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Hydrothermal Eruption Mechanisms and Hot Spring Gold Deposits

CARL E. NELSON AND DAVID L. GILES

Cimarron Exploration, Inc., 66 South Van Gordon, Suite 140, Lakewood, Colorado 80228

Abstract

Episodes of gold mineralization in the shallow hot spring environment are related in time to hydrothermal eruption events and in space to the resulting vent breccias and peripheral stockwork zones. It is proposed that large but short-lived overpressures in a geothermal reservoir, probably triggered by sudden magmatic heat fluxes, induce hydraulic fracturing which then evolves into hydrothermal eruptions if driven through to the surface. The maximum available energy in hot, shallow reservoirs appears easily sufficient to drive such eruptions, particularly if CO_2-rich fluids are involved. In mineralized systems, gold-bearing fluids are subsequently channeled into the outflow conduit where they flood the permeable vent breccia and peripheral stockwork. Gold is lifted into this hot spring environment above a boiling level that is elevated by high flow rates to within several hundred meters of the surface and is precipitated with abundant quartz, pyrite, and adularia, along with a distinctive and steeply zoned trace element suite. These sequential events probably occur as a continuum, which if repeated cyclically in a single vent zone, can result in ore-grade mineralization.

Introduction

> *Events that cannot happen according to received wisdom rarely gain respectability by simple accumulation of evidence for their occurrence: they require a mechanism to explain how they can happen.*
>
> Stephen Jay Gould
> 1980

HYDROTHERMAL eruption products have been recognized with increasing frequency in recent years from both active and fossil hot spring environments. Such deposits were first described by White (1955) at Lake City, Colorado, and by Lloyd (1959) at Waiotapu, New Zealand. More recently, hydrothermal eruptions have been reported by Muffler et al. (1971) at Yellowstone, by Lloyd and Keam (1976) at Waimangu, New Zealand, and by Nairn and Wiradiradja (1980) at Kawerau, New Zealand. Several historic examples have been caught on film, including eruptions at Karapiti, New Zealand, in 1981 and at Waimangu, New Zealand, in 1904.

The term hydrothermal eruption is used here as defined by Lloyd (1959) and Hedenquist (1983). It refers to an eruption, lasting from minutes to hours, which appears to have been driven wholly by the energy contained in the geothermal reservoir. Other proposed terms for the phenomenon include "hydrothermal explosion" (Muffler et al., 1971) and "mud volcano eruption" (White, 1955).

Similarities between epithermal ore deposits and active geothermal systems have long been advocated by White (1955, 1967, 1974, 1981) and others (e.g., Ewers and Keays, 1977). Evidence supporting a genetic link between geothermal activity and precious metal mineralization continues to accumulate, both from research on active geothermal systems (Henley and Ellis, 1983) and from studies of precious metal deposits in hot springs environments. Recent articles dealing with precious metal deposition in the shallower parts of geothermal systems include Weissberg et al. (1979), White (1981), Berger and Eimon (1982), Giles and Nelson (1982), Silberman (1982), White and Heropoulos (1983), and Bonham and Giles (1983).

Hot spring gold deposits are those which show compelling field evidence for mineralization within the upper few hundred meters of the paleosurface. For example, siliceous sinter has been recognized in close proximity to ore at the McLaughlin gold deposit in California and at the Hasbrouck (Graney, 1984) and Buckhorn (Monroe, 1984) deposits in Nevada. Siliceous sinter, fumarolic precipitates, and geyserite are present at three Japanese hot spring gold deposits collectively known as the Nansatsu group (e.g., Saito and Sato, 1978). Geyserite consists of rounded balls of concretionary silica. Layers of silica accumulate on rock fragments in the throats of geysers forming balls up to several inches in diameter. Individual balls are periodically erupted during geyser activity and are found interbedded with sinter near geyser vents. Additional evidence for a shallow origin includes widespread advanced argillic alteration. Widespread cristobalite-alunite-kaolinite is present above an inferred paleoground-water table at Borealis, Nevada (Strachan et al., 1982; Reid, 1984). Other gold deposits with reported evidence for near-surface gold mineralization include Pueblo Viejo in the Dominican Republic (Kesler et al., 1981), Wau in Papua New Guinea (Sillitoe et al., 1984), Cimola in British Columbia (Limbach et al., 1981), Milestone in Idaho (Barrett, 1985), and McGinness Hills in Nevada (Wendell et al., 1984).

This brief paper reviews various aspects of fossil hot spring systems with established gold mineralization. Since hydrothermal eruptions are considered by

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us to be essential for the occurrence of economic quantities of shallow gold mineralization, they are discussed in some detail. In keeping with the theme of the introductory quotation, we propose herein a hydrothermal eruption mechanism that, we are hopeful, synthesizes much of the empirical evidence gathered to date and contributes to an understanding of gold mineralization in hot spring environments.

Hydrothermal Eruption Breccias

The ejected fragmental products of hydrothermal eruptions form massive, poorly sorted units which thicken toward the vent. Individual fragments typically reflect repeated brecciation, hydrothermal alteration, and silicification prior to eruption. Hydrothermal eruption products show no evidence for a direct contribution of magmatic energy. Eruptions intermediate between purely hydrothermal and purely magmatic end members contain fresh scoria, pumice fragments, or glass shards and are best referred to as phreatomagmatic or hydrovolcanic eruptions (Sheridan and Wohletz, 1981).

The largest hydrothermal eruption craters and ejecta blankets presently known are those reported by Nairn and Wiradiradja (1980) from the Kawerau geothermal field in New Zealand. Hydrothermal eruption debris at Kawerau forms units up to 8 m in thickness near a series of coalescing craters 300 to 500 m in diameter. Fragment size and unit thickness decrease abruptly over a radial distance of 1.5 km. Most reported hydrothermal eruption blankets are of much smaller areal extent and were erupted from vents measuring but a few tens of meters in diameter.

These eruption products are commonly eroded away in fossil hot spring systems. One is likely instead to encounter steep-walled hydrothermal eruption vents from which fragmented material has been partially ejected. Contacts of in situ vent breccia with the vent walls are steep, typically irregular in shape, and can be either sharp or gradational through several meters of densely fractured wall rock. Hydrothermal vent breccias are variably clast to matrix supported, suggesting transport, at least locally, in a fluidized medium. Vent breccia fragments are angular to subrounded, depending on distance and velocity of transport and on the character of prior alteration and silicification.

Matrix material in vent breccias is finely comminuted rock flour. Progressive replacement by sulfidic microcrystalline silica often follows brecciation producing a distinctive pattern of interlocking fragments separated by narrow channels of silicified matrix. This feature has been referred to as mosaic, jigsaw puzzle, and explosion texture. Such matrix replacement material contains fine-grained pyrite or marcasite and elevated precious metal values in mineralized systems. Vent breccias from hot spring gold deposits are shown in Figure 1. The pattern of interlocking fragments may form either by a chemical brecciation mechanism such as that proposed by Sawkins (1969) or, alternatively, by local silica replacement along fractures. In either mechanism, silicification, mineralization, and development of the mosaic texture postdate the hydrothermal eruption event. More intensely silicified breccias (those with more complete replacement of matrix material) in a given deposit show better developed mosaic texture (more interlocking fragments). Such texture is absent in unsilicified vent breccias.

Overpressured Geothermal Environments

An overpressured geothermal fluid is one whose pressure exceeds its hydrostatic boiling pressure for a given temperature (White et al., 1975). Typical geothermal fluids are overpressured by a few bars as a result of the tendency for freely convecting fluids to adjust internally to a pressure approximating the cold water hydrostatic head (Cathles, 1977). Hydrothermal eruption, however, requires that fluid pressure exceed the sum of rock tensile strength plus lithostatic load. Therefore, some extraordinary condition must pertain, since hydrostatic pressure increases much more slowly than lithostatic pressure with increasing depth.

Overpressures approaching lithostatic could develop beneath a self-sealed cap rock, a term introduced by Facca and Tonani (1967) for a phenomenon recently invoked by numerous authors as a precursor to hydrothermal eruptions. An effective near-surface barrier to fluid convection would allow overpressures to develop. Given a seal, maximum-contained fluid pressures can be calculated for liquid-dominated and vapor-dominated geothermal reservoirs. Maximum fluid pressure for a liquid-dominated reservoir is the saturated vapor pressure for solutions at various temperatures which contain variable dissolved carbon dioxide.

Overpressure in a vapor-dominated system would develop somewhat differently. Pressure exerted by the cold water hydrostatic head at the boiling depth is transferred through the vapor-dominated column to a shallower point of impeded vertical flow. Maximum fluid pressure for liquid-dominated and for vapor-dominated reservoirs is shown in Figure 2. Comparison of the curves indicates that maximum fluid pressures in vapor-dominated systems are higher and are influenced to a greater extent by dissolved carbon dioxide than those associated with liquid-dominated systems. Overpressures approaching these maxima could exceed lithostatic load to depths of roughly 1,000 m.

Overpressure is a necessary precursor to hydrothermal eruptions. However, we feel that it is questionable whether a specific rock unit or zone of mineral precipitation can effectively seal and contain

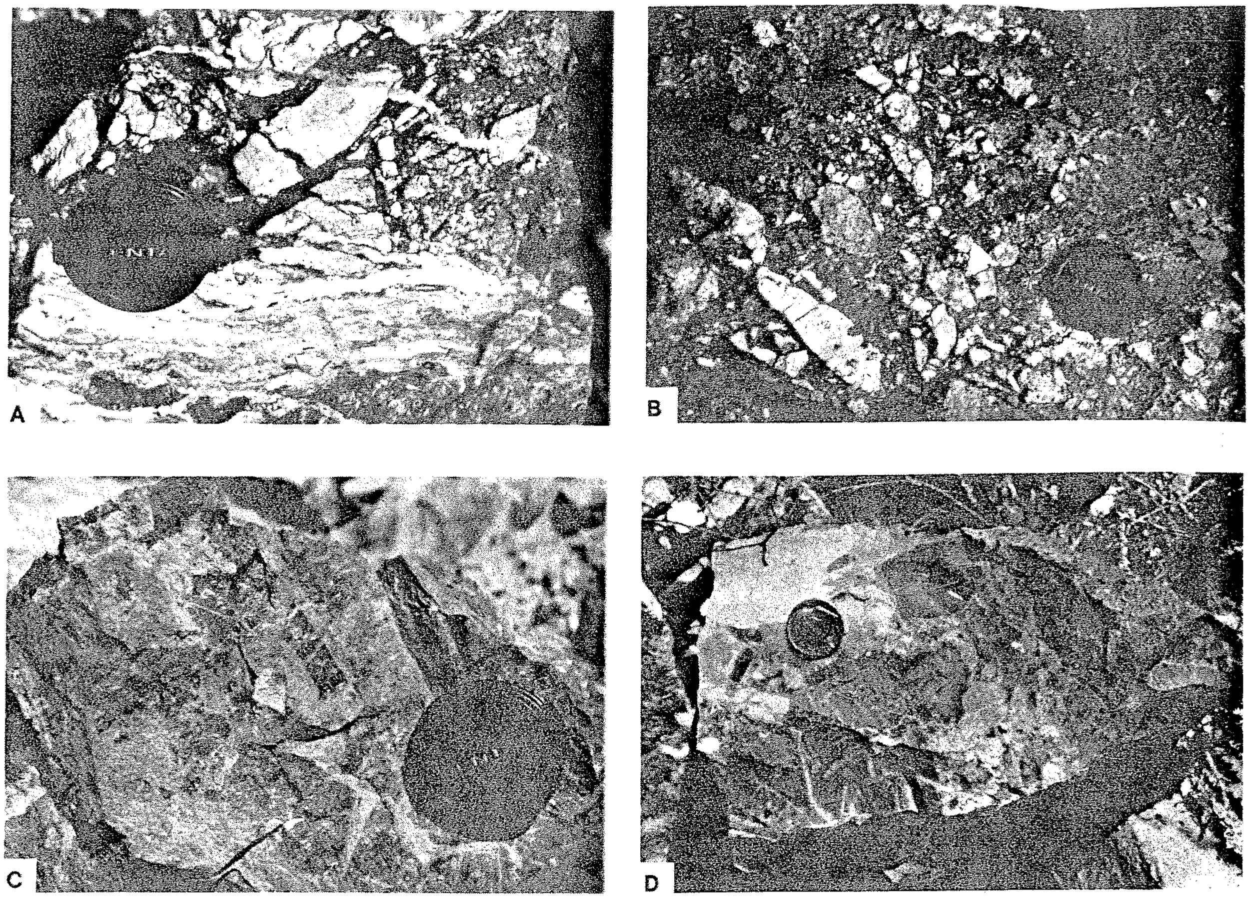

FIG. 1. Hydrothermal vent breccias. Each contains previously silicified rock fragments (basalt-dacite) in a matrix of cryptocrystalline quartz, fine-grained pyrite-marcasite and micron-size native gold. A. McLaughlin deposit, California. Late veins are of gold-bearing chalcedonia silica. B. Kasuga deposit, Japan. Matrix also contains enargite. C. Borealis deposit, Nevada. Both fragments and matrix are mineralized. D. Hasbrouck deposit, Nevada. Lighter colored matrix contains less sulfide and less gold than A, B, or C.

convecting geothermal fluids, particularly near the surface. Pressure measurements to date beneath impermeable portions of active systems are much below lithostatic. For example, overpressures of 2 bars were reported by Bolton (1970) at Wairakei, New Zealand, up to 9 bars by White et al. (1975) at Yellowstone, and 9 bars by Smith (1970) at Kawerau, New Zealand.

These measured overpressures, if they accurately reflect the order of magnitude to be expected in natural geothermal systems, are insufficient to trigger all but the shallowest of hydrothermal eruptions. Pressures are apparently bled off in geothermal systems either by geyser activity as at Yellowstone or by the natural tendency for contained fluids to migrate and find alternate routes to the surface. Effective seals to fluid convection are not indicated by measurements to date in active geothermal systems. This does not pertain to any deep self-sealed zone that might develop close to a magmatic intrusion (Fournier, 1983).

Hydrothermal Eruption Mechanisms

Despite the foregoing arguments against an effective seal, the direct and indirect evidence cited above from active and fossil geothermal systems convincingly indicates that hydrothermal eruptions do in fact occur. Fluid pressure in excess of lithostatic must pertain even if only for short periods of time. Thus, it is necessary to consider mechanisms that might overwhelm the capacity of convecting fluids to adjust pressure internally.

Likely mechanisms are suggested by accounts of hydrothermal eruptions in active geothermal fields. Champagne Pool at Waiotapu, New Zealand, the site of amorphous sulfide precipitates that are anomalous in gold (Weissberg, 1969), occupies one of a series of hydrothermal eruption vents which formed during a period of nearby rhyolite dome emplacement roughly 900 years ago (Lloyd, 1959; Cole, 1976).

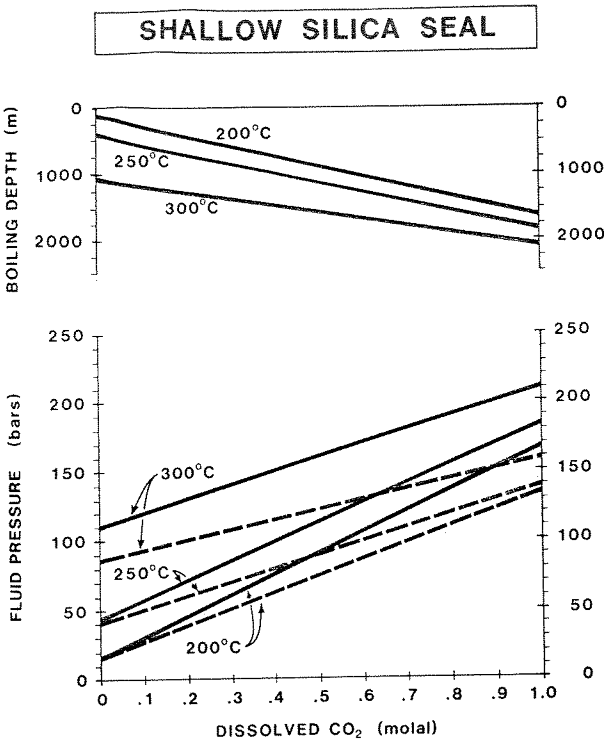

FIG. 2. Maximum fluid pressure in liquid-dominated geothermal systems (dashed line) and vapor-dominated geothermal systems (solid line) confined by a shallow silica seal.

Hydrovolcanic and hydrothermal eruptions have occurred intermittently at the nearby Waimangu geothermal field since 1886 when basaltic scoria was erupted from the Tarawera rift (Nairn, 1979).

Apparently, magmatic intrusion and possibly related seismic activity provide a trigger for hydrothermal eruptions. Any rapid contribution of magmatic heat and/or volatiles to the geothermal reservoir and surrounding rock could produce a greatly overpressured geothermal fluid. Hydraulic fracturing commences with pore-fluid expansion and fracture propagation at the depth at which fluid pressure just exceeds the sum of confining pressure plus rock strength (Norton, 1984). The transition from hydraulic fracturing to hydrothermal eruption then becomes largely a function of available energy in the geothermal reservoir. Available energy in shallow environments, under a small lithostatic load, is more likely to drive hydraulic fracturing through to the surface, resulting in a hydrothermal eruption. Hydraulic fracturing at depths of 2 to 5 km will probably not result in hydrothermal eruptions; energy would be dissipated within fractures and by stockwork brecciation. However, portions of the overpressured fluid may flash and drive toward the surface along cracks, perhaps resulting in features such as pebble dikes. This deeper environment is the likely locale for the formation of breccia pipes and related phenomena, depending on temperatures, pressure gradients, and permeabilities (Phillips, 1973; Henley and McNabb, 1978; Fournier, 1983).

Energy available to drive a hydrothermal eruption is equal to the heat supplied by the cooling rock plus fluid mixture minus the heat absorbed by the vaporizing fluid. Available energy, in units of calories per cubic centimeter, is plotted in Figure 3 for fluids of varying temperature and dissolved carbon dioxide content. The calculation procedure is that of Muffler et al. (1971), modified to account for the contribution of carbon dioxide. An example of the calculation procedure used to construct Figure 3 is provided as an appendix to this report. The appendix is available from the authors upon request.

The available energy calculation assumes, first, that fluid and wall rock are in thermal equilibrium prior to eruption and, second, that all of the heat stored in the fragmented rock mass is contributed to the fluid during an eruption. Nairn and Wiradiradja (1980) provide some justification for this assumption by showing that the time required for thermal equilibration of rock and fluid is roughly equivalent to the duration of an eruption event. However, since larger fragments will be less efficient in transferring heat, the results shown in Figure 3 represent maximum available energy. A 20 percent reduction in the efficiency of heat transfer from rock to fluid during an

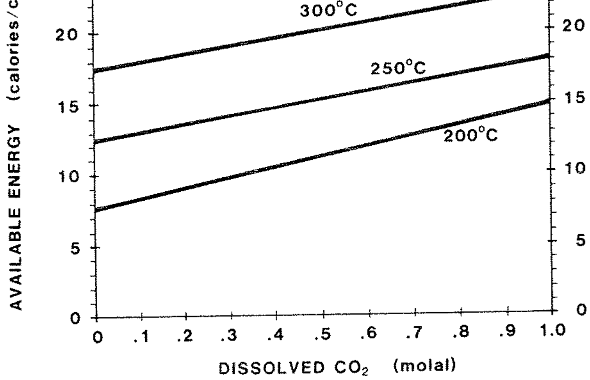

FIG. 3. Maximum available energy in geothermal reservoirs of varying temperature and dissolved carbon dioxide content.

eruption would reduce the total available energy by over 80 percent. Factors affecting the transfer of heat energy from rock to fluid, therefore, have a significant effect on eruption mechanics. The distribution of particle size is one such factor.

The energy required for hydrothermal eruption (as opposed to the energy available in the geothermal reservoir) can be estimated in several ways. Scherkenbach (1982) provides estimates based on energy expended in blasting and by large cratering explosions. Excavating rock by blasting requires 0.3 to 0.5 calories per cubic centimeter of excavated material (Brown, 1956). Large cratering explosions require an estimated 0.5 to 50.0 calories per cubic centimeter (Murphy and Vortman, 1961).

Another method of calculating required energy adds comminution energy to the energy required to lift and accelerate erupted material. Muffler et al. (1971) have calculated this required energy at roughly 2 calories per cubic centimeter for hydrothermal eruptions at Yellowstone that drew material from a depth of up to 70 m.

Given the estimates of required energy and the calculated maximum available energies plotted in Figure 3, it appears that the available energy in a geothermal reservoir is easily sufficient to drive hydrothermal eruptions, especially those which were triggered by hydraulic fracturing at shallow depths and involve hot, carbon dioxide-rich fluids. Figure 3 can therefore be used to assess the relative contribution of temperature and dissolved carbon dioxide and to estimate a maximum focal depth for a hydrothermal eruption.

Hot Spring Gold Mineralization

Mineralized hydrothermal vent breccias are present in all of the hot spring gold deposits recognized to date. However, not all hot spring systems with hydrothermal eruption breccias contain shallow gold mineralization. This section explores the various mechanisms by which hydrothermal eruptions might lead to the formation of a hot spring gold orebody.

Shallow gold mineralization occurs in permeable hydrothermal vent breccias and peripheral fractured stockworks that developed in the vicinity of an outflow conduit. Mineralization appears to have taken place when gold-bearing geothermal fluids flood this zone, a period described as the "prograde regime" by Blakestad and Stanley (1984). Such periods are characterized by high heat flow, high vertical fluid flow rates, and large volumes of focused fluid throughput. Focused fluid throughput refers to the channeling of fluid discharge in an active geothermal system through zones of high permeability. Permeable fluid conduits in active geothermal fields (e.g., Waiotapu, New Zealand) include hydrothermal vent breccias and zones of open or closely spaced fracturing. Ellis (1979) summarizes evidence suggesting that periods of high fluid throughput are brief, on the order of thousands of years, relative to the lifespan of typical geothermal systems which is hundreds of thousands to several million years.

Formation of orebodies during relatively brief periods requires elevated precious metal concentrations in solution. Concentrations of tenths or hundredths of a part per billion as at Broadlands, New Zealand, are well below the solubility limit of gold and are too low to form an ore deposit within a reasonable time frame. Concentrations of hundredths of a part per million in solution, however, may be conducive to mineralization. A fluid flow rate of 10 to 100 g/cm^2/yr could conceivably deposit 0.1 to 1.0 ppm gold per century assuming a vertical mineralized interval of 100 m and an initial gold concentration in solution of 0.10 ppm.

Such elevated gold concentrations in solution imply a combination of source rock and geothermal reservoir chemistry conducive to gold mobilization. Characteristics of gold source rocks are discussed by Keays and Scott (1976) and Keays (1984), who point out that gold accessibility to solution transport depends on total gold content, mineralogical siting, timing of gold liberation, metamorphic and deformational history, and bulk chemical composition of the source rock. A good gold source contains an anomalous amount of accessible gold but may not be particularly anomalous in terms of its absolute gold content.

Gold solubility as a function of solution chemistry has been treated in a number of studies since the pioneering work of Henley (1973) and Seward (1973). Results to date indicate that gold solubility in geothermal reservoirs at 100° to 300°C is enhanced by high temperature, reducing Eh, and neutral to alkaline pH.

Given a source rock for gold and a reservoir fluid capable of transporting gold, it remains to consider an appropriate mechanism for gold precipitation within a structurally prepared, near-surface environment. One gold precipitating mechanism involves mixing of different fluid types, as documented and advocated by Henley and McNabb (1978), Henley and Ellis (1983), and Hedenquist (1983). Passive mixing of acid, near-surface waters with rising neutral to alkaline solutions results in destabilization of bisulfide complexes and formation of precious metal-bearing precipitates. The evidence available from hot spring deposits, however, suggests that boiling is the primary mechanism responsible for gold precipitation. Boiling causes cooling, oxidation, and an increase in solution pH, all very effective mechanisms for gold deposition (Romberger, 1983). The presence of base metal sulfides and adularia, coprecipitated with gold and abundant silicia in hot spring deposits, also suggests a boiling rather than a mixing mechanism for precious metal mineralization (e.g., Fournier, 1983).

Rapid and near-instantaneous channeling of fluids through permeable vent breccias and peripheral fractured stockworks over a heated geothermal reservoir increases the pressure gradient in a convecting fluid column. Donaldson (1968) has shown that the boiling level in a column of given permeability can be raised to within a few hundred meters of the surface by increasing the flow rate toward a limit set by the cold water hydrostatic head. Precious metals are thereby lifted into the hot spring environment above an elevated boiling level and precipitated along with pyrite and marcasite, quartz, and a distinctive trace element suite consisting of shallow, enriched Hg, Tl, Sb, and Ba grading with depth through the gold (plus As) orebody to deep, enriched Cu, Pb, Zn, W, and Ag. Repeated hydraulic fracturing and hydrothermal eruptions followed by periods of high fluid throughput combine to produce economically recoverable gold deposits in the shallow parts of geothermal systems.

Conclusions

A distinctive group of epithermal gold deposits have formed (and probably are forming) within a few hundred meters of the earth's surface. Evidence for their shallow origin includes siliceous sinter, fumarolic mineral precipitates, and wall-rock alteration for a paleoground-water table, all in close spatial and temporal proximity to ore. Mineralization in each of the deposits was preceded by hydrothermal eruptions. This paper has brought together a discussion of hydrothermal eruptions and a model for shallow gold mineralization.

Hydrothermal eruptions in active geothermal systems typically appear to be triggered by an external event such as heat flux from a magmatic intrusion that suddenly increases pore pressure in the geothermal reservoir. Large overpressures must exist, if only for a brief period, in order to trigger an eruption event. Hydraulic fracturing is induced at a level where fluid pressure exceeds the sum of lithostatic load plus rock tensile strength. Available energy then becomes the critical factor in determining whether hydraulic fracturing will evolve into a hydrothermal eruption or simply dissipate energy by injection of fluid along preexisting zones of weakness. Once hydraulic fracturing has taken place, geothermal fluids are channeled through permeable solution conduits.

It is proposed that hot spring gold deposition occurs during relatively brief periods of high fluid throughput along permeable breccia conduits and through peripheral fractured wall rock. There is probably a continuum of events beginning with a magma-induced heat pulse causing a suddenly overpressured fluid reservoir and continuing through hydrothermal eruption, fluid focusing, elevated boiling level, and shallow precious metal precipitation. Cyclical repetition of this continuum of events within a given rock volume results in ore-grade mineralization. Micron-size native gold occurs with silica, adularia, several percent pyrite plus marcasite, and a distinctive, steeply zoned (telescoped) trace element suite. The irregularly shaped ore stockwork surrounds the mineralized vent breccia conduits and becomes progressively narrower and locally higher in grade with depth.

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REFERENCES

Barrett, R. A., 1985, The geology, mineralization and geochemistry of the Milestone hot-spring silver-gold deposit near the Delamar silver mine. Owyhee County, Idaho: Idaho Geol. Survey Tech. Rept. (in press).

Berger, B. R., and Eimon, P. I., 1982, Comparative models of epithermal silver-gold deposits: Am. Inst. Mining Metall. Petroleum Engineers Preprint 82-13, 25 p.

Blakestad, R. B., and Stanley, W. R., 1984, Monotonic and prograding geothermal systems and precious metal mineral deposits [abs]: Exploration for Ore Deposits of the North American Cordillera Symposium, Assoc. Explor. Geochemists, Reno, March 1984, Abstracts with Programs, p. 34.

Bolton, R. S., 1970, The behavior of the Wairakei geothermal field during exploitation: Geothermics Spec. Issue 2, v. 2, pt. 2, p. 1426–1439.

Bonham, H. F., Jr., and Giles, D. L., 1983, Epithermal gold/silver deposits: The geothermal connection: Davis, California, Geothermal Resources Council Spec. Pub. 13, p. 257–262.

Brown, F. W., 1956, Determination of basic performance properties of blasting explosives: Colorado Sch. Mines Quart., v. 51, p. 170–188.

Cathles, L. M., 1977, An analysis of the cooling of intrusives by ground-water convection which includes boiling: ECON. GEOL., v. 72, p. 804–826.

Cole, J., 1976, Rhyolitic volcanism on Mt. Tarawera: Internat. Geol. Cong., 25th, Sydney, 1976, Excursion guide 55 A and 56 A, p. 19.

Donaldson, I. G., 1968, The flow of steam-water mixtures through permeable beds: A simple simulation of a natural undisturbed hydrothermal region: New Zealand Jour. Sci., v. 11, p. 3–23.

Ellis, A. J., 1979, Explored geothermal systems, *in* Barnes, H. L., ed., Geochemistry of hydrothermal ore deposits: New York, John Wiley and Sons, p. 632–683.

Ewers, G. R., and Keays, R. R., 1977, Volatile and precious metal zoning in the Broadlands geothermal field, New Zealand: ECON. GEOL., v. 72, p. 1337–1354.

Facca, G., and Tonani, F., 1967, The self-sealing geothermal field: Bull. Volcanol., v. 30, p. 271–273.

Fournier, R. O., 1983, Active hydrothermal systems as analogues of fossil systems: Davis, California, Geothermal. Resources Council Spec. Rept. 13, p. 263–284.

Giles, D. L., and Nelson, C. E., 1982, Epithermal lode gold deposits of the circum-Pacific rim, *in* Transactions, Circum-Pacific

Energy Mineral Resources Conf., 3rd Honolulu, August, 1982: Tulsa, Oklahoma, Am. Assoc. Petroleum Geologists, p. 273–278.

Graney, J. R., 1984. Controls of alteration and precious metal mineralization in a fossil hydrothermal system, Hasbrouck Mountain, Nevada [abs.]: Geol. Soc. America Abstracts with Programs, v. 16, p. 523.

Hedenquist, J. W., 1983, Waiotapu, New Zealand: The geochemical evolution and mineralization of an active hydrothermal system: Unpub. Ph.D. thesis, Univ. Auckland, 243 p.

Henley, R. W., 1973, Solubility of gold in hydrothermal chloride solutions: Chem. Geology, v. 11, p. 73–87.

Henley, R. W., and Ellis, A. J., 1983, Geothermal systems ancient and modern: A geochemical review: Earth-Sci. Rev., v. 19, p. 1–50.

Henley, R. W., and McNabb, A., 1978, Magmatic vapor plumes and ground-water interaction in porphyry copper emplacement: ECON. GEOL., v. 73, p. 1–20.

Keays, R. R., 1984. Archean gold deposits and their source rocks: The upper mantle connection, *in* Foster, N. P., ed., Gold 82—The geology, geochemistry and genesis of gold deposits: Rotterdam, A. A. Valkema, p. 17–51.

Keays, R. R., and Scott, R. B., 1976, Precious metals in ocean-ridge basalts: Implications for basalts as source rocks for gold mineralization: ECON. GEOL., v. 71, p. 705–720.

Kesler, S. E., Russell, N., Seaward, M., Rivera, J., McCurdy, K., Cumming, G. L., and Sutter, J. F., 1981, Geology and geochemistry of sulfide mineralization underlying the Pueblo Viejo gold-silver oxide deposit. Dominican Republic: ECON. GEOL., v. 76, p. 1096–1117.

Limbach, F. W., Cruson, M. G., and Brooks, R. H., 1981, Geology of stockwork gold deposits as exemplified by the Cinola deposit: Denver, Colorado, Mining Assoc. Yearbook, p. 122–129.

Lloyd, E. F., 1959, The hot springs and hydrothermal eruptions of Waiotapu: New Zealand Jour. Geology Geophysics, v. 2, p. 141–176.

Lloyd, E. F., and Keam, R. F., 1976, Waimangu hydrothermal field: Internat. Geol. Cong., 25th, Sydney, 1976, Excursion guide 55A and 56A, p. 23–26.

Monroe, S. C., 1984, Geology and geochemistry of the Buckhorn mine, Eureka Co., Nevada [abs.]: Geol. Soc. America Abstracts with Programs, v. 16, p. 599.

Muffler, L. J. P., White, D. F., and Truesdell, A. H., 1971, Hydrothermal explosion craters in Yellowstone National Park: Geol. Soc. America Bull., v. 82, p. 723–740.

Murphy, B. F., and Vortman, L. J., 1961, High-explosive craters in desert alluvium, tuff and basalt: Jour. Geophys. Research, v. 66, p. 3389–3404.

Nairn, I. A., 1979, Rotomahana—Waimangu, 1886: Base surge and basalt magma: New Zealand Jour. Geology Geophysics, v. 22, p. 363–378.

Nairn, I. A., and Wiradiradja, S., 1980, Late Quaternary hydrothermal explosion breccias at Kawerau geothermal field, New Zealand: Bull. Volcanol., v. 43, p. 1–13.

Norton, D. L., 1984, Theory of hydrothermal systems: Ann. Rev. Earth Planet. Sci., v. 12, p. 155–177.

Phillips, W. J., 1973, Mechanical effects of retrograde boiling and its probable importance in the formation of porphyry ore deposits: Inst. Mining Metallurgy Trans., sec. B., v. 82, p. B90–B98.

Reid, R. F., 1984, The geology of the Borealis deposit [abs.]: Geol. Soc. America Abstracts with Programs, v. 16, p. 632.

Romberger, S. B., 1983, The behavior of precious metals in geothermal fluids [abs]: Geothermal Systems and Ore Deposits Symposium Am. Assoc. Mining Metall. Petroleum Geologists Ann. Mtg., Los Angeles, Abstracts with Programs, p. 12.

Saito, M., and Sato, E., 1978, On the recent exploration at the Iwato gold mine: Mining Geology, v. 29, p. 191–202.

Sawkins, F. J., 1969, Chemical brecciation, an unrecognized mechanism for breccia formation?: ECON. GEOL., v. 64, p. 613–617.

Scherkenbach, D. A., 1982, Geologic, mineralogic, fluid inclusion and geochemical studies of the mineralized breccias at Cumobabi, Sonora, Mexico: Unpub. Ph.D. thesis, Univ. Minnesota, 190 p.

Seward, T. M., 1973, Thio complexes of gold and the transport of gold in hydrothermal ore solutions: Geochim. et Cosmochim. Acta, v. 37, p. 379–399.

Sheridan, M. F., and Wohletz, K. H., 1981, Hydrovolcanic explosions: The systematics of water-pyroclast equilibration: Science, v. 212, p. 1387–1389.

Silberman, M. L., 1982, Hot spring type large tonnage, low grade gold deposits: U. S. Geol. Survey Open-file Rept. 82-795, p. 131–143.

Sillitoe, R. H., Baker, E. M., and Brook, W. A., 1984, Gold deposits and hydrothermal eruption breccias associated with a maar volcano at Wau, Papua New Guinea: ECON. GEOL., v. 79, p. 638–655.

Smith, J. H., 1970, Geothermal development in New Zealand: Geothermics Spec. Issue 2, v. 3, pt. 1, p. 232–247.

Strachan, D. G., Pettit, P. M., and Reid, R. F., 1982, The geology of the Borealis gold deposit, Mineral Co., Nevada [abs.]: Geol. Soc. America Abstracts with Programs, v. 14, p. 627.

Weissberg, B. G., 1969, Gold-silver ore-grade precipitates from New Zealand thermal waters: ECON. GEOL., v. 64, p. 95–108.

Weissberg, B. G., Browne, P. R. L., and Seward, T. M., 1979, Ore metals in active geothermal systems, *in* Barnes, H. L., ed., Geochemistry of hydrothermal ore deposits: New York, John Wiley and Sons, p. 738–780.

Wendell, D. E., Casaceli, R. J., and Hoisington, W. D., 1984, Geochemical patterns in an altered area at McGinness Hills, Lander County, Nevada [abs.]: Exploration for Ore Deposits of the North American Cordillera Symposium, Assoc. Explor. Geochemists, Reno, March 1984, Abstracts with Programs, p. 34.

White, D. E., 1955. Violent mud-volcano eruption of Lake City hot springs, northeastern California: Geol. Soc. America Bull., v. 66, p. 1109–1130.

—— 1967, Mercury and base metal deposits with associated thermal and mineral waters, *in* Barnes, H. L., ed., Geochemistry of hydrothermal ore deposits: New York, Holt, Rinehart and Winston, p. 575–631.

—— 1974, Diverse origins of hydrothermal ore fluids: ECON. GEOL., v. 69, p. 954–973.

—— 1981, Active geothermal systems and hydrothermal ore deposits: ECON. GEOL., 75TH ANNIV. VOL., p. 392–423.

White, D. E., Fournier, R. O., Muffler, L. J. P., and Truesdell, A. H., 1975, Physical results of research drilling in thermal areas of Yellowstone National Park, Wyoming: U. S. Geol. Survey Prof. Paper 892, 70 p.

White, D. E., and Heropoulos, C., 1983, Active and fossil hydrothermal-convection systems of the Great Basin: Geothermal Resources Council Spec. Rept. 13, p. 41–53.