LECTURES ON GEOTHERMAL GEOLOGY AND PETROLOGY

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PREFACE

The material in this report covers four of five lectures that I gave during a wonderful visit to Iceland in the summer of 1983.

These lectures were prepared for an audience mainly comprising United Nations University geothermal fellows and members of the staff of Orkustofnun. The purpose of the first part of this report is obvious, but the second, however, requires a word of explanation. This section is based upon a paper which I published in 1978. While I include much new material here, shortage of time and space has meant that I have had to omit a great deal also; this includes, for example, the detailed and pioneering studies of surface and subsurface alteration by Japanese geologists. Nor have I discussed much about Icelandic geothermal systems - to do so would be as useful as sending your lamb to New Zealand.

Many people helped to make our visit to Iceland an enjoyable and unforgettable one, I am grateful to them, and especially Hrefna Kristmannsdottir, Jens Tomasson, Hjalti Franzson, Mary and Peter Linton, Stefan Arnorsson and Gudmundur Omar Fridleifsson. It is a pleasure to acknowledge also the help of Sigurjon Asbjornsson and the Director of the UNU Geothermal Training Programme, Ingvar B. Fridleifsson (and his family); their many kindnesses made our stay a very rewarding one and a delight. I thank them.

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1 INTRODUCTION

These lectures are intended to;

a) review briefly the current state of geothermal geology and petrology in New Zealand,

b) describe the role of petrological studies in geothermal investigations and how hydrothermal alteration mineralogy can assist in developing a four dimensional model of a geothermal reservoir.

Although these lectures are geological in content I emphasize at the start that it is a mistake to isolate geology from other earthsciences and, indeed, many engineering disciplines. Our experience shows that the best results in assessing a prospect, or helping to solve a particular geothermal problem, are achieved where more than lip service is paid to "cooperation and exchange of information"; this view is also that held here in Iceland. There is, for example, an increased "blurring" of the boundaries between petrology, fluid geochemistry and geology. Thus minerals are only naturally occurring chemical compounds whose stability and identity are controlled by kinetic and ambient physico-chemical conditions. Chemists can no more afford to ignore their presence than geologists can to treat them as names without chemical reality. This is not to say, however, that areas of responsibility during geothermal investigations need not remain distinct.

PART A

2 THE PRESENT NEW ZEALAND GEOTHERMAL SCENE

Drilling in New Zealand's geothermal areas is the responsibility of the Ministry of Works and Development (MWD) and the scientific services are undertaken by another government agency, the Department of Scientific and Industrial Research (DSIR), in particular, six of its divisions; the New Zealand Geological Survey, Chemistry, Geophysics, Applied Mathematics, Physics and Engineering Divisions and the Institute of Nuclear Sciences. DSIR geothermal work is co-ordinated by a small Head Office group presently headed by Dr. W.A.J. Mahon. The New Zealand Energy Department, incorporating the Electricity Division, is responsible for generating electricity and distributing it to local supply authorities.

Following the founding of the Geothermal Institute, geothermal research increased at Auckland University, within the Departments of Geology, Physics and the School of Engineering; this work has been greatly assisted by help, especially on an individual basis, from scientists and engineers with DSIR and MWD. Occasional geothermal based theses have been produced by research students at the Victoria University of Wellington.

Some private engineering and geological consultants also undertake geothermal work, the largest of which, Kingston, Reynolds, Thom and Allardice (KRTA) and Geothermal Energy New Zealand Ltd. (GENZL), operate mainly outside New Zealand.

As described by Freeston (1982), most geothermal investigations in New Zealand are carried out to produce electricity although, at present, this is achieved only at Wairakei (~175 MW) and Kawerau(~10 MW). The Ohaaki (Broadlands) power station, 20 km northeast of Wairakei, is scheduled for completion in 1986. No significant investigation work has taken place at Tauhara and Rotokawa since

Freeston's (1982) lectures, but a further well has been drilled at Kawerau. Drilling ceased at Ngawha in June with the completion there of a deep (2380 m) well (Ng13); this is expected to be a good producer with the highest temperature encountered being 287°C. Several other recent wells here are also good and it is likely that the field could produce 100-MW_e; since Ngawha is located in an energy-poor part of New Zealand I expect that drilling will resume here again (this is the third hiatus).

A recent bright spot on the geothermal scene has been the results of investigations at Mokai, about 20 km northwest of Wairakei; four wells have so far been drilled down to 1650 m and the GC-350 rig is scheduled to drill a deeper hole there shortly. At least two holes are good producers (~15 MW_e) and temperatures encountered exceed 295°C.

Non-electrical use of geothermal energy is still relatively minor but includes; lucerne drying (Broadlands), timber treatment (Broadlands, Waimangu, Kawerau), greenhouse cultivation (Wairakei, Rotorua), air conditioning (Rotorua) and domestic and light industrial heating (Rotorua). Since Freeston's (1982) lectures the Rotorua situation has deteriorated even further resulting in a great deal of controversy about the continued exploitation of this field and the effects of drawdown on geyser and spring activity at nearby Whakarewarewa. To date, the main features, including Pohutu Geyser remain unaffected but their early demise is predicted by some.

2.1 Geologists role during drilling

When a well is being drilled in a N.Z. field, a geologist at the closest branch (either Rotorua, one hours drive to all systems in the Taupo Volcanic Zone, or Auckland, 4 hours drive to Ngawha) of the NZ Geological Survey logs the well and keeps in close contact with the well-site engineer; the geologist also usually selects cores and cuttings for petrological and, if needed, micropaleontological or palynological, examination. Well sitting is rare as the geologist makes both regular visits and when requested to do so should a geological problem arise. The petrologist's report and other data are then used to prepare a final report for a well which then helps in interpretating the stratigraphy, structure, etc. of the field.

2.2 Cores and cuttings

Cores, up to 3 m long, are taken with a cheap core bit and used mainly by the geologist, petrologist and geophysicist. We regard coring as an essential part of geothermal investigations. Cores are mainly taken because they provide a sample (minus the fluid) of the reservoir from a known location. We have had poor success in taking side wall cores or in interpreting the results of downhole geophysical techniques.

2.2.1 Uses of cores

In New Zealand cores are used to;

a) Identify the reservoir rocks. This assists in determining the stratigraphy, structure and hydrology of a field. Rock identifications based upon cuttings are obviously much less reliable.

b) Examine the hydrothermal aleration produced by fluid/rock interaction (PART B); it is important to know, however, textural relationships, e.g. in vesicles, of the hydrothermal and primary minerals. These are much more difficult, or often impossible, to resolve using cuttings.

c) Analyze for major and trace elements and so learn about mass transfer processes taking place in a reservoir. The effort required to make accurate and precise analyses is only justified, in my opinion, on cores.

d) Make physical measurements and apply these results to interpret reservoir conditions. For example, we determine core porosity, dry, wet and particle density, magnetic susceptibility, thermal conductivity.

2.2.2 How often should cores be taken?

This is a difficult question to answer and we have no firm rules. Obviously the final decision is taken by the drilling or project engineer who, however, always listens to (even if he doesn't always follow the advise of) the project geologists and petrologist. Some of the factors considered in deciding an appropriate coring interval for a well are:

a) Cost. Coring is expensive and the deepest cores cost the most because of the time taken in tripping; where suitable, MWD take a core after a tricone bit has worn out thus saving one return trip.

b) The stage of the investigation. More cores are taken from a first well in a field than from a well in a field whose stratigraphy is well known. For example, the first wells at Broadlands provided 40-60 cores each but only 5-6 were recovered from recent wells there.

c) The stratigraphy of the field. Fewer cores are taken if it is known that a drillhole will encounter a fairly homogenous sequence, e.g. at Ngawha, and especially where the reservoir rocks are fine-grained.

d) To help answer a specific problem or for a special reason. For example, to recover samples from a newly encountered formation or to test the strength of a cap rock as happened at Tauhara where some wells were drilled with continuous cores.

e) Where circulation losses occur, or if cuttings are not being returned, then there is a case for taking more cores.

f) MWD drillers commonly take a bottom hole core.

Obviously these factors require a coring programme that allows for considerable flexibility.

2.2.3 Uses of cuttings

Cuttings are not a substitute for cores but in New Zealand we use them, together with the drillers' logs, to:

a) identify subsurface formations where this can be done,

b) locate formation boundaries,

c) recognise permeable zones,

d) provide material for fluid inclusion and stable isotope studies.

2.2.4 Problems with cuttings

a) Although we find cuttings very useful, they have limitations because the grain size of a formation's components may be larger than the cutting chips and it is often difficult to identify rocks from small samples. It is also harder to determine textures in amygdales and veins that have been damaged and broken in cuttings.

b) Cutting chips may not be representative of the rocks from which they derive. For example, the softer material may be ground away and only the harder constituents reach the surface; the cuttings collected over a stated interval comprise only a small proportion of the rocks actually ground up by the bit.

c) Considerable time may elapse between the time the cuttings are produced and when they reach the surface and are collected; thus some cuttings may derive from a shallower depth than that indicated. Obviously this is more of a problem for cuttings from greater depths.

d) Cavings from shallower in the well. Unlike cores one can never be really sure that the cuttings come from their indicated depth. Material from higher up the well often falls down and is returned to the surface with the deeper cuttings; in some cases this can be recognised, e.g. where the collapsing formation is significantly different. Usually caving chips derive from softer, granular or swelling rocks but occasionally material from hard, but fractured formations caves also. Caliper logs can be used to measure the diameter of the hole and thus the extent of caving estimated but this is usually done after the well has been completed.

3 SOME CHARACTERISTICS OF GEOTHERMAL SYSTEMS IN THE TAUPO VOLCANIC ZONE

All the known hot fields, except Ngawha, are located in the Taupo Volcanic Zone. (Figure 1) Some general characteristics of this zone and N.Z. fields are:

(a) The Taupo Volcanic Zone is a region of tension; it is 30-50km wide and about 250km long. Normal faults are common here and the region is still subsiding.

(b) About 15.000km3 of calcalkaline volcanic rocks have erupted here in the past 1.8 million years or so; these are mainly rhyolite pyroclastics and lavas with very minor amounts of dacite, andesite and basalt. The last eruption occurred in 1886.

(c) The 'basement' rocks of the region are Mesozoic greywackes and argillites; these have been penetrated by drillholes only at Broadlands and Kawerau.

(d) The geothermal systems are not usually associated with volcanoes.

(e) The extent of their surface activity and alteration varies greatly and the field boundaries are best located by resistivity surveys.

(f) Hydrothermal eruptions are common (two occurred in 1981), some as a result of drawdown, others due to local "overpressuring".

(g) The geothermal systems are dynamic; many have been destroyed by eruptions or become extinct. The oldest seem to occur in the western part of the Taupo Volcanic Zone.

(h) The deep fluids are of near neutral alkali chloride type with low salinity (upto 3000 ppm); only gas concentrations (CO_2, H_2S) vary significantly from field to field.

(i) Measured drillhole temperatures are between 230 and 300°C and generally follow the boiling point for depth curve.

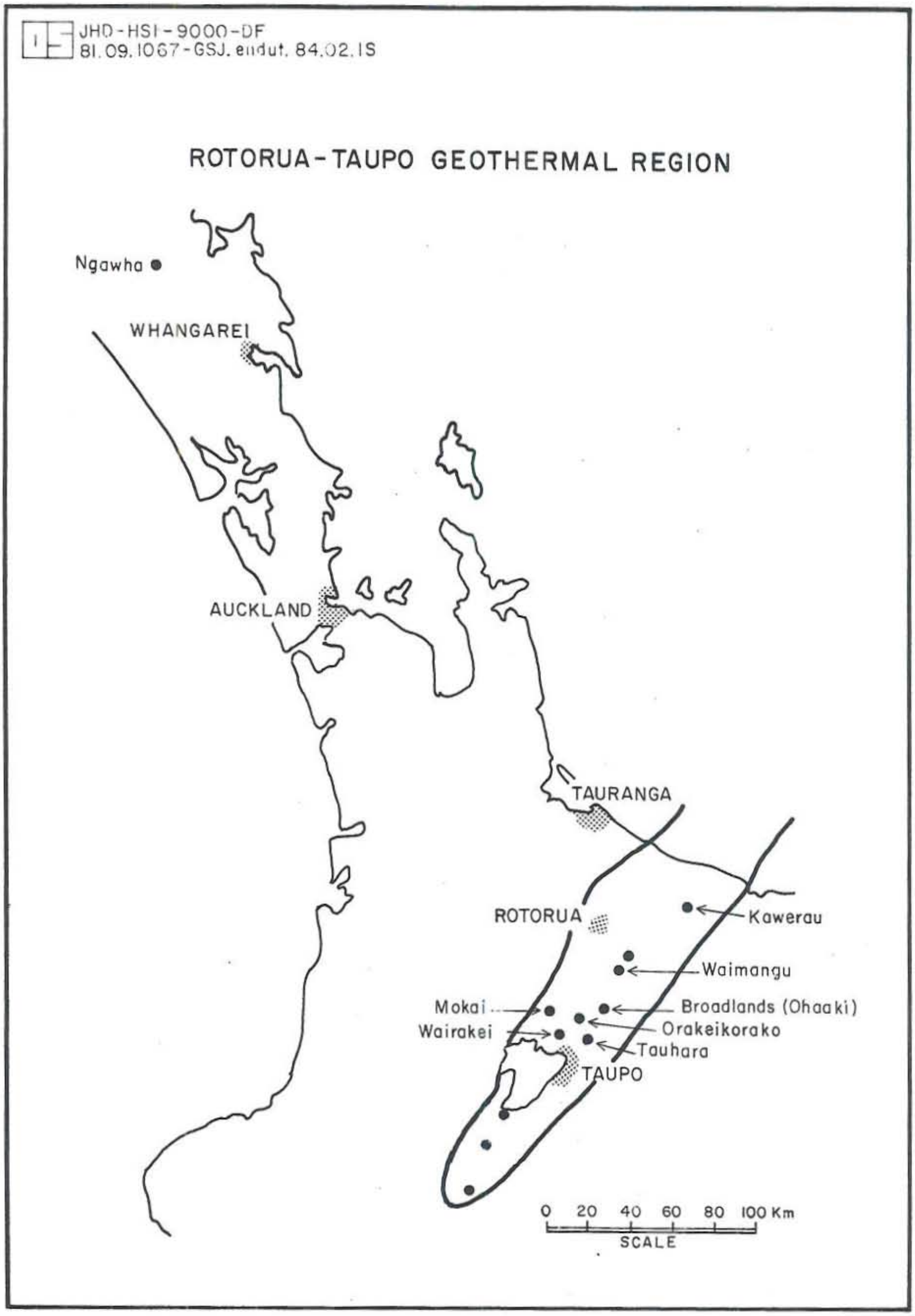

Fig. 1. Location of the thermal areas referred to in the text in the North Island, New Zealand.

4 SUMMARY NOTES ON RECENT GEOLOGICAL INVESTIGATIONS

4.1 Geothermal Inst., Geology Dept. Auckland University

Several M.Sc. and Ph.D. students have been, and are, undertaking research work on cores recovered from wells drilled several years ago. This includes: Hedenquist (Waiotapu), Cox (Ngawha), Christenson (Kawerau), Kakimoto (Tauhara) on New Zealand field plus Youngman (El-Tatio, Chile), Scott, Knox and Arevalo (Tongonan, Philippines).

4.2 Mokai, Taupo Volcanic Zone

The Mokai field is characterised by minor surface activity consisting of a few, aligned, steam-heated fumaroles, and springs and eroded explosion craters; these occur at higher elevation than a spring discharging near neutral alkali chloride water at 59°C. The field, which was defined by resistivity surveys, is located at the southwest margin of the Mokai Caldera but is cut by a northeast striking fault zone. The surface rocks include widespread superficial tephras (Taupo and Oruanui), the Mokai Ignimbrite and topographically prominent rhyolite lavas (LLoyd, 1978). Four drillholes penetrated these rocks and at least two other Ignimbrites, one of which probably correlates with the Whakamaru ignimbrite (~300.000 years old) now exposed outside the caldera (C.P. Wood, pers. comm.). The type of permeability encountered is uncertain but some production seems to be obtained from contacts. Hydrothermal alteration of cores and cuttings is extensive and this has produced a suite of secondary minerals including, laumontite, wairakite, yugawaralite, epidote, calcite, chorite, quartz, adularia, pyrite and possibly an amphibole (C.P. Wood, pers. comm.). Drilling has confirmed a model for the field whereby hot (~300°C), alkali chloride water, very similar in salinity and composition to that at Wairakei, ascends along the caldera margin until it reaches a depth and permeable horizon where it migrates northward, towards the Waikato River; this it does in response to the regional hydraulic gradient. As it moves, separated steam and gases ascend slowly, through the fairly impermeable Mokai Ignimbrite, condensing and mixing at the surface to produce the observed hydrothermal features. The Mokai prospect looks a good one and, at this stage, environmental problems seem unlikely to inhibit its development.

4.3 Ngawha, Northland

The extent of this field is still uncertain, (but it is probably large in area), despite there now being 15 holes there. Descriptions of scientific aspects of the field are available in the DSIR (1981) report, which is soon to be updated, and in several papers published in recent Proceedings of the New Zealand Geothermal Workshop. The field has a stratigraphy quite unlike that of geothermal fields located in the Taupo Volcanic Zone, in that the drilled section consists entirely of sedimentary rocks. However, young (Holocene) volcanic rocks (basalt, andesite and comendite) occur on the northern margin of the field and basalt south and west; the relationship between the Ngawha system and this volcanism still remains to be established. The top 500-700m of the field mostly consist of strongly deformed, sheared and brecciated claystones, siltstones, mudstones, argillaceous limestones and occasional sandstones; they are of late Tertiary to Cretaceous age but several age reversals occur. These rocks are regarded by almost everyone as cap rocks since they are impermeable to the extent that they are penetrated only by ascending steam and gas that separates from near neutral alkali chloride water in the reservoir (aquifer) rocks below. These consist of dense, Mesozoic argillites and greywackes of great but unknown thickness, that allow fluid to ascend and circulate through them by way of several joint systems, themselves probably related to faults. The hydrothermal alteration of the cap rocks is dominated by carbonates and clay minerals, the latter having pronounced variations in their thermal characteristics (Browne and Gardner, 1982). The reservoir rocks show widespread veining, that formed during several episodes, the most recent, of course, being the present cycle of geothermal activity; hydrothermal minerals present are similar to those which occur in the volcanic systems and include quartz, chlorite, illite, epidote, prehnite, pumpellyite, pyrite and pyrrhotite. Zeolites, apart from thomsonite, are absent and hydrothermal K-feldspar is rare. Calcite is abundant, however, specially above a depth of about 1000m where gas separation occurs. There are high gas flows at Ngawha and the waters are high in boron, ammonia and bicarbonate (about 1000, 50 and 400 ppm respectively; Sheppard and Lyon, 1981). The present low discharge rate (~ 1 litre/sec) of heated groundwater from the field was probably greatly exceeded in the past as is indicated by the presence of old sinters on the northern and southern margins of the field.

5 PERMEABILITY IN NEW ZEALAND GEOTHERMAL FIELDS

Many prospects and wells in New Zealand fail because of poor rock permeability and some effort is currently being made to understand the nature of permeability and fluid flow in our systems. The discussion that follows summarizes, and describes, some aspects of the subject from a geologist's view point.

5.1 Introduction

Geothermal fields which can be successfully exploited have the following qualities:

(a) they are able to store an appreciable amount of thermal energy in fluids and rocks of the reservoir i.e. heat.

(b) a significant pressure gradient and good permeability which ensures that hot fluids can move through the reservoir to be tapped by drillholes. Here permeability can be considered as the capacity of a rock and its adjacent surrounding for transmitting fluid.

In N.Z., fields which have good permeability are rarer than those which are hot but impermeable; for example, the Waiotapu Field has not been exploited, mainly because of its poor permeability despite subsurface temperatures in excess of 295°C. Similarly, poor permeability has restricted development of fields at Reporoa and Te Kopia.

On a local scale, more drillholes at Broadlands failed through encountering poorly permeable rocks than by not meeting high temperatures.

However, permeabilities in a geothermal field are very difficult to measure in situ and impossible to predict prior to drilling; permeability measurements made on cores in the laboratory are not a reliable guide to subsurface permeability. Some promise is offered, on the other hand, by the success achieved through downhole pumping tests and injecting radioactive and chemical tracers and monitoring their return in surrounding drillholes.

5.2 Primary Permeability

'Primary permeability' is used here to include permeable features whose formation predated the onset of geothermal activity.

All rocks, of course, are somewhat permeable but permeability as a physical parameter depends on the structure, shape, and occurrence of cavities and small and large scale fractures within a given volume of reservoir rock. Permeability is therefore not linearly related to porosity.

Porosity can be classified into several types; for example intergranular (mainly sediments, including tephra), joint (most rock types) and vesicular (or vug) type. The last variety is common in Iceland where basaltic rocks predominate but, by contrast is non-existent at Ngawha (NZ),

In the typical New Zealand geothermal reservoir, comprised of volcanic rocks, intergranular and vesicular porosity (5-25%) are significant whereas joint porosity is low (≈1%). Standard laboratory techniques do not allow for separate measurement of these types of porosity; if no intergranular porosity exists, as in the greywackes /argillite reservoir rocks at Ngawha, joint porosity (~3%) provides the permeability. However, here joints have also formed as a result of geothermal activity and thus both primary and secondary joint permeability occurs.

Other examples of primary permeability include bedding planes, unconformities and ancient solution features. Joints and faults, whose formation predated the onset of geothermal activity, should also be classed in this category as should dykes and fractures resulting from the intrusion of a magma not related to the present geothermal activity. This may be difficult to demonstrate, however.

New Zealand geothermal reservoirs are typically composed of rocks whose primary permeability varies because of differences in lithology; in a hydrological sense, these fields consist of an alternating sequence of aquifers and aquitards. The main aquifer at Wairakei consists of silicic pumice breccias (Waiora Formation) plus minor brecciated andesite overlain, and partly capped, by lacustrine sediments of the Huka Falls Formation. At Broadlands, two aquifers consisting of pyroclastic rocks (Upper Waiora and Rautawiri Breccia Formation) are effectively capped by rather impermeable rhyolite (Ohaaki) in the west and dacite (Broadlands) in the east. Impermeable rocks need not always be unwelcome since they reduce convective heat loss and form a barrier to descending cold meteoric water. Thus the Waikato River is prevented from swamping the Wairakei system.

5.3 Secondary Permeability

This broad grouping includes permeable features whose formation immediately predated, or is contemporaneous with, the present geothermal activity.

5.3.1 Faults

In New Zealand there has been a long debate about the reality and significance of faults in geothermal fields. Do they serve as permeable channels. However, it is surely no accident that most of the world's geothermal areas occur in places where active faulting is common. Drillholes at Wairakei, for example, were sited to intersect steeply dipping, normal faults at depths where temperatures are sufficiently hot to supply usable thermal fluid (Grindley, 1965).

One would expect that faults formed as a result of tension (normal faults) would be more permeable than those resulting from the forces of compression (reverse or thrust) and this seems to be the case since the latter type of faults are rare in active geothermal fields. Major transverse or lateral faults and their associated splinter faults occur in several fields (Tongonan, Philippines and Imperial Valley, California) where they have effectively broken up the reservoir rocks to generate good secondary permeability.

Obviously the age of a fault is important as one would expect the youngest faults to be the most permeable; at Ngawha, faults have been encountered by drillholes that are now too old, and hence sealed, to transport fluid. Wan and Hedenquist (1981), identified many young faults in several N.Z. geothermal fields and showed that the fields are located where dominantly northeast striking faults intersect those with a northwest orientation.

Possibly only segments of faults provide passage for thermal fluids, especially since there is some evidence (Grant, 1981), that the average horizontal permeability in many hot water fields is an order of magnitude greater than their vertical permeability. It must also be admitted that many mapped faults have only slight surface expression and that the displacement of subsurface units and formations through faulting is sometimes minor; however, exploration drilling in N.Z. based upon a model whereby faulting generates good channel permeability, has been remarkably successful. Further, measurements by Whitehead (1980) of natural radon discharge from Wairakei shows that higher rates of discharge occur in linear zones that coincide with the locations of faults.

5.3.2 Hydraulic Fracturing

Some reconciliation between the pro- and anti-fault model proponents is made possible by considering the nature and role of natural hydraulic fractures in places associated with normal faulting. An excellent paper describing the formation and characteristics of hydraulic fractures has been published by W.J. Phillips (1972). Grindley and Browne (1976) applied his work to active geothermal systems and noted that many wells drilled into active fault-zones at both Wairakei and Broadlands encoutered strongly silicified breccias adjacent to the open fissures. These breccias comprise angular fragments, mostly derived from the enclosing rocks, cemented by quartz, adularia, and usually pyrite; at Broadlands they are occasionally accompanied by calcite and base-metal sulfides. For a long time it was thought that these were silicified fault breccias; however, this interpretation is incorrect because normal faults do not produce thick breccia zones, due to lack of friction on the fault plane. Furthermore, similar breccias have been encountered directly below aquicludes where a direct fault association can be ruled out.

We believe that hydraulic fracturing produces these silicified breccias. Theoretical studies (such as that by Phillips) indicate that active normal faults in water-impregnated strata propagate upwards, either along the inclined plane of a fault or vertically, by natural hydraulic fractures following accumulation of hydrothermal fluids at the apices of "blind faults". Where the fluids are prevented from reaching the surface, either by self-sealing at the top of a fault or below an aquiclude,and pore fluid pressures exceed the least principal stress by an amount equal to the tensile strength of the rocks (usually 20 to 30 bars) fissures will propagate by natural hydraulic fracturing.

Laboratory experiments and numerous case histories of induced hydrofracturing in oilfields have shown that the rate of fracturing is extremely fast, about the speed of sound in fluid. Rapid extension of the fissures takes place in a vertical direction perpendicular to the least principal stress. At the top of geothermal aquifers, pore fluids pressures may increase sufficiently to fracture the rock repeatedly and produce numerous vertical fissures.

Hydraulic fracturing is most effective where the reservoir rocks are competent; they do not propagate as effectively through rocks such as unwelded tuffs or soft sediments with a tendency towards plastic deformation.

Rapid extension of fissures by hydraulic fracturing results in a sudden drop in fluid pressure, especially at the ends of fissures. The rate at which pressure restores is proportional to rock or fissure permeability and is inversely proportional to fluid viscosity and the square of the distance from a reservoir of fluid at constant pressure. If recharge is rapid, fluid pressures may be maintained above the minimum values needed for fracturing and some fissures may extend further. Usually, however, fluid pressures are not maintained by recharge, the rocks remain dilatant, and those adjacent to fissures fail explosively to form angular breccias within the fissure. The fragments of rock filling the fissures keep them open to serve as fluid channels; later movements may transport fragments within these fissures. Eventually, the clasts become cemented to the fissure walls by hydrothermal minerals; successive accumulations in long-active fissures may reach several meters in thickness as at Wairakei, Broadlands, and the fossil hydrothermal fields of the Coromandel Peninsula (N.Z.) (Grindley and Browne, 1976).

On occasions where hydraulic fractures reach the surface, hydrothermal eruptions occur and these events, we believe, are typical features of some active hydrothermal systems. Although their form is not usually preserved for long, prehistoric eruptions have occurred at Kawerau, Rotorua, Tikitere, Waiotapu, Waimangu, Waikite, Orakeikorako, Mokai, Rotokawa, Wairakei and Tauhara. Note however, that these eruptions have a different genesis from shallower focus eruptions that occur due to withdrawal of fluid from a reservoir.

5.3.3 Secondary Joints

Joints form readily in indurated rocks, such as welded tuffs, greywacke and basalt lava.

Although very narrow, where joints are present in large numbers, they are capable of transporting large quantites of fluid before they seal. I believe, for example, that fluids move through the dense greywacke and argillite rocks of the Ngawha reservoir by way of closely spaced (more than 15 per m^2), narrow (0.5 to 1.0 mm wide) discontinous joints orientated in at least three different directions (Browne, 1980).

One would expect joints to become fewer and narrower with increasing depth but this is not necessarily the case. For example, thermal stresses, proportional to the local temperature gradient,are likely to cause fracturing where gradients are high (~1°C/m); this should occur in the vicinity of heat sources if high heat transfer takes place through a thin shell (perhaps a hornfels zone). Another zone with high thermal gradients is a condensate layer which caps a 2-phase reservoir. During the life of a geothermal system it will undergo many changes in its temperature gradients and development of joints can be expected if the rocks are competent enough.

5.3.4 Hydrothermal Leaching

Because the concentration of dissolved mineral constituents in geothermal fluids is governed by water-rock equilibria, channels into the fields should widen due to dissolution of silica from the walls by heating recharge water, although calcite or anhydrite might precipitate. In some places, aquifer permeability may increase by fluids dissolving, but not replacing, primary minerals as in sandstones at Tatun, Taiwan, and also in parts of the Broadlands Field (e.g. BR28, 990 m) where primary feldspar has disappeared. Silica specimens, including quartz, lowered to the bottom of a Wairakei well for a week, lost weight by dissolution, but calcite was not appreciably dissolved (Ellis and Mahon, 1977). This experiment shows that in some active faults and fissures in upflow zones, hydrothermal solutions leach elements that are the mostly deposited at higher levels. However, over the long life of a hydrothermal system, water-rock equilibrium should be attained, at least close to flow channels. This would be less easily achieved, however, in fields like Tongonan and Kawah Kamojang, where the matrix of the andesitic lavas is usually sealed off, by early deposited vein minerals, from pervasive continued fluid penetration.

5.4 Recognition of reservoir permeability

Permeable zones in rocks penetrated by drillholes are mainly recognised by circulation losses during drilling, well completion tests, drawdown, and relative differences in downhole heating rates.

In addition, petrographic examination of ejected rock fragments can also help by matching lithologies and hydrothermal mineral assemblages with those in recovered cores and cuttings. Where channel permeability is present, ejected rocks usually derive from fissured walls and indicate production zones. In poorly permeable holes, substantial drawdown leads to caving of the producing formations; ejected rocks in this situation derive mainly from the weakest rocks.

Interpretations of hydrothermal mineral assemblages in cores (and cuttings) often provide information about subsurface temperature and permeability and such studies have proved useful in recognising permeable zones in geothermal fields in New Zealand and elsewhere (see PART B).

PART B

6 OCCURRENCE AND SIGNIFICANCE OF HYDROTHERMAL ALTERATION IN ACTIVE GEOTHERMAL SYSTEMS

6.1 Introduction

Fluids and reservoir rocks in geothermal systems commonly react together, with the result that the composition of both solid and fluid phases change; these changes can be viewed as straightforward chemical reactions, although in many cases temporary intermediate products must form. The identity and abundance of hydrothermal minerals produced during these fluid/rock interactions depends upon several factors, particularly temperature, fluid composition (especially its pH), availability of fluid (permeability), and whether or not boiling occurs.

6.2 The minerals

A wide range of hydrothermal minerals has been recognised in active geothermal fluids (Table 1); some are rare, such as buddingtonite, others, for example, aegirine and lepidolite are perhaps unexpected in geothermal environments. The more common ones are;

Carbonates
    Calcite, aragonite, siderite

Sulphates
    Anhydrite, alunite, natroalunite, barite

Sulphides
    Pyrite, pyrrhotite, marcasite, sphalerite, galena, chalcopyrite

Oxides
    Hematite, magnetite, leucoxene, diaspore

Table 1. Distribution of some hydrothermal minerals in active geothermal fields.

	Cerro Prieto, Mexico	Yellowstone, Wyoming	The Geysers, Calif.	Pauzhetsk, Kamchatka	Matsukawa, Japan	Otake, Japan	Philippines fields	Kawah Kamojang, Java	N.Z. Volcanic Zone	El Tatio, Chile	Low temp., Iceland	High temp., Iceland	Larderello, Italy	Olkaria, Kenya
Quartz	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Cristobalite		x		x	x	x	x	x	x	x				
Kaolin group	x	x	x	x	x	x	x	x	x	x				
Smectite	x	x		x	x	x	x	x	x	x	x	x		x
Interlayered illite/mont.	x		x	x	x	x	x	x	x		x	x		x
Illite	x	x	x	x	x	x	x	x	x	x			x	x
Biotite	x			x			x		x					x
Chlorite	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Celadonite		x	x	x						x	x			x
Alunite, Natroalunite			x	x	x	x	x		x					
Anhydrite				x	x	x	x	x	x	x		x	x	x
Sulphur			x	x	x		x		x					
Pyrophyllite					x	x	x	x						
Talc							x					x		
Diaspore					x	x	x	x						
Calcite	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Aragonite							x		x		x	x		
Siderite			x	x			x		x	x	x			
Ankerite	x			x			x							
Dolomite	x						x							
Analcime		x		x					x		x	x		
Wairakite	x		x	x		x	x	x	x			x	x	x
Erionite		x												
Laumontite		x		x	x	x	x		x	x	x	x		x
Phillipsite				x							x	x		
Scolecite				x							x	x		
Chabazite				x							x			
Thomsonite				x							x			
Epistilbite									x		x			
Heulandite		x		x		x	x		x		x	x		
Stilbite							x				x	x		
Mordenite		x		x					x		x	x		x
Prehnite	x		x	x			x		x	x		x		x
Amphibole	x			x			x		x			x		
Garnet	x		x	?			x							
Epidote	x		x	x		x	x	x	x	x		x	x	x
Clinozoisite							x	x	x	x				
Pumpellyite										x				
Sphene	x		x	x			x	x	x	x			x	x
Adularia	x	x	x	x		x	x	x	x	x			x	x
Albite	x			x		x	x	x	x			x	x	x
Leucoxene			x	x	x		x	x	x					
Magnetite							x					x		x
Hematite	x	x	x	x			x	x	x	x	x	x	x	x
Pyrite	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Pyrrhotite			x				x		x			x	x	
Marcastite							x		x			x		
Base-metal sulphides			x				x		x	x			x	
Fluorite		x					x	x						x
Wollastonite												x		
Hedenbergite												x		

References:

Cerro Prieto: Elders, et al., (1981); Reed, (1976); Yellowstone: Fenner, (1936), Honda and Muffler, (1970), Barger et al., (1973), Keith and Muffler, (1978), Keith et al., (1978); The Geysers: Steiner (1958), McNitt, (1964), Sternfeld, (1981), Pauzhetsk, Naboko, (1970), Matsukawa, Sumi, (1968), Otake, Hayashi, (1973); Philippines: Reyes and Tolentino, (1981); Kawah Kamojang: Browne, (1978); N.Z. Volcanic Zone: Browne, (1978), C.P. Wood (pers. comm.); El Tatio: Browne, (1978), K. Youngman (pers. comm.); Iceland: Sigvaldason, (1963), Tomasson and Kristmannsdottir, (1972), Kristmannsdottir and Tomasson (1974, 1976) Kristmannsdottir, (1981), Palmason et al., (1979); Larderello: Marinelli, (1969), Cavaretta et al., (1980).

Phosphate

Apatite

Halide

Fluorite

Silicates - Ortho- and Ring

Sphene, garnet, epidote

Silicates - chain

Tremolite, actinolite

Silicates - sheet

Illite, biotite, pyrophyllite, chlorite, kaolin group, montmorillonite, prehnite

Silicates - framework

Adularia, albite, quartz, cristobalite, mordenite, laumontite, wairakite

6.3 Application of petrology to geothermal investigations

Because hydrothermal minerals do form, their identity can be used to comment upon present and past conditions in geothermal reservoirs. Thus, for example, mineralogical estimates of subsurface temperatures, can become available shortly after samples are recovered and this information can be used to help in making decisions and planning without the long waiting period needed for a well to stabilize thermally. More important, perhaps, mineralogical deduced temperatures provide information on the thermal stability of a field. All that is required to recognise this is to compare the 'mineral temperature' with the stable well temperature.

There are also some minerals that can be related to subsurface permeability and an early indication of well permeability can often be obtained by petrographic examination of cores and cuttings. This could be used in deciding casing depth, for example, or learning whether or not, or perhaps where, well damage has occurred.

The close relationship between fluid composition and mineralogy has been recognised for many years. Mineralogical information can therefore be used to draw conclusions about the nature of the altering fluid e.g. is it high in dissolved CO_2, H_2S,? Is it acid? Is it high in fluorine? What will happen if we reinject spent fluid into the reservoir? Is the drillhole in a single or two phase zone? Where has boiling occurred? Is the well located in a recharge or discharge zone? Sometimes, a few of these questions may be answered.

6.4 Methods of study

Easily the most important tool needed for geothermal petrology is the petrographic microscope. Not only can it be used to identify many minerals but it is the only device that provides information, on a detailed scale, about important textural relations. It is also portable and cheap although some additional equipment is needed to make thin sections; even so, many minerals can be identified from their optical characteristics using crushed grain mounts and a set of refractive index liquids. The main disadvantage of the method is that petrographic training is needed to use a microscope effectively.

An X-Ray Diffractometer (XRD) is an extremely valuable instrument for identifying minerals and can be used, under some circumstances, to provide quantitative information.It is especially good for identifying clay and zeolite minerals but is expensive and scarcely portable.

Differential Thermal Analysis (DTA) has been used in geothermal work in New Zealand for many years (Browne and Gardner, 1981); it is capable of detecting subtle differences in the thermal characteristics of clays and is very sensitive to the presence of sulphides and carbonates. Several abundant and otherwise easily identified silicate minerals (e.g. feldspars) are nearly thermally inert so that little sample preparation, other than crushing, is needed.

Infrared spectrometry has also been used in New Zealand and is very useful for identifying clay, zeolite and feldspar minerals; it can be developed as a quantitive tool and requires little sample (appr. 10 mg).

In the past few years fluid inclusion geothermometry has become an important and powerful tool in geothermal petrology. A standard heating stage can be fitted to most microscopes and the homogenisation temperatures of inclusions in many minerals easily measured on double polished crystals. A freezing stage can be used to estimate the salinity of fluids trapped in these inclusions but this is often a slightly more difficult procedure, especially when it comes to interpreting results.

Other instruments that can be, and have been, used in geothermal petrology include an electron microprobe and a scanning electron microscope. Their importance will increase but, to date, they have been mainly used as research tools. Another group of instruments which I expect will be used more extensively in future are those that provide information about rock chemistry; in particular, X-ray Fluorescence and Atomic Absorption.

6.5 Intensity and rank of alteration

We distinguish between these two paramters. The intensity of alteration, I_a, is a measure of how completely a rock has reacted to produce new (i.e. hydrothermal) minerals. For example, a unit volume of rock which has not been affected by hydrothermal solutions has zero intensity of alteration ($I_a = 0.00$), whereas one in which all the primary phases have been replaced has an alteration intensity of 100%. (i.e. $I_a = 1.00$). Intensity of alteration can be estimated by pointcounting with a microscope or using a semiquantitative X-Ray Diffraction technique. Note that intensity has nothing to do with the identity of the new minerals formed, only their total abundance.

Alteration rank, on the other hand, depends upon the identity of the new minerals and is based upon their significance in terms of subsurface conditions, i.e. it is an empirical and more objective parameter than intensity and derives from macroscopic and microscopic examination assisted by X-Ray Diffraction, Differential Thermal Analysis, or some other instrumental technique. Adularia, for example, has a high rank when considering permeability, and epidote a high rank on any temperature scale. It is possible, therefore, to get rocks which have a mineralogy symptomatic of high rank but low intensity (in hot, impermeable zones) and others of low rank but high intensity, generally where cooler, permeable conditions prevail.

6.6 Types of hydrothermal alteration

6.6.1 Direct deposition

This type of alteration is very common, and most hydrothermal minerals found in geothermal fields can deposit directly from solution. To be able to do so requires, of course, that the reservoir rocks contain passages along which the depositing fluids can move. These could include features such as joints, faults, hydraulic fractures, unconformities, vugs, pores and fissures (see PART A).

Quartz, calcite and anhydrite readily form veins and fill vugs, but chlorite, illite, adularia, pyrite, pyrrhotite, hematite, wairakite, fluorite, laumontite, mordenite, prehnite and epidote have also been observed to occur in places where they could only have deposited directly from a fluid. Deposition in drillpipes and discharge channels also occurs; calcite, aragonite and silica commonly form this way but also, in places, (e.g. Krafla and Broadlands) iron bearing minerals can form scales (Arnorsson, 1981).

6.6.2 Replacement

Most rocks contain some primary minerals which are unstable in a geothermal environment, and these have a tendency to be replaced by new minerals that are stable, or at least metastable, under the new conditions (Tables 2 and 3).

The rate at which this replacement occurs is highly variable and depends upon permeability; for example, an active volcano in the Bay of Plenty, White Island, discharged andesite ash that was completely altered (I_a = 1.00) within one year of its deposition on the island through the action of volcanic gases. By contrast, portions of an impermeable, welded ignimbrite occurring within the (Broadlands) Ohaaki Field have been at a temperature of about 250°C for perhaps 300,000 years without being altered to an intensity of more than 0.1.

Where replacement is not complete (intensity is less than 1.00), i.e. fluid/mineral equilibrium is not achieved, reactions are, in effect, preserved or frozen in cores and able to be seen unete (intensity is less than 1.00), i.e. fluid/mineral equilibrium is not achieved, reactions are, in effect, preserved or frozen in cores and able to be seen under the microscope. In reservoir rocks which are volcanic it is usually fairly easy to distinguish between the primary and secondary (hydrothermal) minerals but this is more difficult where sediments or low-grade metamorphic rocks occur (e.g. The Geysers, Larderello, Kizildere, Imperial Valley, Cerro Prieto, Ngawha). The reason for this is that many of the primary minerals which these rocks contain (e.g. quartz, feldspars, calcite, prehnite, illite, epidote, etc.) are also stable in

TABLE 2. USUAL ORDER OF REPLACEMENT OF PRIMARY MINERALS IN SOME SYSTEMS (Browne, 1982)

Primary Mineral	N.Z. fields (TVZ)	Ngawha	Olkaria	Philippines & Indonesia	Iceland
volcanic glass	1	absent	1	1	1
magnetite) titanomagnetite) ilmenite)	2	1	5	2	
pyroxene) amphibole) olivine)	3	absent	2	3	2-3
biotite	4-5	2	absent	rare?	
Ca plagioclase	4-5	albite present	3	4	4
microcline) sanidine) orthoclase)	absent	3	4	absent	absent
quartz	not affected	not affected	not affected	absent	absent

geothermal environments where alkali chloride waters occur. In fact it is rather the minerals which control the composition of the waters, although not their salinity.

6.6.3 Leaching

This process takes place at, or indeed outside, the margins of geothermal fields, and for this reason is not generally seen in recovered cores and cuttings. It occurs, for example, where steam condensate, acidified by oxidation of H_2S, attacks rocks, dissolving primary minerals but without replacing the voids thus produced (see PART A).

6.6.4 Ejecta

Quartz, wairakite and other minerals are sometimes blasted from fissures penetrated by some drillholes when these are first opened; on occasions bladed, euhedral calcite that has no obvious place of attachment to a rock surface is also ejected. Perhaps these crystals form in the turbulence of a boiling zone (see Tulloch, 1982)?

TABLE 3. TYPICAL ALTERATION REPLACEMENT PRODUCTS (from Browne, 1982)

Original mineral	Replacement products
volcanic glass	zeolites (e.g. mordenite, laumontite), cristobalite, quartz, calcite, clays (e.g. montmorillonite)
magnetite/ilmenite/titano-magnetite	pyrite, leucoxene, sphene, pyrrhotite, hematite
pyroxene/amphibole/olivine/biotite	chlorite, illite, quartz, pyrite, calcite, anhydrite
calcic plagioclase	calcite, albite, adularia, wairakite, quartz, anhydrite, chlorite, illite, kaolin, montmorillonite, epidote
anorthoclase/sanidine/orthoclase	adularia

7 CHANGES TO THE HOST ROCK PRODUCED BY HYDROTHERMAL ALTERATION

Hydrothermal alteration often produces changes to the properties of rocks by altering their density (increase or decrease), porosity, permeability (decrease or increase), magnetic strength (usually decrease but occasionally an increase) and resistivity (decrease). Concurrently with these changes, of course, related and/or unrelated events may occur, such as faulting and formation of joints that affect the alteration process.

Replacement, leaching and deposition also causes chemical changes whose extent and nature vary greatly but which are obviously a function of the mineralogy; some replacement can proceed isochemically, at least on a local scale, but many studies show that some constituents are actually added or removed during alteration, even taking into account the effects of "element dilution" by density changes.

Table 4 describes the typical behaviour of the traditionally quoted major constituents during the hydrothermal alteration of volcanic reservoir rocks reacting with a thermal fluid. A similar table could also be constructed for many trace elements.

Table 4. Typical behaviour of major element oxides during hydrothermal alteration processes within a geothermal reservoir (discharge region). After Browne (1982).

Oxide	Typical behaviour	Hydrothermal minerals containing this oxide or constituent
SiO_2	added	quartz, cristobalite, silicates
TiO_2	unchanged?	sphene, leucoxene
Al_2O_3	added and removed	many silicates, some oxides
Fe_2O_3)	added and removed	chlorite, pyrite, pyrrhotite
FeO)		siderite, epidote, hematite
MnO	unchanged?	
MgO	removed	chlorite, biotite
CaO	added and removed	calcite, wairakite, epidote, prehnite, anhydrite, montmorillonite, sphene, fluorite, other zeolites, wollastonite
Na_2O	added and removed	albite
K_2O	added	adularia, illite, alunite, biotite
CO_2	added	calcite, siderite
S, SO_3	added	anhydrite, alunite, pyrite, pyrrhotite, barite
H_2O	added	all clays, epidote, prehnite, all zeolites, diaspore, pyrophyllite, amphiboles
P_2O_5	unchanged or added	apatite
Cl	removed	
F	added or unchanged	fluorite

8 FACTORS THAT AFFECT THE FORMATION OF HYDROTHERMAL MINERALS IN ACTIVE GEOTHERMAL FIELDS

The approach described here is an empirical one and is based upon experience gained from the petrological study of cores and cuttings recovered from drillholes, supplemented by results of hydrothermal experiments, elementary thermodynamic considerations and observations made on fossil geothermal fields, such as those associated with hydrothermal ore deposits. Because we use an essentiallly practical approach some observations and comments are made without substantiation; reasons are still being sought to explain many observations.

The factors which can affect the formation of hydrothermal minerals are (Browne, 1978);

(a) temperature; (b) pressure - mainly because it controls the depths at which boiling occurs; (c) parent rock type; (d) reservoir permeability; (e) fluid composition; (f) duration of activity

Of these, the most useful information which a petrologist can supply are mineral deduced estimates of subsurface temperatures and permeability and some comments on any temporal changes that may have occurred within the reservoir (Browne, 1970; Elders, 1977). The problem is that some of the factors listed above are so intimately interconnected that, at first sight, it is difficult to separate one from another. However, I will attempt to do so.

9 HYDROTHERMAL ALTERATION AND TEMPERATURE

With a few exceptions the minerals that provide information on their formation temperatures are these that contain in their structure, either (OH) or n·H_2O; these include some clays, zeolites, prehnite and amphiboles. In addition, fluid inclusion geothermometry, also provides information on mineral formation temperatures.

9.1 Clay minerals

These can be useful despite the difficulty of identifying them precisely.

9.1.1 Kaolin group

Where acidic waters interact or deposit kaolin there may be a change from kaolinite and halloysite, which occur below about 120°C, to dickite; this is itself replaced by pyrophyllite at temperatures above about 250°C. However, this sequence is only tentative because there are very few (fortunately) fields in which hot acid fluids now circulate.

9.1.2 Chlorite group

Chlorite minerals are very common and widespread in active geothermal fields (Table 1); they have a great range in composition, however, which is not obviously related to temperature and certainly not in the New Zealand geothermal fields. However, Kristmannsdottir (1977) has shown that smectites from Reykjanes, Iceland are present as a discrete phase where temperatures are below 200°C; they become randomly interlayered with chlorite where temperatures are between 200°C and 270°C, but above 270°C non-swelling chorite is the only clay mineral present.

9.1.3 Montmorillonite, illite and interlayered illite-montmorillonite

These clays, plus chlorite, are the common clay minerals found in the New Zealand geothermal fields. Montmorillonite is stable to about 140°C and illite above 220°C; in the interval between interlayered illite/montmorillonite commonly occurs, the proportions of the end members changing with temperature (Table 5). These changes are shown whereby, with increasing temperature, the clays show:

(i) an increase in their birefringence; (ii) an increase in their crystallinity, as revealed by the sharpness of their (001) basal reflections; (iii) a decrease in their basal spacings (Table 5); (iv) commonly a change in the Differential Thermal Analysis (DTA) reactions whereby the higher (500-600°C) endothermic peak increases relative to the lower (120-180°C) one, i.e. there is a systematic change in the hydration characteristics of these clays with temperature. This is not apparent using X-Ray diffraction methods so that DTA has been used routinely (Browne and Gardner, 1982) to recognise alteration in the cap rocks at Ngawha. Here heat moves by conduction with little mass transfer except for addition of H_2O, CO_2 and perhaps sulphur.

9.1.4 Biotite

This mineral is a common primary constituent in many geothermal fields but is rare as a hydrothermal mineral. However, it does occur as pale brown crystals in one core from Ngawha, at Tongonan (both occurrences where the temperatures are above 220°C, and in the hotter (> 325°C) parts of the Cerro Prieto reservoir (Browne and Gardner, 1982; Reyes and Tolentino, 1981; Elders et. al, 1981).

Table 5. X-ray diffraction characteristics of interlayered illite/montmorillonite.

Mineral	Position of (001) Reflection (Å)	Well temperature (°C)
montmorillonite	>13.0	<140
interlayered illite - montmorillonite	11.0 - 12.5	140 - 180
	10.9 - 11.0	180
	10.6 - 10.8	180 - 200
	10.5	200 - 220
	10.3	220
illite	9.8 - 10.3	>220

Based on Steiner (1968), C.P. Wood (pers. comm.), Browne 1978 and unpublished data.

9.2 Calc-silicates and related minerals

The formation of zeolites is strongly temperature dependent and because of this their identity is a useful guide to their deposition temperature. There is a range of zeolites that occur at low temperatures (<110°C) best seen in the low temperature Iceland fields (Kristmannsdottir and Tomasson, 1978). Among the minerals (Figure 2) that occur at higher temperatures, epidote seems to be the most reliable and consistent temperature guide, first appearing at 250°C in many fields, irrespective of the lithology of the host rock. Some variation exists regarding prehnite (> 220°C in New Zealand compared with > 300°C at Cerro Prieto), probably due to differences in the pH and calcium contents of the circulating fluids, but this mineral should also be a useful temperature indicator when applied to samples from bores located in a single field.

A good example of the application of calc-silicate mineralogy for understanding the thermal regime of a geothermal system is that of Cerro Prieto. Elders et.al (1981) used these minerals to delineate the section of this reservoir where maximum heating has occurred.

9.3 Vitrinite reflectance geothermometry

A very interesting study of the reflectance properties of organic material in the sediment samples recovered from the Cerro Prieto drillholes has been made by C.E. Barker and W.A. Elders (1979).

Minute particles of organic material are very common and widespread in sedimentary rocks, although a microscope is needed to see them; it is possible to measure their reflectance (brightness in reflected light). Barker and Elders show that samples from the reservoir increase in mean reflectance from 0.12% at 240m depth to 4.1% at 1700m; downhole temperatures measured over this interval increase from 60°C to 340°C. The correlation between reflectance

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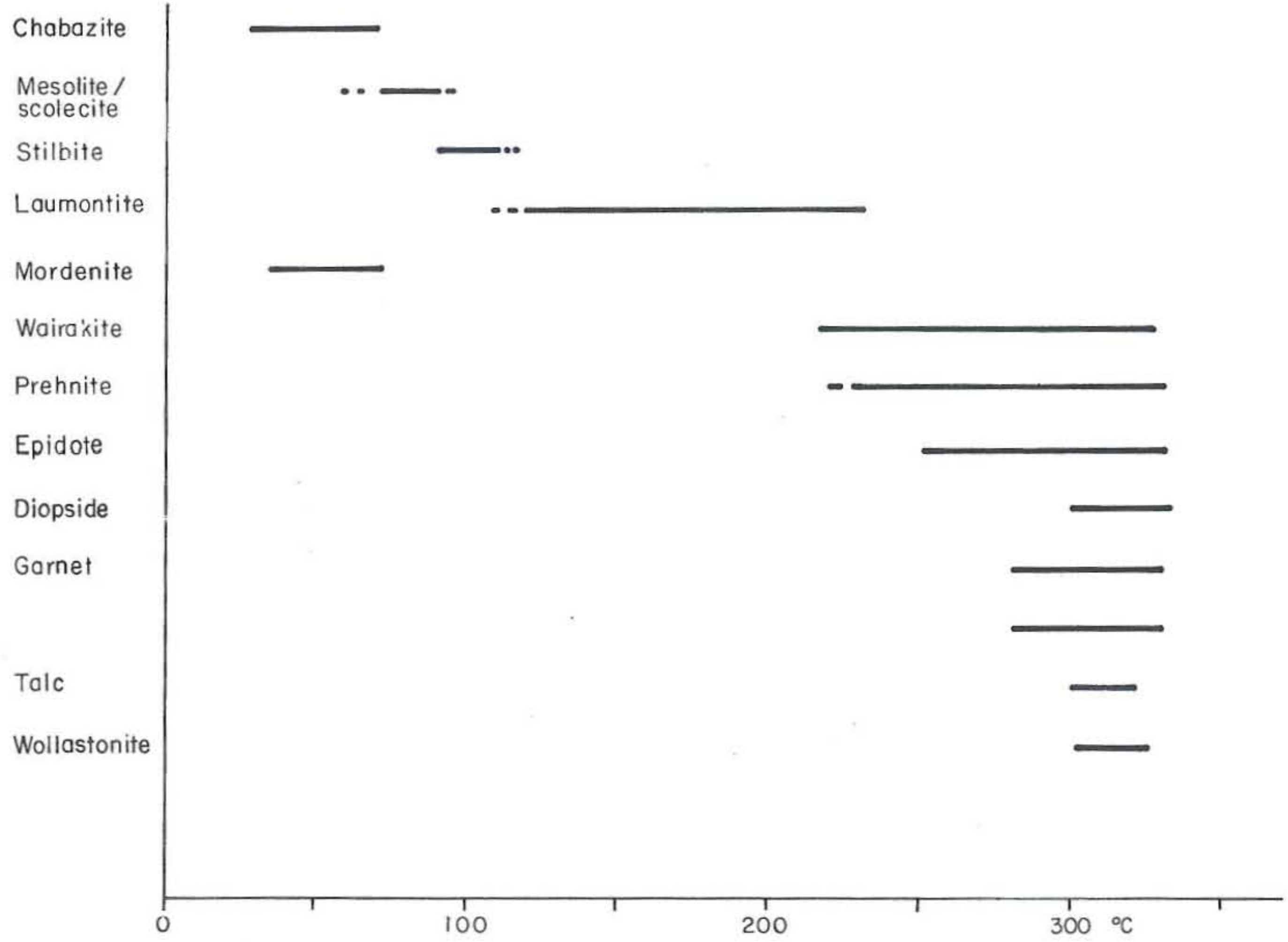

Fig. 2. Calcium silicates.

values and the measured downhole temperatures is very close and the fact that the method is now "calibrated" means that good estimates of the downhole temperatures can be made immediately after future samples are recovered. Although this method is not applicable to geothermal systems located wholly in volcanic fields, sedimentary rocks do occur in most NZ and some Philippine systems, and the technique has potential at both places. A preliminary study by Struckmeyer (1983) on organic matter in lacustrine sediments of the Huka Falls Formation, NZ, confirmed this potential.

9.4 Fluid inclusion geothermometry

When hydrothermal minerals grow, or recrystallize, in a fluid environment, tiny growth irregularities trap small amounts of the depositing fluid within the solid crystal. The sealing off of these irregularities yields primary inclusions, or by their later crystallization, perhaps in response to a microfracture, secondary inclusions. If the fluid trapped is a single phase this will separate into a vapour (bubble) and a liquid at lower temperatures (two phase boundary). Heating of an inclusion on a heating stage will restore this fluid to a single phase, at its homogenisation temperature; this is, therefore, taken to be its entrapment temperature (pressure corrections for geothermal samples are unusually small). The value of fluid inclusion geothermometry in helping determine the thermal history of a reservoir is obvious but only recently has the method been used routinely (e.g. Leach, 1982). In my opinion it is a cheap but powerful aid to geothermal petrology; fluid inclusions are common and abundant in many hydrothermal minerals (e.g. quartz, calcite, anhydrite are best but epidote, wairakite and sphalerite have been used also). The literature covering fluid inclusion geothermometry itself is by now very extensive but a good starting place to learn the method, and theory are some of the many excellent papers written by Roedder (e.g. 1967, 1979). He discusses some of the assumptions inherent in the method including:

(i) That the fluid trapped was a single phase; this assumption is not always met but a two-phase zone can be readily recognised (e.g. parts of Broadlands and Kawah Kamojang) by looking at many inclusions.

(ii) That the inclusion cavity does not change in volume after it is sealed, e.g. by crystallization of a daughter mineral; however, many daughter minerals dissolve upon heating and they are rare in geothermal samples. This assumption causes no real problems.

(iii) That the inclusions have not leaked; quartz, because it has no cleavage rarely leaks.

9.4.1 Example: Broadlands

Two phase H_2O inclusions occur in hydrothermal quartz and sphalerite crystals recovered from several bores in this field. Primary inclusions typically occur in the clear tips of euhedral crystals whereas vast numbers of tiny secondary inclusions occur near their base (Browne et al., 1976; Weissberg et al., 1979).

The homogenisation temperatures of 177 primary fluid inclusions in hydrothermal quartz and sphalerite crystals varied between 201°C and 293°C; they ranged from 13°C below to 37°C above measured well temperatures but averaged 8°C above. One hundred and ninety-eight secondary inclusions had a wider temperature spread than the primary ones and averaged 6°C below the corresponding well temperature, demonstrating among other things, that as many measurements as possible should be made. Many vapour-rich (i.e. steam) inclusions occur in all samples from two bores (Br 24 and Br 7 -845 m) because at the time of their deposition, the fluid was boiling somewhere in the system. It was hard to measure the precise homogenisation temperature of fluids in these inclusions but in many cases the samples derive from depths where the bore temperature is below the boiling temperature of pure water. Reasons for the difference might be: (a) changes in fluid pressure (about 30 bars) assuming that the temperature and fluid composition have not changed; (b) the temperature has dropped some 30°C since the time of the formation of the inclusions; (c) assuming that temperature and pressure have remained constant, then changes in fluid composition have lowered the boiling temperature by about 30°C. The effect of dissolved NaCl is to increase the downwell boiling temperatures above that for the boiling temperature of pure water. However, the presence of even small amounts of CO_2 in solution has a spectacular effect in lowering the boiling temperature. Thus the Br 24 samples plot on, or close to, the vapour pressure curve for 0.60 moles CO_2 indicating that this concentration would have been sufficient to cause vast boiling at the present bore temperatures. Similarly, vast CO_2 concentrations of 0.85 mole % would have been sufficient to cause boiling at the measured bore temperatures in Br 7. The amounts of dissolved CO_2 varies throughout the field (for example, the total discharge from Br 24 and Br 7 contains 5.4 and 2.1 mole % CO_2 respectively) and this seems to be the most likely explanation (Browne, Roedder and Wodzicki, 1976).

However, the close agreement between the homogenisation and bore temperatures both validates the method and demonstrates that Broadlands is thermally stable.

9.4.2 Other examples

Another example is Cerro Prieto where work on vein samples (Elders, et al., 1981) shows that there is generally a close correspondance between the bore and the fluid inclusion homogenisation temperatures; indeed the method is claimed to be a more accurate guide to reservoir temperatures than temperatures measured in the bores. The method is now also used widely in Japan and on the material from the Philippine fields.

A prime value of fluid inclusion geothermometry lies in its ability to identify those fields, or parts of fields, where heating or cooling have, and are, occurring (see Section 15.4).

10 HYDROTHERMAL ALTERATION AND PRESSURE

Compared with metamorphic environments, fluid pressures in geothermal areas, down to the depths drilled, are low and seldom exceed 200 bars. In liquid-dominated fields they are usually close to, or slightly above, hot hydrostatic pressure but must locally have exceeded lithostatic pressure to have triggered the hydrothermal eruptions common in many geothermal fields. In vapour dominated fields, of course, fluid pressures are always well below that of the corresponding hydrostatic pressure.

The main effect of pressure on hydrothermal alteration is an indirect one in that it controls the depth at which boiling occurs.

11 HYDROTHERMAL ALTERATION AND BOILING

11.1 Boiling

Boiling has a traumatic effect on the chemistry of a fluid and this is reflected in the deposition of hydrothermal minerals that precipitate to minimize any resultant changes. Loss of even a small amount, of steam also removes, a high proportion of dissolved gases especially H_2S and CO_2. Loss of carbon dioxide is especially important since it causes calcite to precipitate and the remaining liquid to become more alkaline, i.e. the reaction:

$$Ca^{+2} + 2HCO_3^- = CaCO_3\downarrow + CO_2\uparrow + H_2O$$

moves to the right

This can be represented by Figure 3 which shows the relationship between kaolin, K-mica (muscovite) and adularia (K-feldspar). Boiling of a liquid represented by composition point A(most NZ systems) at 260°C will increase the pH of the remaining liquid (i.e. H^+ will decrease) and cause a slight cooling.

As a result, the fluid composition will change in the direction indicated by the arrow; as this happens K-feldspar (adularia) will precipitate as the system responds by removing K^+ from solution.

Point A is located upon the K-feldspar/muscovite (see Section 13.1) boundary but, by contrast a hypothetical fluid represented by point B lies well within the muscovite field. In this case boiling causes this mineral to precipitate (as illite) as the fluid becomes more alkaline. In practice it is really the pH of the boiling fluid which determines which potassic mineral will precipitate in response to the process. For similar reasons albite and/or paragonite could be expected to precipitate when one considers the Na_2O-SiO_2-Al_2O_3-H_2O-HCl system but this does not seem to happen, possibly for some kinetic reason.

Boiling is a self-limiting process, at least locally, both because of the resultant cooling and a reduction in permeability, due to mineral deposition.

11.2 Recognising boiling zones

Because of these processes, boiling zones, at least where the deep fluid is of the usual near neutral alkali chloride type, may be recognised by:

(a) veins of adularia, formed due to the liquid becoming more alkaline; (b) calcite crystals deposited due to CO_2 loss (not cooling); this type of calcite almost always has a bladed morphology (Tulloch, 1982); (c) abundant quartz, deposited due to cooling; (d) vapour and two phase fluid inclusions in the same sample; however, liquid-rich inclusions can still form in boiling zones (J.W. Hedenquist, pers. comm.) so that many inclusions may need to be examined to recognize a boiling zone using this criterion. The fluid inclusion evidence for boiling needs to be evaluated carefully on a case by case basis.

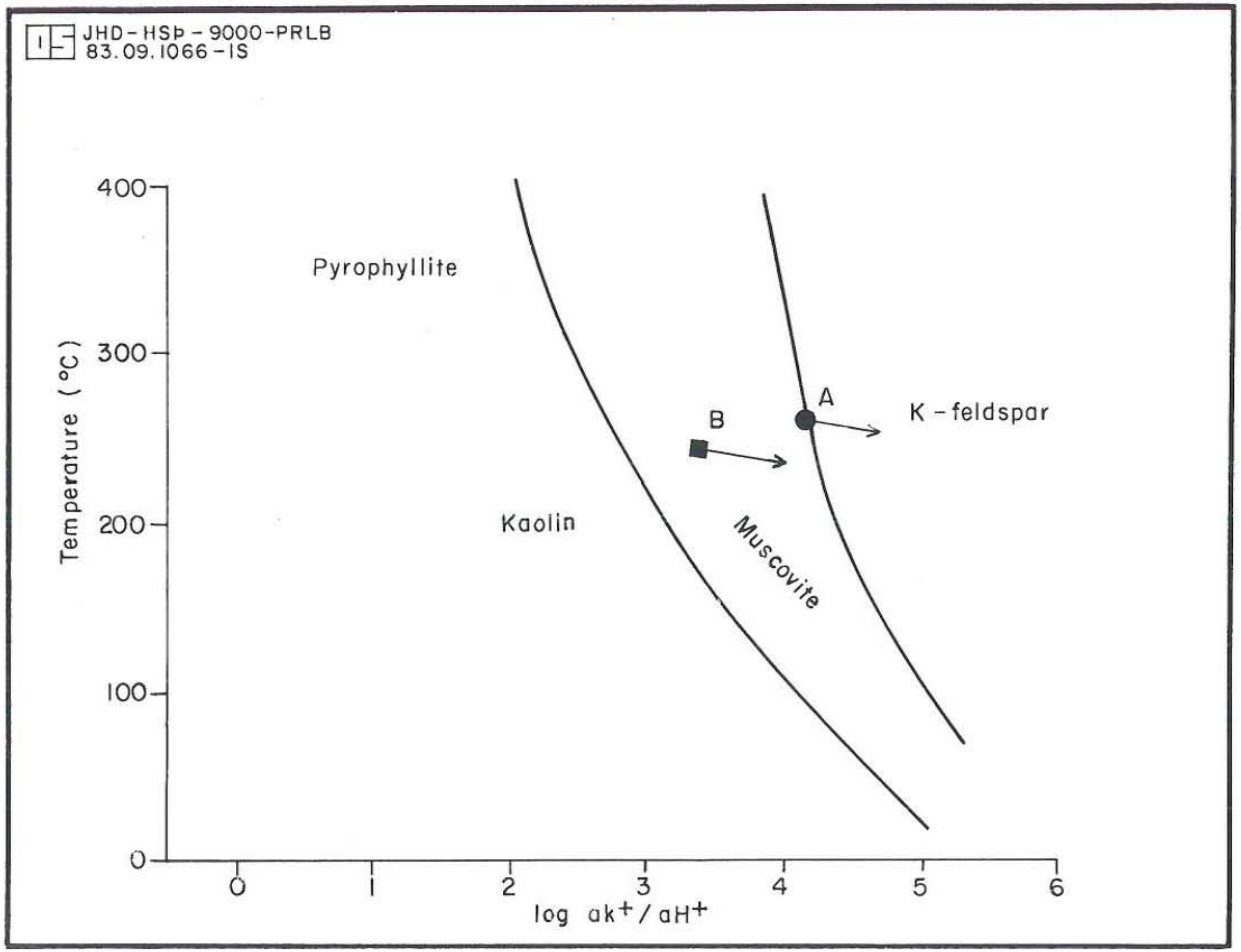

Fig. 3. General form of the effect of temperature on the $K_2O - SiO_2 - Al_2O_3 - H_2O - HCl$ system. (after Montoya and Hemley, 1975 and Rose and Burt, 1979); (see text for discussions.)

12 HYDROTHERMAL ALTERATION AND RESERVOIR ROCK TYPE

The parent rock influences hydrothermal alteration mainly through its control of permeability by texture. The initial mineralogy of the host rocks seems to have little effect on the alteration assemblages occurring in discharge zones. For example a typical equilibrium assemblage at 260°C is albite quartz, chlorite, epidote, calcite +/- pyrite, +/- adularia, +/- illite irrespective of whether the rocks are andesites (Philippines and Indonesia), calc-alkaline rhyolites (New Zealand), alkaline lavas (Kenya) or sediments (Cerro Prieto). Only the basalts of the active Icelandic fields do not, so far as is known, contain appreciable K-mica or K-feldspar (Arnorsson et al., 1983) but adularia occurs in at least one fossil (Geitafell) system (Fridleifsson, 1983).

At lower temperatures the nature of the parent material clearly influences the alteration product. High-silica zeolites like mordenite, are common in rhyolitic fields at Yellowstone and New Zealand whereas lower silica zeolites like chabazite, thomsonite, scolecite, occur in the basalts of Iceland and andesites of Kamchatka. Further, the quantity of a mineral which forms may reflect the nature of the parent rock, e.g. hydrothermal calcite is widespread in Kizildere and Ngawha where the reservoir rocks include limestone.

One might also expect that the composition of a hydrothermal mineral whose structure allows for considerable amount of element substitution, in some way reflects the composition of the mineral it replaces. Little information is so far available on this point, but this does not, however, appear to be the case. For example, chlorite replacing plagioclase and pyroxene in the same samples from Heber (Browne, 1976) have very similar compositions. On the other hand illite clay which replaces biotite in NZ fields is invariably better crystallized than illite replacing plagioclase in the same sample.

However, another way of approaching this subject is to recognise that the rock mineralogy in the recharge parts of the fields controls fluid composition (including its pH). This is true for the New Zealand fields, including Ngawha where the deep "basement" greywackes and argillites through which the fluids move have a primary mineralogy that includes quartz, K-feldspar, albite, K-mica, epidote, calcite and pyrite.

It is therefore no surprise that;

(a) the deep waters of New Zealand systems are of very similar composition with significant variations in only carbon dioxide, boron and infinitely soluble constituents such as chloride,

(b) the above minerals, plus some others, are also those that precipitate in the discharge zones of the same fields.

Of course the salinity of the fluid itself is controlled by the amount of chloride it contains, which in turn depends upon the quantity of chlorine that fluids remove from the rocks through which they travel (Arnorsson et. al, 1983). Further, other minerals also occur in the discharge zone, notably wairakite plus rarer zeolites and montmorillonite but it is possible that the zeolites are metastable (Giggenbach, 1981).

13 HYDROTHERMAL ALTERATION AND FLUID COMPOSITION

The close relationship between fluid composition and hydrothermal alteration was clearly recognised as early as 1934 when Fenner published his first account of the results of drilling at Yellowstone National Park; subsequent work has confirmed and greatly strengthened his conclusion.

In general the geothermal petrologist observes and identifies the hydrothermal minerals that occur in cores and cuttings and then predicts, or comments upon, the composition of the altering fluids. However, it is worth remembering that other factors, such as temperature are also important and this is certainly true of the calc-silicates (Section 4.2).

The close relationship between fluid composition and hydrothermal mineralogy may be seen in superficial alteration; thus acid ($pH < 3$) fluids deposit sulphur, alunite and kaolin, and near neutral alkali chloride water precipitates silica sinter. Anhydrite and other sulphates commonly form where sulphate bearing fluids circulate (e.g. in many Philippine and some sea-water systems).

Where ammonia concentrations are high, such as in the fumaroles at Ketetahi, New Zealand, ammonium minerals deposit. Lithium mica (lepidolite) has been observed in shallow drillholes at Yellowstone where it presently precipitates from waters with high fluorine (~30 ppm) and an "extraordinarily high" ratio of Li to K (1 ~2) (Barger et al., 1973). Other examples are fluorite deposited at Olkaria and datolite at Larderello, where fluids are high in fluorine and boron respectively. The amounts of carbon dioxide and hydrogen sulphide dissolved in the reservoir fluids have a very direct affect on whether or not, and to what extent, carbonate and sulphide minerals precipitate in the reservoir rocks.

One important conclusion about the affect of fluid composition on hydrothermal alteration is that the absolute concentration of dissolved constituents is much less important than the ratios of the activities of the major ions. This is demonstrated by the fact that the highly saline brines of the Salton Sea (~ 250,000 ppm) react with their reservoir rocks to produce an assemblage consisting of quartz, calcite, epidote, K-feldspar, albite, K-mica and chlorite; this is also the equilibrium mineral assemblage in hot (260°C) parts of several geothermal fields where the altering fluids are very dilute (< 3000 ppm TDS).

13.1 Activity Diagrams

These are useful tools for summarising the relationship between hydrothermal minerals and fluids. Typically they plot log ion activity ratios on the axes of a diagram at a useful temperature and in the presence of excess silica (where the activity of an ion is defined as its effective concentration in a reaction). Figure 4 is one such diagram and shows the relationships between sodium and potassium minerals at 260°C in the presence of quartz, by a plot of log (a_{Na^+}/a_{H^+}) on one axis and a plot of log (a_{K^+}/a_{H^+}) on the other. The boundaries are drawn where two or more phases coexist in equilibrium; their positions were determined by experiment, by calclulation and by observations of mineral relations in geothermal fields. For example, the K-feldspar - K-mica boundary can be expressed by the equation:

$$3KAlSi_3O_8 + 2H^+ = KAl_3Si_3O_{10}(OH)_2 + 6SiO_2 + 2K^+$$

whose equilibrium constant depends upon the ratio of a_{K^+}/a_{H^+}; this value has been determined, for a useful temperature range in a long series of experiments by Hemley and Jones (1964). Similarly, transformation of K-mica to kaolin also involves a_{K^+}/a_{H^+} and this equilibrium constant also varies with temperature.

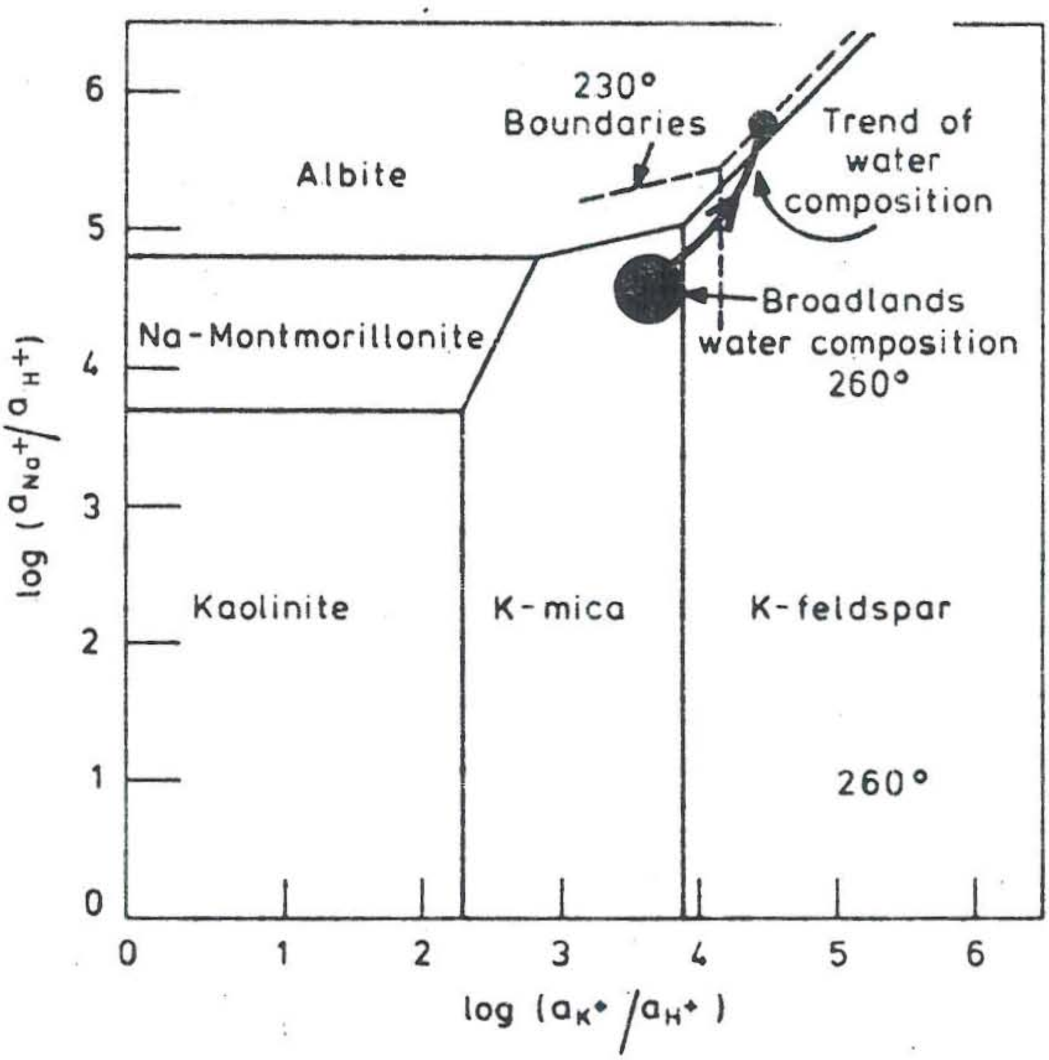

Fig. 4. Activity diagram for sodium and potassium in the presence of quartz at 260°C in terms of ion activity ratios - see text (from Browne and Ellis, 1970).

This reaction is

$$2KAl_3Si_3O_{10}(OH)_2 + 3H_2O + 2H^+ = 3Al_2Si_2O_5(OH)_4 + 2K^+$$

However, representation of the reaction between albite and K-mica can be used to calculate the slope (but not the position) of the boundary between these two phases and this is plotted (Figure 4) as a straight line with a slope of 1/3. Thus:

$$3NaAlSi_3O_8 + K^+ + 2H^+ = KAl_3Si_3O_{10}(OH)_2 + 6SiO_2 + 3Na^+$$

for which the equilibrium constant, K, is = $a^3_{Na^+}/a_{K^+} \cdot a^2_{H^+}$

The effect of pressure on the position of the boundaries is not great for the low pressures encountered in geothermal systems but Hemley and Jones's experiments were conducted at 1 kbar; however, a slight adjustment of the value of log (a_{K^+}/a_{H^+}) for K-mica, K-feldspar equilibrium proportional to a 0.15 increase per 1000 bars is allowed for in these figures. Much more important is the effect of temperature as is clearly shown by the relocation of the same mineral equilibrium boundaries at 230°C (Browne and Ellis, 1970).

The value of activity diagrams, such as these, is that one can also plot the composition of a thermal fluid on the same diagram; to do this we need to know the concentration of K^+, Na^+ and H^+ in the deep water and to use the appropriate activity coefficients; these vary, depending especially on the charges on the cations and the ionic strength of the thermal fluid, but can often be calculated using the Debye-Huckel theory of activity coefficients. (see Ellis and Mahon, p. 129-131) and a brief and simple summary account in Garrels and Christ, 1965, p. 61-67 . Deep waters of Broadlands composition (ionic strength about 0.04 m and taking activity coefficients for Na^+ and K^+ of 0.70) plot as shown by the black circle on Figure 4. Thus at 260°C this water is in equilibrium with K-mica (illite or sericite) but very close to a_{Na^+}/a_{H^+}, a_{K^+}/a_{H^+} values where albite and K-feldspar (adularia) also coexist.

From Figure 4 we can, for example, also follow what happens in the rocks when steam is lost from the system. The main trend is an increase of pH and a slight cooling (we can ignore concentration due to evaporation). As this occurs the fluid deposits K-feldspar to remove excess K^+ ions from the solution and the composition of the fluid moves away from, and out of, the K-mica stability field towards the albite-K-feldspar mineral equilibrium boundary.

Similar diagrams can be constructed for other aluminosilicate minerals using different combinations of cations (Browne and Ellis, 1970; Bird et. al, 1984) and these are of a reliability that depends on the quality of the appropriate thermodynamic data (see Helgeson, et al., 1978). In many of the older activity diagrams, minerals which commonly form solid solution series (e.g. epidote, chlorite, garnet) are treated as stoichiometric end members; recent work by Bird et al. (1980, 1981) goes a long way to taking the effects of solid solution into account. Optimum use of their diagrams requires, however, a knowledge of the chemistry of the key mineral usually obtained by using an electron microprobe. Note also that figures 4 and 5 do not allow interlayered clays to be treated as a single phase.

13.2 The $CaO - Al_2O_3 - SiO_2 - K_2O - H_2O$ +/- CO_2 system

This is a very important system because calcium minerals are so common in active geothermal fields (Table 1). Figure 5 is an activity diagram showing the relationships between some of the calcium phases; more comprehensive figures, for higher $a_{Ca^{2+}}/a_{2H^+}$ and different temperature are available in Bird et. al 1984.

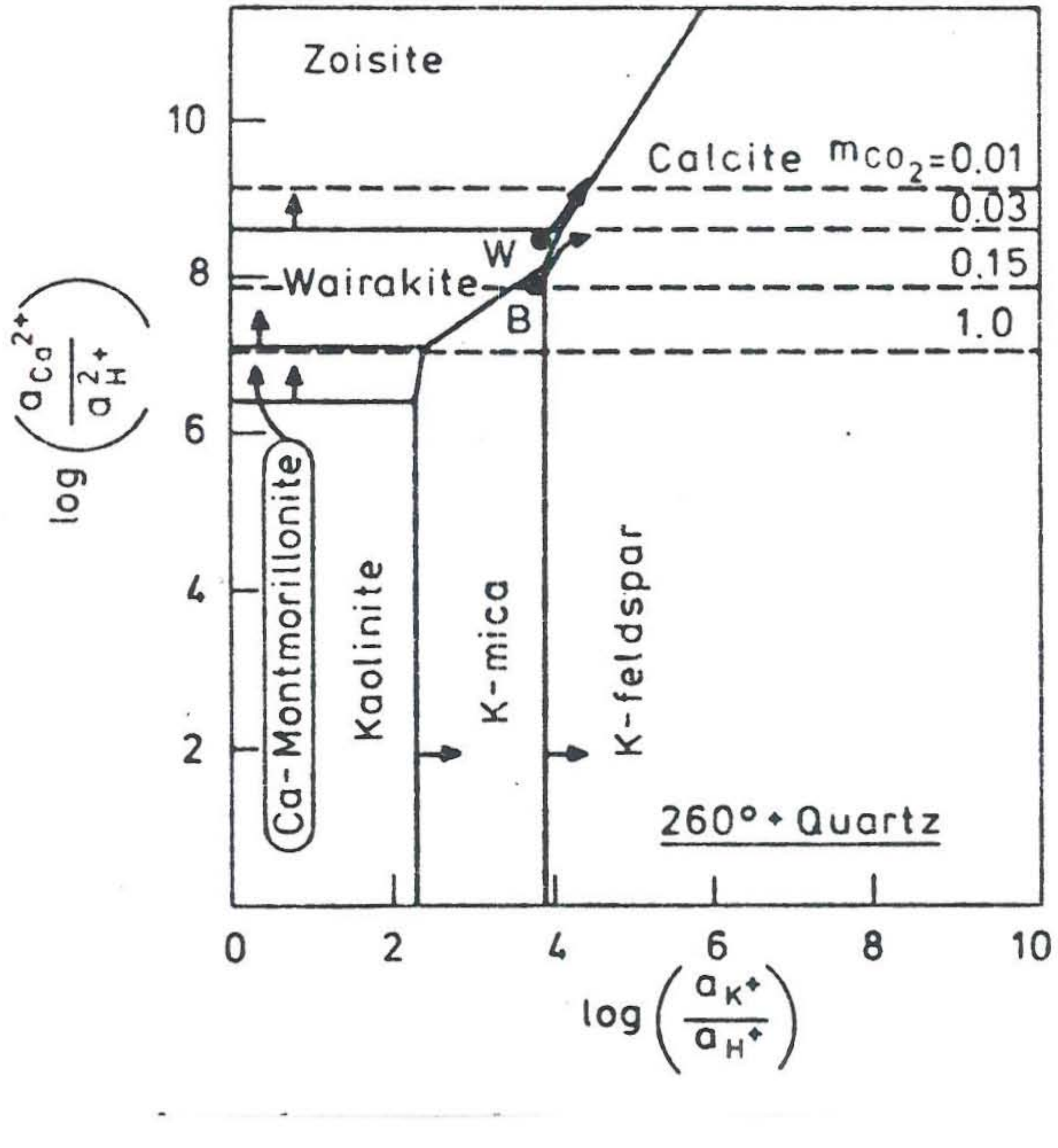

Fig. 5. Activity diagram for calcium and potassium minerals at 260°C in terms of ion activity ratios (from Browne and Ellis, 1970); B represents the composition of the deep Broadlands water, W that of Wairakei.

An important component in this system is carbon dioxide which results in the formation of HCO_3^- and CO_3^{2-} ions. For a particular value of mCO_2 in a hydrothermal solution, a horizontal line drawn on an activity diagram with log $(a_{Ca^{2+}}/a^2_{H^+})$ as an axis (e.g. Figure 5) represents the value at which calcite precipitates. Above this line calcium silicate phases are unable to precipitate (which they can do in a CO_2-free system) and calcite does so instead. For higher concentrations of CO_2 the calcite blind is lowered to lower log $(a_{Ca^{2+}}/a^2_{H^+})$ values. For example for $mCO_2 = 0.15$, zoisite (or epidote) cannot form and for $mCO_2 = 1.0$ very little of the wairakite field is open. This explains observed differences in alteration at Wairakei and Broadlands; calcite is common at the latter field whereas epidote and wairakite are rare. The opposite situation applies at Wairakei where the water, with $mCO_2 = 0.01$, is near to the wairakite, epidote, K-feldspar coexistence point. This composition is shown as W on Figure 5 and is appreciably below the calcite solubility line that occurs at $\log(a_{Ca^{2+}}/a^2_{H^+}) = 9.1$ for 0.01 mCO_2 concentration. Slight separation of steam and the consequent pH rise would bring the water on to the epidote (zoisite), K-feldspar stability line, and an appreciable steam separation would be required before the composition point reached the calcite line (Browne and Ellis, 1970).

Broadlands, however, has a deep water composition point close to the K-mica, K-feldspar, and wairakite coexistence point and to the calcite line. With steam separation, calcite would be precipitated, and although the composition point would subsequently follow along the wairakite, K-feldspar boudary, it is unlikely that wairakite would nucleate and grow with any efficiency while calcite was forming (Browne and Ellis, 1970).

For a given temperature there is a particular concentration of carbon dioxide in the underground water above which calcite will precipitate as the calcium mineral phase with first steam separation, and below which a calcium silicate phase forms or remains stable with steam separation. At

260°C this carbon dioxide concentration is about the Broadlands value (0.1m). Calcium silicate phases rarely deposits in drill-pipes, whereas calcite can do so. Slight calciting of drillholes occurs at Broadlands but not at Wairakei. Reasoning along these lines may make it possible to recognise, at an early stage, geothermal fields which could give trouble with calcite or aragonite precipitation caused by flashing. The CO_2 concentration of the deep water is a key factor and it is an easy matter for a petrologist to recognise whether calcite or calcium silicate phases predominate in cores and cuttings.

13.3 Iron sulphide and oxide phases

Pyrite and pyrrhotite are the common iron minerals which occur in active geothermal fields (Table 1) although very occasionally traces of marcasite and chalcopyrite are also present. The stabilities of pyrite and pyrrhotite can be related to the partial pressure of H_2S and H_2 and to the temperature (see Ellis and Mahon p. 108 - 109). Pyrrhotite may form where steam collects after it has separated from the liquid phase (which consequently has a higher pH_2S/pH_2 ratio). Thus a petrologist should be able to comment on the relative concentrations of H_2S and H_2 in the subsurface waters (and this can vary even within a single field). Since the amount of H_2S which a field will discharge during production is a very important (and expensive) consideration, an early qualitative opinion about the likely concentration of H_2S may be given from examination of the iron oxide and sulphides present. In some fields (e.g. those in El Salvador, Olkaria, Kenya and El Tatio, Chile) iron oxides such as hematite are common although they may locally coexist with pyrite but seldom pyrrhotite. This implies that the fluids here contain relatively more oxygen but less hydrogen sulphide than fluids of the New Zealand systems.

14 HYDROTHERMAL ALTERATION AND PERMEABILITY

14.1 Introduction

Several studies of hydrothermal alteration mineralogy (e.g. Steiner, 1977; Browne, 1970; Elders, et. al, 1981; Reyes and Tolentino, 1981) have drawn attention to the control that permeability has on the deposition of hydrothermal minerals; these papers also show how a knowledge of the mineralogy of cores and cuttings recovered from a drillhole can be used to interpret, qualitatively at least, subsurface permeability.

Mineral reactions are seldom isochemical and extensive alteration and hydration needs more than pore water to proceed. In many cases, at least carbon dioxide and sulphide species must be added to rocks from solutions.

In rocks of low permeability, equilibrium between minerals and the reservoir fluid is seldom achieved and primary minerals or glass can persist to high temperatures. On the other hand the marked zoning of hydrothermal minerals about fluid channels in several Japanese geothermal fields shows clear structural control of alteration and demonstrates how the intensity and type of alteration reflects permeability (Sumi, 1968). Permeable fissure channels at Wairakei, Broadlands, Waiotapu, Kawerau (all in New Zealand), and Tongonan (Philippines) are characterised by veins of adularia, usually occurring together with quartz and calcite.

14.1.1 Example: Systems of the Taupo Volcanic Zone, N.Z.

14.1.2 Mineralogy changes

Here the observed hydrothermal alteration is used to:

(a) help in determining the relative permeability of the subsurface formations i.e. identify cap and aquifer rocks.

(b) make qualitative estimates of bore outputs before their initial discharge (which often takes place months after drilling).

(c) locate production zones. The most important minerals relating permeability to alteration are the feldspars. With increasing permeability the feldspar mineralogy changes as follows:

Primary andesine
Albite
Albite and adularia
Adularia

(a) The hydrothermal behaviour of the common formations and lithological units (Table 6) shows that the main production, away from zones where faulting or natural hydraulic fracturing dominates, are the Waiora pumice breccias and a coarse pyroclastic unit, the Rautawiri Breccia. Cores of these units taken from good bores commonly contain adularia which, by contrast, is absent from the less permeable rocks.

(b) Browne (1970) showed that the first 12 bores drilled at Broadlands had outputs that reflected the identity of the feldspars in their production zones; bores which discharged more than 150,000 lb/hr of steam were characterised by adularia whereas albite plus adularia were the common feldspars in bores with steam outputs between 40 and 110 thousand lb/hr; poor, or non-producers,(e.g. Br 12,1 and 7) have cores that contain albite and/or andesine. Subsequent drilling at Broadlands, Kawerau and elsewhere has confirmed this empirical relationship.

(c) Following from these points we see that the producing horizons of the good bores contain adularia and quartz whereas albite and adularia occur in zones of lower output, so far as the location of these are known. By contrast, rocks containing primary andesine or andesine plus albite

Table 6. Subsurface formations at Broadlands, their typical alteration and usual hydrological function (after Browne, 1970).

Formation or unit name	Content	Thickness (m)	Adularia, usually present in high output bores	Usual function (except where fractured)
Huka Falls	Lacustrine sediments, tuffs, grits	30-350	No	Barrier
Ohaaki Rhyolite	Pumiceous and spherulitic rhyolite	0-450	Occasional	Partial barrier
Waiora	Pumiceous tuff-breccia	0-250	Yes	Aquifer
Broadlands Dacite	Dense lavas	0-500	No	Barrier
Rautawiri Breccia	Vitric-crystal-lithic tuff and tuff-breccia	160-500	Yes	Aquifer
Rangitaiki	Locally densely welded lithic-vitric-crystal ignimbrite	30-400	No	Barrier
Ohakuri	Bedded pyroclastics	~500	Yes	Aquifer

never serve as producing horizons. Sometimes however, a precise correlation between feldspar mineralogy and the location of production zones is not always possible, partly due to the difficulty of recognising,independently, where multiple feed zones occur. Figure 6 is an example of downwell alteration in cores from Broadlands drillhole, Br 25;

This shows adularia to be both widespread and abundant in recovered cores and one would expect the well to have a good output. This it certainly does since it discharges steam at a rate of 100 tons/hr at a wellhead pressure of 15 bars g. The reservoir rocks are highly permeable and tests show that the permeable levels are at about the depth of the middle adularia peak (750m) and from 1150 to 1190 m where adularia occurs without albite. The well is cased to 600 m but the distribution of adularia, peaking also at about 500 m, indicates that useful quantities of thermal fluid could also be derived by perforating the casing here. (See also Ellis and Mahon, p. 96, 1977).

14.1.3 Chemical changes

As may be expected from these mineral changes we find that cores from good production zones have high potassium and low sodium contents so that their K_2O/Na_2O ratios are high (>4). In the "medium grade" production zones where both albite and adularia occur, the K_2O/Na_2O ratio commonly varies from 0.5 to 4 but in non-producing zones it is usually less than 1. This relationship is shown clearly in Figure 6 where the downwell K_2O content parallels the abundance of adularia and Na_2O that of albite.

However, in bores whose cores also contain abundant illite or interlayered illite-montmorillonite the distribution of K_2O does not so closely match that of adularia.

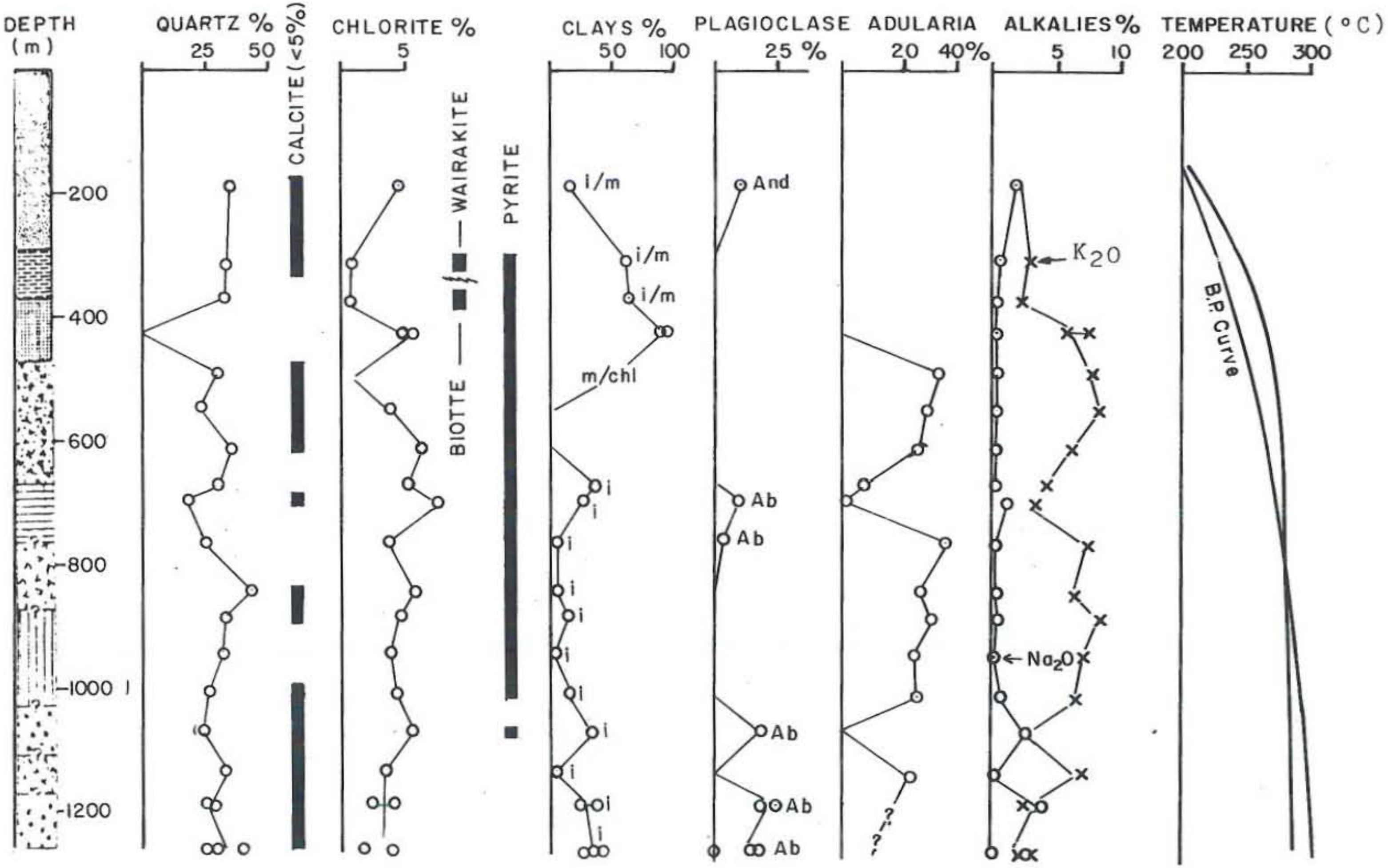

Fig. 6. Downwell alteration in Br 25; O represents cores; i = illite; Ab = albite; And = andesine; m = montmorillonite; chl = chlorite. Rocks encountered are rhyolitic lavas and tuffs.

Figure 7 shows the general relationship between the alkali and calcium contents of feldspars and permeability in New Zealand fields.

14.1.4 Pyrrhotite - an indicator of poor permeability?

Cores from some hot but impermeable zones at Broadlands contain pyrrhotite (usually about Fe_8S_9 but variable). It is most abundant in lake beds where organic matter persists causing local reducing conditions, and its formation may be due to the trapping of separated steam in the impermeable rocks. Pyrite, on the other hand, can precipitate from residual water from which steam has separated (Ellis and Mahon, 1977).

14.1.5 Example: Systems in the Philippines

Reyes and Tolentino (1981) summarised and described how petrology has assisted in geothermal exploration in the Philippines. Among other applications,hydrothermal alteration is used to predict well permeabilities prior to completion tests and to decide on casing depths, particularly that of the 9 5/8 inch. In addition, petrologists there are sometimes able to determine the position of a well in relation to upflow regions of the field (Reyes and Tolentino, 1981; Yock, 1982).

A comparison of the common hydrothermal minerals in the production zones, as recognised by circulation losses during drilling and well completion tests (Table 7), shows that quartz and adularia occur in the permeable zones of four systems composed of igneous rocks whereas anhydrite and illite are present in three. The significance and cause(s) of anhydrite deposition in active geothermal fields are poorly understood, however; whether illite or adularia precipitates in permeable zones mainly depends on fluid pH and its potassium content.

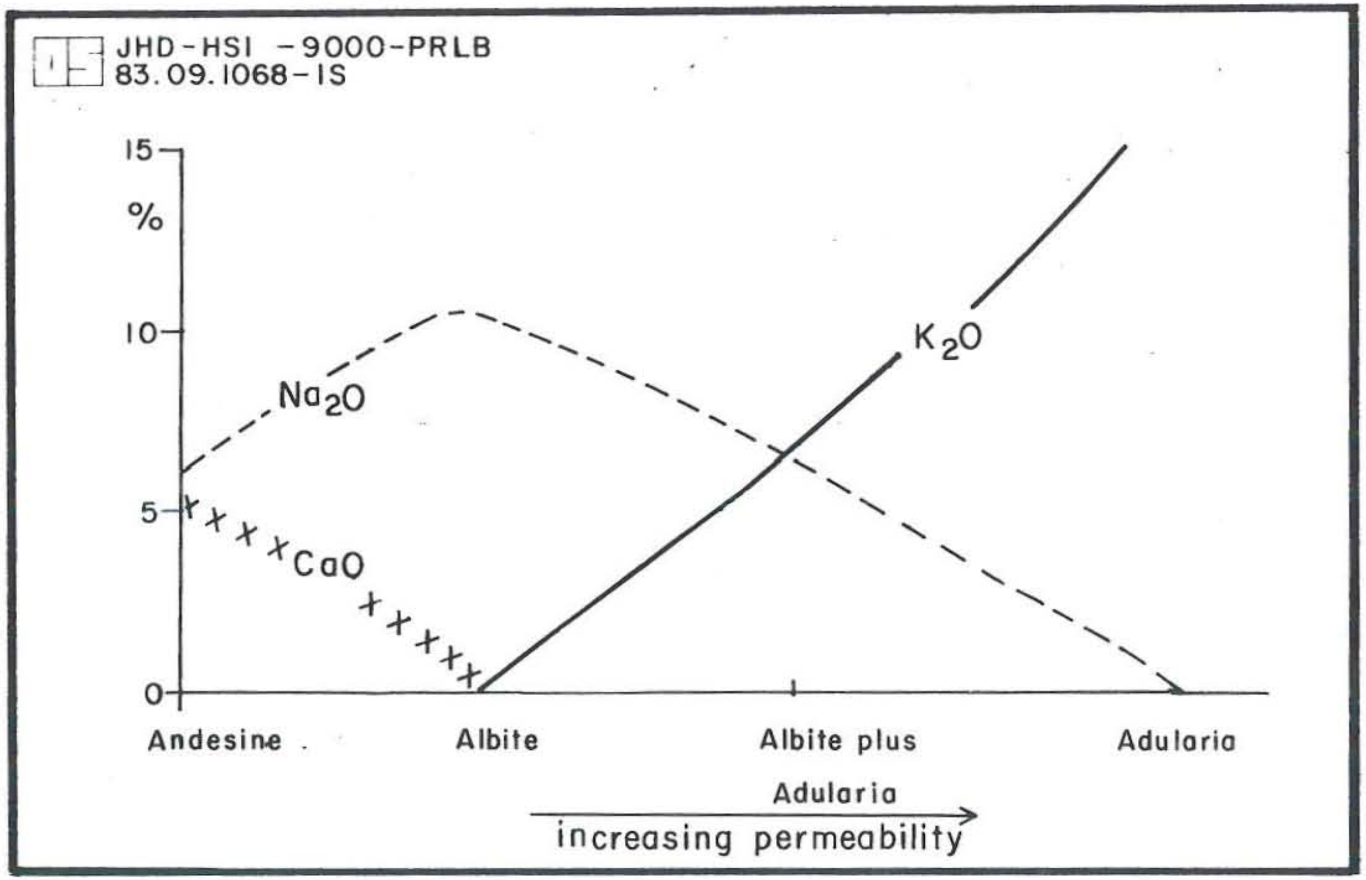

Fig. 7. Chemical changes of feldspars with increasing permeability.

Table 7. Common hydrothermal minerals found in permeable zones of Philippines geothermal systems (Reyes and Tolentino, 1981).

Selected geothermal areas	Lithology of permeable zones	Alteration minerals commonly found in permeable zones
Tongonan, Leyte	Andesite lavas and volcaniclastics	Quartz, adularia, illite, anhydrite, wairakite, calcite
Southern Negros	Andesite volcaniclastites, diorite pluton	Quartz, adularia, illite, anhydrite
Daklan Benguet	Volcanics and calcareous sediments	Quartz, adularia, illite, calcite
Northern Negros	Limestone	Quartz, calcite
Bac-Man Albay	Andesite lavas and volcaniclastics	Quartz, adularia, anhydrite pyrite

15 TIME AND THE DURATION OF GEOTHERMAL ACTIVITY

15.1 Introduction

Geothermal systems have a beginning. And they have an end. We do not know much about what happens between these two events but it is a matter which is well worth thinking about. Let us consider some possible scenarios for the evolution of a geothermal system. (Figure 8).

We may presently be looking at a very narrow time slit on one of these curves in terms of permeability, temperature, salinity and the location of activity and know a little about general hanges when we monitor a field during its exploitation. It is not likely, however, that geothermal systems would undergo the same (induced) changes if they were left undisturbed. In this lecture we will attempt to put numbers along the axes of Figure 8.

15.2 The time axis

Two approaches: (a) indirect estimates about the ages of active systems themselves.

(b) from analogy with measurements, or estimates, made on the duration of activity in fossil geothermal systems, chiefly hydrothermal ore deposits.

Direct dating, for example, by K/Ar or Rb/Sr methods, of hydrothermal minerals might seem at first sight, to be an obvious method of getting some idea about the age of an active field. However, so far as I know, worthwhile results have been obtained only on biotite, muscovite and hornblende samples from Larderello (Del Moro et al., 1982). These workers concluded that this system reached a thermal peak more than 3 million years ago but that it has cooled

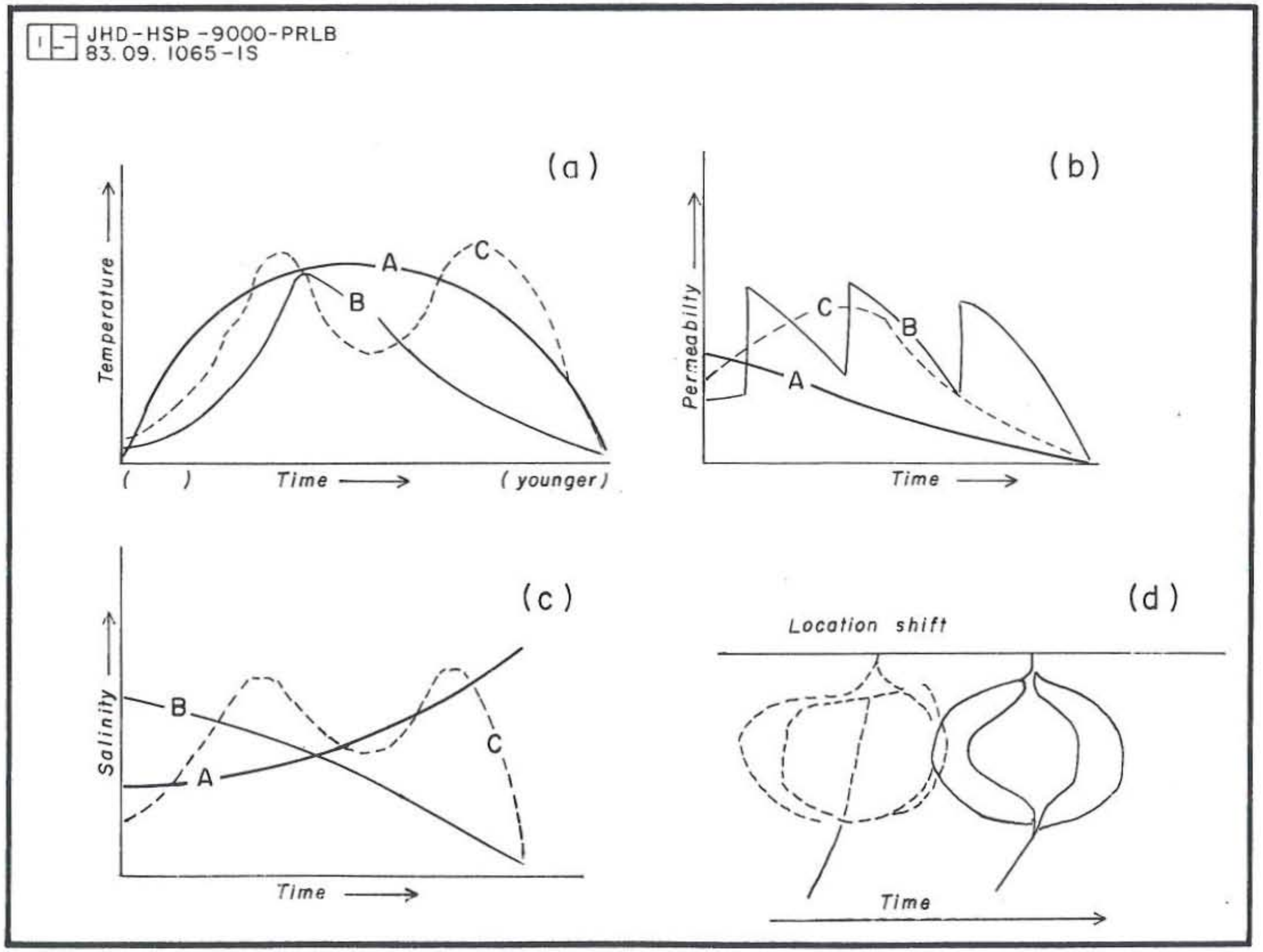

Fig. 8. Scenarios for changes in temperature, permeability, and location during the life of a geothermal system.

by about 120°C since then. The ages of other fields are known only from indirect evidence, except for Waimangu New Zealand, where surface activity began in June 1886.

15.2.1 Ages of active systems

Table 8 summarises estimates made of the duration of activity of several geothermal fields; the relevant evidence is of variable quality but some additional remarks follow.

Kawerau:

This geothermal system has been active, in some form or another for over 200,000 years. The relevant evidence is circumstantial but convincing (Browne, 1979).

A core of locally-derived hydrothermal explosion breccia was recovered from a depth of 487 m in KA25. Because of the low temperature and permeability here hydrothermal alteration is slight; however, a greywacke fragment in the breccia contains veins of wairakite and prehnite, which do not penetrate into the matrix of the rock. Clearly the veins formed before the hydrothermal explosion occurred that erupted the rock, i.e. there was hydrothermal activity in greywackes nearby prior to this eruption. The problem then becomes one of dating the eruption - not an easy task. However, we know that the breccia is overlain by one of the important and widespread ignimbrites which occur in the area (Matahina Ignimbrite) and this has been dated, by B.P. Kohn using the fission track method, as being 200,000 +/- 20,000 years old. Therefore, since the explosion breccia is older than the ignimbrite we can conclude that geothermal activity began at Kawerau over 200,000 years ago.

Table 8. Summary of the duration of activity at selected fields.

Field	Age (years)	Remarks	Reference
Kawerau, N.Z.	>200,000	Surface activity	Browne (1979)
Waimangu, N.Z.	97	big error likely	Weissberg et al., (1978)
Broadlands, N.Z.	370,000		
Wairakei, N.Z.	500,000	Probably maximum	Grindley (1965)
Orakeikorako, N.Z.	>>20,000	Probably oldest active system in New Zealand	LLoyd (1972)
Nesjavellir, Icel.	? 2,000	Certainly young	Kristmannsdottir and Tomasson (1974); Stefansson et al., (1983)
Valles, N.Mex.	1.2 million	"probably not continous"	White, (1979)
Long Valley, Calif.	-300,000	"perhaps intermittent"	White, (1979)
Steamboat Springs, Nevada	-3 million	"probably intermittent"	White (1979)
Yellowstone, Wyoming	? 600,000		White (1979)
The Geysers, Calif.	>57,000	Stored heat calculation	White, (1979)
Sulfur Bank, Calif.	-27,000		White, (1979)
Larderello, Italy	- 3 million	Rb-Sr and K-Ar ages	Del Moro et al., (1982)

Broadlands:

An order of magnitude calculation based upon the estimated amount of sphalerite (zinc sulphide) deposited by hydrothermal fluids into the reservoir rocks indicates a very approximate age for this system of 370,000 years. However, this calculation includes some dubious assumptions about, for example, flow rate and a big error is likely. Clearly though, the field is older than 10,000 years but younger than 1 million.

Wairakei: (Grindley, 1965)

(i) The probable coincidence in time of hydrothermal activity, the eruption of andesite flows and the formation of intersecting faults is cited as evidence that the Wairakei system began between 600 and 900 thousand years ago.

(ii) Better evidence is the occurrence of intensely altered fragments of Wairakei Ignimbrite, produced by hydrothermal eruptions, in conglomerates in the western part of the field. Palynological examination of these sediments indicated they were deposited about 500,000 years ago but there is now some uncertainty about the absolute chronology for this period.

Orakeikorako:

C^{14} dating of hydrothermal eruption products and of a 20,000 year old tephra overlying sinter indicate that this field has been active for longer than 20,000 years (Lloyd, 1972).

Icelandic fields:

Geothermal systems in Iceland have been estimated to last for a minimum of 10,000 to 20,000 years (Stefansson and Bjornsson, 1982) and the general view here (I.B. Fridleifsson, pers. comm.) is that they do not persist for longer than about 200,000 years before all usable heat is removed by circulating fluids. The complete sequence of vug fillings seen in cutting samples is consistent with the view that the Iceland fields have a shorter life than those in New Zealand and the USA; in many places, very fast thermal changes have allowed early deposited,low temperature minerals to survive during hotter episodes and perhaps even subsequent cooling (e.g. Geitafell).

This is seen, for example at Nesjavellir (Stefansson et al., 1983), a field where thermal activity may have begun only as recently as a few thousand years ago (H. Franzson, pers comm) and whose youth was recognised during the earlier alteration studies of Kristmannsdottir and Tomasson (1974a). Vesicles in cutting samples recovered from the southernmost well, Ng6, for example, have a distinct and regular filling sequence that varies with depth. In vugs from shallow depths (<600m) the earliest deposited minerals are (in deposition order), poorly-crystallised smectite, chalcedony, well-crystallised smectite, thomsonite, stilbite, analcime and mordenite followed by, in places, laumontite, quartz, calcite and stilbite plus pyrite. The filling sequence in the deepest samples (1000 to 1100m) consists of calcite, clay, quartz, ?, wollastonite, ?, quartz, epidote, prehnite and wairakite. Information of this type allows deductions to be made about some aspects of the chemical and thermal evolution of the system, but it is probably because of its youth that low temperature phases such as smectite and some zeolites, persist at the present temperatures of nearly 300°C at hole bottom (H. Franzson, pers comm).

United States fields

White (1979) reviewed the time spans of geothermal systems in the United States (Table 8); the oldest field is clearly Steamboat Springs, Nevada, but activity both here, and at other fields, has varied greatly in its intensity.

15.2.2 Duration of activity at fossil geothermal systems

There is a little information presently available about the duration of activity at fossil geothermal systems. Activity at the Tui Mine area, New Zealand, may have lasted about 1.4 m years (see White, 1979) and thermal events at four U.S. fields (Bodie, Goldfield, Creede, Silver City) may have lasted between 500,000 and 1.8 million years (White, 1979). By contrast fossil geothermal fields in Iceland probably had active phases lasting about, or less than 200,-300,000 years (I.B. Fridleifsson, 1973; Stefansson and BjΦrnsson, 1982; G.O. Fridleifsson, 1983).

In summary then, the time axis on Figure 8 has dimensions that differ from place to place: at most (Steamboat Springs, Larderello) it is probably about 3 million years but in Iceland fields it is less than 300,000 years; New Zealand and some USA fields may last about 500,000 years. However, the uncertainties inherent in these estimates are obvious and great.

15.2.3 Episodic activity

Continuous thermal activity is not implied in the above discussion, as there is little doubt that fluctuations in temperature, permeability and even fluid compositions have occurred throughout the life of a geothermal field, i.e. activity is episodic. Evidence for this conclusion derives from studies of fossil systems; for example, many ore deposits demonstrate strong structural control of hydrothermal alteration (and hence past fluid flow). Typically, zoning and the textures in veins show that a period of slow mineral deposition, perhaps culminating in the blockage of flow channels, was followed by a tectonic event which fractured the reservoir producing new channels along which more fluids then flowed. The sequence - deposition, sealing, fracture - has occurred many times; it is no coincidence that most of the world's geothermal systems are located in tectonically active areas. Anywhere else their "plumbing" would soon become blocked through mineral deposition, and their lives be of short duration.

15.3 The permeability axis

This axis is very difficult to put numbers on; in fact we are fortunate if we even know whereabouts on a curve (Figure 8) we are today. The approach of Norton and Knight (1977) however does give an idea of permeability dimensions. If episodic activity is typical of geothermal fields then one would expect that permeability changes would occur in the manner indicated by curve B in Figure 8. One would predict that drastic changes in reservoir permeability need not be accompanied by significant fluctuations of temperature and that the reservoir rocks of a geothermal field serve as a thermal buffer.

15.4 The temperature axis

Clearly it is important to know whether a geothermal field is naturally heating up, cooling down or is thermally stable, at least in the short term. It is impossible to predict future temperatures but one way of looking at the past temperatures is through studying the hydrothermal mineralogy. Several proposed mineral geothermometers have been applied to geothermal materials but testing has mainly served to invalidate the methods suggested, e.g. Al - in quartz and the sphalerite geothermometers. The three most promising methods are stable isotope measurements on hydrothermal minerals, sequences in vug fillings and fluid inclusion geothermometry. A discussion of isotopic methods is beyond the scope of these lectures but many such studies on geothermal minerals have been made (e.g. Blattner, 1975; Eslinger and Savin, 1973; Browne, et al., 1975).

Studies of the textural relations of hydrothermal minerals forming veins and filling vugs are common and have been made in the greatest detailed in Iceland (e.g. Franzson, 1983); this work has been used to identify systems that are heating (e.g. Nesjavellir; Stefansson et al., 1983) or have cooled (e.g. Husavik, ; Tomasson, 1983).

Cooling has been recognised from petrographic examination of hydrothermal minerals in cores from holes at the margins of Langano Northern lakes, Ethiopia where hole 2 encountered minerals symptomatic of temperatures up to 280°C; however, this is now 170°C hotter than the maximum measured temperature in this well (Kahsai, 1983) and shows that substantial cooling has occurred here.

Fluid inclusion geothermometry offers a more precise tool for following the thermal evolution of a geothermal system and the technique has been applied to samples from several fields. For example Wairakei, Broadlands and Tauhara are probably thermally stable whereas the northern sector of Waiotapu has cooled by as much as 20°C (Hedenquist, 1983). The northern margin of Ngawha however, seems to be heating up but its central part may have cooled several degrees since the formation of the observed hydrothermal minerals.

Fluid inclusion geothermometry, and other evidence suggests that Cerro Prieto has never been hotter (Elders, et al., 1981) whereas some fields in the nearby Imperial Valley have cooled substantially; for example Heber by about 40°C (Browne, 1977), East Mesa by 15°C and parts of the Salton Sea by as much as 80°C (Freckman, 1978).

15.5 The salinity axis

Changes in the compositions of circulating fluids (Figure 8) are sometimes recorded by the hydrothermal minerals whereby older veins are sealed off and not penetrated by younger fluids of different composition; this sometimes preserves a record of the earlier fluids. There are several generations of veins in some of the sediments at Ngawha, the youngest of which are characterised by abundant calcite showing that solutions saturated with carbon dioxide circulated most recently. There has also been a change in the chemistry of fluids at Kawah Kamojang recorded by vein relations whereby those composed of wairakite and quartz are transected by those of anhydrite; this probably means that, at an unknown stage during the evolution of this field, sulphate-dominated water replaced that of an alkali chloride type.

Where sufficient material is available it is possible to crush hydrothermal minerals thereby releasing the thermal fluids trapped in the inclusions; this fluid can then be analysed. This method has been applied to calcite cuttings from Ngawha drillhole, Ng5 (A. Christie, pers. comm.) but the method is seldom attempted on geothermal minerals because of the scarcity of suitable material. However, an idea of the gross salinity of the fluids (expressed as NaCl equivalent) trapped in the inclusions can be gleaned by measuring their freezing temperatures under the microscope using a freezing stage. To do this one makes use of the long established relationship between the concentration of NaCl dissolved in a fluid and the temperature at which it freezes. This procedure is much simpler than it sounds and the usual procedure is to freeze the liquid in the inclusion and measure the temperature at which it melts (Roedder, 1962). However, constituents other than NaCl also help to lower the freezing temperature; the most important of these, in geothermal samples, is carbon dioxide so that a knowledge of the CO_2 - H_2O - NaCl phase relations is needed before accurate results can be obtained, at least on inclusions containing dilute solutions. These relations, however, will be described in a forthcoming paper by J.W. Hedenquist (DSIR, Wairakei). However, the techique has been successfully applied to inclusions of the highly saline fluids circulating in geothermal systems in the Imperial Valley (Freckman, 1978) where there is generally good agreement between salinity estimates based upon the depression of freezing point of fluid inclusions and the measured salinity of fluids at the depths from which they derive.

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