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PETROLOGIC EVIDENCE FOR BOILING TO DRYNESS IN THE KARAHA-TELAGA BODAS GEOTHERMAL SYSTEM, INDONESIA

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ABSTRACT

Karaha-Telaga Bodas is a recently discovered geothermal system located adjacent to Galunggung Volcano in western Java. Drilling to depths of 3 km has documented an extensive vapor-dominated zone and a deeper liquid reservoir with measured temperatures up to 350°C and salinities of 1-2 weight percent.

The vein parageneses record the evolution of a high temperature liquid-dominated system that developed before the onset of vapor-dominated conditions. The change to a vapor-dominated regime was marked by the deposition of chalcedony and quartz. Fluid inclusions trapped in quartz and the mineral relationships indicate that the pressure gradients at the start of this change greatly exceeded hydrostatic and that much of the chalcedony was deposited at temperatures >250°C. The deposition of chalcedony at these temperatures implies catastrophic depressurization and boiling of the reservoir liquids.

As pressures within the reservoir declined, downward percolating steam condensate deposited anhydrite, calcite, fluorite, pyrite, and chlorite. At depths greater than 800 m, boiling off of the condensate resulted in increasing salinities with increasing temperature. Hypersaline fluids with salinities of 31 weight percent NaCl equivalent were trapped at 300°C in anhydrite. Rock surfaces were subsequently coated with scales of NaCl, KCl, $FeCl_x$, and Ti-Si-Fe as the condensate boiled to dryness.

The age of chalcedony deposition is constrained by ^{14}C dating of lakebeds from a depth of 988.8 m. These deposits, which predate the formation of chalcedony and the carbonate and propylitic veins, yielded an age of 5910 ± 76 years BP. The depressurization and boiling that caused deposition of the chalcedony could have been triggered by massive slope failure of the volcanic edifice ~4200 years ago. This event produced the crater at Kawah Galunggung. The low salinities of the modern reservoir fluids imply that mixtures of meteoric water and condensate are now recharging the system.

INTRODUCTION

Vapor-dominated geothermal regimes may be a common, but transient feature of andesitic volcanoes. Low- to moderate-temperature vapor-dominated regimes frequently form parasitic caps over liquid-dominated regions. More extensive vapor-dominated regimes like those found at Darajat-Kamojang and Karaha-Telaga Bodas, Indonesia have temperatures of ~240° to >350°C. Allis and Moore (2000) suggested that these zones might develop from higher temperature magmatic chimneys (Reyes et al., 1983) due to the combined effects of low permeabilities, high-heat flow and open-system boiling.

In this paper, we use mineral relationships, fluid inclusion measurements and ^{14}C dating from Karaha-Telaga Bodas to characterize an early, over pressured liquid system close to a magma body, and its evolution to a vapor-dominated regime.

GEOLOGIC SETTING OF THE KARAHA-TELAGA BODAS GEOTHERMAL SYSTEM

The geothermal system at Karaha-Telaga Bodas was discovered in the mid 1990s by the Karaha Bodas Co. LLC. The field is located beneath a north-trending ridge extending from Kawah Galunggung (Fig. 1). Galunggung Volcano erupted 5 times between 1822 and 1984. The main vent is a horseshoe-shaped crater, Kawah Galunggung, that is thought to have formed ~4200 years ago as a result of massive slope failure (Katili and Sudradjat, 1984). The resulting debris-avalanche flow produced extensive hummocky deposits southeast of the crater, forming the Ten Thousand Hills of Tasikmalaya.

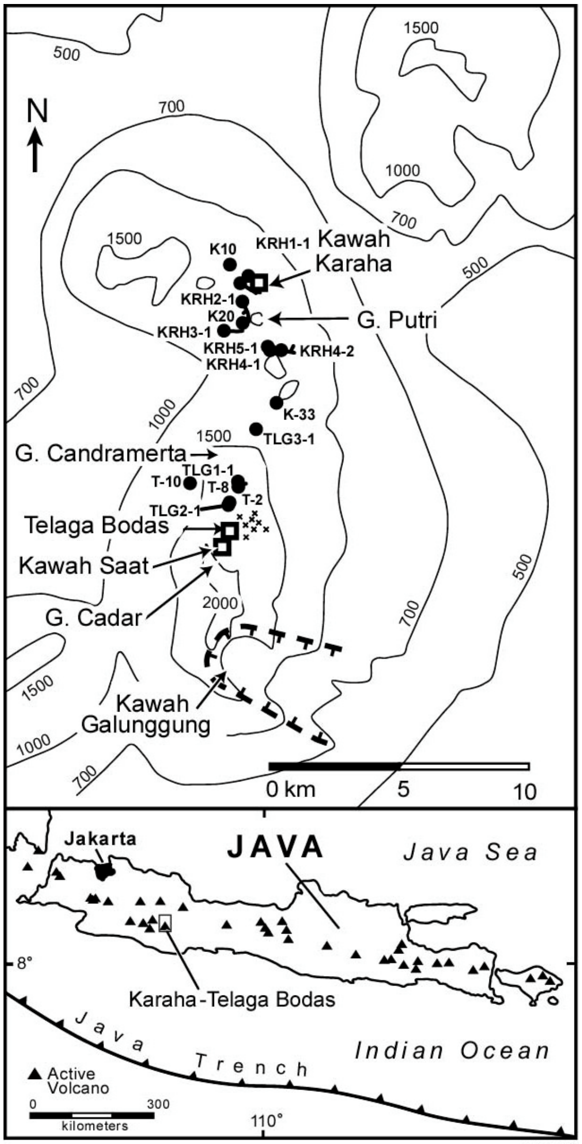

Fig. 1. Map illustrating the distribution of volcanic features, thermal manifestations (Telaga Bodas, Kawah Saat, thermal springs (x's), and Kawah Karaha), and geothermal wells (filled circles) at Karaha-Telaga Bodas. Contour lines show surface elevations (masl). Kawah Galunggung is the main vent of Galunggung Volcano.

Exploration at Karaha-Telaga Bodas has focused on a portion of the volcanic ridge between the Telaga Bodas and Kawah Karaha thermal areas. The fumarole field at Kawah Saat, a shallow acidic lake, (Telaga Bodas), and chloride-sulfate-bicarbonate springs occur at the southern end of the prospect, approximately 5 km from Kawah Galunggung. A smaller fumarole field, Kawah Karaha, occurs at the northern end of the prospect. Elevations along the ridge range between 1500 and 2000 masl and are ~1 km above the adjacent valleys.

Drilling to depths of 3 km indicates that the geothermal system consists of an extensive vapor-dominated regime and an underlying liquid-dominated resource (Fig. 2; Allis et al., 2000). Vapor-dominated conditions extend laterally for at least 10 km beneath the ridgeline and to depths below sea level. The deep drill holes demonstrate that this regime is continuous beneath the area between Telaga Bodas, where it is thickest, and Kawah Karaha. The underlying liquid resource is characterized by fluids with salinities of 1-2 weight percent and temperatures up to 350^{o}C (Powell et al., 2001).

The reservoir is developed mainly in a thick sequence of andesitic flows, pyroclastics, and breccias. Fine-grained lakebeds that formed intermittently in locally developed basins are interbedded with the volcanic rocks. Deposits from a depth of 988.8 m in T-8 have provided sufficient organic material for ^{14}C dating. The age of this deposit provides important constraints on the time-temperature-composition history of the geothermal system.

Heat may be supplied to the system by a cooling quartz diorite encountered in several of the wells at a depth of ~3 km. These intrusive rocks underlie much of the field. Although the quartz diorite was not encountered in the deep wells near Telaga Bodas, temperatures of 350^{o}C and a conspicuous circular gravity anomaly suggest that the intrusive rocks may be present at relatively shallow depths beneath the southern thermal area (Tripp et al., 2002).

HYDROTHERMAL ALTERATION

Hydrothermal mineral assemblages indicate that the present vapor-dominated regime evolved from a larger liquid-dominated system. Moore et al. (2000a) documented successive episodes of argillic alteration, silicification, and propylitic/potassic alteration within the reservoir rocks. In T-8, veins of chlorite + pyrite and later carbonate + hematite that cement hydrothermal breccias are found as shallow as 771 m (top of core hole). Propylitic assemblages characterized by epidote first occur at a depth of 830 m and at 885 m, actinolite joins the assemblage. At depths greater than ~1125 m in T-2 and T-8, potassic assemblages that also contain biotite, clinopyroxene, pyrite, and Cu-Fe sulfides are present.

The carbonate and silicate vein assemblages were followed by the deposition of chalcedony and quartz. Silica deposited as chalcedony (or amorphous silica)

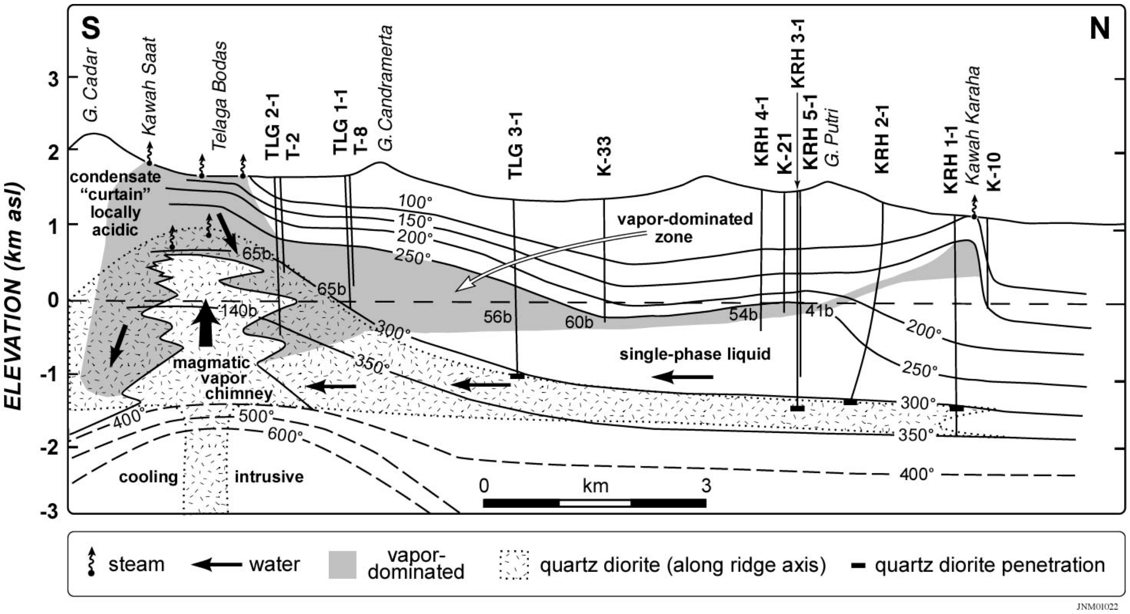

Fig. 2. North-south cross section through the geothermal system. Modified from Allis et al. (2000) and Tripp et al. (2002).

can be recognized by the presence of botryoidal textures (Fig. 3). Even though the chalcedony has been converted to quartz because of the high reservoir temperatures, the original botryoidal textures are frequently defined by small fluid inclusions.

At shallow depths, chalcedony replaces the carbonate veins and fills the remaining open spaces (Fig. 3a). In the deeper parts of the system, chalcedony encapsulates epidote and actinolite and forms the cores of quartz crystals (Figs. 3b, c). The deposition of chalcedony occurred over a large portion of the field. It has been found in all three core holes studied in detail (T-2, T-8, and K-33) to depths of at least of 1848 m.

Quartz overgrowths on chalcedony cores display a variety of unusual morphologies. Twinning, curved “c” axes, epitaxial growth and complex extinction patterns are common (Fig. 3d). Significantly, no evidence of a hiatus, marked by dissolution or mineral precipitation, has been observed between the early botryoidal forms of the chalcedony and the later quartz. Thus, our petrographic observations suggest that the deposition of chalcedony and quartz represents a single continuous event.

Quartz crystals from K-33 display evidence of dissolution (Fig. 4). This is an uncommon feature of geothermal systems, but similar textures have been reported from The Geysers. Moore et al. (2000b) argued that the descent of silica undersaturated steam condensate in a vapor-dominated regime was the most viable mechanism for producing fluids capable of quartz dissolution. They showed that quartz dissolution occurred near the base of the overlying cap rock and that both dissolution and deposition could occur at the same site in response to episodic boiling and condensate drainage.

The deposition of quartz was commonly followed by the successive appearance of veins dominated by: 1) chlorite; 2) anhydrite $\pm$ tourmaline $\pm$ quartz; 3) pyrite; 4) wairakite; 5) calcite; and 6) fluorite. In addition, native sulfur is found to depths of ~440 m in T-2, where it postdates stage 5 carbonate (dolomite). Chlorite and wairakite are found mainly in the deeper parts of the wells. Chlorite occurs below 680 m and wairakite is found primarily below 850 m. Anhydrite, calcite, and fluorite have retrograde solubilities and most veins containing these minerals are essentially monomineralic. These features suggest that the veins were deposited in response to the heating of descending steam condensate. Pyrite and chlorite can also precipitate from descending waters (Moore et al., 2002). The deposition of wairakite may reflect the effects of boiling and mixing between the reservoir fluids and steam condensate.

Chemical precipitates represent the youngest stage of hydrothermal activity. An SEM image of a Ti-rich scale on anhydrite was presented by Moore et al. (2000a). Scales consisting of NaCl, KCl, $FeCl_x$, and Ti-Si-Fe have been documented in core samples from T-2 and T-8 and X-ray and electron microprobe analyses confirm that halite is present. The occurrence of these scales demonstrates that parts of

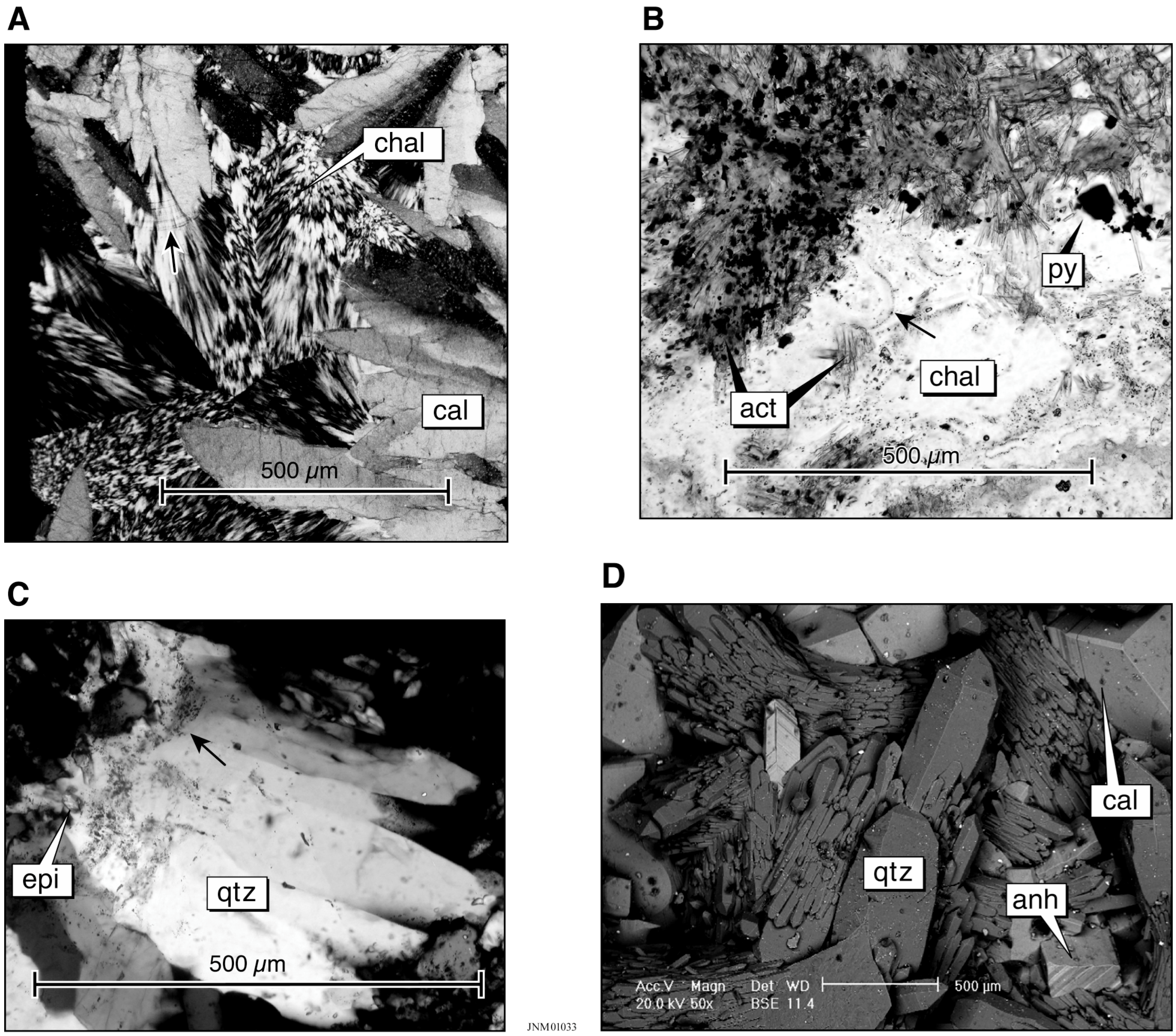

Fig. 3. Textures displayed by silica deposited as chalcedony and quartz. Arrows point to botryoidal textures. Abbreviations: act = actinolite; anhy = anhydrite; cal = calcite; chal = chalcedony; epi = epidote. A) Chalcedony filling open spaces in a carbonate vein from 802 m in T-8. B) Chalcedony after actinolite and epidote from 968.1 m in T-8. C) Quartz crystals with cores of chalcedony from 1203 m in T-8. D) Unusual growth forms of quartz containing cores of chalcedony from 793.5 m in T-2.

the geothermal system now hosting the vapor-dominated regime had dried out prior to drilling.

FLUID INCLUSION SYSTEMATICS

Fluid inclusions trapped in quartz, anhydrite, calcite, and fluorite were studied. The occurrence of daughter crystals and the high salinities of the liquids distinguish these inclusions from the more typical low salinity inclusion fluids characteristic of most volcanic-hosted geothermal systems. Temperature-depth relationships are shown in Figure 5; homogenization temperatures and salinities of the liquid-rich fluid inclusions are shown in Figure 6.

The most common inclusions in the quartz crystals are vapor-rich. In many samples, primary and secondary vapor-rich fluid inclusions dominate the inclusion population. Primary vapor-rich inclusions provide direct evidence of precipitation from boiling fluids. Multiphase liquid-rich inclusions are found in quartz crystals from 1102.2 m in T-8. Fluorite is the most common daughter mineral, occurring as small rounded grains that form octahedral crystals as the inclusions are heated and cooled. Although we have not yet identified the other solid phases present in the inclusions, their morphologies and optical properties indicate that halite and sylvite, which are commonly found up to several hundred meters from intrusions, are not present.

Fig. 4. Quartz with strongly corroded crystal faces from 1214.3 m in K-33. The complex extinction pattern suggests deposition from boiling fluids. The quartz is surrounded by later calcite.

Most of the quartz-hosted inclusions from T-2 and T-8 yielded homogenization temperatures of 235^o to ~350^{o}C. The minimum fluid inclusion temperatures, however, are lower than the measured downhole temperatures and the maximum temperatures recorded by later anhydrite and calcite.

These inclusions trapped fluids with an exceptionally broad range of salinities. The ice-melting temperatures of the fluids ranged from -0.1^{o}C (0.2 weight percent NaCl equivalent) to -27^{o}C (~24 weight percent NaCl-$CaCl_2$ equivalent). However, eutectics as low as -50^o to -60^{o}C indicate that divalent cations (e.g. Mg, Ca or Fe) are present and that the fluids were chemically complex. Extensive boiling and concentration of the fluids is the most likely explanation for the broad range of salinities recorded by these inclusions.

In contrast to T-2 and T-8, fluid inclusions from K-33 have lower salinities, ranging from 0 to 2.1 weight percent NaCl equivalent. These low salinity fluids may represent mixtures of condensate and the original reservoir fluids. Salinities of vapor-rich inclusions show a similar range from 0 (primary inclusions from T-8-1203 m) to 4.7 weight percent NaCl equivalent (secondary inclusions from T-8-1151.7 m). Nil salinity fluids are interpreted as condensate; higher salinities represent the accidental trapping of the boiling liquids.

Fluid inclusions in anhydrite, calcite and fluorite from T-2 and T-8 record the evolution of the descending steam condensate. Both vapor- and liquid-rich inclusions are common. Only the anhydrite-hosted inclusions from 1044.9 m in T-8 contained daughter crystals. Liquid-rich fluid inclusions from T-2 record temperatures similar to the present-day measurements whereas those from T-8 display a broader range of temperatures, with the lowest temperatures being closest to the downhole measured values (Fig. 5). Two distinct temperature-salinity trends are displayed by these inclusions (Fig. 6). At depths of less than 800 m, the apparent salinities of the inclusion fluids decrease as temperatures increase from 160^o to 205^{o}C and then remain approximately constant as temperatures increase further to 235^{o}C. This trend is the result of mineral precipitation from the sulfate-rich steam condensates (Moore et al., 2002). At greater depths, salinities increase with increasing temperature. Evidence of boiling is found in both anhydrite and calcite. Secondary vapor-rich inclusions are common in the anhydrite and primary coeval liquid- and vapor-rich inclusions occur in the calcite. The deepest sample of anhydrite trapped brines that precipitated daughter crystals of halite. These hypersaline fluids have an average homogenization temperature of 301^{o}C and salinities of 31 weight percent NaCl equivalent. The anhydrite crystals are coated with Ti-Si-Fe precipitates. The presence of these precipitates, the hypersaline brines and the vapor-rich inclusions indicate that the salinity increases were caused by the boiling off of the fluids as they percolated downward.

In summary, there are several important differences in the temperatures and salinities of fluids trapped in quartz compared to those in anhydrite, calcite, and fluorite. First, the minimum temperatures of fluid inclusions in quartz are lower than those in anhydrite and calcite from similar depths. Second, the minimum temperatures of the quartz-hosted inclusions are lower than the present-day measured temperatures but this is not the case for the other minerals. Finally, there is no evidence that the hypersaline fluids trapped in the deep anhydrite were also trapped in late quartz from nearby depths. These observations suggest that the inclusions in anhydrite, calcite, and fluorite record conditions that developed after the fluid inclusions in quartz were trapped. Thus the inclusions appear to document two cycles of heating and cooling; one represented by the inclusions in quartz and a second, younger cycle, represented by inclusions in anhydrite, calcite, and fluorite.

TEMPERATURE OF CHALCEDONY DEPOSITION

Extensive deposition of chalcedony in the high temperature portions of active geothermal systems has not previously been reported. Chalcedony is the stable silica polymorph only at temperatures less than ~180^{o}C (Fournier, 1985). At higher temperatures, quartz is the stable form. The temperatures of chalcedony deposition cannot be determined directly

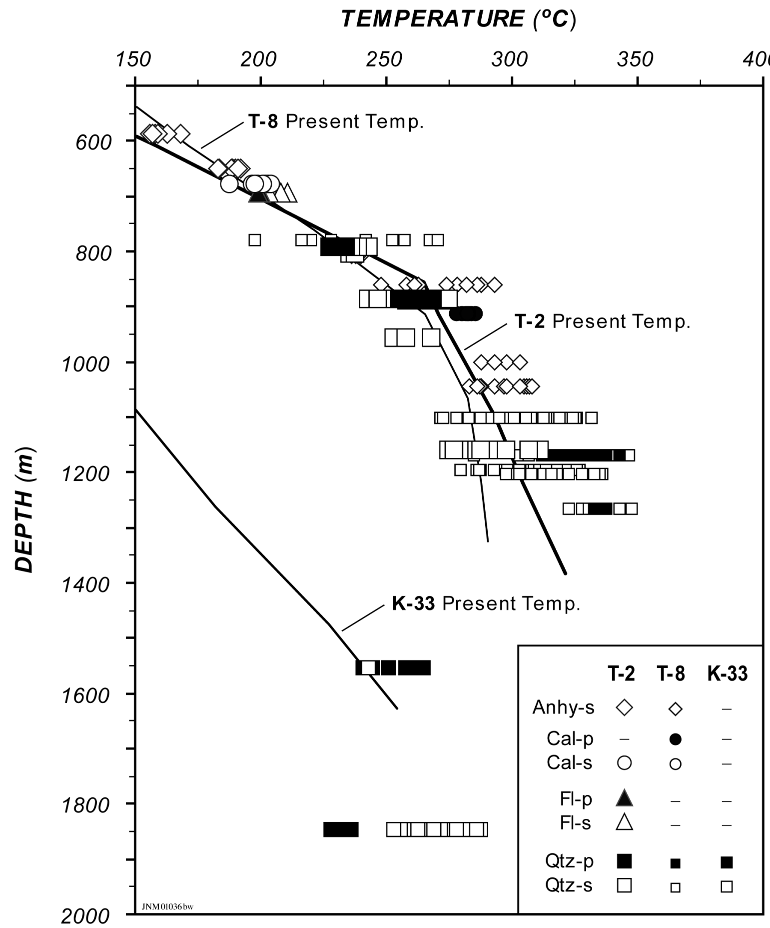

Fig. 5. Homogenization temperatures vs. depth of inclusions from T-2, T-8, and K-33. Downhole measured temperatures are shown for comparison. Abbreviations: anhy = anhydrite; cal calcite; fl = fluorite; p = primary; qtz = quartz; s = secondary.

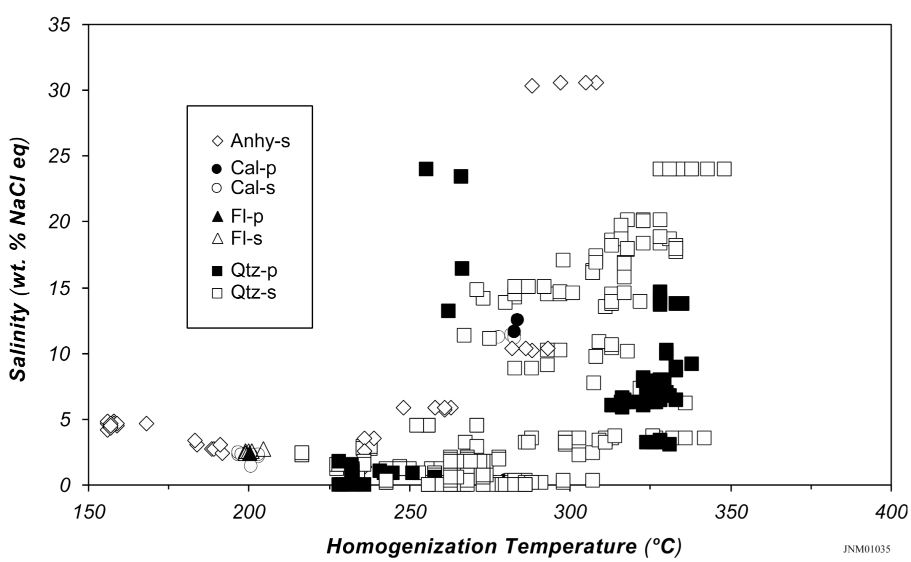

Fig. 6. Temperature-salinity relationships of individual fluid inclusions shown in Fig. 5. Abbreviations as in Fig. 5.

because the fluid inclusions they contain reflect its transition to quartz. However, the absence of any apparent hiatus in the deposition of quartz after chalcedony suggests that inclusions trapped in the quartz overgrowths can be used to estimate temperatures of chalcedony precipitation. Two samples were collected from veins where quartz overgrowths on chalcedony can be clearly demonstrated. Primary inclusions in quartz from 793.4 m in T-2 yielded homogenization temperatures of 228°-234°C (average 232°C; n=30) and an average salinity of 1.1 weight percent NaCl equivalent (n=19). Secondary and pseudosecondary inclusions yielded essentially the same average values and ranges (averages of 235°C and 1.4 weight percent NaCl equivalent). Secondary fluid inclusions in quartz from 1203 m in T-8 yielded temperatures of 298°-335°C (average 318°C; n = 73) and an average salinity of 18.4 weight percent NaCl equivalent (n=27). However, numerous primary vapor-rich inclusions are present in the sample from 1203 m, indicating that the fluid was boiling at the time of mineral deposition.

The mineralogic relationships support the conclusion that temperatures were high during chalcedony deposition. The presence of epidote + actinolite in the deeper samples indicates that peak temperatures exceeded 250°-300°C before deposition of the chalcedony and quartz (Henley and Ellis, 1983), while fluid inclusions in the quartz document similarly high temperatures during and after quartz deposition. Minerals that would indicate temperatures had declined to <~200°C prior to the deposition of chalcedony or later quartz such as chlorite (after actinolite) or wairakite are notably absent, although they are present later in the paragenesis. As shown by Fournier (1985), the deposition of chalcedony at temperatures much greater than 200°C requires extreme supersaturation of silica with respect to quartz. The most likely mechanism for producing these conditions is catastrophic decompression and boiling.

AGE OF CHALCEDONY DEPOSITION

The paragenetic relationships demonstrate that the formation of chalcedony represents a unique event in the evolution of the Karaha-Telaga Bodas geothermal system. We have constrained the timing of chalcedony deposition by ^{14}C dating of the lakebeds encountered at a depth of 988.8 m in T-8. The overlying volcanic rocks have been altered to actinolite, epidote, and later chalcedony and quartz. Chalcedony and quartz occur at depths at least as shallow as 771 m, epidote occurs in rocks deeper than 830 m, and actinolite is present at depths greater than 885 m in this well. These minerals were followed by the deposition of anhydrite, pyrite and calcite.

The organic matter in the lakebeds was dated by accelerated mass spectrometry (AMS). Material suitable for dating consisted of dispersed cellular organic matter with a few rare polypodiaceous fern spores that were oxidized and not identifiable as to species. No pollen was recognized. Because so few spores were present, the bulk organic material, including charcoal, was used for radiocarbon dating. The sample was crushed and treated with: 1) hot dilute HCl to remove carbonates; 2) room temperature HF to remove silica, silicates and metal oxides; 3) hot concentrated and dilute HCl to remove fluorides; 4) HNO_3 to convert decomposed organic material into alkali-soluble humic acids; and 5) KOH to eliminate contamination by younger humic acids. The remaining sample was sieved, combusted to CO_2, reduced to graphite, and dated. Although the 75.28 g sample contained only 0.03 weight percent dateable organic residue, it yielded sufficient graphite (70 micrograms) to date and produce a finite result. A $\delta^{13}C$ value of -23.7 per mil obtained directly on the CO_2 indicates that the material is typical of terrestrial organic matter.

The sample yielded a conventional radiocarbon age of 5910 +/- 76 years BP. This date is younger than expected. It indicates that significant volcano building, shallow carbonate veining, high-temperature hydrothermal alteration and the development of the vapor-dominated regime occurred since the lakebeds were buried.

Several different mechanisms could have produced the catastrophic decompression and boiling that affected a large portion of the geothermal reservoir and led to the deposition of chalcedony. However, no massive volcanic eruptions are known to have occurred during the last 6000 years and it is clear from the paragenetic record that the individual volcanic eruptions leading to the formation of Galunggung Volcano were of insufficient strength to depressurize the reservoir. Alternatively, depressurization could have been related to the formation of the large horseshoe-shaped crater at Kawah Galunggung that is thought to have formed ~4200 years ago in response to massive slope failure (Katili and Sudradjat, 1984). The crater has an estimated volume of 1,736,948,000 m^3 (G. Nash, written comm. 2001).

PRESSURE DISTRIBUTIONS

Additional information about the hydrothermal changes that have occurred at Karaha-Telaga Bodas system can be gleaned from the pressure trends derived from the fluid inclusion homogenization temperatures and salinities (Fig. 7). These pressures are minimum values because they assume trapping under two-phase conditions. The common occurrence of vapor-rich inclusions in most samples suggests that this is a reasonable assumption. It is also assumed that the fluids are gas free. At the high salinities of most inclusions, dissolved gases will not significantly contribute to the freezing point depression. At low salinities, the true salinity may be overestimated, resulting in inferred trapping pressures that are too low. Given the variations in the homogenization temperatures, we believe the uncertainty in pressure is ~± 5 bars.

Quartz-, anhydrite-, calcite-, and fluorite-hosted inclusions from wells T-2 and T-8 form a linear trend that extends from ~1100 masl and 5 bars (absolute), to ~120 bars at 400 masl. This is the maximum depth of T-2. It is not possible to distinguish between the trend based on inclusions in quartz from the trend defined by later anhydrite-, calcite-, and fluorite-hosted inclusions. However, the maximum anhydrite pressure is 77 bars, whereas the quartz pressure trend ranges from 32 to 118 bars. When the inclusion pressure trends are compared to present-day trends based on well data (Allis et al., 2000), it appears that the trend recorded by inclusions in anhydrite, calcite, and fluorite reflects present-day conditions in the condensate zone at shallow depths near T-2 and T-8. There is no shallow well data in this area, so conditions have to be inferred from the hydrology and chemistry of the hydrothermal features. Telaga Bodas is assumed to be perched and fed by rising steam and gas from about 1100 masl. The occurrence of advanced argillic alteration assemblages in T-2 help explain why the lake is hydrologically perched and sealed at its base. We presume that while steam and gas is flowing up beneath the Telaga Bodas thermal area, condensate and meteoric water is flowing downwards to the underlying steam reservoir. The high-pressure gradient between 1100 masl and the top of the steam zone at 700 masl indicates poor vertical permeability, and in effect, a capping structure.

The quartz inclusion data define a gradient of ~20 bars/100 m. This gradient is at least double a hot hydrostatic gradient, and is between 60 and 75% of the lithostatic gradient. At the time the quartz was

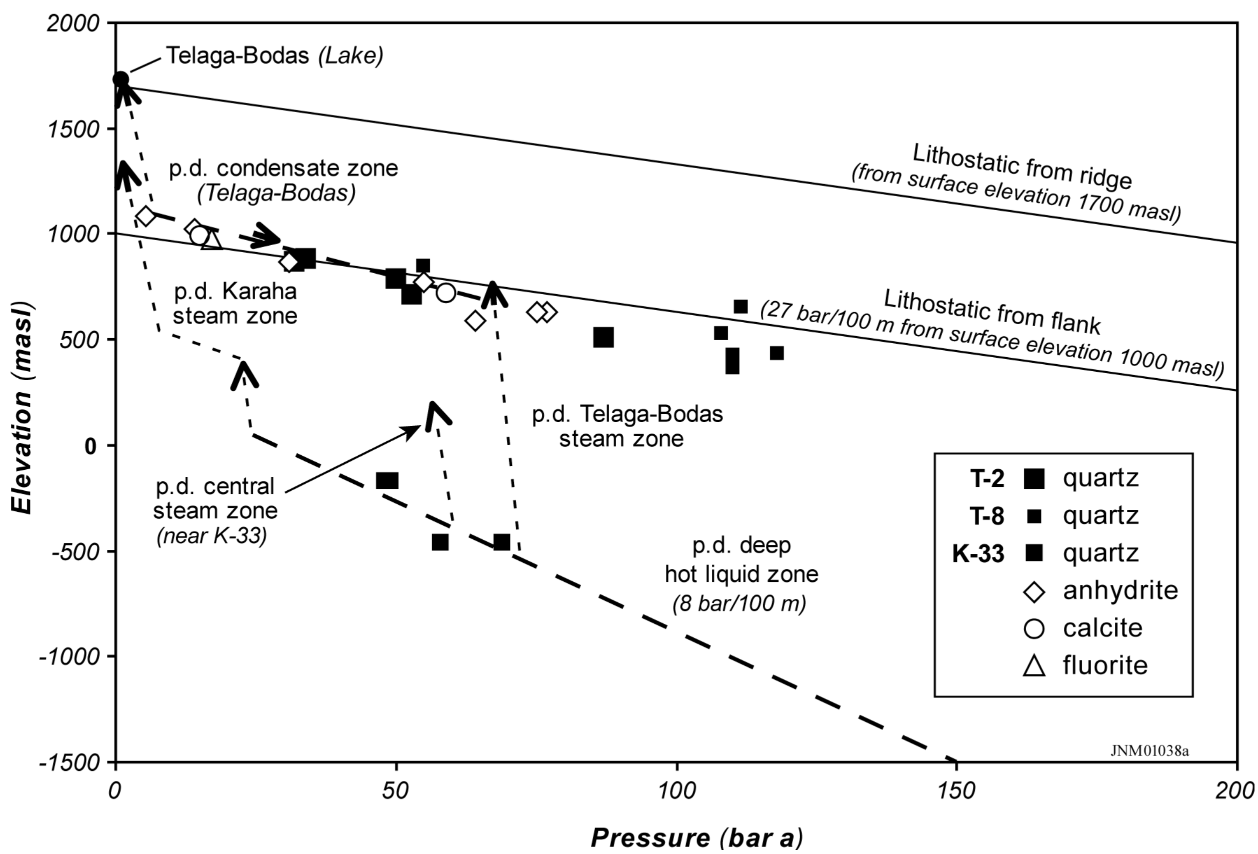

Fig. 7. Pressure trends based on fluid inclusion homogenization temperatures and downhole measurements. Modified from Allis et al. (2000). Only homogenization temperatures of primary inclusions or the highest temperature secondary inclusions were used to determine pressures.

being precipitated, relatively high fluid pressures were present at depth beneath the volcanic complex. The temperature gradient of 200°C/km defined by these inclusions is similarly high and suggests that the Telaga Bodas area lies close to a magma body.

Although the quartz postdates the chalcedony, we believe the quartz-hosted inclusions in T-2 and T-8 provide a close approximation of the conditions at the start of the transition to a vapor-dominated regime. This conclusion is based on the observation that a relatively small pressure drop (e.g. a few bars) can cause significant dry out of a low porosity rock (Grant et al., 1982) and therefore result in chalcedony deposition. In contrast to T-2 and T-8, pressures based on quartz-hosted inclusions from K-33 are close to existing conditions in the reservoir. Quartz in K-33, which occurs near the base of the steam zone, may have been precipitated recently.

The pressure gradient defined by the quartz-hosted inclusions from T-2 and T-8 could have pushed hot water to ~1800 masl if there was good permeability at shallow depths, although this is not indicated by our observations. Alternatively, these pressures could have produced hydrothermal eruptions on the flanks. At 1000 masl (surface elevation) the fluid pressure is almost the same as the lithostatic load and catastrophic slope failure is possible.

EVOLUTION OF THE GEOTHERMAL SYSTEM

Mineral assemblages, fluid inclusions and ^{14}C dating document the development of a very young vapor-dominated regime at Karaha-Telaga Bodas. The ^{14}C dating indicates that lakebeds encountered at a depth of 988.8 m were deposited 5900 years ago. These lakebeds and the overlying volcanic rocks were subsequently heated to temperatures in excess of 300°C. Epidote, indicative of temperatures >250°C, formed below 830 m while actinolite, which is stable only above ~300°C, was deposited at depths below 885 m. These assemblages document development of a liquid-dominated system that appears to have formed in response to the intrusion of quartz diorite emplaced at a depth of <3 km near the Telaga Bodas thermal area.

The transition from a high-temperature liquid-dominated system to vapor-dominated conditions is represented by the widespread deposition of chalcedony. Throughout most of the reservoir, chalcedony was deposited at temperatures in excess of 250°C. At these temperatures, extreme supersaturation of silica with respect to quartz is required. Massive slope failure of Galunggung Volcano ~4200 years ago (Katili and Sudradjat, 1984) could have caused catastrophic depressurization and the boiling that resulted in silica supersaturation and chalcedony deposition.

As the silica concentrations moderated, overgrowths of quartz were deposited on the chalcedony without any apparent hiatus. Continued boiling resulted in a progressive increase in the salinities of the residual fluids. Inclusions with salinities up to 24 weight percent $NaCl\text{-}CaCl_2$ equivalent were trapped in T-2 and T-8. Lower salinity fluids (0 to 2.1 weight percent NaCl equivalent) that may represent mixtures of the original reservoir fluids and steam condensate were trapped in quartz from K-33.

Steam condensate percolated downward as liquid levels and pressures within the reservoir declined. Interactions between the condensate and wall rocks produced advanced argillic alteration assemblages. Anhydrite, calcite, pyrite, fluorite, and chlorite were deposited in veins.

Homogenization temperatures of the fluid inclusions trapped in anhydrite and calcite from depths greater than 800 m suggest that temperatures increased again after deposition of the quartz. As the condensate percolated downward and boiled, its salinity increased. Hypersaline brines with salinities of 31 weight percent NaCl equivalent were trapped at 300°C in anhydrite from 1044.9 m in T-8. Continued boiling of the fluids resulted in the deposition of NaCl, KCl, $FeCl_x$, and Ti-Si-Fe scales. The presence of these precipitates demonstrates that the rocks had dried out prior to drilling. In T-8, complete dry out may have occurred shortly after trapping of the hypersaline brines at 300°C.

The waters encountered today beneath the vapor-dominated region in the deep production wells could not represent the residual liquids left after deposition of the chalcedony and quartz. The residual fluids would have had much higher salinities than the 1-2 weight percent of the present-day waters. These low salinity waters may represent mixtures of meteoric recharge and downward percolating condensate.

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