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25 June 1979; revised 23 October 1979

Geothermal System at 21°N, East Pacific Rise: Physical Limits on Geothermal Fluid and Role of Adiabatic Expansion

Abstract. *Pressure-volume-temperature relations for water at the depth of the magma chamber at 21°N on the East Pacific Rise suggest that the maximum subsurface temperature of the geothermal fluid is about 420°C. Both the chemistry of the discharging fluid and thermal balance considerations indicate that the effective water/rock ratios in the geothermal system are between 7 and 16. Such low ratios preclude effective metal transport at temperatures below 350°C, but metal solubilization at 400°C and above is effective even at such low ratios. It is proposed that the 420°C fluid ascends essentially adiabatically and in the process expands, cools, and precipitates metal sulfides within the upper few hundred meters of the sea floor and on the sea floor itself.*

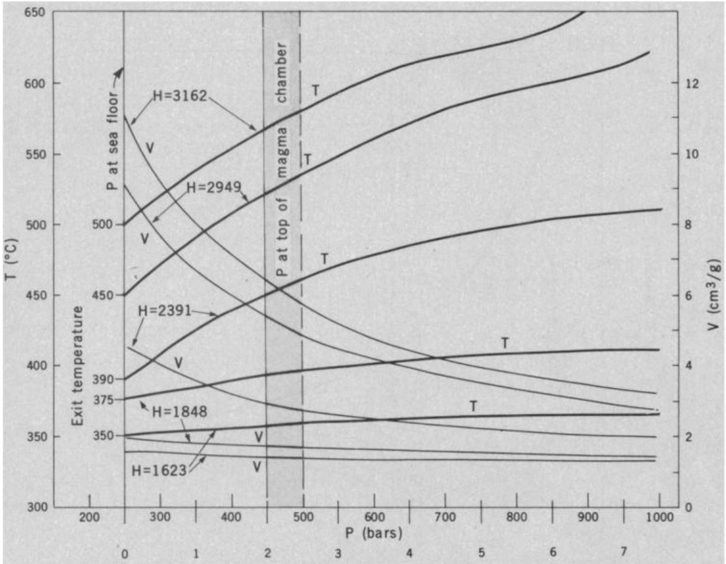
The discoveries by manned submersible of actively forming massive sulfide deposits at 21°N on the East Pacific Rise (*1*, *2*) substantiate the long-held belief that such deposits are associated with sea-floor spreading. The discharging fluids are apparently actively precipitating a deposit largely composed of sphalerite, chalcopyrite, anhydrite, and barite (*3*, *4*).

The temperature of the fluid is unexpectedly high (*2*), and a reevaluation of the conditions of fluid generation is warranted. The purpose of this report is to consider the possible pressure and temperature limits on the fluid in the subsurface and the implications these limits might have for the mechanisms of metal transport and deposition.

It has been commonly held that convecting seawater cools the upper portion of new oceanic crust and in the process alters the rocks and leaches heavy metals at maximum temperatures of about 350°C. Mineral deposition was believed to result from cooling of this fluid.

This model does not easily apply to the system at 21°N. Experimental studies have shown that during cooling of heated seawater, silica should overwhelmingly dominate the precipitation products (*5*), yet silica is a relatively minor component of the observed assemblage at 21°N (*3*, *4*). Moreover, adequate concentrations of heavy metals are solubilized at 350°C only when water/rock ratios during interaction are large (*6*), whereas experimental results show that the water/rock ratio does not control metal solubilization above about 400°C (*7*).

An estimated exit temperature of 390° ± 10°C was reported for a single vent by Speiss *et al.* (*2*) from dives carried out with the submersible *Alvin* during the spring of 1979 at 21°N. During the dives in the fall of 1979 the temperature directly observed at several vents was 350°C (*8*), suggesting that the earlier estimate might be spurious. However, later examination of the chart record of *Alvin*'s temperature probe indicated two measurements during the final dive in excess of 400°C (*9*). During the fall 1979 dives samples of the hydrothermal fluid were taken but were contaminated by various amounts of seawater. Plots of



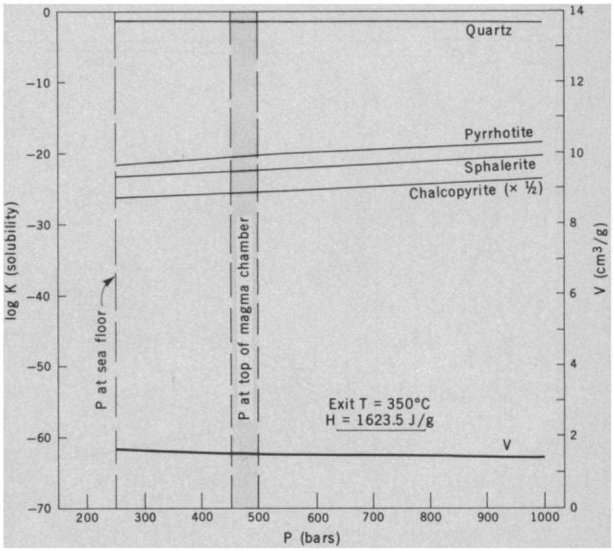


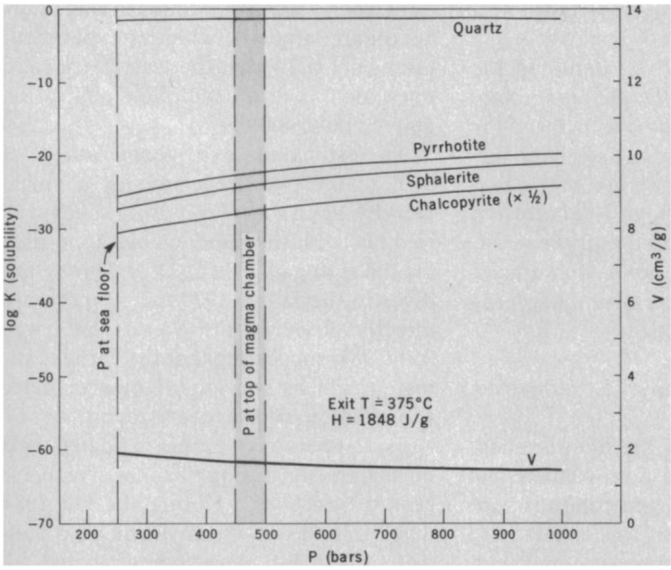
Fig. 1 (left). Changes in water temperature *T* and specific volume *V* as a function of pressure at constant enthalpy *H* (adiabatic compression). Relations calculated from data in (*27*) for sea-floor exit temperatures of 350°, 375°, 390°, 450°, and 500°C. Fig. 2 (right). Solubility constants of minerals and specific volume of water as a function of pressure at a constant enthalpy (adiabatic) of 1623 J/g, corresponding to a sea-floor exit temperature of 350°C. In Figs. 2 through 6, data for water volume are from (*27*) and solubility constants were calculated by SUPCRT (*17*).

dissolved components against temperature for these samples indicate that the projected salinity of the hydrothermal fluid differs by no more than 10 percent from that of normal seawater, and that the projected concentrations of Mg and SO_4 are essentially zero at 350°C (*8*). If the fluid contains no Mg, then the system is rock-dominated with respect to Mg (*6*), and the fluid at this temperature should be incapable of containing large amounts of metals and H_2S in a stable condition. That the fluid is actively precipitating metal sulfides implies that the subsurface fluid temperature is closer to 400°C.

Another line of evidence that suggests higher subsurface temperatures is the presence of magnesium oxysulfates in the chimney deposits (*4*). Because the fluid lacks Mg and SO_4, precipitation of this phase is attributed to local heating of bottom water by injection and partial mixing with hydrothermal fluid. Magnesium oxysulfate begins to precipitate during the heating of seawater at a temperature between 250° and 300°C (*10*). It is difficult to visualize how the injection of 350°C hydrothermal fluid into 2°C bottom water could induce precipitation of this phase.

From the foregoing, and if the cooling model holds, it follows that the fluid at depth must be at some higher temperature. Because water is highly expandable at temperatures above the critical temperature (374°C), very much higher temperatures are unlikely unless pressures (and therefore depth of water penetration) are very great. What are the pressure limits on the circulating fluid? Water depth all along the crest of the East Pacific Rise ranges from 2500 to 2900 m (*11*), corresponding to a lower limit of hydrostatic pressure of 250 to 290 bars for the discharging fluids.

Seawater circulation at 21°N corresponds to Lister's (*12*) "active" phase of hydrothermal cooling of oceanic crust. During this phase, seawater extends to a zone immediately above the magma



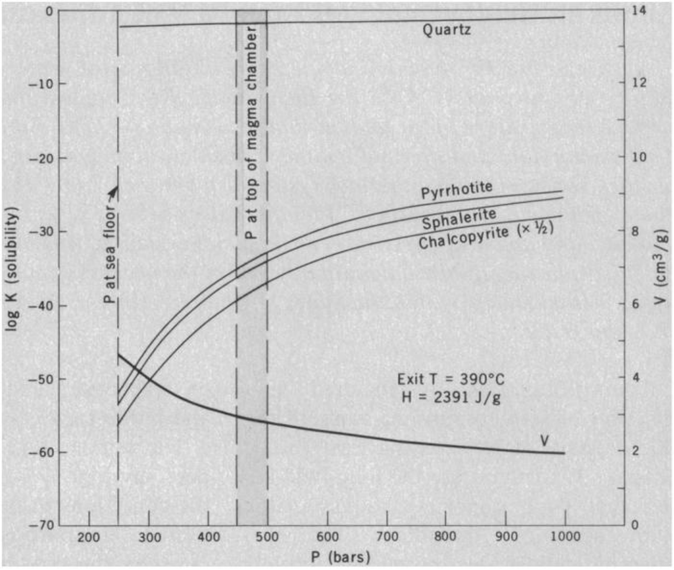
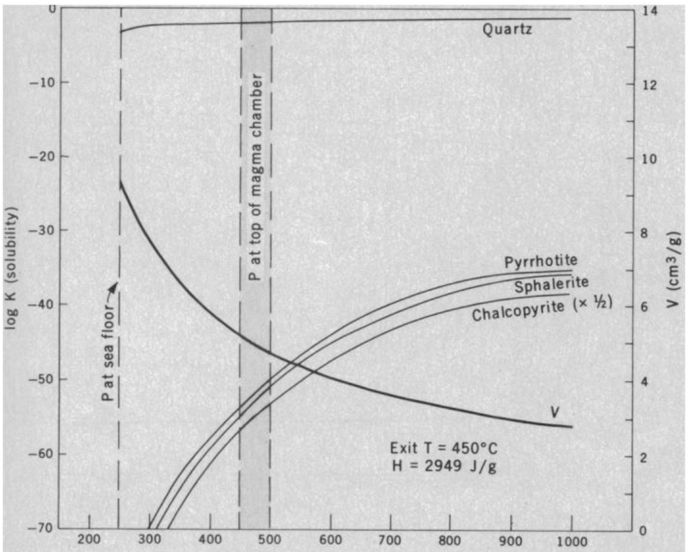


Fig. 3 (left). Solubility constants of minerals and specific volume of water as a function of pressure at a constant enthalpy (adiabatic) of 1848 J/g, corresponding to a sea-floor exit temperature of 375°C. **Fig. 4 (right).** Solubility constants of minerals and specific volume of water as a function of pressure at a constant enthalpy (adiabatic) of 2391 J/g, corresponding to a sea-floor exit temperature of 390°C.



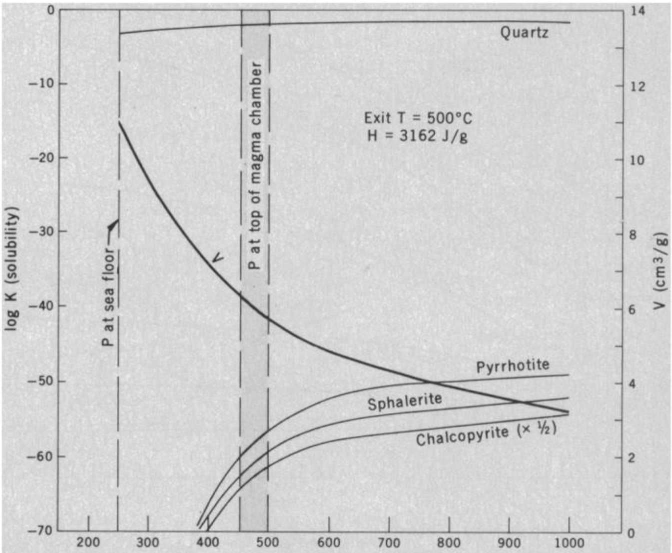


Fig. 5 (left). Solubility constants of minerals and specific volume of water as a function of pressure at a constant enthalpy (adiabatic) of 2949 J/g, corresponding to a sea-floor exit temperature of 450°C. **Fig. 6 (right).** Solubility constants of minerals and specific volume of water as a function of pressure at a constant enthalpy (adiabatic) of 3162 J/g, corresponding to a sea-floor exit temperature of 500°C.

chamber, penetrating downward along a "cracking" front. The depth and effective bottom boundary temperature of this front are related to the temperature at which the rock is sufficiently stressed to crack. Unfortunately, this cracking temperature is not yet predictable from current knowledge of the mechanical behavior of rocks (*12*).

During cooling, the cracking front moves downward and eventually extends through the depth of the original magma chamber, presumably represented by the cumulate gabbros in ophiolite sequences (held to be former oceanic crust). As the lower depth is approached, intensity of circulation decreases because most of the heat has been transported by this time, and the hydrothermal system enters the "passive" stage. Isotopic and mineralogical evidence from the Oman ophiolite indicates that at this late stage the volume of seawater reaching this depth is very small (water/rock ratios ≤ 0.3), and that seawater reaches temperatures of 400° to 500°C (*13*) at a maximum depth of about 4 km below the sea floor (hydrostatic pressures of 650 to 700 bars).

Thus, at any given segment of rise crest, the maximum depth of seawater penetration must be somewhat above the top of the magma chamber. Seismic data from 21°N show a low-velocity zone, believed to represent the magma chamber, with its top approximately 2 km below the sea floor (*14*). This would suggest that 21°N is undergoing active-phase hydrothermal cooling with a maximum depth of circulation of about 2 km (hydrostatic pressure approximately 450 to 500 bars). Thus, the pressure limits for the circulating fluid are confined between approximately 250 and 500 bars.

Under these conditions, what might be the maximum fluid temperatures? Because the hydrothermal fluids and normal seawater are very close in salinity, they can be treated as the same fluid. The pressure-volume-temperature (PVT) relations for seawater, essentially a 3.5 percent NaCl solution, differ very little from those for pure water (*15*). Because the pressure at the sea floor slightly exceeds the critical pressure, subsurface temperature for the circulating fluids is not constrained by boiling.

During the active phase of hydrothermal circulation, discharge is very rapid and is confined to narrow conduits, so that little heat is lost by conduction to the overlying rocks during the fluid's ascent (*16*). However, some expansion of the fluid must take place and result in an adiabatic temperature drop. Thus, minimum temperatures at depth can be estimated from the sea-floor exit temperatures.

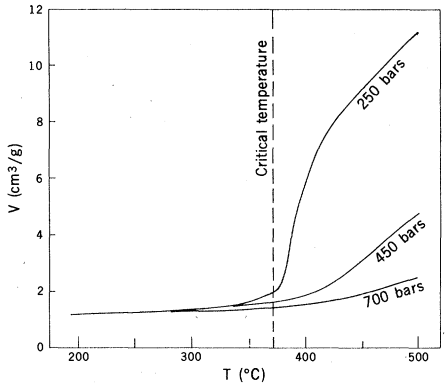


Fig. 7. Specific volume of water as a function of temperature for pressures of 250, 450, and 700 bars. Data are from (*27*).

As a first approximation, therefore, and as a minimum, it is assumed that the fluid has conserved enthalpy during ascent. Enthalpies for possible exit temperatures ranging from 350° to 500°C at 250 bars for pure water are shown in Fig. 1 along with corresponding values for specific volume. Temperature and volume as a function of pressure at constant enthalpy are then plotted to illustrate the magnitude of temperature change with depth. Figure 1 shows that even at the pressures of the sea floor, water is greatly expanded at temperatures above critical (374°C); specific volumes range from 4.6 cm³/g at 390°C to 11.1 cm³/g at 500°C. At these expanded volumes water is more like a gas than a liquid, and as such would be a very poor solvent for heavy metals.

The effects of compressing the fluid to pressures corresponding to the top of the magma chamber are minimal for exit temperatures below 375°C, but are profound for those above (Fig. 1). For example, adiabatic compression from 250 to 450 bars of a fluid with a discharge temperature of 375°C (enthalpy, 1848 J/g) increases its temperature to 394°C while decreasing its specific volume from 1.97 to 1.72 cm³/g. Similar compression of a 390°C fluid (enthalpy, 2391 J/g) results in a temperature increase to 450°C and a specific volume decrease from 4.61 to 2.9 cm³/g. In the latter case, this fluid changes from being more liquidlike at the top of the magma chamber to being more gaslike at the sea floor, a change that may have strong implications for metal transport and precipitation. Fluids with exit temperatures of 450°C and above are still essentially gases even at the pressures at the top of the magma chamber (specific volume is 5.4 cm³/g at 450 bars) and are unlikely to be effective solvents for metal sulfides during interaction with oceanic crust. Indeed, the great expansibility and increased buoyancy of water above about 420°C at 450 bars must act as a barrier to greater heating and deeper penetration.

In the absence of experimental data, it is possible to approximate the effect of adiabatic expansion on heavy-metal solubilization theoretically. Solubility constants for chalcopyrite, sphalerite, pyrrhotite, and quartz were calculated at various pressures and temperatures with the SUPCRT computer program of Helgeson *et al.* (*17*).

Plots of solubility as a function of pressure at constant enthalpy for various exit temperatures are shown in Figs. 2 to 6. For an exit temperature of 350°C (Fig. 2), pressure has very little effect on the solubility constants, as might be expected from the small changes in specific volume. At 375°C, however, solubilities of

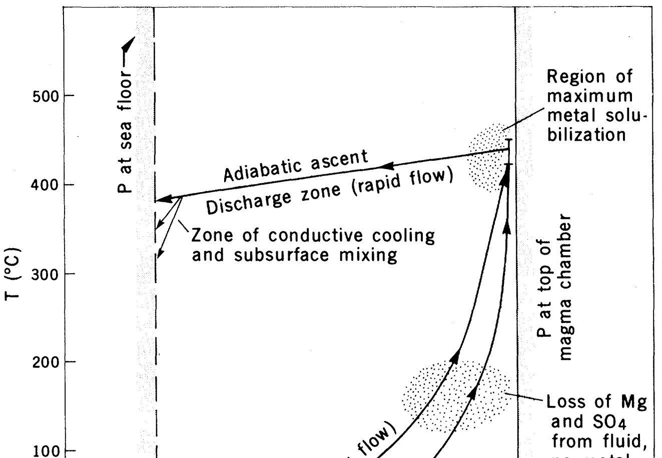


Fig. 8. Suggested pressure-temperature relations for cycle of geothermal fluid at 21°N.

the sulfides decrease by about 2 log units during expansion from 450 to 250 bars. Over the same range, quartz solubility is virtually unchanged. At an exit temperature of 390°C, corresponding to the temperature at 21°N reported by Speiss *et al.* (*2*), solubilities of the sulfides decrease by 7 to 10 log units for the same expansion (Fig. 4), and that of quartz decreases by only 0.7 log unit. These relations suggest that adiabatic expansion is the mechanism of precipitation of the mineral suite observed at 21°N.

At higher exit temperatures (Figs. 5 and 6), similarly dramatic changes in mineral solubilities are seen for the same pressure drop, but absolute solubilities at 450 bars are much lower than at an exit temperature of 390°C because at the higher temperature the fluids are highly expanded. Fluids with exit temperatures of 450°C or above, at 450 bars, are unlikely to solubilize metals effectively. Mineral solubilities for these fluids are greatly increased at pressures of 650 to 700 bars (Figs. 5 and 6), a fact that may have importance for later stages in the cooling of the magma chamber.

Apparently, the solubilities of ionic species such as sulfides are much more strongly dependent on the density of the fluid than are those of molecular species such as dissolved silica. Precipitation by adiabatic expansion, therefore, would explain the paucity of quartz and other silica phases in the deposits at 21°N.

What limits can be placed on the maximum temperature of the fluid? The degree of expansion the fluid can undergo during the recharge cycle of the hydrothermal system is limited by buoyancy, as first suggested by Truesdell and Fournier (*18*) for the Yellowstone geothermal system. Such buoyancy is in fact the driving force for discharge and therefore must be the limiting factor for the maximum temperature the fluid acquires. A plot of specific volume against temperature (Fig. 7) shows that water at 250 bars expands precipitously at temperatures above about 380°C, and water at 450 bars expands markedly above about 420°C. The actual value of the specific volume that limits hydrothermal flow is related to the magnitude of the driving forces of the individual systems and is therefore difficult to specify quantitatively. However, judging from the rapidity of water expansion above 420°C at 450 bars, the maximum possible temperature cannot be much above 420°C.

Another consequence of expansion of the fluid is the deposition of minerals, primarily quartz, that may seal the passageways. Truesdell and Fournier (*18*) suggested that self-sealing by quartz deposition was the primary control on the higher temperature of geothermal waters in Yellowstone. The temperature at which quartz has its maximum solubility increases with pressure. At 500 bars, quartz solubility reaches a maximum at 420°C and falls steeply at higher temperature (*19*), presumably as a consequence of rapid fluid expansion. A fluid equilibrating with rocks at 420°C and 500 bars would quickly deposit quartz with further heating, because the rate of quartz deposition is rapid at these temperatures. When this occurs, permeability would be abruptly decreased at the same time that buoyancy is rapidly increasing.

With increased pressure, the point of maximum quartz solubility becomes less pronounced, but only at 750 bars and above does it disappear altogether. This pressure corresponds to a depth of 5 km below the sea floor, which would place it within the lower portion of the layered gabbro, the bottom of the presumed magma chamber. Thus, the point of maximum quartz solubility may control the maximum temperature of hydrothermal circulation through most of the cooling history of the magma body. Note, however, that as the cooling front moves downward and into higher pressure regimes, maximum permissible temperatures increase as well. If we consider eventual circulation to be as deep as suggested by isotopic data for the Oman ophiolite, 4.5 to 5 km below the sea floor (*13*), then pressures as great as 700 bars may be experienced by the fluid. At this pressure (Fig. 7), significant fluid expansion does not occur until about 475° to 500°C. The waning of the maximum in quartz solubility at this pressure removes the quartz sealing barrier as well.

Thus, the following model is proposed for the present conditions at 21°N (depicted in Fig. 8). Convecting seawater harvests heat from the top of a recently emplaced magma chamber, reaching depths no greater than about 2 km below the sea floor. The recharge area for the seawater is very large compared to the area of discharge, and recharge flow is low. Integrated water/rock ratios during interaction will be in the range of 7 to 16 (*20*) and the system will be rock-dominated with respect to Mg. The descending seawater loses its Mg and SO_4 early in the cycle, probably at temperatures no greater than 100° to 150°C. Metals and H_2S will be initially solubilized from the rocks during the Mg removal stage, but will be reprecipitated into alteration products soon after Mg removal is complete and *p*H begins to rise. The fluid reaches its maximum enthalpy and temperature somewhat above the magma chamber, a temperature limited to about 420°C by the abruptly increasing expansion and buoyancy of the fluid and by self-sealing processes at 450 bars. Under these conditions, even though rock-dominated, the seawater solubilizes tens of parts per million of heavy metals and H_2S from the rocks. At this point the fluid is close to equilibrium with quartz and Fe and base-metal sulfides. The fluid now ascends, and because it is confined to a few localized and narrow channels, the flow now must be rapid and heat loss is negligible. The fluid ascends adiabatically until it reaches the vicinity of the sea floor (*16*). Adiabatic expansion will result in a continuous temperature drop to about 385°C at the sea floor, accompanied by abrupt and increasing supersaturation of the sulfide minerals by 20 orders of magnitude (Fig. 4). Any voids in the upper few hundred meters of the channel ways will become filled with sulfide minerals, and self-sealing and rechannelization of the fluid will be frequent.

Because of space constraints, the largest deposits of sulfides will be near or on the sea floor, in cavities, lava tubes, and evacuated lava lakes in the top few tens of meters below the sea floor. Precipitation by cooling and mixing will also take place on the sea floor, but the resulting deposits should have much silica in addition to sulfides. Anhydrite and barite will form as the hydrothermal fluid mixes with the sulfate-rich bottom waters. The concentration of Fe should greatly exceed those of Zn and Cu in the fluid (*7*), so the observed predominance of sphalerite in the deposits suggests that either pyrite or pyrrhotite predominates in the subsurface, or Fe is discharged directly into the bottom water to precipitate as colloids with the excess silica and seawater Mg. This precipitate will have the composition of highly stable metalliferous sediment (*21*) and will be dispersed throughout the crestal area.

The resulting massive sulfide deposit on the sea floor is markedly metastable with respect to oxygenated seawater. Barite is apparently stable with respect to seawater and will likely survive (*22*), but anhydrite will dissolve and iron sulfide will undergo autocatalytic oxidation, producing self-destroying sulfuric acid. Unless the deposit is soon sealed off and covered by later volcanism, it will completely alter to a sea-floor gossan, perhaps analogous to the ochers overlying massive sulfide deposits in Cyprus (*23*).

The sea-floor deposits at 21°N contrast in many ways with massive sulfide deposits in Cyprus and other ophiolite complexes. In these deposits, pyrite overwhelmingly dominates the sulfide mineralogy and primary sulfate minerals are absent. At 21°N, however, we see only the very top of the system, which may not be representative of the whole and may be geologically ephemeral.

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5 October 1979; revised 21 December 1979

Cytochrome P-450: Localization in Rabbit Lung

Abstract. *Cytochrome P-450–dependent monooxygenase systems, which metabolize endogenous as well as foreign compounds, are found in hepatic and several extrahepatic tissues of mammals, including humans. A form of cytochrome P-450 is localized in the nonciliated bronchiolar epithelial cells (Clara cells) of the small airways of rabbit lung. The apparent high concentration of the cytochrome in this pulmonary cell type compared to liver may be an important determinant in the susceptibility of the lung to a number of toxic chemicals that undergo metabolic activation.*

Cytochrome P-450–dependent monooxygenase systems provide a major pathway in a number of tissues for the oxidative metabolism of many chemicals present in the environment. These enzyme systems are bound to the endoplasmic reticulum and are composed of a flavoprotein (NADPH–cytochrome P-450 reductase; E.C. 1.6.2.4) and a family of hemoproteins called cytochrome P-450. Several forms of the cytochrome that have overlapping, but different, substrate specificities have been purified and characterized (*1, 2*). The monooxygenase systems generally produce metabolites that are less toxic than the parent compounds, which include a number of pesticides, drugs, and direct-acting carcinogens and mutagens. However, many procarcinogens, promutagens, and other toxic substances are activated by this metabolic process and are converted from relatively nontoxic compounds to extremely toxic metabolites.

Cytochrome P-450–dependent metabolism is required for the activation of several pulmonary toxins that produce different deleterious effects. For example, activation of 4-methyl-benzaldehyde results in the destruction of pulmonary cytochrome P-450 (*3*), activation of benzo[*a*]pyrene is involved in pulmonary carcinogenesis (*4*), and activation of 4-ipomeanol ("lung edema factor" isolated from mold-infected sweet potatoes) leads to pulmonary edema, congestion, and hemorrhage (*5*). Histopathologic and autoradiographic studies in rat and mouse (*6, 7*) have shown that the toxicity and covalent binding of 4-ipomeanol are both associated with the nonciliated bronchiolar epithelial cell (Clara cell) of the small airways of the lung. These findings suggest that the Clara cell is the site of activation of this toxin. In addition, the Clara cell contains large amounts of endoplasmic reticulum relative to other pulmonary cell types (*8*).

By using antibodies against a form of cytochrome P-450, we have demonstrated directly the presence of this enzyme in the nonciliated bronchiolar epithelium of rabbit lung. The cytochrome, P-450$_\mathrm{I}$ (*9*), was purified to apparent homogeneity, as determined by sodium dodecyl sulfate (SDS)–polyacrylamide gel electrophoresis, by the method of Wolf *et al.* (*10*). Antibodies were elicited in goats by intermittent injections of the purified protein. The specificity of the antibody was demonstrated by the Ouchterlony technique of double immunodiffusion with purified enzymes and microsomal preparations (*11*). The antibody was used with the techniques of immunofluorescence (*12*) and antibody peroxidase-antiperoxidase (PAP) (*13*) in the study reported here. Sections of tissue containing small airways were examined.

Figure 1 shows that incubation of rabbit pulmonary tissue with the specific antiserum (1000-fold dilution), but not with the normal goat serum, resulted in most of the fluorescein isothiocyanate–conjugated rabbit antiserum to goat immunoglobulin G being bound in the Clara cell. No fluorescent staining was observed in the ciliated bronchiolar epithelium or in the alveolar epithelium. Similar results were obtained by the PAP method (Fig. 2). Reaction product was observed in the luminal pole of the cell cytoplasm with the greatest intensity at the perimeter of the cytoplasm when either method was used. Endogenous peroxidase activity was reduced by exposing the tissue to hydrogen peroxide (3 percent) prior to treatment with the antiserum when the PAP method was used. Further dilution of the antiserum (2500-fold) resulted in a decreased intensity of staining but did not alter the location of the reaction.

Our results suggest that cytochrome P-450$_\mathrm{I}$ is an important factor in the pulmonary-specific effects of some toxic chemicals even though this enzyme is present in the livers of untreated rabbits at ap-