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Geothermal systems on the island of Bali, Indonesia

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²H and ¹⁸O isotope

1. Introduction

The island of Bali (Indonesia) hosts several geothermal systems and some are of interest to geothermal exploitation. The Bedugul geothermal field, located near Lake Bratan, covers an area of approximately 8 km² with an estimated annual electric energy potential of 80 MWe for 30 years (e.g., Hochstein et al., 2005; Mulyadi et al., 2005; Hochstein and Sudarman, 2008). However, the development was suspended due to environmental and cultural concerns. In addition to Bedugul, other geothermal prospects on Bali are the Batur, Banyuwedang and Banjar geothermal systems. To estimate the geothermal potential of a given geothermal system its reservoir temperature needs to be known. Ultimately this temperature is measured directly in a geothermal well, but prior to drilling, solute geothermometers are used to aid in geothermal exploration (e.g., Giggenbach, 1991). However, one of the prerequisites for their application is information about the

composition and type of the geothermal host rock (e.g., Giggenbach, 1991).

Bali is dominantly covered by volcanic rocks, overlying the Tertiary carbonate rocks that outcrop in the southern and western part of the island (Hadiwidjojo et al., 1998). The Bedugul geothermal field was reported producing brines where the gas phase was dominated by CO₂ of approximately 97 wt.% and thus it was thought that reservoir was in carbonate rocks (Mulyadi et al., 2005). To the contrary a thermobarometric study indicated that the reservoir could be in volcanic rocks and that the carbonate basement reacted with shallow magmatic intrusions in the Batur and Agung volcanoes (Geiger, 2014). The assimilation (thermal decomposition) of carbonate rocks by magma releases large amounts of CO₂ gas due to the breakdown of CaCO₃ into CO₂ and CaO (Allard, 1983; Gertisser and Keller, 2003; Chadwick et al., 2007; Marziano et al., 2007; Marziano et al., 2009; Deegan et al., 2010). The CO₂-rich volatile magma subsequently ascends and promotes phase separation in the geothermal reservoir, which in turn produces a CO₂-rich vapor phase (Lowenstern, 2001).

It is possible that geothermal systems on Bali could be hosted by carbonate rocks, but CO₂ content alone is not sufficient to allow that conclusion. Carbonate rocks, such as limestone and dolomite, for example,

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were identified as the reservoir rock in some volcanic areas in Italy, e.g., Vicano-Cimino and Sabatini-Tolfa (Cinti et al., 2011; Cinti et al., 2014). There the thermal water is characterized by a $(Ca^{2+} + Mg^{2+})/HCO_3^-$ molar ratio of ~1 as a result of calcite and/or dolomite dissolution (Gemici and Filiz, 2001; Levet et al., 2002; Grassa et al., 2006; Capaccioni et al., 2011; Cinti et al., 2011; Cinti et al., 2014). Another characteristic of thermal waters hosted by carbonate rocks can be a pronounced positive ^{18}O -shift due to the heavier $\delta^{18}O$ composition of carbonate rocks compared to magmatic rocks (Craig, 1966; Arana and Panichi, 1974; Gemici and Filiz, 2001). Although a depletion of $\delta^{18}O$ is also possible if there is isotope exchange between hydrothermal water and CO_2 , as demonstrated for some geothermal systems in Italy (Grassa et al., 2006; Cinti et al., 2011; Cinti et al., 2014).

This paper presents new physicochemical and isotope (^{18}O , 2H and ^{11}B) data for hot springs and shallow thermal wells on Bali with the

objective to investigate the host rock of the geothermal systems and to determine the most applicable geothermometer for geothermal exploration. Additionally, boron isotopes were applied to identify sea-water input and solute geothermometers to predict the reservoir temperatures.

2. Geological setting

Bali is a part of the Sunda-Banda volcanic islands arc, which extends for approximately 4700 km east to west, from the island of Damar to the island of Sumatera. The arc is caused by the convergence of the Indo-Australian and Eurasia plates, with a rate of 6 to 7 cm/a (Hamilton, 1979; Simandjuntak and Barber, 1996). This process drives volcanism on Bali since the late Tertiary (Van Bemellen, 1949; Hamilton, 1979; Hadiwidjojo et al., 1998) and produced a vast distribution of volcanic

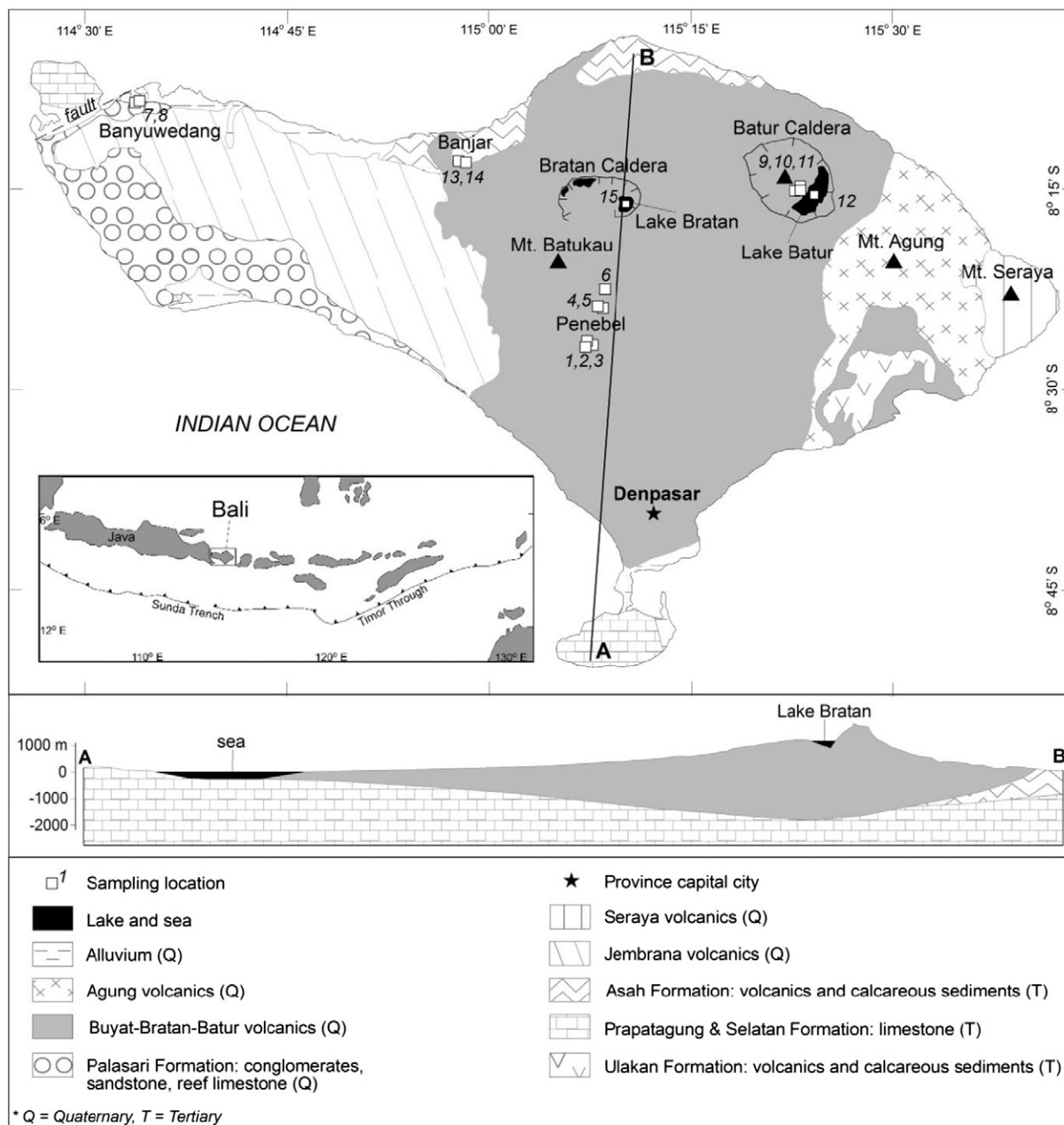


Fig. 1. Geological map of and cross section across (modified from Hadiwidjojo et al. (1998)). The sampling locations are indicated. Volcanic roots for the Batur and Bratan volcanic complexes were not indicated, due to a lack of geophysical mapping data. For scale: the distance between points A and B of the cross section is 80 km.

rocks. The Jembrana volcanic complex occupies the western part of the island, the Buyat–Bratan–Batur volcanic complex the central part, and the Agung and Seraya volcanic complexes the eastern parts. Underlying the volcanic rocks are sedimentary rocks of Tertiary age, which are minimally exposed in the east, south and west part of the island (Fig. 1) (Hadiwidjojo et al., 1998).

The volcanic rocks on Bali are calc-alkaline and characterized by intermediate contents of K (Whitford et al., 1979; Nicholls and Whitford, 1983). Based on the thermobarometric results of clinopyroxene and plagioclase at Batur volcano, Geiger (2014) suggested the existence of a shallow magma chamber in 2 to 4 km depth. This is corroborated by InSar satellite data indicating shallow magma at 2 to 4 km depth (Chaussard and Amelung, 2012) and by earthquake focal zones in 1.5 to 5 km depth (Hidayati and Sulaeman, 2013).

Explosive eruptions formed two large calderas on Bali: Batur and Bratan (Wheller and Varne, 1986; Reubi and Nicholls, 2004; Watanabe et al., 2010) and both are geothermal prospects. Although a geothermal reservoir was confirmed beneath the Bratan lake (Mulyadi et al., 2005), surface features, such as hot springs, are virtually absent in the Bratan caldera. Outside the caldera to the south, however, several hot springs are present in the Penebel area (Fig. 1). On the northwestern side of the Buyat–Bratan volcano, the Banjar hot spring is located at the contact between the Buyat–Bratan–Batur volcanic complex and the Tertiary Asah Formation. At the western end of the island, the Banyuwedang hot spring is located in carbonate rocks of the Prapatagung Formation.

Following the approach of Purnomo and Pichler (2014), which was developed for geothermal systems on Java, the geothermal systems on Bali were divided into volcano-hosted and fault-hosted geothermal systems based on their geologic association. The former is a geothermal system related to a volcanic complex and the latter is a geothermal system located in a fault zone. Thus, solely based on their geological setting geothermal systems on Bali can be divided into two groups, volcano-hosted are Batur (B9, B10 and B11) and Penebel (B1, B2, B3, B4 and B6), while Banyuwedang (B7) and Banjar (B13) are fault-hosted. Following the interpretation of Purnomo and Pichler (2014) only Banyuwedang (B7) seems to be a truly fault-hosted geothermal system.

3. Sampling and analysis

In October and November 2013 water samples were collected on Bali form hot springs and shallow wells, cold springs and freshwater lakes. Temperature, pH, conductivity, ORP and alkalinity were measured in the field by either probe or acid titration. The samples were filtered through a 0.45 µm nylon membrane and stored in polyethylene bottles for anion, cation and isotope (²H, ¹⁸O and ¹¹B) analyses. The split sample for cation and ¹¹B isotope was acidified to 1% concentrated HNO₃. Cations (Ca²⁺, Mg²⁺, Na⁺, K²⁺ and Sr²⁺), metals (Al, Fe and Mn) and trace elements (B and Li) were measured by inductively coupled plasma-optical emission spectrometry (ICP-OES) using an Optima 7300 instrument (Perkin Elmer). Anions, Cl⁻ and SO₄²⁻, were analyzed by ion chromatography using an IC Plus Chromathograph (Metrohm). ²H and ¹⁸O isotope were measured using an LGR DLT-100 laser spectrometer (Los Gatos Research). Boron isotopes were analyzed by multi-collector inductively plasma mass spectrometry following Wang et al. (2010) at the National University of Taiwan. ²H and ¹⁸O isotopes are reported in δ per mil (‰) relative to VSMOW and ¹¹B isotope relative to SRM NBS 951. The analytical uncertainty of δ²H was ±1‰, δ¹⁸O ± 0.2‰ and δ¹¹B < 0.2‰.

4. Results

The field and laboratory measurements for the water samples from Bali are presented in Table 1. The temperatures of thermal waters ranged from 37.2 to 45.2 °C, while the selected cold waters ranged from 24 to 30.1 °C. The thermal waters had slightly acid to neutral pH, while cold springs were neutral and lake waters were slightly alkaline.

Table 1 Sampling locations, physicochemical and stable isotope (²H, ¹⁸O and ¹¹B) data of thermal and cold waters on Bali. The pCO₂ was calculated using the computer code PHREEQC (Pichler et al., 1999).

Sample ID	Location	T (°C)	pH	ORP (mV)	Cond. (µS/cm)	TDS (mg/L)	Ca	Mg	Na	K	Cl	HCO ₃	SO ₄	Si	Al	Li	B	Sr	δ ¹⁸ O (‰)	δ ² H	δ ¹¹ B	Log pCO ₂
<i>Hot springs</i>																						
B1	Yeh Panas 1, Penebel	38.8	6.5	25	3016	2238	122.4	161.6	263.4	50.1	377.0	1466.4	<0.1	76.0	0.2	1.3	8.0	0.4	-6.5	-33.2	n.d.	-0.5
B2	Yeh Panas 2, Penebel	38.8	6.6	0	3052	2272	122.5	164.0	270.3	51.2	363.7	1525.0	<0.1	76.0	0.2	1.3	7.6	0.4	-6.1	-36.4	n.d.	-0.5
B3	Yeh Panas 3, Penebel	42.6	6.4	-41	3282	2453	135.1	161.2	309.4	58.4	443.9	1555.5	<0.1	80.8	0.2	1.4	9.1	0.4	n.d.	n.d.	10.4	-0.4
B4	Belulang, Penebel	41.8	6.5	-47	3402	2555	211.5	243.8	234.5	67.2	61.2	2235.0	111.7	72.8	0.2	1.4	4.3	0.8	-6.8	-40.7	4.0	-0.2
B6	Angseri, Penebel	45.2	6.1	1	1363	943	54.3	81.7	123.0	40.0	16.6	634.4	166.0	97.3	0.1	0.7	5.4	0.2	-5.8	-33.3	n.d.	-0.6
B7	Banyuwedang	44.6	7.8	-310	3453	2600	51.3	51.6	526.8	14.8	902.1	31.7	200.2	11.7	0.2	1.2	1.1	0.3	-6	-36	22.5	-3.2
B13	Banjar	37.2	6.2	-41	1265	873.6	68.4	66.2	109.2	23.6	17.3	773.5	2.2	73.1	0.1	0.6	1.9	0.2	-6.1	-37.1	1.7	-0.6
<i>Thermal shallow wells</i>																						
B9	Toya Bongkah, Batur	39.9	7.5	163	2122	1523	46.0	75.8	294.2	24.0	159.3	463.6	370.3	56.6	0.2	1.3	1.9	0.1	-6.4	-42.4	1.3	-1.8
B10	Tirta Husada, Batur	43.1	7.3	191	2007	1430	46.6	69.7	277.8	22.8	136.2	458.7	325.2	62.0	0.2	1.3	2.0	0.1	-6	-41.7	n.d.	-1.6
B11	Toya Devasya, Batur	40.6	7.4	163	2055	1478	47.4	71.1	281.6	22.8	147.0	488.0	328.7	57.2	0.3	1.3	2.0	0.1	-6.8	-41.9	n.d.	-1.7
<i>Volcanic lakes</i>																						
B12	Batur Lake	27.2	8.5	135	2122	1525	31.8	67.2	321.2	26.0	188.6	336.7	488.6	1.3	0.3	1.4	1.7	0.0	-1.7	-16.4	n.d.	-3.0
B15	Bratan Lake	24	8.7	195	4912	32.2	4.6	0.3	<0.1	0.8	<0.1	19.5	2.7	<0.1	0.2	0.6	0.4	<0.01	-2.3	-14.5	n.d.	-4.4
<i>Cold springs and shallow wells</i>																						
B5	Belulang	30.1	7.0	36	1057	727.9	48.9	49.5	48.6	13.4	4.3	575.8	28.0	27.5	0.1	0.6	1.0	0.2	-6	-30	n.d.	-1.3
B8	Pejarakan	29	7.4	93	4974	3891	40.0	74.5	874.9	68.9	1025.7	761.3	361.5	31.0	0.2	1.2	1.1	1.0	-5.7	-36.9	n.d.	n.d.
B14	Banjar	30.9	7.7	143	876.2	598.5	55.4	41.2	64.2	15.6	2.6	561.2	4.7	52.3	0.1	0.6	1.1	0.3	-5.4	-31.4	n.d.	-1.9

*n.d. = not determined.

The thermal and cold waters had relatively similar ranges of TDS, Cl^- , Na^+ and K^+ concentrations. Meanwhile the thermal waters had wider ranges of Ca^{2+} , Mg^{2+} and HCO_3^- contents compared to the cold waters. The Ca^{2+} concentration of the thermal waters ranged from 51.3 to 211.5 mg/L, Mg^{2+} from 51.5 to 243.8 mg/L and HCO_3^- from 31.7 to 2235 mg/L, while for cold waters the Ca^{2+} and Mg^{2+} were lower than 100 mg/L and HCO_3^- ranged from 19.5 to 761.3 mg/L. The Cl^- content of thermal waters ranged from 17.3 to 902.1 mg/L and for cold waters from under detection limit to 1025.7 mg/L. The highest Cl^- concentrations in cold waters were found in B12 (Batur Lake) and B8 (Pejarakan), with Cl^- contents of 188.6 and 1025.7 mg/L, respectively. The thermal waters B1, B2, B3, and B4 had Ca^{2+} , Mg^{2+} and HCO_3^- concentrations a magnitude higher than the other thermal waters. However, B4 differed from B1, B2 and B3 due to its SO_4^{2-} content of 111.7 mg/L and Cl^- content of 61.2 mg/L, compared to SO_4^{2-} of below detection limit and Cl^- ranged from 377 to 444 mg/L in B1, B2 and B3. Thermal waters B6 and B13 had a TDS of <1000 mg/L and Cl^- <20 mg/L, while the other thermal waters varied between 1430 and 2600 mg/L and between 61.2 and 902.1 mg/L, respectively. The thermal water B7, which was located near the coastline had the highest TDS and Cl^- and the lowest HCO_3^- concentrations.

The thermal waters had $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values ranging from -42.4 to -33.2% and from -6.8 to -5.6% , respectively. The $\delta^2\text{H}$ of cold springs and a shallow well ranged from 36.9 to -30.0% and $\delta^{18}\text{O}$ ranged from -6.0 to -5.4% . In contrast, cold waters from two freshwater lakes, Batur and Bratan, had heavier $\delta^2\text{H}$, ranging from -16.4 to -14.6% , and $\delta^{18}\text{O}$, from -2.3 to -1.7% . The $\delta^{11}\text{B}$ compositions ranged from $+1.3$ to $+22.5\%$. The heaviest $\delta^{11}\text{B}$ value was found in sample B7, which also had the highest Cl^- content.

5. Discussion

5.1. Geochemistry of thermal waters

A geothermal system generally produces three types of hot springs, neutral chloride, acid sulfate and bicarbonate waters, but mixtures between the individual groups are common (White, 1957; Hedenquist, 1990; Nicholson, 1993; Hochstein and Browne, 2000). The discharge composition of thermal springs is controlled by two sets of processes: 1) deep reservoir conditions (deep reservoir = reaction zone immediately above the heat source), and 2) secondary processes during ascent. In the deep reservoir, host-rock composition, temperature, direct magmatic contributions and residence time are the controlling factors. During ascent a drop in pressure and temperature can initiate phase separation and mineral precipitation, causing a dramatic change in fluid composition. Mixing with other hydrothermal fluids and/or groundwater is possible at any depth. In near-shore and submarine environments mixing with seawater cannot be ruled out. The chemical composition of a hydrothermal fluid, sampled at the surface, generally contains an imprint of its subsurface history and chemically inert constituents (tracers) provide information about their source, whereas chemically reactive species (geoindicators) record physicochemical changes (Ellis and Mahon, 1977; Giggenbach, 1991; Nicholson, 1993). Classification of thermal waters on Bali using the $\text{Cl}-\text{SO}_4-\text{HCO}_3$ ternary diagram (Chang, 1984; Giggenbach, 1991; Giggenbach, 1997) indicated a bicarbonate (HCO_3^-) type for B1, B2, B3, B4, B6 and B13, a mixing type for B9, B10 and B11, and a neutral chloride (Cl^-) type for B7 (Fig. 2). Neutral chloride waters are usually thought to represent the deep reservoir fluid, while acid sulfate and bicarbonate waters form by

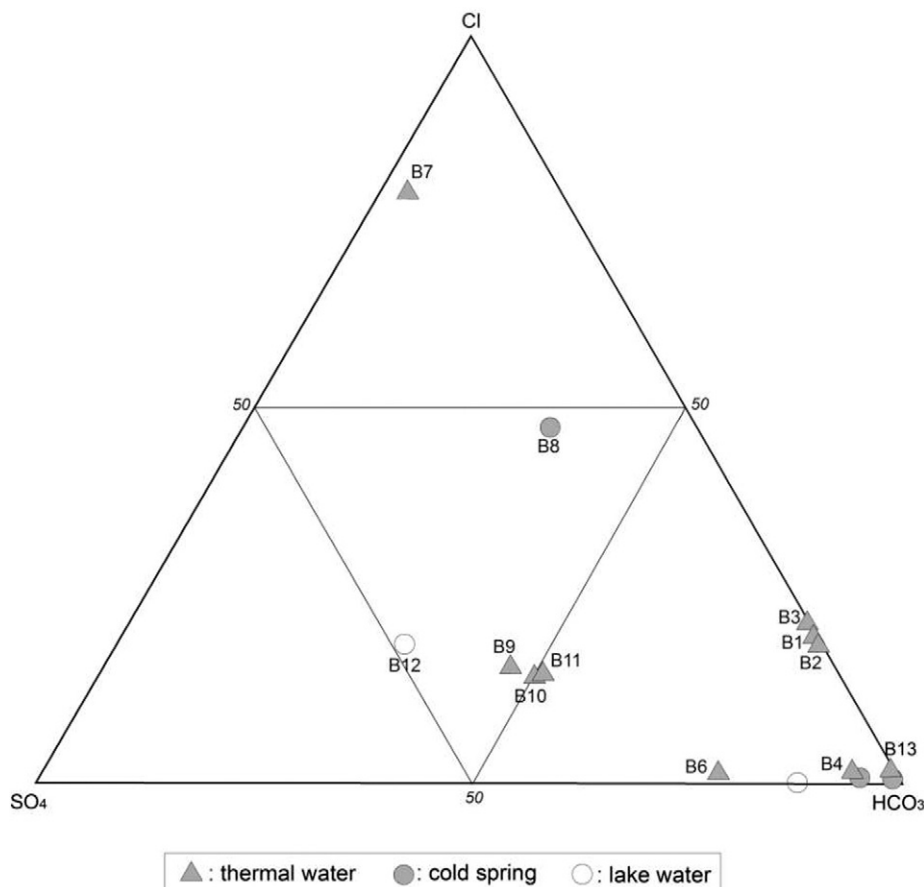


Fig. 2. $\text{Cl}-\text{SO}_4-\text{HCO}_3$ ternary diagram (Giggenbach diagram). Most of the thermal waters were of the bicarbonate type. The positions of the samples from Batur (B9, B10 and B11) indicate mixing. Sample B7 plots close to Cl, due to seawater input.

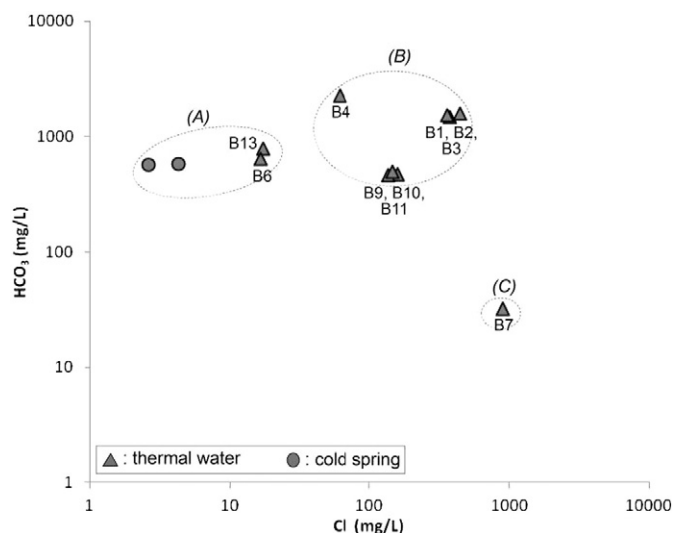


Fig. 3. HCO_3^- vs. Cl diagram. Groups A and B are volcano-hosted thermal waters from the 'primary neutralization' zone with different degrees of dilution by shallow groundwater. Sample B7 is from a fault-hosted geothermal system.

underground absorption of vapors separated from a neutral chloride water into cooler ground water. Whether acid sulfate or bicarbonate waters are formed depends on the gas content of the vapor and redox conditions in the shallow subsurface (Ellis and Mahon, 1977; Henley and Ellis, 1983; Hedenquist, 1990; Giggenbach, 1997).

The TDS value of 1525 mg/L of B12 was relatively similar to those of the Batur thermal waters (B9, B10 and B11), probably due to major

discharge of thermal water into the lake. However, it should be noted that B12 was sampled from a site relatively close to of Batur thermal waters, hence considering the large dimension of the lake of approximately 6.6 km length and 2.5 km width, the sample probably does not represent the general chemistry of the lake water. Meanwhile, the higher Cl^- concentration of the shallow well (B8) was likely caused by seawater input due to its location close to sea.

Solely based on their geological setting geothermal systems on Bali can be divided into volcano-hosted and fault-hosted. Following the interpretation of Purnomo and Pichler (2014) only Banyuwedang (B7) was considered a truly fault-hosted geothermal system (Fig. 3), because its thermal water had a low HCO_3^- content. The volcano-hosted thermal waters were further divided into two groups, A and B. Group A samples represent diluted thermal waters and group B samples are from the margin of the 'primary neutralization' zone as defined by Giggenbach (1988). In this zone, HCO_3^- is produced by hydrolysis of CO_2 and subsequent mixing with groundwater controls the Cl^- concentration of thermal waters. The HCO_3^- content of thermal waters was well correlated ($R^2 \sim 0.9$) with those of Ca^{2+} , Mg^{2+} , Sr^{2+} and K^+ (Fig. 4), thus suggesting rock dissolution triggered by the formation of carbonic acid (H_2CO_3), at temperatures below 300 °C due to hydrolysis of CO_2 in groundwater (Bischoff and Rosenbauer, 1996; Giggenbach, 1997; Lowenstern, 2001). The $(\text{Ca}^{2+} + \text{Mg}^{2+})/\text{HCO}_3^-$ molar ratios of the thermal waters were approximately 0.4 (Fig. 5), a ratio which is significantly lower than a ratio of 1, which was reported for thermal waters from carbonate rock hosted geothermal systems (e.g., Cinti et al., 2014). Another line of argumentation against a carbonate reservoir is the low molar $\text{Ca}^{2+}/\text{Mg}^{2+}$ ratio of approximately 0.5 in the geothermal waters from Bali (Table 1). This ratio has to be above 1 for a limestone/dolomite reservoir. Accordingly, a carbonate host-rock reservoir for the geothermal systems on Bali is not fully supported and thus we suggest calc-alkaline magmatic rocks

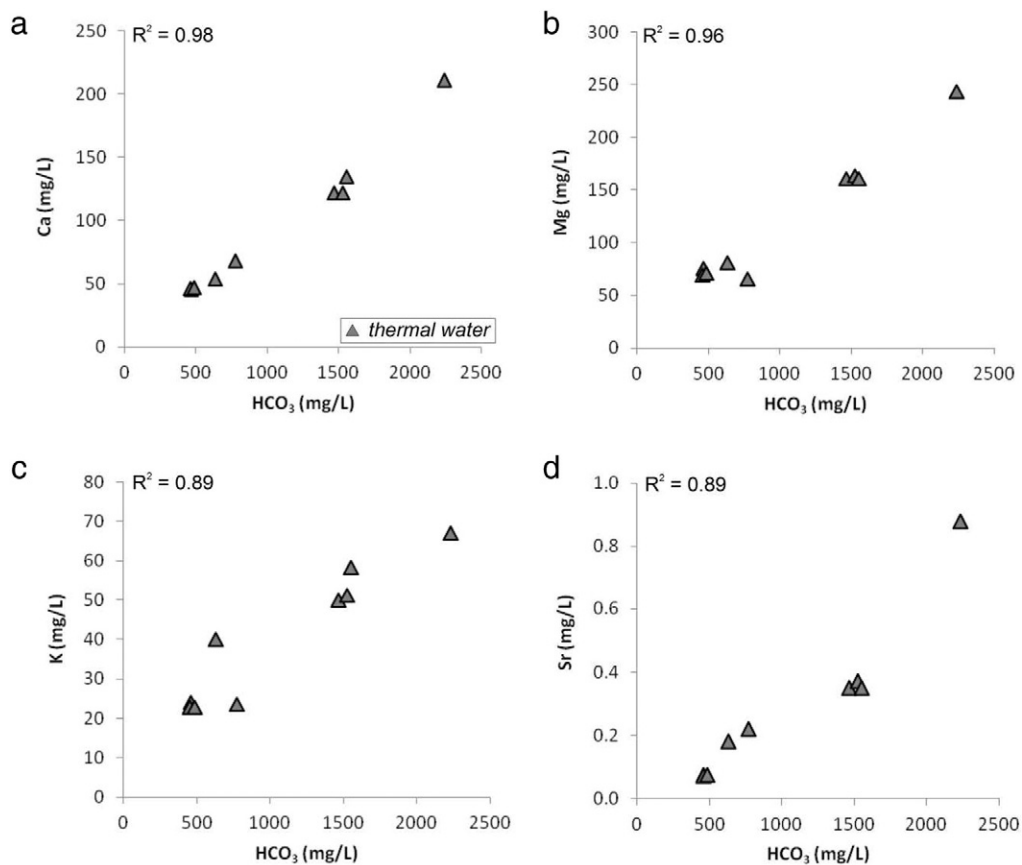


Fig. 4. a) Ca vs. HCO_3^- , b) Mg vs. HCO_3^- , c) K vs. HCO_3^- and d) Sr vs. HCO_3^- diagrams. The apparent linear correlation between HCO_3^- and Ca^{2+} , Mg^{2+} , K^+ and Sr^{2+} suggests water-rock interaction in the presence of carbonic acid.

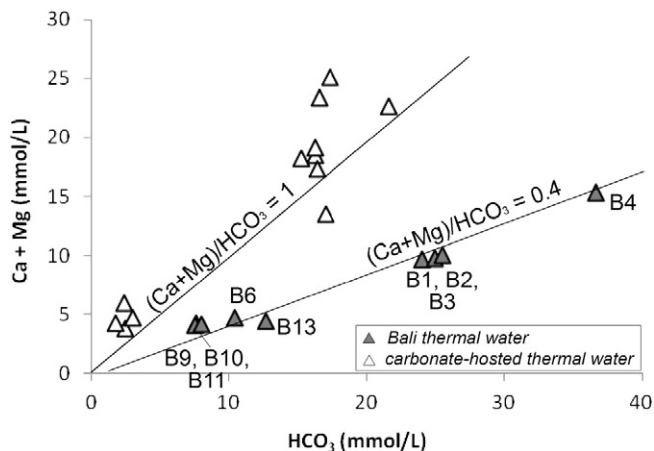


Fig. 5. The $(Ca^{2+} + Mg^{2+})/HCO_3^-$ molar ratios of Bali thermal waters compared to those of thermal waters from carbonate-hosted geothermal systems. The data is from Cinti et al. (2014), Capaccioni et al. (2011) and Levet et al. (2002).

as the likely geothermal reservoir, as also suggested in the K/Mg vs. Na/K diagram (Fig. 6). This type of magmatic rock is generally produced between the tholeiitic and high-K calc-alkaline magmatic regions in the center of a volcanic island arc (Whitford et al., 1979).

The final pH of the thermal waters was controlled mainly by addition of CO_2 and thus H_2CO_3 formation, followed by water-rock interaction. While addition of CO_2 lowers the pH, water-rock interaction causes an increase due to the consumption of H^+ . The effect of CO_2 addition to thermal waters in the Penebel group (B1, B2, B3, B4 and B6) and in Banjar (B13) dominated that of water-rock interaction and thus those samples had a slightly acidic pH (Fig. 7a). The Batur group (B9, B10 and B11) samples had a pH above 7, which would either indicate a relatively larger degree of water-rock interaction or less addition of CO_2 . Those samples were saturated with respect to calcite, thus being well buffered (Fig. 7b).

5.2. Phase separation and seawater input

The B/Cl ratio and $\delta^{11}B$ of thermal waters can be used to identify water-rock interaction, steam separation and seawater input in the subsurface of a geothermal system (Arnorsson and Andresdottir, 1995; Valentino and Stanzione, 2003; Purnomo and Pichler, 2014). During phase separation B is preferentially partitioned into the vapor phase,

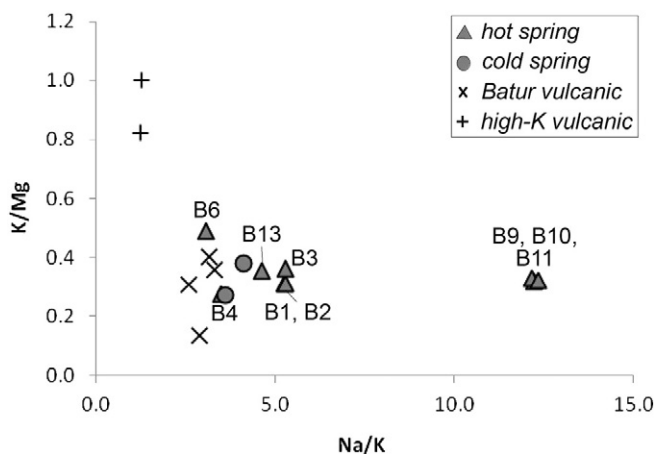


Fig. 6. K/Mg vs. Na/K diagram where hot springs and volcanic rocks of the Batur volcano plot together indicating water-rock interaction in a geothermal reservoir hosted by calc-alkaline volcanic rocks of the Batur volcanic complex. Data of calc-alkaline volcanic rocks of Batur are from Reubi and Nicholls (2004) and high-K volcanic rocks of Muria volcano are from Whitford et al. (1979).

while Cl tends to be enriched in the liquid phase (Truesdell et al., 1989; Arnorsson and Andresdottir, 1995) and thus their ratio can be applied to unravel part of the physicochemical history of a geothermal fluid. A high ratio would indicate phase separation, while a low ratio would indicate seawater input. The Cl vs. B and diagram of thermal waters from Bali indicated water-rock interaction with andesitic rocks for the thermal waters of the Penebel group (B1, B2 and B3) and the Batur group (B9, B10 and B11). Either B depletion or seawater input was indicated for Banyuwedang (B7) and phase separation for Belulang (B4), Angseri (B6) and Banjar (B13) (Figs. 8 and 9). Phase separation was also confirmed by direct observation of a steam phase in the reservoir of the Bedugul geothermal field in geothermal exploration wells (Hochstein et al., 2005).

The heavy $\delta^{11}B$ composition of seawater of +39.61‰ (Foster et al., 2010) can be used to investigate seawater input or B adsorption onto minerals to explain the B/Cl ratio observed for the Banyuwedang (B7) hot spring. This method was successfully applied for some geothermal systems on Java island, Indonesia (Parangtritis and Krakal), Iceland (the Reykjanes and Svartsengi) and three areas in Japan (the Izu-Bonin arc, Kusatsu-Shirane area, and Kagoshima) (Kakihana et al., 1987; Musashi et al., 1988; Oi et al., 1993; Aggarwal and Palmer, 1995; Aggarwal et al., 2000; Millot et al., 2009). Adsorption of B by minerals also increases the $\delta^{11}B$ composition of a liquid phase due to the preferential fractionation of ^{10}B into the solid phase (Schwarcz et al., 1969; Palmer et al., 1987; Xiao et al., 2013). The magnitude, however, is lower than what can be observed due to seawater input. Banyuwedang (B7) had a $\delta^{11}B$ composition of +22.5‰ and a B/Cl ratio of 806; therefore, it plots close to the mixing line between seawater and thermal water in the $\delta^{11}B$ vs. B/Cl diagram (Fig. 9). In addition this sample also had the B/Cl ratio closest to that of seawater (Fig. 8), although total concentrations of both elements were lower due to mixing with groundwater. However since groundwater contains relatively more B than Cl the sample did not plot exactly on the B/Cl line in Fig. 8. The Na/Cl ratio in sample B7 is more or less identical to that of seawater, i.e., 0.57 vs. 0.58 (calculated in mg/L), which further corroborates seawater input.

5.3. Oxygen and hydrogen isotope considerations

The deuterium and oxygen isotopic composition of thermal waters has been successfully applied to investigate the fluid origin, i.e., meteoric, marine or magmatic, mixing and physicochemical processes, such as, water-rock interaction and water- CO_2 isotope exchange (Craig et al., 1956; Arnason, 1977; Giggenbach et al., 1983; Gemici and Filiz, 2001; Pichler, 2005; Cinti et al., 2014; Purnomo and Pichler, 2014). The δ^2H and $\delta^{18}O$ composition of water is generally a good indicator of its origin. Hydrothermal fluids normally plot to the right of the LMWL due to the exchange of ^{18}O during water-rock interaction (Craig, 1966) or due to subsurface mixing with an andesitic water (Giggenbach, 1992). Particularly in carbonate reservoirs this shift is pronounced compared to silicate reservoirs, because carbonates are comparably enriched in ^{18}O with values close to 29‰ VSMOW vs. approximately 8‰ to 10‰ VSMOW for silicates (Clark and Fritz, 1997).

In the δ^2H vs. $\delta^{18}O$ diagram all thermal waters, except the two lake waters, plot close to the local meteoric water line (LMWL) and weighted mean annual value for precipitation in the region, indicating local rainwater as the ultimate fluid source (Fig. 10). With the exception of sample B10 the thermal waters from Bali did not shift to the right of the LMWL indicating neither water-rock isotope exchange nor substantial mixing with an andesitic water. Neither process, however, can be completely ruled out because an initial ^{18}O -shift to the right may have been later reversed by a subsequent isotope exchange between CO_2 and H_2O . Such isotope shifts were observed in several CO_2 -rich aquifers and hydrothermal waters (Vuataz and Goff, 1986; Chiodini et al., 2000; Grassa et al., 2006; Cinti et al., 2011; Cinti et al., 2014). The Tirta Husada hot spring (B10) from the Batur group plots slightly right shifted from

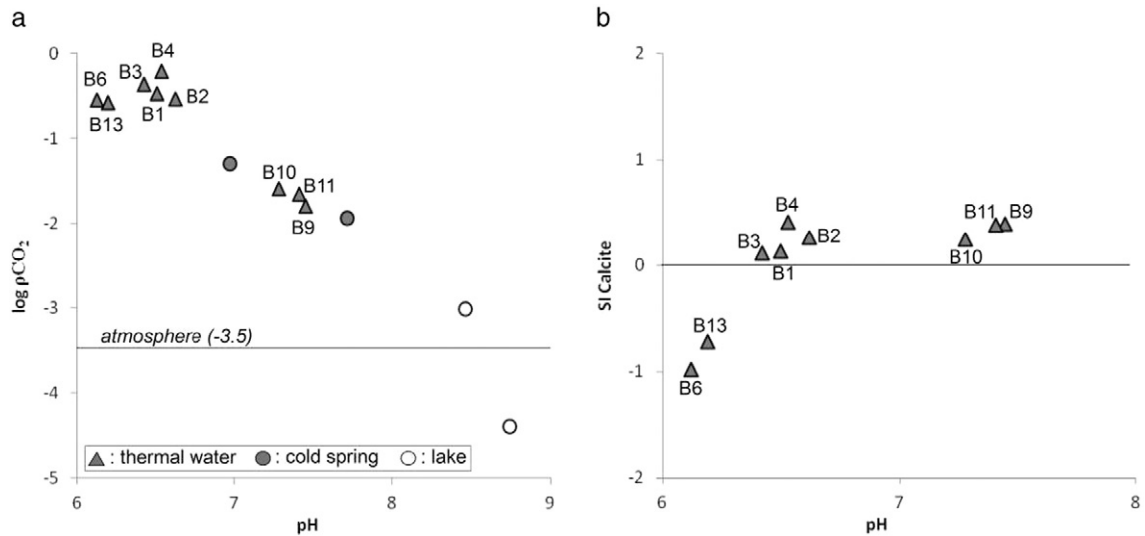


Fig. 7. a) Diagrams of $\log p\text{CO}_2$ vs. pH and b) calcite saturation index (SI) vs. pH. Thermal waters with high $p\text{CO}_2$ have lower pH, which is typical for CO_2 -fed thermal waters. The water pH probably was also buffered by calcite precipitation.

Toya Devasya (B11), potentially signaling some water–rock interaction. However, the TDS of B10 was relatively similar to B11 and B9 (Fig. 11). The absence of a pronounced horizontal, $\delta^{18}\text{O}$ -only shift to heavier values also points towards a non-carbonate reservoir, because due to water–rock interaction heavier $\delta^{18}\text{O}$ values are generally observed in carbonate reservoirs (e.g., Arana and Panichi, 1974; Gemici and Filiz, 2001; Levet et al., 2002; Grassa et al., 2006; Capaccioni et al., 2011; Cinti et al., 2011; Cinti et al., 2014).

The elevated $\delta^2\text{H}$ and $\delta^{18}\text{O}$ values in Lake Batur (B12) and Lake Bratan (B15), which were studied in detail by Varekamp and Kreulen (2000), were a result of evaporation from a cold lake. B12 was slightly enriched in ^{18}O compared to B15, likely due to more input of thermal waters into the lake that elevated its temperature. A lake with a higher temperature would produce a heavier $\delta^{18}\text{O}$ due to evaporation (Gonfiantini, 1986; Varekamp and Kreulen, 2000).

5.4. Geothermometry

The reservoir temperatures of geothermal systems on Bali were calculated using the solute geothermometers SiO_2 , Na–K–Ca, Na/K and Na/Li (Table 2) (Fournier and Truesdell, 1973; Fournier, 1977;

Fournier, 1979; Kharaka and Mariner, 1989). These geothermometers are based on temperature-dependent equilibrium reactions, hence application of multiple geothermometers can be used to evaluate secondary processes during thermal water ascent from the reservoir to the surface. These processes include dilution/mixing, conductive cooling, adiabatic cooling, mineral precipitation, water–rock interaction and re-equilibration (Fournier, 1977; Kaasalainen and Stefánsson, 2012). Such an evaluation has been successfully applied, for instances, on Java, Indonesia and Ambitle island, Papua New Guinea (Pichler et al., 1999; Purnomo and Pichler, 2014).

The silica geothermometer calculated lower reservoir temperatures compared to the Na/K, Na–K–Ca and Na/Li geothermometers, which is common for samples taken at the surface from hot springs, rather than directly from the geothermal reservoir. That geothermometer predicted reservoir temperatures ranging from 44 to 136 °C, while those predicted by the Na/K thermometer ranged from 257 to 773 °C, those predicted by the Na–K–Ca thermometer from 130 to 236 °C and those

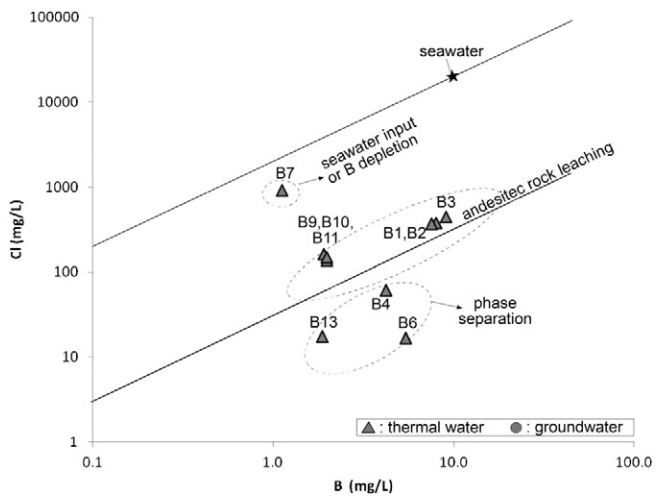


Fig. 8. Cl vs. B diagram illustrates four processes in the sub surface, i.e., a steam phase separation for B4, B6 and B13; an andesitic rock leaching for B1 to B3 and B9 to B11; and either a B depletion or seawater input for B7.

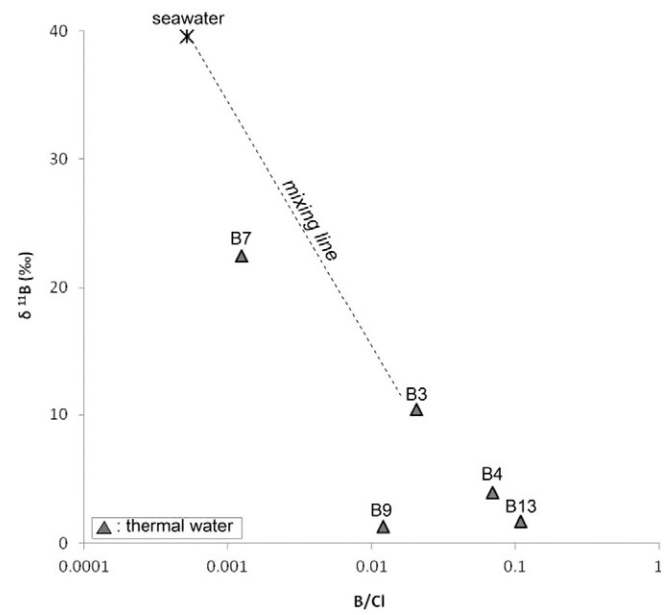


Fig. 9. The plot of Banyuwedang close to the mixing line of seawater–thermal water in the $\delta^{11}\text{B}$ vs. B/Cl diagram confirms seawater input.

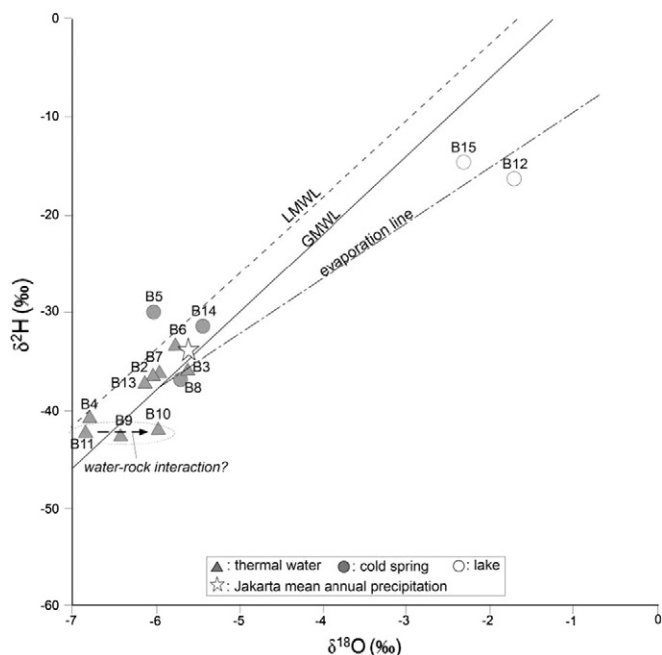


Fig. 10. $\delta^2\text{H}$ vs. $\delta^{18}\text{O}$ diagram with the local meteoric water line (LMWL) from Wandou (2001), GMWL from Craig (1966) and the mean annual value for precipitation in Jakarta from IAEA/WMO (2015).

predicted by the Na/Li thermometer from 162 to 254 °C (Table 2). The calculation of silica geothermometry is based on absolute silica content, hence sensitive to boiling, precipitation and dilution (e.g. Nicholson, 1993). This deficiency can be overcome by calculating the silica parent using the silica mixing model of Fournier (1977). However, this method could not be used on Bali because the thermal waters in a given geothermal system had relatively similar temperatures and silica contents, for example the Penebel group (B1, B2, B3, B4 and B6). Meanwhile, the Na/K temperatures are likely overestimations caused by competition of Ca^{2+} , Na^+ and K^+ during ion exchange (Nicholson, 1993). The use of the Na/Li geothermometer resulted in reservoir temperatures ranging from 235 to 254 °C for the Penebel thermal waters (B1, B2, B3, B4 and B6), which were relatively similar to actual reservoir temperatures at 1800 m below ground of the nearby Bedugul geothermal field of 243 °C (Mulyadi et al., 2005). This indicates the applicability of the Na/Li geothermometer as has been proposed by, e.g., Fouillac and Michard (1981). Based on this, the reservoir temperatures of the other geothermal systems on Bali, with an exception of Banyuwedang (B7),

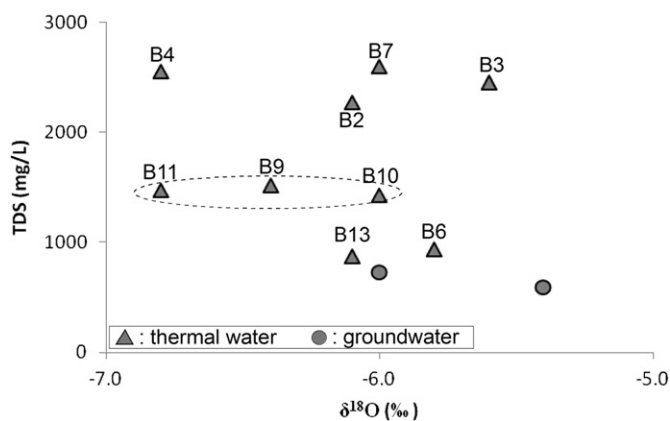


Fig. 11. TDS vs. $\delta^{18}\text{O}$ diagram shows relatively similar TDS for the Batur thermal waters, B9, B10 and B11, hence indicating an insignificant ^{18}O enrichment due to water-rock interaction.

Table 2
Calculated reservoir temperatures using solute geothermometers.

Location	Sample ID	Geothermometers (°C)			
		Silica (Truesdell et al., 1989)	Na-K-Ca (Fournier and Truesdell, 1973)	Na/K (Fournier, 1979)	Na/Li (Kharaka and Mariner, 1989)
Penebel	B1	122	209	568	244
	B2	122	209	567	242
	B3	126	211	566	235
	B4	120	227	688	254
	B6	136	236	733	254
	Banyuwedang	B7	44	130	257
Batur	B9	108	172	395	236
	B10	122	171	396	240
	B11	108	171	394	239
Banjar	B13	120	205	602	255

were predicted using Na/Li geothermometer. Therefore, the reservoir temperature of the Batur geothermal system was approximately 240 °C and Banjar was 255 °C. However, due to the input of seawater in Banyuwedang (B7), a geothermometer based on Na^+ content is unreliable. The silica geothermometer predicted a temperature of 44 °C for B7, similar to the discharge temperature and hence a likely underestimation. Therefore, the reservoir temperature of B7 could not be reliably calculated, but probably is lower than 100 °C, a temperature similar to most of the fault-hosted geothermal system on Java (Purnomo and Pichler, 2014).

Calcite precipitation during thermal water ascent was predicted by comparing the result of Na/Li and Na-K-Ca geothermometers. Precipitation of calcite during thermal water ascent reduces the Ca^{2+} concentration and thus should result in lower calculated Na-K-Ca temperatures. Calculated temperatures ranged from 209 to 236 °C for the Penebel group of thermal waters and thus were similar to those calculated with the Na/Li geothermometer, which would indicate insignificant calcite precipitation during fluid ascent. In contrast, the difference between 170 °C (Na-K-Ca) and 240 °C (Na/Li) for the Batur group of thermal waters indicates that calcite precipitated during ascent. This is corroborated by the lower ρCO_2 and higher pH compared to the Penebel thermal waters (Fig. 7).

6. Conclusions

Two types of geothermal systems are present on Bali. The Banyuwedang geothermal system was considered the fault-hosted and the Penebel, Batur and Banjar geothermal systems were considered volcano-hosted. Contrary what was previously suggested (Mulyadi et al., 2005), we may conclude that, although Bali is underlain by a carbonate basement, the geothermal systems there are hosted by calc-alkaline magmatic rocks. Steam phase separation occurred in the Penebel and Banjar geothermal systems, while seawater input was confirmed for the fault-hosted geothermal system of Banyuwedang. The hydrogen and oxygen isotopic composition indicated that the geothermal reservoirs are fed by meteoric water without significant water-rock interaction and/or mixing with an andesitic magmatic fluid. The Na/Li thermometer provided the best results for geothermal systems on Bali. Using this thermometer, the following reservoir temperatures were calculated: (1) Penebel (Bedugul) from 235 to 254 °C, (2) Batur 240 °C and (3) Banjar 255 °C.

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