ISOTOPE SHIFTING CAPACITY OF ROCK - NOTATION AND APPLICATION TO NGAWHA GEOTHERMAL SYSTEM

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ABSTRACT

Any oxygen isotope shifted rock volume exactly defines a past throughput of water. An expression is derived that relates the throughput of an open system to the isotope shift of reservoir rock and present-day output. The small isotope shift of Ngawha reservoir rock and the small, high $\delta^{18}O$ output are best accounted for by a magmatic water source.

INTRODUCTION

Rocks in general consist to about 50 weight percent of oxygen, and by its isotope ratio the oxygen of the mineral and fluid phases of a geothermal system constrains (1) the range of oxygen isotope compositions of minerals and fluids with which they may have interacted in the past, and (2) the temperature at which such interaction may have taken place. In fact, because of the omnipresence of oxygen, the natural tagging of waters ('meteoric water line'), and the large range of oxygen isotope compositions of natural rocks, oxygen isotope ratios of rocks cannot fail to provide a record of former interactions. A driving force for isotope exchange exists wherever a rock-water system is out of equilibrium, and the isotope record is quantitative, since the oxygen masses of interacting rock and fluid, and the respective changes in concentration of the trace isotope (so-called 'isotope shifts'), are related by mass balance.

NOTATION AND MASS BALANCE

The oxygen isotope composition of a sample s is usually given as a δ value with respect to a reference sample 'ref' (usually SMOW):

$$\delta^{18}O_{s\ vs\ ref}(^{o}/_{oo}) = \left[\frac{(^{18}O/^{16}O)_s}{(^{18}O/^{16}O)_{ref}} - 1\right] 10^3,$$

so that differences between δ values of different samples are close to proportional concentration differences of the trace amounts of ^{18}O; the natural abundance of ^{18}O is ca. 0.2 percent. The parameters of exchange between rock and a given batch of water can be set out as follows: The driving force or shifting potential P is defined as the initial 'distance' from equilibrium

$$P = \delta_{ri}-\delta_{wi}-\delta_{rf}+\delta_{wf}=\delta_{ri}-\delta_{wi}-f(T),$$

where r means rock, w water, i initial, and f final. The sensitive temperature function f(T) is the oxygen isotope equilibrium fractionation between water and a given solid,

$$\delta_{rf}-\delta_{wf} = f(T).$$

If σ are the respective isotope shifts $(\delta_f-\delta_i)$,

$$P = \sigma_w-\sigma_r.$$

With w and r equal to the oxygen masses of water and rock, the mass balance equation governing the equilibration is

$$w\sigma_w = -r\sigma_r.$$

The ratio w/r is the 'water-to-rock ratio'. In practice one may define the shifting capacity $-r\sigma_r$ of a rock to isotopically shift water as

$$C_w = v_w\sigma_w \simeq -1.3\ v_r\sigma_r, \qquad (1)$$

the v's being the volumes of rock and STP water (Units: permille-km3).

SPENT SHIFTING CAPACITY OF ROCK AS A MEASURE OF CUMULATIVE WATER THROUGHPUT

In many geothermal fields the rock altered by geothermal fluids is accessible enough to allow an estimate of its mass r and its isotope shift σ_r. The additional knowledge of the δ's of input and output water (and therefore of σ_w) then potentially allows to estimate the total water throughput by eq (1). However, in an open system with considerable water throughput, the rock gradually approaches equilibrium with input water, and the temporary σ_w, as well as the shift for the cumulative water throughput σ_W, decreases with time. Therefore, in determining total throughput W from a spent shifting capacity C_W and the present-day isotope shift for the water, σ_{wz}, one overestimates worse with increasing W/r and decreasing degree of mixing in the reservoir.

To gain control of this error source, an idealised 'mixed' system is considered, through which pass n discrete batches of water, so that W/r equals n times the w/r of a single batch (= closed system) situation (Blattner and Bunting 1980 [1]).

For such an open system

$$\sigma_r = P\left[\frac{1}{(1+W/rn)^n} - 1\right],$$

and if n = ∞,

$$\sigma = P\,(e^{-W/r} - 1),$$

an expression which may alternatively be given in terms of six δ values (Taylor 1979 [2]). Also from this approach, we have

$$\sigma_W/\sigma_{wz} = W/r\left[(1+W/rn)^n - 1\right]$$

and, substituting in the material balance equation (1) (using again v_W/v_r = 1.3 W/r), an expression for the desired total throughput of water by volume v_W, in terms of the present-day σ_{wz} instead of σ_W, which is not accessible to measurement

$$v_W = 1.3\ v_r\ n(\sqrt[n]{1+\sigma_r/\sigma_{wz}} - 1). \qquad (2)$$

Actual values are tabulated in Fig. 1, where the column for n=1 represents a closed system. Values in this column, when compared to those for larger n's, show the degree to which v_W would have been overestimated if σ_W had been assumed equal to σ_{wz}. Errors are less than ~20% for v_W/v_r <0.5.

HIGH $\delta^{18}O$ OF NGAWHA OUTPUT, AND POSSIBLE RECHARGE WATERS

The natural geothermal output at Ngawha is small and suffers from considerable dilution, so that surface sampling has allowed only an extrapolated estimate of the present reservoir output δ_{wz} (>5‰, Giggenbach and Lyon [3]). On the other hand, drillcore contains abundant calcite that has been deposited in the reservoir, below the cover sequence, by geothermal solution (Blattner and Bunting [1]). Since diffusional exchange is unlikely to be important for existing crystals of >0.1 mm [1] these may represent any stages in the history of the Ngawha system. Their tight grouping (Fig. 2) gives no hint of a significant change of δ_W during the stages actually represented, although the one outlier with respect to $\delta^{13}C$ shows that crystallisation did occur under changing conditions; an average δ_W of 7.5±1‰ may reasonably be assumed (for 270±20°C).

In the absence of hard information on a source of recharge water we face, thanks to the natural labelling of waters, a clear-cut choice between three recharge waters and respective σ_{wz} values (Table 1):

σ_r/σ_{wz} \ n	1	2	5	10	10^4	
0.1	.130	.127	.125	.125	.124	
0.2	.260	.248	.241	.240	.237	1.1
0.3	.390	.364	.350	.346	.341	
0.4	.520	.476	.45	.45	.44	
0.5	65	.584	.55	.54	.53	
1	1.3	1.08	.97	.93	.90	1.5
2	2.6	1.9	1.6	1.5	1.4	2
5	6.5	3.8	2.8	2.6	2.3	
10	13	6	4.0	3.5	3.1	10
100	130	24	9.9	7.6	6.0	

Fig. 1: $v_W/v_r = 1.3\ n(\sqrt[n]{1+\sigma_r/\sigma_{wz}} - 1)$. The shaded area is 'forbidden' for porosities <10 percent. Contours give the overestimate of v_W if σ_{wz} is used in eq (1).

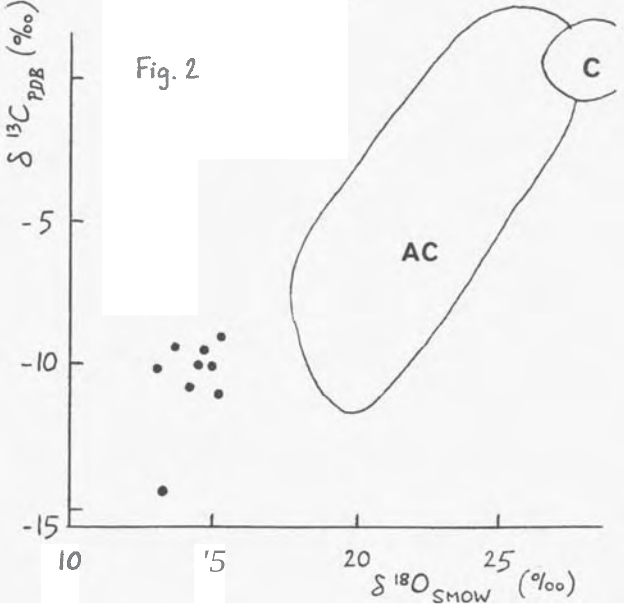

Fig. 2: Isotopic composition of secondary (geothermally precipitated) calcite from Ngawha basement. "C" marks field for cover rocks, "AC" for altered cover rocks. From Blattner and Bunting [1].

Table 1

	δ	σ_{wz}
(1) Fresh Northland surface water	-5	12.5
(2) Sea water	0	7.5
(3) Magmatic or metamorphic water	>5	<2.5

If we can estimate the volume v_r and isotope shift σ_r of the geothermally altered rock at Ngawha by analysis of minerals and rock, eq (2) will translate this choice into one between three different total water throughputs, v_W.

SURFACE RECHARGE OR MAGMATIC WATER SOURCE?

The rocks of the Ngawha reservoir consist of the altered part of a high-6^{18}O Tertiary cover sequence of ca. 0.5 km thickness, that acts as a low-permeability cap rock, and a potentially large volume of greywacke basement.

Cumulative throughput by spent capacity of cover sequence only

The normal marine δ^{18}O of about 7 km^3 of cover rock has been shifted negatively by an average of 5‰ in the geothermal area [1], so that the consumed C_W equals 1.3 x 35 = 45 permille-km^3. Physically, surface water could be visualised to enter a heated reservoir at the base of the cover, in which the water would be shifted mainly by cover rock. We then arrive, by eq (2), at the data of Table 2.

Table 2

Recharge source	σ_r/σ_{wz}	v_W (km^3) n=5	n=10	n>20	Age of system (years)*
(1) Northland surface water	.40	3.5	3.2	3.1	50,000
(2) Ocean water	.67	X	4.8	4.7	75,000
(3) Magmatic water	>2	X	X	>10	>160,000

*For a 'deep-water' output of 2 ℓ/sec [3]

It is seen that the volumes and deduced ages for options (1) and (2) are unrealistically low. Together with the fact that the 6 of the reservoir water does not rise above essentially an equilibrium value for the greywacke, and thus seems buffered by greywacke, this suggests substantial basement involvement.

System involving basement

To the total output v_W, and ages, of Table 2 we can add water shifted by basement, until a minimal reasonable output is reached. Because of the limited C_W of the altered cover rock, greater throughputs v_W of surface water must mean a negative shift of basement (in terms of the cover rock 'subsystem' this is similar to supplying a pre-shifted water, approaching option 3 of Table 2). By rock and mineral analyses, any such shift is smaller than about 1‰ [1], leading to $v_W/v_r \simeq 0.1$ for a meteoric water input. By mass balance between v_W of required throughput and C_W's of cover and basement, the following basement volumes have to be added to the previous, cover-only, subsystem, if the water source is to be meteoric:

Water throughput (km^3)	3	7	10	20
Basement involved (km^3)	0	>35	>65	>170

To follow through the idea of a meteoric water recharge, it is noticed that the low W/r ratio necessitated by the combination of the small used shifting capacity of cover rock and a negligible isotope shift of basement, requires either

- an almost absurdly short-lived, low-throughput near surface model for the Ngawha system, or
- vast dispersal of a modest recharge, through more than about 100 km^3 of basement.

CONCLUSION

The unusually high ^{18}O content of the Ngawha geothermal reservoir water, the required low water-to-rock ratio and considerable dispersal of the water, and the small geothermal discharge rate, all favor a magmatic source for the Ngawha hot water. Put differently, if recharge took place from the surface, so much water would potentially be available, that special reasons are required to explain the large dispersal and the fact that relevant parameters match requirements for a magmatic water source perfectly.

On gravity and petrological evidence, Cordon and Hochstein (1979) [4] and Heming (1979) [5] have argued for the presence of cooling magma at several km depth at Ngawha. Crystallising magma at that or greater depth could well be supplying the heat and fluids for the Ngawha geothermal system. The oxygen isotope evidence would alternatively also agree with a water source from underlying greywacke suffering regional metamorphism (cf. Blattner and Cooper [6], Paterson [7]). If Ngawha reservoir fluids were to be drawn off for power production, both a magmatic or metamorphic water source would severely limit possible recharge rates.

(P.T.O)

APPENDIX: OXYGEN ISOTOPE SHIFT OF WATER BY STEAM LOSS

Apart from isotope exchange with rock, boiling and steam loss might also cause a positive oxygen isotope shift of geothermal water. At 224°C boiling is not accompanied by a change in hydrogen isotope ratios D/H, so that a shift by interaction with rock may be simulated. While Craig [8] considered it improbable that steam loss would occur at this preferred temperature in different geothermal systems, it later appeared that the rather similar temperature of 236°C is the stable temperature for vapour dominated reservoirs (James [9], White et al. [10]). In addition, vapour rich reservoirs provide extensive opportunity for isotope exchange between ascending steam and descending condensate, and therefore for isotopic fractionation according to the Rayleigh equation

$$\delta_{residue} - \delta_{initial} = 10^{3} \left[f^{n(1-\alpha)} - 1 \right] ,$$

where f stands for the fraction of water remaining, α is the equilibrium isotope fractionation factor between water and vapour, and n equals the number of theoretical plates (e.g., Craig, 1969 [11]).

Although in nature n will usually be small, it is possible that in some vapour dominated geothermal systems, oxygen isotope shifts of recharge water at constant D/H may be due to this cause.

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