Proc. 8th NZ Geothermal Workshop, 1986

OXYGEN ISOTOPE FINE STRUCTURE AND FLUID THROUGHPUT OF THE TONGONAN GEOTHERMAL FIELD, PHILIPPINES - A PRELIMINARY REPORT

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ABSTRACT. Oxygen isotope ratios for 40 reservoir rocks from the plutonic basement and overlying andesitic rocks, and 14 separated geothermal quartz samples from the volcanics, range from 2.5 to **9.9** per mil. The lowest $\delta^{18}O$ values (average 2.9 per mil) in diorite cores from wells 401, 407 and 410 are located in the most productive northwest (Mahiao) sector of the field. In the Malitbog sector, the average $\delta^{18}O$ values for basement rocks are higher (c. 4.6 per mil). Plutonic rock samples from the Mamban (well MN1) sector, located outside the present-day field margin, are only slightly altered (6 per mil) except possibly near the contact zone between the basement and overlying volcanics. The highest cumulative fluid/rock ratios are calculated for the Mahiao sector, whereas Malitbog is possibly a relatively recent extension of the field. Relatively shallow (Bao Formation) quartz has $\delta^{18}O$ values suggesting past tectonic uplift.

1. INTRODUCTION

Oxygen isotope studies in active geothermal areas have provided important information about the origin and flow of fluids, the ratio of fluid to host rock, equilibrium and disequilibrium relationships between mineral assemblages, and the thermal regime within each geothermal system (Craig, 1963; Clayton et al., 1968; Clayton and Steiner, 1975; Norton and Taylor, 1979; Blattner, 1985). The purpose of this work is to explore the isotope fine structure of the Tongonan field and thereby provide evidence for the extent of high temperature meteoric water-rock interaction, and flow direction of meteoric recharge waters.

2. GEOLOGIC SETTING

The Tongonan Geothermal Field is located on the island of Leyte, Republic of the Philippines (Fig. 1) The reservoir comprises three principal rock types: a plutonic basement (Mahiao Plutonic Complex, MPC); a thick (ca. 2 km) volcanic sequence (Bao Volcanic Formation); and thinner (ca. 20 m) volcanic erosion deposits (North Central Leyte Formation, NCLF).

The basement is composed of diorites (c. 70%), quartz diorites (15%), aranodiorites (10%). and aranites (5%). These rocks come-from several plutons within the MPC which were intruded during the late Miocene (5-7 m.y. ago) and the Quaternary (<3 m.y. ago) periods, based on preliminary K/Ar dates from hornblendes (J.R. Hulston, pers. comm.). The older K/Ar dates are broadly compatible with ages determined from foraminifera in mid- to late-Miocene limestone and shale beds intercalated within the overlying hydrothermally altered andesitic breccias, tuffs and lavas of the Bao Formation. The preservation of igneous rock chemical trends, and semi-quantitative XRD and point count modal data suggest that the basement rocks are only slightly hydrothermally altered (<10% in terms of chemical composition; Scott, 1983).

The NCLF is composed of laharic, colluvial, and alluvial terrace deposits which formed during the Pliocene and Quaternary when rapid uplift occurred along the NW-SE trending Philippine Fault. Vertical fault movement of about 1 km also raised a Miocene diorite basement and permitted intrusion of granitic rocks (the inferred heat source) which appears to control the current position of the Tongonan Geothermal Field.

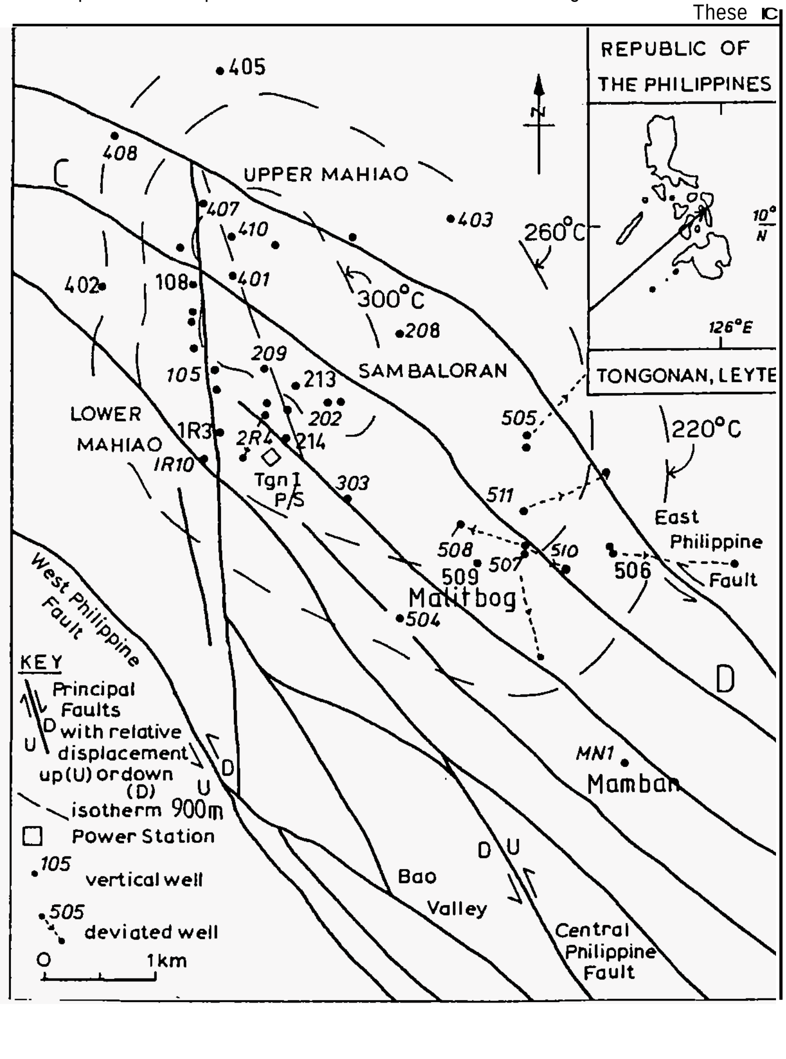


Fig. 1:

Sketch map of the Tongonan geothermal Field showing the principal faults (modified after unpublished airphoto interpretation by G.W. Grindley, NZGS), ipproximate location of isotherms at 300 m below sea level, and well locations.

3. METHODS, SAMPLING AND RESULTS

Twenty-eight total rock and 14 quartz samples were taken from cores and cuttings from 14 deep (ca. 2 km), exploratory wells. Apart from two surface samples, further whole rock data were obtained from cores at depths shown in Table 1. Five quartz and six feldspars were also analysed from pluton core samples (Table 1). The samples of geothermal quartz located within the Bao volcanics were picked from cuttings (Table 2). Oxygen was extracted from 10-15 mg quartz and whole rock samples by reaction with BrF_5, using a method modified after Clayton and Mayeda (1963). The liberated oxygen was converted to CO_2 by reaction with heated carbon. $^{18}O/^{16}O$ ratios for CO_2 were obtained using an NAA RMS-8 mass spectrometer. Oxygen isotope compositions are expressed as per mil deviations from the $^{18}O/^{16}O$ ratio of Standard Mean Ocean Water (SMOW) as:

$$‰\ \delta^{18}O = \left[\frac{(^{18}O/^{16}O)_{sample}}{(^{18}O/^{16}O)_{standard}} - 1\right] 10^3$$

As a reference, NBS-28 quartz with a normalised value of 9.6 per mil SMOW was used.

Table Ia: Diorites (s.l.) of the Mahiao Plutonic Complex. Depths rel. to sea level.

Sample	Well no.	Level (m)	Temp (°C)	$\delta^{18}O$ (‰)
Mahiao, Sambaloran				
Granite	105	-1390	310	4.3
Diorite	202	-1030	305	3.2
Diorite	208	-1143	310	3.0
Oiorite	209	-1472	320	3.2
Oiorite	209	-1873	330	2.9
Feldspar				2.9
Quartz				4.3
Diorite	2R4	-1188	275	3.3
Oiorite	303	-1266	260	3.3
Diorite	401	-1247	310	2.5
Diorite	403	-1254	310	4.2
Diorite	407	-1070	280	2.7
Diorite	408	-2056	310	4.7
Diorite	410	-1822	330	2.7
Malitbog				
Diorite	504	-1872	270	4.4
Diorite	507	-1540	220	4.0
Diorite	507	-1775	210	4.6
Granod.	507	-2427	230	4.0
Feldspar				3.9
Quartz				5.2
Granod.	508	-2155	260	4.7
Feldspar				3.8
Quartz				7.6
Granod.	510	-1822	280	5.2
Granod.	511	-1796	310	5.1
Mamban				
Diorite	MN1	-1686	195	6.5
Feldspar				7.0
Quartz				8.9
Oiorite	MN1	-1881	190	6.1
Feldspar				5.7
Quartz				8.9

4. DISCUSSION

4.1 Oxygen isotope shifts

The distribution of oxygen isotopes among minerals and water at equilibrium follows well known rules. Fractionations

$$\Delta_{phase1-phase2} = 10^3 \ln \left[\frac{(^{18}O/^{16}O)_{phase1}}{(^{18}O/^{16}O)_{phase2}}\right] \approx \delta_{phase1} - \delta_{phase2}$$

increase from zero at very high temperatures

Table Ib: Bao Formation andesites

Sample	Well no.	Level (m)	Temp (°C)	$\delta^{18}O$ (‰)
Andesite	504/509	+400	20	7.3
Andesite	214	+500	20	7.5
Tuff	1R3	-1467	65	4.0
Andesite	202	-296	260	3.5
Breccia	213	-147	240	6.1
Tuff	105	-212	230	5.6
Breccia	402	-1886	240	9.9
Andesite	405	-679	255	4.0
Andesite	405	-1602	235	3.6
Andesite	505	+49	110	6.8
Andesite	505	-830	215	3.7

typically to tens of per mils at room temperature. Ouartz is most enriched in ^{18}O, and usually minerals poor in SiO_2 also have low $\delta^{18}O$. Oxygen isotopes provide one of the most useful indicators of progressive water-rock interaction in geothermal systems (McKibbin and Absar. this volume; Blattner, 1985, and this volume). The meteoric waters of the Tongonan area have approximately $\delta^{18}O$ = -6.5 per mil and discharge waters from wells have -1 ± 1 per mil, i.e. could be meteoric waters shifted positively by about 5.5 per mil (Hulston et al.. 1982). This positive shift of the water (σ_W) requires an opposing negative oxygen isotope shift of rock (σ_R). For a continuous flow of meteoric water through a well mixed simplified reaction zone of a geothermal system, the reservoir rock would, over tens or hundreds of thousands of years, gradually be pulled towards an isotope composition in equilibrium with the meteoric input water. In the initial stages of this development the water throughput itself would suffer a much more visible isotope shift and this could be the obvious explanation of the +5.5 per mil shift at Tongonan. A corollary of that interpretation would be that the reservoir rocks would, correspondingly, show a relatively small oxygen isotope shift, but approach isotope equilibrium with the shifted meteoric water.

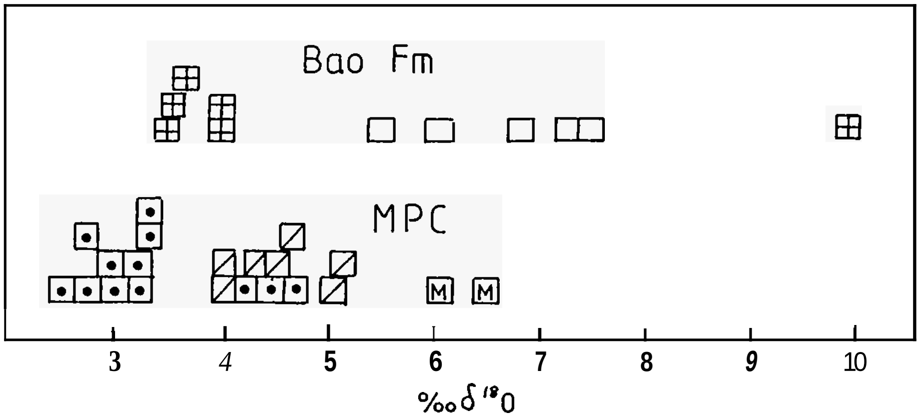
4.2 Original oxygen isotope composition of reservoir rocks

In order to determine by how much a reservoir rock has been isotopically altered, its original isotope composition must be known. Based on systematics from the literature, plutonic rocks of oceanic affinity could range from c. 6‰ $\delta^{18}O$ (at low silica values, e.g. gabbro) to about 7.5‰ (granites), with quartz as the most ^{18}O enriched mineral, reaching about 8.5‰ (Garlick, 1966). Of the analysed samples one diorite with 6.5‰, from well Mamban 1, shows almost concordant quartz and feldspar values and could represent an almost unaltered primary value (Table 1). In addition, weathered surface samples from Malitbog and Sambaloran have $\delta^{18}O$ values of only about +8 per mil, even though they may have undergone deuteric exchange with low temperature groundwater (<200°C). It will therefore be assumed that the original oxygen isotope composition for andesites and diorites ranged from 6 to 7 while the silica-rich plutonic rocks varied from 7 to 8 per mil $\delta^{18}O$.

Table 2: Geothermal quartz

Well no.	Depth (m)	$\delta^{18}O$(‰)	Temp °C
105	-250	9.4	250
105	-309	7.4	255
108	-430	7.4	275
208	-88	8.2	190
209	+215	5.9	150
209	-575	7.1	270
209	-1873	4.3	330
214	-212	10.1	260
214	-212	6.3	260
2R2	-947	6.7	270
407	-505	6.4	285
407	-484	7.7	285
407	-520	9.9	285
506	+72	8.4	80

Fig. 2: Distribution of $\delta^{18}O$ values for Mahiao Plutonic Complex (dots: Mahiao/Sambaloran, diagonals: Malitbog, M: Mamban), and Bao volcanics. For the Bao volcanics the deeper samples (crosses< -300 m) show compositions similar to the MPC (one exception), whereas the shallow samples (> -200 m) seem little affected by interaction with water (Levels rel. s.l.).



4.3 Present cycle alteration of rocks?

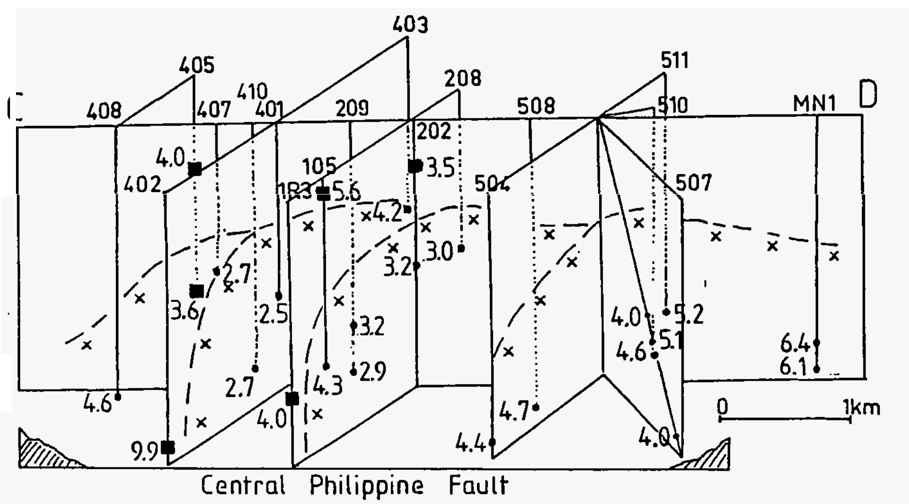
As shown in Tables Ia and Ib, the oxygen isotope compositions of altered rocks vary from 2.5 to 7.5 per mil with one value at 9.9 per mil. Excepting Mamban, the plutonic samples are confined to the range between 2.5 and 4.6 per mil in Mahiao and Sambaloran (ave.3.5) and between 4.0 and 5.2 per mil in Malitbog (ave. 4.6). The histogram of Fig. 2 shows the remarkably clear distinction between $\delta^{18}O$ values of various groups of rock and settings.

For the purposes of a first approximation, we shall later consider the Mahiao/Sambaloran sector to be the main zone of water-rock interation. Here the oxygen isotope shift from the estimated original c.7 down to 3.5 per mil gives σ_R = -3.5 per mil. Is it possible that this extensive isotopic alteration of the Mahiao reservoir rocks was induced by the present geothermal cycle? This would make sense, because the Mahiao sector is in fact the main producing area, but it would require the altered rocks to approach isotope after feldspar has been completely exchanged (Taylor and Forester, 1979). This leads to frequent discordances in geothermal systems, between $\delta^{18}O$ of primary unexchanged quartz and exchanged feldspar (e.g. Blattner, 1985, Fig. 6). The two primary quartz-feldspar pairs from Malitbog are excellent examples of that situation. The quartzes, along with those from Mamban, confirm initial unaltered diorite compositions near 7.5‰ $\delta^{18}O$. In contrast, the picked quartz from well 209 appears exchanged or recrystallised, but cannot be in equilibrium with the analysed coexisting feldspar, nor with water near -5‰ at less than about 400 C, and needs detailed confirmation.

4.4 Geothermal hydrology

Fig. 3 gives an overview of the oxygen isotope compositions of rocks. Regardless of the modelling used, the considerable isotope shift σ_R of many samples, especially the whole Mahiao sector, requires a considerable past throughput of meteoric water.

Fig. 3: Total rock $\delta^{18}O$ values set in three-dimensional view of Tongonan geothermal reservoir with Mahiao Plutonic Complex (crosses), overlain by Bao Volcanic Formation. Squares refer to Bao Fm. and dots to Mahiao Intrusives (whole rock analyses; section C-D of Fig. 1). Values decreasing froin Mamban toward NW, and lowest in Mahiao/Sambaloran. Top of diagram is sea level; no vertical exaggeration.



equilibrium with present day deep throughput water. A low 3/00 for rock with the approximate isotope properties of andesite feldspar would require water of c. -1‰ at 290°C or c. -0.5‰ at 310°C for equilibrium (cf. Taylor. 1979; Matsuhisa et al., 1979), which is sufficiently close to the measured $\delta^{18}O$ values of Hulston et al., especially in view of uncertainties in exact feldspar compositions, water sampling, and thermometer calibration. (However, if many Mahiao cores had overshot equilibrium by more than one per mil or so, then. a previous higher temperature or lower-$\delta^{18}O$ alteration event would have to be invoked.)

Relict primary quartz. Quartz is slow in exchanging oxygen when compared with feldspar, and, below 400°C at least, often preserves its original $\delta^{18}O$ value even

Box model. While a closed 'box' is obviously unrealistic for a geothermal system on account of porosity limits, an open or partially open 'mixed' box may have considerable relevance because most systems have pronounced high temperature reaction zones. Applying a perfectly open model to Mahiao and using $\sigma_R \simeq -3.5$ and $\sigma_W \simeq 5.5$ from above, we obtain the following cumulative water-to-rock ratio by volume

$$V_W/V_R \simeq 1.3 \ln\left(1 - \frac{\sigma_R}{\sigma_{wf}}\right) = 0.64$$

(Blattner, 1985). That is for every km³ of exchanged rock, c. 0.64 km³ of meteoric water would have passed through until now; an outer maximum could be calculated (for the closed box) at ~0.83 km³. This corresponds to a throughput rate of 10 l/sec per 10km^3 of altered reservoir for a period of about 20,000 years, or of 1l/s for 200,000 years. In any event, the cumulative discharge would put the Tongonan system less than half way to Wairakei in the evolutionary series of the open mixed box model.

Limits of system. Malitbog. The seemingly less exchanged $\delta^{18}O$ values of altered rock from the Malitbog sector could initially suggest that we approach the lower temperature and lower permeability margins of the main system. However, in that case the δ values of discharge water should be correspondingly somewhat higher, too, except for temperature correction. In fact, Hulston et al. find -2 to -3 per mil for the 500 series wells, plus low chloride values. This suggests near-surface admixture of short-circuited meteoric waters that have not taken part in the main cycle, and indeed concurs with the fact that production zones (from where the measured discharge waters must have come) are relatively shallow in this area. The lack of a corresponding increased down-shift of the altered rocks may indicate slower reaction rates at lower temperature, and, more importantly, relatively recent invasion of a former lower-permeability reaction zone by near-surface waters.

4.5 Bao Formation geothermal quartz

The Bao Formation contains numerous veins of secondary (geothermal) quartz and a random sample of these has been analysed (Table 2). The results are plotted in Fig. 4. Boiling point for depth (BPD) curves for the ^{18}O values for quartz are also plotted on Fig.4 . These curves are calculated using an average fiald pressure profile of 81 bars/100 m (Sarit, 1983), steam tables (Keenan et al., 1969), and the quartz-water fractionation equation of Matsuhisa et al. (1979). Temperatures at saturated vapour pressures are first converted to fractionation factors, then $\delta^{18}O$ values for quartz are calculated using these factors and two assumed $\delta^{18}O$ values (0 and -1 per mil) for present-day geothermal water, the band denoting BPD conditions for pure water. Less than half of the quartzes plot within or close to the BPD band, suggesting that up to 70 percent of the quartz is relict. The relict samples, which plot outside the BPD band, may also have precipitated from boiling fluids, but the water-table of these older fluids would have been higher (by up to ca. 1 km !?) than present,possibly because of tectonic uplift of the region in the Quaternary, or drops in the water table for other reasons. In some cases (wells 407, 214) different generations of quartz are present, but without evidence of relative ages. It is noted that all these quartzes are from relatively shallow levels,where fluids are subject to the isotope effects of evaporation, condensation and surface mixing more than at deeper levels.

5. SUMMARY AND CONCLUSIONS

In spite of the still limited amount of work to date, the Tongonan geothermal field is by now probably one of the isotopically better known fields of the Philippines and a number of conclusions can be drawn.

(1) A large fraction of the Mahiao Plutonic Complex to more than 1 km below the contact is geothermally altered and approaches oxygen isotope equilibrium with present-day fluids discharging from wells. The oxygen isotope shifts of the rocks, of c. -3.5‰ in the most productive part of the field are likely to be due to the present geothermal cycle . The total water throughput from a meteoric source can be calculated for Mahiao as about 0.7 km^3 STP for each km^3 of altered rock, much less than for example at Wairakei, but considerably more than at Ngawha in New Zealand.

(2) The slightly lower-temperature Malitbog sector shows higher $\delta^{18}O$ values for the rock and lower $\delta^{18}O$ for the water than Mahiao. This suggests cool groundwater admixture in a previously lower fluid throughput, slightly marginal zone of the field. One could ask whether the deepest Malitbog fluids might not in fact have enriched $\delta^{18}O$ in comparison to Mahiao, if sampling techniques for water could be improved.

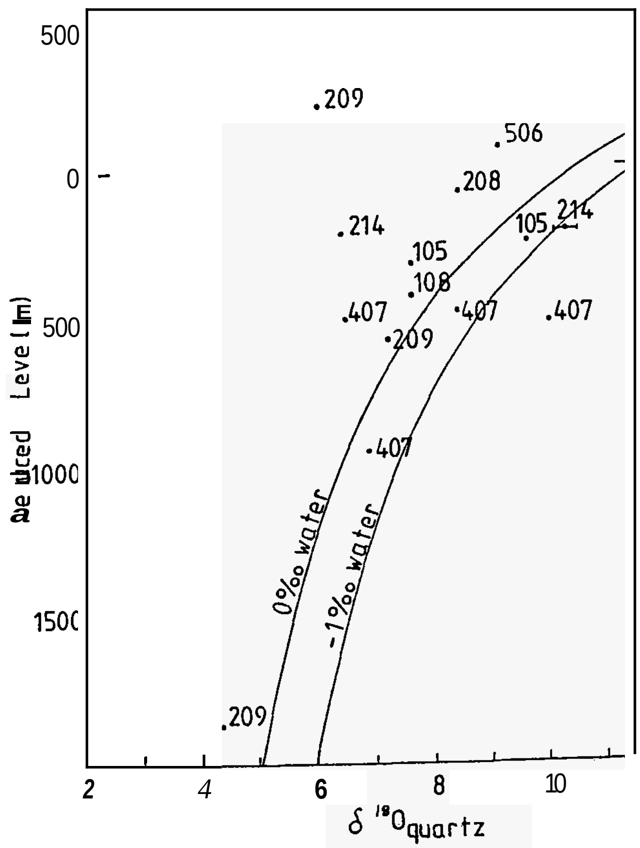


Fig. 4: Plot of reduced level versus $\delta^{18}O$ values for geothermal quartz in the Bao Formation. The solid curves represent present-day boiling point for depth curves for a reasonable range of present-day discharge waters.

(3) The main surface recharge area may lie close by in the hills north of Mahiao (Philippine Fault?), and is perhaps less likely to be found in the south east, since fluid would then have to come in below the cooler and almost unexchanged Mahanagdong and Mamban areas.

(4) About 253 of shallow geothermal quartz in the Bao Formation may be ascribed to deposition from boiling fluids during the present cycle. The remainder suggest a gradual drop of the water table during the geothermal activity, possibly due to tectonic uplift.

The large cumulative throughput in evidence for the Mahiao sector would inspire confidence that a relatively large and long-lasting source has been tapped.

ACKNOWLEDGEMENTS

G. Scott acknowledges the support, encouragement and continued interest of Professor P.R.L. Browne during his Ph.D. project, early reviews by P.R.L. Browne, M.P. Hochstein and A. Absar, and the financial assistance provided by the Auckland University Grants Committee. Both Scott and Blattner thank the Philippine National Oil Company (PNOC) and Mr B. Tolentino in particular, for assistance and core samples. S. Healy and C. Houston have provided part of the data and S.A. Stuart did the typesetting.

P.T.O.

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