO_3^-$) concentration was measured using sulfuric acid titration. However, the $HCO_3^-$ concentration was almost zero for a few samples with pH of 2.6–4.3. Thus, for the samples at pH of 4.3 or less, the titration was omitted.

4. Results

4.1. Water Chemistry

Figure 4 shows stiff diagrams of $Na^+$, $K^+$, $Ca^{2+}$, $Mg^{2+}$, $Cl^-$, $HCO_3^-$ and $SO_4^{2-}$ concentrations (mEq/L), and the pH and EC25 (mS/m) values for the lake and stream waters. These water samples were taken on non-rainfall days of (a)–(c) 21 August 2019, (d) 18 October 2019 and (e) 4 September 2020. The waters commonly have very high $Ca^{2+}$ and $SO_4^{2-}$ concentrations with a high acidity of pH = 3.2–4.3, though the two stream waters (Figure 4c,d) into Okama exhibit relatively low acidity (pH = 4.3 and 4.0 for pH = 3.4 of lake water) with an ionic balance rich in $K^+$ and $Na^+$. The EC25 values appear to consistently increase as the $Ca^{2+}$ and $SO_4^{2-}$ concentrations increase.

![Stiff diagrams](image-url)

**Figure 4.** Stiff diagrams of (a) lake surface water, (b) lake water at 20 m depth, (c) Goshiki Stream water, (d) Western Stream water and (e) Nigori River water, and their pH and EC25 values. (a–c) were sampled on 21 August 2019, and (d,e) were sampled on 18 October 2019 and 4 September 2020, respectively.

The Nigori River water (Figure 4e) was sampled in the catchment neighboring the Okama catchment (Figure 1) and, thus, appears to be unaffected by the surface water and
groundwater in the Okama catchment. However, its high acidity (pH = 3.2) suggests that the Nigori River catchment is also affected by sulfate substances as part of Zao Volcano.

Figure 5 shows relations between the EC25 and (a) SO$_4^{2-}$ concentration, (b) Ca$^{2+}$ concentration, (c) Ca$^{2+}$ + SO$_4^{2-}$ concentration or (d) the pH for sampled lake and stream waters. There are clear linear relationships with high correlations of $R^2=0.746–0.914$. Especially, the relation between the EC25 and SO$_4^{2-}$ concentration (Figure 5a) is highly correlated at $R^2=0.914$. These relations indicate that, if the EC25 is measured by the EC logger or the profiler, then the Ca$^{2+}$ and SO$_4^{2-}$ concentrations and pH are also numerically obtained with high confidence levels. Here, the pH appears to be controlled by the SO$_4^{2-}$ concentration (Figure 4).

![Figure 5. Relations between EC25 (mS/m) and (a) SO$_4^{2-}$ concentration (mg/L), (b) Ca$^{2+}$ concentration (mg/L), (c) Ca$^{2+}$ + SO$_4^{2-}$ concentration (mg/L) or (d) pH for lake and stream waters.](image)

Figures 6 and 7 show the vertical profiles of water temperature, EC25, SO$_4^{2-}$ concentration and pH at site MD, obtained three times each in June–October 2019 and July–October 2020. The pH and SO$_4^{2-}$ concentration profiles were acquired using the regression lines in Figure 5a,d. The vertical distributions of the water temperatures exhibit the development of thermocline in the heating season (from June to mid-August) and the attenuation and descent of thermocline in the cooling season from late August to October. The developed thermocline on 21 August 2019 and the vertically uniform water temperature and EC25 on 18 October 2019 (Figure 6a) suggest that when the descending thermocline arrives at the bottom, vertical whole circulation is established.
Figure 6. Vertical distributions of (a) water temperature, (b) electric conductivity at 25 °C (EC25), (c) SO$_4^{2-}$ concentration and (d) pH at site MD in 2019.

Figure 7. Vertical distributions of (a) water temperature, (b) EC25, (c) SO$_4^{2-}$ concentration and (d) pH at site MD in 2020.

When the thermocline reached the bottom, the bottom temperature should have increased abruptly from 6.8 to 9.5 °C in 2019 or from 4.9 to 9.2 °C in 2020. This is because, during the cooling season, the thermocline descends to the bottom, while the upper layer keeps thermally uniform.
The EC25 values (and therefore the SO$_4^{2-}$ concentration) greatly increased downward at or below the thermocline. These great spatio-temporal increases in the EC25 appear to have been produced by the density underflow from the sediment runoffs of inflowing streams or the surface flow from rainfalls. In fact, such a sediment runoff was observed on the lake’s shore after heavy rainfall (169 mm/day) on 28 July 2020 (Figure 8). The water temperature, EC25 and pH of the surface water flowing over the exposed silt layers were then measured at 10.5–12.7 °C, 388–495 mS/m and 2.3–2.5, respectively, thus showing very high EC25 and low pH. The bulk density of the surface water was then calculated at 1001.6–1001.9 kg/m$^3$, since the EC25 at 388–495 mS/m corresponds to total dissolved solid (TDS) at 1.9–2.5 g/L, using the conversion coefficient 5.00 (mg/L)/(mS/m) of the EC25 to TDS for the portable EC meter of HORIBA, Co., Ltd. Meanwhile, the Okama bottom water on 31 July 2020 (Figure 7) had the bulk density of 1000.4 kg/m$^3$ for water temperature at 5 °C, and the EC25 was 87 mS/m (equivalent to TDS at 435 mg/L). Thus, the surface water is consistently heavier than the whole lake water. This means that, after flowing into Okama, the surface water flows down the bottom slope as a density underflow [20].

Figure 8. Inflow of milky turbid water observed after heavy rainfall (169 mm/day) on 28 July 2020 (photo on 31 July 2020); this is due to sediment erosion from the surface water flowing over the gray silt layers distributed along the lake shore on the left side of the Goshiki Stream. The marked inflowing points of milky water are represented by blue arrows.

Then, the suspended sediment concentration (SSC) at 50–100 mg/L in the surface water could contribute slightly to the magnitude of the bulk density. However, during and just after a heavy rainfall of more than 100 mm/day, the SSC could increase greatly and produce a turbid layer at the bottom of the lake, contributing to the sedimentation. During development of the thermocline, the EC25 increased even more greatly at 0.2 m or less above the lake bottom. This suggests that the deposition of suspended sediment is slow enough to produce a turbid layer, such as the nepheloid layer often observed near the ocean bottom [21].

Figure 9 shows a time series of the hourly water temperature, EC25 and pH from the EC logger at 2 m above the bottom of site MD (Figure 3) and hourly rainfall at site M for the time period from 15:00, 31 July to 12:00, 19 October 2020. The continual rainfall gradually increased the EC25 and decreased the pH. This appears to reflect the integrated stagnation of the high EC25 and low pH water from the intrusion of density underflow into the lower layer (Figure 7b,d). During this intrusion of the underflow, mixing with lake water could occur. Thus, the EC25 at the logger may be lower than that of the surface water. Meanwhile, the water temperature gradually increased with slight decreases of 0.03–0.05 °C, being highly responsive to each rainfall.
Figure 9. Time series of the hourly EC 25, pH and water temperature at 2 m above the bottom of site MD and the hourly rainfall at site M (Figure 1).

4.2. Thermal Variations in Okama

Figure 10 shows temporal variations in the hourly water temperature (a) at lake bottom, 2 m (b-2 m) and 10 m (b-10 m) above the bottom, (b) at 15 m above the bottom (b-15 m) and 5 m depth, and (c) at surface, 1 m depth and 5 m depth at site MD from 21 August 2019 to 18 May 2021. Figure 10d is an enlargement of the red arrow period in Figure 10c. The five black arrows in Figure 10a–c present the measurement times of the vertical profiles in Figures 6 and 7. In mid-October 2019 and late October 2020, the water temperature increased abruptly at the bottom and 2 m above the bottom, which was preceded by an increase at 10 m above the bottom at or around the end of September (Figure 10a). This indicates the descent and bottom arrival of the thermocline during the cooling season (Figures 6 and 7). At the bottom and 2 m above the bottom, the temperature appears to slightly increase throughout the heating and cooling seasons before this abrupt temperature increase (Figure 9).

After the abrupt temperature increase, Okama was thermally uniform, as shown by the 18 October profile in Figure 6a. Thereafter, the lake surface cooled to 0 °C and remained completely ice-covered until early May 2020 or early April 2021, as shown by the almost 0 °C temperature at the lake surface (Figure 10c,d). In early January of 2020, the surface temperature became less than 0 °C. This indicates that, following the downward ice growth, the temperature logger just below the surface buoy was enclosed by lake ice. For the completely ice-covered period of 20 December 2019–1 May 2020, a small oscillation of the water temperature was observed at the bottom and 2 m above the bottom. This is probably due to the small vertical movement of the covered ice responding to air pressure variation, since both the surface buoy and the lower buoy in the mooring system (Figure 3) were captured, possibly by the covered ice grown.

Such an ice capture of the two buoys could occur via the consistent decrease in lake water level; in early November, the lake level may have started to decrease due to a lack of rainwater supply (Figure 11). In the completely ice-covered period of December 2020–March 2021, the water temperature at 5 m depth was 0.03–0.09 °C larger than that at 15 m above the bottom (Figure 10b). This means that due to the consistent decrease in the lake level, the logger at 5 m depth had come down to the level as low as the logger at 15 m above the lake bottom (Figure 3). In this situation, the lower buoy could float at the same level as the surface buoy and be captured by the covered ice. On 26 June 2020 and 18 May 2021, when the first survey in the open water season of Okama was performed, the
mooring system was found to have moved to a shallower point 66.5 m northwest and 81.7 m southeast of site MD, respectively. This suggests that the mooring system moved due to a wind-driven drift of the covered ice during May (Figure 10c). The covered-ice drift could have started with the partial ice loss at the lake surface and continued until the abrupt increase in surface water temperature (Figure 10c,d). From 29 March to 26 April 2021, the water temperature at the surface and 1 m depth increased from 0.12 to 0.66 °C and from 0.19 to 1.70 °C at 1 m depth, respectively (Figure 10d). This probably reflects the passage of solar radiation onto the lake water through ice cracks during the season of snowmelt and ice melt.

![Figure 10](image1.png)

**Figure 10.** Temporal variations of the hourly water temperature at (a) the lake bottom and 2 m (b-2 m) as well as 10 m (b-10 m) above the bottom; (b) 15 m above the bottom (b-15 m) and 5 m depth; and (c) surface, 1 m depth and 5 m depth at site MD from 21 August 2019 to 18 May 2021. The red arrow period in (c) is enlarged in (d). The five black arrows in (a–c) show measurement times of the vertical profiles in Figures 6 and 7.

![Figure 11](image2.png)

**Figure 11.** Temporal variations of the daily mean air temperature, lake level and lake surface water temperature for 1 August–30 November 2020; daily mean relative humidity and wind speed for 1 August–18 October 2020; and the diurnal rainfall for 1 July–18 October 2020. The daily mean air temperature and air pressure (at lake level) for 19 October–30 November were evaluated using the high correlations for the data between site M and the Daikokuten observation hut (Figure 12).
During the complete ice cover, the temperature gradually increased between the lake bottom and 15 m above the bottom (Figure 10a,b). At the surface and 1 m depth, such a gradual increase was not observed (Figure 10d), and the temperature at 15 m above the bottom was almost equal to that at 5 m depth (Figure 10b). Hence, the upper limit of the gradual increase could be located at ca. 17 m above the lake bottom, equivalent to ca. 3 m depth. The lake water is at almost 0 °C at the interface of water and ice, and the downward ice growth tends to cool the lower water [22]. Hence, the temperature increase at 0–15 m above the bottom is probably due to the geothermal heat supplied at the lake bottom, which could originate from the hydrothermal reservoir below the Okama bottom basin [23].

4.3. Meteorology at Okama

Figure 11 shows temporal variations of the daily mean air temperature, lake surface water temperature and lake level for 1 August–30 November 2020; daily mean relative humidity and wind speed for 1 August–18 October 2020; and the diurnal rainfall for 1 July–18 October 2020. The air pressure and air temperature during 19 October–30 November 2020 at site M, after the removal of the weather station, were evaluated by the regression lines with high correlations (R² = 0.982 and 0.991, respectively) between site M and the Daikokuten observation hut (Figure 12).

The lake level declined during a non-rainfall period from 18 to 28 August 2020, as well as after the removal of the weather station at site M. Referring to Equation (1), this suggests that during a relatively lengthy non-rainfall period, the groundwater outflow from the lake is larger than the groundwater inflow plus the river inflow minus the evaporation over the lake surface, i.e., \( G_{\text{out}} > G_{\text{in}} + R_{\text{in}} - E_{\text{A0}} \).

In the period from 18 to 28 August 2020, the lake level decreased at an average of 33 mm/day. Meanwhile, using Equations (2) and (3), the mean evaporation at the lake water surface was calculated to be 0.51 mm/day, while the river inflow was 0.0033 m³/s, equivalent to 3.5 mm/day per lake surface area. Hence, the net groundwater inflow decreases the lake level at −36 mm/day. On 24–25 September 2020, a large rainfall occurred totaling 231.6 mm. Thereby, the lake level rose 0.47 m. A decline of the lake level was not observed during rainfalls of more than 30 mm/day. This means that, due to negligibly small evaporation during rainfalls, the absolute value of the net groundwater inflow is equivalent to the 30 mm/day rainfall over lake surface plus the correspondent river inflow. After the removal of the weather station at site M, the lake level consistently decreased at 27 mm/day, reflecting, small rainfalls, if any. As a side note, the monthly rainfall at a weather station 20.3 km east of Okama was only 2.0 mm in November 2020.

The data of meteorology and lake level in Figure 11 can be applied to the hydrological budget estimate for the lake, shown by Equation (1).
The surface water inflow with high EC25, low pH and milky color in Figure 8 appears to have been produced continuously by continual rainfalls of 14–48 mm/day during 1–27 July 2020, in addition to the large rainfall of 169 mm/day on 28 July 2020.

5. Discussion

5.1. Geothermal Heat Flux in Okama

In the completely ice-covered periods of December–April, an increase in the lake water temperature was observed between the lake bottom and 15 m above the bottom at the deepest point (Figure 10). This means that some geothermal heat is supplied at the lake bottom. Hence, using the bathymetric map in Figure 2 and a time series of the daily mean water temperature from Figure 10, a time series of the daily mean heat flux and its 10-day moving average were acquired (Figure 13). Then, the periods of 17 December 2019–27 April 2020 and 15 December 2020–29 March 2021 in Figure 10 were selected for the calculation, since the water temperature at the surface increased after 27 April 2020 and 29 March 2021, indicating the incidence of solar radiation via a partial loss of covered ice from spring ice melt.

![Figure 13. Temporal variation of daily mean or 10-day moving average (red line) heat flux (W/m²) from the bottom for the completely ice-covered periods of (a) 17 December 2019–27 April 2020 and (b) 15 December 2010–29 March 2021.](image)

First, the area, \( A(z) \), at each logger’s height above the lake bottom, acquired by the bathymetric map, was multiplied by the following factor:

\[
ρ_w \cdot c_p \cdot T(z)
\]  

where \( ρ_w \) is the water density (kg/m³), \( c_p \) is the specific heat (J/Kg/K) of the water, and \( T(z) \) is the water temperature (K) at the logger’s height \( z(\text{m}) \) above the bottom. The total heat storage (J) of the lake on a certain day was obtained by summing the following:

\[
∆h[ρ_w \cdot c_p \cdot T(z) \cdot A(z)]
\]

at 0–17 m above the lake bottom, where \( ∆h \) is each height difference between the loggers. Then, a horizontal multi-layer thermal structure in the lake was assumed, i.e., horizontally taking the same temperature at the same water depth. Finally, the heat storage change per day per unit area (W/m²) was calculated by dividing a daily change (J/day) in the total heat storage of the lake by the area \( 6.627 \times 10^4 \text{ m}^2 \) at 17 m above the bottom (Figure 2).
Focusing on the temporal variations of the 10-day moving average, the heat flux varied at a range of $-0.5$–$5.6$ W/m$^2$ in December 2019–April 2020 and $1.5$–$4.7$ W/m$^2$ in December 2020–March 2021. The heat flux averages over these periods were $2.5$ W/m$^2$ in December 2019–April 2020 and $2.9$ W/m$^2$ in December 2020–March 2021. These values are similar to the heat flux (2.8–10.7 W/m$^2$) at Lake Kuttara belonging to the active Kuttara Volcano [9] but larger than that (0.29 W/m$^2$) at Lake Shikotsu, which is surrounded by three active volcanos [24]. By using the audio-frequency magnetotelluric (AMT) method, Ichiki et al. [23] revealed that a low resistivity zone, such as a hydrothermal reservoir, exists with the center at ca. 250 m below the lake bottom. The calculated geothermal heat flux in Figure 13 appears to reflect thermal leakage from the hydrothermal reservoir.

### 5.2. Estimating Groundwater Inflow and Outflow in Okama

Using Equations (1)–(7), the groundwater inflow and outflow in Okama were calculated in the daily database, where one day was adopted as the fundamental budget time period $\Delta t$ in Equations (1) and (5). Here, the $\text{SO}_4^{2-}$ concentration was taken as the representative ionic concentration for the chemical budget of Okama because the $\text{SO}_4^{2-}$ concentration highly correlated to the EC25 and pH (Figure 5). Then, the net depositional flux S in Equation (5) was assumed to be neglected for $\text{SO}_4^{2-}$. The EC25 record of the EC logger at 2 m above the bottom of site MD was converted to a daily series of the $\text{SO}_4^{2-}$ concentration, where the daily series of the volume-averaged $\text{SO}_4^{2-}$ concentration as $C_L$ in Equation (5) was obtained using a relation between the EC25 values in the logger and the three vertical profiles in Figure 7b (Figure 14). Then, assuming a horizontal multi-layered structure of the chemistry in the lake, the volume-averaged EC25 values from the three profiles (vertical axis in Figure 14) were calculated using the bathymetric map in Figure 2. Additionally, the water surface area $A_0$ was temporally changed following the lake level change in Figure 11 and the bathymetry in Figure 2. The total stream inflow measured on the non-rainfall days was given as the river inflow $R_{in}$ in Equations (1) and (5). The mean EC25 and $\text{SO}_4^{2-}$ concentration values (177 mS/m and 823 mg/L, respectively) of pore water in the bottom sediment at site MD were adopted as those of the groundwater inflow.

![Figure 14](image)

**Figure 14.** Relation between the volume-averaged EC25 from the three profiles in Figure 7b and the EC25 from the EC logger at 2 m above the bottom of site MD in 2020.

Table 1 shows the net groundwater inflow ($G_{in} - G_{out}$), as well as the separated groundwater inflow $G_{in}$ and groundwater outflow $G_{out}$, calculated for the six non-rainfall periods in Figure 11. The correspondent river flow observed is also shown. For the non-rainfall periods, the first day after a rainfall was omitted because the water supplied by the rainfall appears to raise the lake level one day later (Figure 11). All the calculated results of the net groundwater inflow are negative. Thus, the groundwater outflow is larger than the groundwater inflow. Of all the calculated results, Periods (1), (2) and (6) include more
than three non-rainfall days. The calculated results for these three periods could thus be more reliable. $G_{\text{in}} - G_{\text{out}}$, $G_{\text{in}}$ and $G_{\text{out}}$ averaged for the three periods were $-0.028$ m$^3$/s, $0.012$ m$^3$/s and $0.039$ m$^3$/s, respectively. This negative net groundwater inflow decreases the lake level at $-0.029$ m/day, and thus is comparable in magnitude to the observed declining rates of $-0.033$ m/day and $-0.027$ m/day (Figure 11).

Table 1. Measured river inflow $R_{\text{in}}$; calculated net groundwater inflow ($G_{\text{in}} - G_{\text{out}}$); as well as separated groundwater inflow $G_{\text{in}}$ and outflow $G_{\text{out}}$ for the six non-rainfall periods in Figure 11.

| No. | Period              | $R_{\text{in}}$ (m$^3$/s) | $G_{\text{in}} - G_{\text{out}}$ (m$^3$/s) | $G_{\text{in}}$ (m$^3$/s) | $G_{\text{out}}$ (m$^3$/s) |
|-----|---------------------|-----------------------------|---------------------------------------------|---------------------------|-----------------------------|
| (1) | 3–6 August 2020     | 0.00332                     | $-0.020$                                    | 0.016                     | 0.036                       |
| (2) | 19–28 August 2020   | 0.00332                     | $-0.035$                                    | 0.012                     | 0.047                       |
| (3) | 7–8 September 2020  | 0.00271                     | $-0.043$                                    | 0.018                     | 0.061                       |
| (4) | 16–17 September 2020| 0.00271                     | $-0.072$                                    | 0.019                     | 0.091                       |
| (5) | 21–22 September 2020| 0.00271                     | $-0.031$                                    | 0.011                     | 0.042                       |
| (6) | 30 September–4 October 2020 | 0.00212               | $-0.028$                                    | 0.007                     | 0.035                       |

If the heat flux $H_{\text{Gin}}$ (W/m$^2$) by the groundwater inflow $G_{\text{in}}$ at $0.012$ m$^3$/s is equal to the geothermal heat flux, 2.5 W/m$^2$ in 2019–2020 and 2.9 W/m$^2$ in 2020–2021 at the lake bottom, the inflowing groundwater temperature $T_{\text{Gin}}$ is estimated using the following equation:

$$H_{\text{Gin}} = \rho_w c_p G_{\text{in}} (T_{\text{Gin}} - T_L)$$

(10)

where $\rho_w$ is the groundwater density, $c_p$ is the specific heat (J/kg/K) of water and $T_L$ is the representative lake water temperature (K) (273.15 K = 0 °C + 273.15). Here, if $T_L$ is supposed to be the bottom water temperature 275.45 K (2.3 °C) and 275.25 K (2.1 °C) averaged over the periods of 2019–2020 and 2020–2021, respectively, $T_{\text{Gin}}$ is given at 278.75 K (5.6 °C) and 279.05 K (5.9 °C), respectively. These values are larger than the annual mean air temperature 277.95 K (4.8 °C) at Okama inferred from those at Daikokuten. In general, the shallow groundwater in the catchment without any geothermal heat source tends to take the water temperature near to the local, annual mean air temperature [25,26]. The geothermal heat flux in Okama is thus significantly larger than that in the non-geothermal catchment. This suggests that, though Okama stays relatively calm, thermal leakage from the hydrothermal reservoir below the lake basin exists. In the completely ice-covered periods, the infiltration of snowmelt water at the snow-ground interface should be much smaller than that of rainwater in the rainfall season [27,28]. This means that the groundwater inflow to Okama in the ice-covered period is smaller than that in the rainfall season. For example, if $G_{\text{in}} = 0.010$ m$^3$/s at the calculated $H_{\text{Gin}} = 2.5$ and 2.9 W/m$^2$, Equation (10) gives $T_{\text{Gin}} = 279.45$ K (6.3 °C) and 279.85 K (6.7 °C), respectively. Hence, the calculated temperatures 5.6 °C and 5.9 °C appear to correspond to the lowest values in the ice-covered period.

6. Conclusions

In order to explore the geothermal effect on Okama Crater Lake in the active Zao Volcano, as well as the hydrological and chemical cycles in or around the lake related to the underground geothermal heat source, the lake water temperature was monitored at eight depths throughout the year and the hydrological and chemical budgets of the lake were estimated. The temperature monitoring revealed that, during the completely ice-covered period, the water temperature consistently increases between the lake bottom and 15 m above the bottom. This temperature increase is probably due to geothermal leakage from the hydrothermal reservoir below the lake bottom. The geothermal heat flux was calculated at 2.5 W/m$^2$ and 2.9 W/m$^2$ during the completely ice-covered seasons of 2019–2020 and 2020–2021, respectively.

Meanwhile, estimates of the hydrological and chemical budgets for Okama provided the groundwater inflow at $0.012$ m$^3$/s and the groundwater outflow at $0.039$ m$^3$/s, averaged over the three non-rainfall periods of 2020. In the chemical budget estimate, the $\text{SO}_4^{2-}$...
concentration was selected because $\text{SO}_4^{2-}$ was dominant in the water chemistry and highly correlated with the electric conductivity at 25 °C (EC25) as well as the pH of lake water and stream water. The negative groundwater inflow at $-0.028 \text{ m}^3/\text{s}$ can produce a decline in the lake level of $-0.029 \text{ m/day}$ during non-rainfall periods.

A coupling of the geothermal heat flux and groundwater inflow, which was obtained by two independent methods, evaluated the temperature of inflowing groundwater. Then, it was assumed that the geothermal heat flux calculated at 2.5 W/m$^2$ or 2.9 W/m$^2$ is supplied by the groundwater inflow estimated at 0.012 m$^3$/s. The negative groundwater inflow at $-0.028 \text{ m}^3/\text{s}$ can produce a decline in the lake level of $-0.029 \text{ m/day}$ during non-rainfall periods.

A coupling of the geothermal heat flux and groundwater inflow, which was obtained by two independent methods, evaluated the temperature of inflowing groundwater. Then, it was assumed that the geothermal heat flux calculated at 2.5 W/m$^2$ or 2.9 W/m$^2$ is supplied by the groundwater inflow estimated at 0.012 m$^3$/s. The groundwater temperature was calculated at 5.6 °C or 5.9 °C. These values are significantly larger than the annual mean air temperature (4.8 °C) at Okama. In the catchment without a geothermal heat source, the shallow groundwater is supposed to take a temperature near to the local annual mean air temperature. Hence, geothermal leakage from the underground hydrothermal reservoir is judged to be significant in the lake, and, therefore, it is necessary to continue monitoring Okama.

As a next step, an estimate of the thermal budget in the lake, including the heat flux at lake surface, is needed to determine whether the geothermal heat flux obtained by this study is reasonable or not in magnitude, as well as to understand how the inflowing and outflowing groundwaters are related to the hydrothermal flow system below Okama.

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